

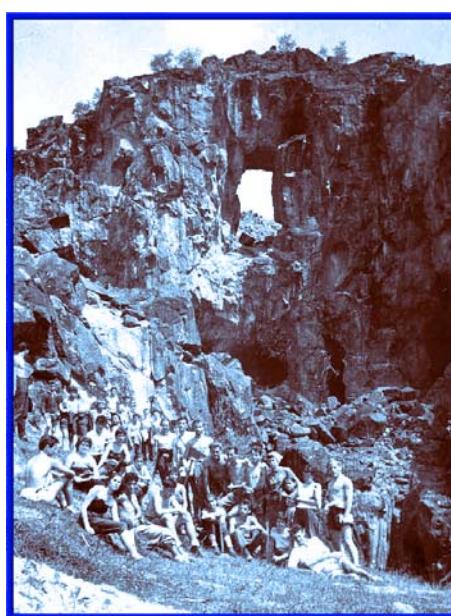
GEOLOGICAL SOCIETY OF ROMANIA



GEO 2005

PROCEEDINGS

**OF THE
ANNUAL SCIENTIFIC SESSION
AND
FIELD TRIP GUIDE**



Rosia Montana, 20-21 May 2005



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GEO 2005 - PROCEEDINGS

Annual Scientific Session of the Geological Society of Romania – Rosia Montana, 20-21 May, 2005



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FOREWORDS



GEOLOGIA SI PROVOCARILE UE

NICOLAE ANASTASIU

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Tratatul de aderare a României la Uniunea Europeană a fost semnat!

Un moment demult asteptat și o speranta – a multora dintre noi – au fost împlinite. Portile Occidentului s-au deschis și pentru noi: cu regulile lui clare, dar severe, cu economia de piata în plina diversificare și cu un tutore intransigent: *legea profitului*. Cu alte cuvinte, un lant solid: *scopuri precise - mijloace adevărate - procesari rapide și efecte de calitate* cu valoare de piata sau utilitate socială, patrimonială. Un lant solid în mijlocul unei „poieni” în permanenta verde. Altfel spus, pentru semenii nostri, o *dezvoltare durabilă*.

Vom intra într-o casa rostuită, decorată cu obiecte traditionale și instalatii electronice de ultima generație. O strategie clara în care există prioritati precise și, dacă vom asculta cu atenție, o să auzim o „limba nouă” și un „limbaj nou”, greu de patruns, chiar dacă vorbim engleză, franceză, germană sau italiană. Este vorba de *normele, procedurile și legile* pe care Consiliul UE le-a stabilit, deja.

La poarta UE, comunitatea geologică – ca și multe alte comunități profesionale, medicale, economice, ale științelor exakte, ingineresti, sociologice – își cauta locul și încearcă să-si revendice identitatea și tradiția. În România, peste un an vom aniverza centenarul Institutului Geologic al României și am sărbatorit, deja, cu 10 ani în urma, centenarul Catedrei de Mineralogie. De asemenea, sunt 75 de ani de când s-a constituit Societatea Geologică a României.

Rasfoind însă anuarul profesiilor din UE și urmarind identificarea domeniilor care, cel puțin, nouă, geologilor, ne erau familiare – *paleontologie, petrografie, stratigrafie, zacaminte, cartografie, prospectiune, explorare* – o să descoperim, cu insatisfacție, lipsa lor. Aceasta surpriza este generată de perceptia noastră istorică, iar contactul brutal cu noua realitate devine o provocare a capacitatii noastre de adaptare, din mers, adică de reorganizare, restrukturare și diversificare, de respectare a noilor standarde într-un cadru în care competiția – succesul celui mai tare și celui mai bun – este unul din motoare, iar motivația lucrului facut – utilitatea lui – este un altul .

Ce a fost cu GEOLOGIA în România ultimilor 50 de ani stim cu totii, cu cele bune și cele rele...

Ce va fi și cum va fi cu GEOLOGIA în viitorii ani, nu mai stim: zonele în care ea s-a facut simtita, institutiile și sectoarele în care ea a avut un cuvânt de spus – în învățământ, în cercetare, în industria extractivă, extrem de complexă, dar și de stufoasa în anii ce-au trecut – vor sta cu totul altfel.

O succinta trecere în revista a provocarilor pe care UE ni le pune în față în legătura cu ÎNVATAMÂNTUL GEOLOGIC, CERCETAREA STIINTIFICA și ORGANIZAȚIILE NEGUVERNAMENTALE aferente o facem în continuare.



Ministerul Educatiei si Cercetarii, cel care a dezbatut si negociat aquis-ul comunitar în domeniu a început sa-si faca temele. Ordonantele si Hotărârile de Guvern emise au deja putere de lege. Din toamna anului 2005, sistemul de învățământ se transforma.

Domeniile si specializarile din învățământul superior – compatibile cu cele europene (în numar de 80-90) s-au redus la 250 (de la cca. 650) si s-au stabilit, pentru fiecare dintre ele, *cicluri* specifice de *învățământ continuu*, care cuprind *licenta*, *masterul* si *doctoratul*, în doua forme adecvate gradului de complexitate: 3-2-3 sau 4-1-3 (ani de studiu finalizati prin obtinerea licentei (diplomei în....), a titlului de Master of Science (MSc) si, respectiv, al celui de Doctor în Stiinte (PhD).

Aria curriculara a GEOLOGIEI se va regasi în doua domenii fundamentale:

1. Domeniul Stiintele naturii, cu specializarile Biologie, Geografie, **Geologie-180cr** (Univ. din Cluj si Iasi), **Geofizica-240cr** (Univ. din Bucuresti), Stiinta mediului si
2. Domeniul Stiinte ingineresti, cu specializarile: **Geologie tehnica-240cr** (Univ. din Bucuresti), **Geologie miniera-240cr**(Univ. din Petrosani), **Geologia petrolului-240cr** (Univ. din Ploiesti).

O a doua treapta a reformei vizeaza introducerea *creditelor transferabile* (activitati desfasurate de un student la fiecare din disciplinele planului de învățământ) odata cu *compatibilizarea* programelor de invățământ prin reducerea disciplinelor de studiu la un semestru si a numarului de ore afectate la $2c+2lp$. Aceasta formula va permite *echivalarea* diplomelor între oricare dintre tarile UE si, astfel, deschiderea pietei muncii pentru orice absolvent care a cumulat creditele obligatorii domeniului ales.

Procesul de învățământ va fi dublat, prin profilul si cerintele disciplinelor si prin criteriile de acordare a licentei, de *activitatile de cercetare stiintifica* – interdisciplinara sau interuniversitara - derive din grant-urile (proiectele) câștigate de profesorii îndrumatori.

Evaluarea si asigurarea calitatii la nivelul Institutiei de învățământ superior va fi ordonata prin legi si norme specifice, care vizeaza monitorizarea anuala a fiecarui departament sau unitate de învățământ, monitorizarea periodica a activitatii fiecarui cadru didactic si, respectiv, evolutiei fiecarui program de studiu (disciplina). Standarde clare vor impune: existenta în fiecare unitate si institutie de invățământ a unor strategii si proceduri pentru asigurarea calitatii, a revizuirii periodice a programelor si activitatilor desfasurate; evaluarea rezultatelor învățarii pe baza unor criterii obiective si transparente; evaluarea periodica a calitatii corpului profesional; asigurarea resurselor adecvate pentru învatare, asigurarea transparentei informatiilor de interes public cu privire la programele de studii, certificatele si diplomele, respectiv calificarile oferite.

Programe de tip CEEPUS, ERASMUS, SOCRATES asigura mobilitatile universitare: schimburile de profesori si studenti între universitatile partenere din Europa, specializarile profesorilor, integrarea lor în programe comunitare de educatie si cercetare.

O alta componenta a analizei pe care o propunem este CERCETAREA STIINTIFICA.

Cercetarea geologica – în sensul ei fundamental si, partial, aplicativ – este în prezent acoperita în cadrul Academiei Române, la Institutul de Geodinamica, în cadrul Institutelor Nationale de Cercetare-Dezvoltare (INCD), la IGR si GEOECOMAR si prin proiectele lansate de MEC-CNCSIS la Universitatile din Bucuresti, Cluj si Iasi. Finantarea este partial bugetara, partial contractuala, iar tematica nu se inscrie într-o strategie clara. Ea reflecta, adesea, afinitatile specialistilor fata de subiecte alese conjectural.



O analiza corecta în spiritul tendintelor europene ar trebui sa raspunda la un sir de întrebări constituite într-un lant de conexiuni: ce institutii guverneaza si administreaza pârghiile cercetarii geologice în România, care sunt legaturile lor cu institutii similare din Europa, care sunt resursele pe care se bazeaza cu privire la documentare si echipamente performante (deci, infrastructura operationala, resursa umana – specializarile si vîrstă cercetatorilor), în ce tip de programe se înscrie tematica curenta a acestor institutii, care sunt beneficiarii proiectelor si cum sunt disseminate rezultatele cercetarilor (prin publicatii în reviste de nivel european, prin colocvii si simpozioane, participari la congrese internationale), în fine, la ce tip de evaluare au fost supuse proiectele realizate si care a fost, în final, utilitatea lor.

Intr-o *economie bazata pe cunoastere* – asa cum UE o declara – astfel de conexiuni conditioneaza competititia dintre cercetatori si asigura calitatea investitiilor. In intentia de a se constituie o *arie europeana a cercetarii*, la 11 martie 2005 a fost lansata **Carta Europeană a Cercetarii si Recrutarii de cercetatori**. Organizatiile neguvernamentale de profil au schitat, deja, statutul „eurogeologist”-ului.

Programul de cercetare cu cel mai mare impact si cu cele mai multe fonduri este Programul cadru 6 (PC6) cu 205 mil. EURO si urmeaza a fi lansat Programul cadru 7 (PC7) cu 39 mld. EURO, din care 2.5 mld. EURO pentru Energie si 2.2 mld. EURO pt. Mediu). Agentia CORDIS (Community Research & Development Information Service) defineste prin aceste programe prioritatile tematicale ale UE :

- Sisteme de energie durabila (Schimbari globale si ecosisteme), Ciclul apei, Biodiversitate, Mecanisme de desertificare si dezastre naturale, Managementul zonelor costiere, Schimbari climatice, Cercetari complementare. Lor li se mai adauga (în special, prin PC7):
 - Sanatate;
 - Alimentatie, agricultura si biotehnologie;
 - Informatica si tehnologia comunicarii;
 - Nanostiente si nanotehnologii;
 - Mediu (aici fiind inclusa si problema resurselor, inchiderile de mine, haldele);
 - Transport;
 - Stiinte umaniste si socio-economice;
 - Securitate si Spatiu extraterestru.

Unde este locul nostru, al geologilor si al geologiei traditionale?

Este evident ca trebuie sa identificam noi cai de apropiere de spiritul acestor programe, sa dovedim flexibilitate si adaptabilitate la tendintele conturate. Intr-o tematica asa de diversa, proiectele noastre au putine brese (mediu, energie, schimbari globale...).

O raza de speranta – pentru conservatori – apare atunci când urmarim «diseminarea rezultatelor cercetarii » si rolul activ, mereu activ, pe care îl au asociatiile profesionale de tip ONG (IUGS, EGGA,IMA, IAS, IAH, IAP....).

Uniunea Internationala a Stiintelor Geologice (IUGS) de ani buni, cu o consecventa care trebuie apreciata, ne ofera prin fiecare Congres International pe care-l organizeaza numeroase solutii. Dezbaterile în simpozioane si colocvii periodice releva rolul pe care geologii si institutiile care-i grupeaza îl au în mai buna înțelegere a proceselor geologice.



În vara anului 2004 a avut loc la Florenta al 32-lea Congres International de Geologie. Tematica sa – cuprinsă în volumele care comentau evenimentul – este variată și surprinde direct spiritul Congresului și „privirile spre viitor” pe care acesta le arunca: „Din zona Mediteranei spre renasterea geologică globală”. Principalele sale obiective s-au identificat cu ideea de „abordare alternativă” și de adaptare a geologilor la noile cerinte ale societății în care trăim. Câteva din dezideratele acestei prestigioase manifestări științifice ne apar ca foarte semnificative:

- Renasterea Geologiei prin proiecția unei imagini pozitive a geostiintelor pentru a demonstra cum servesc ele societății și drepturile omului;
- Cunoașterea și înțelegerea Pamântului prin popularizarea și diseminarea progreselor înregistrate de geologi în înțelegerea proceselor de bază care au loc în interiorul și la suprafața Pamântului;
- Strânsa legătura dintre geostiintele fundamentale și cele aplicative pentru o mai bună perceptie a proceselor geologice legate de :
 - cauzele hazardelor naturale și antropice;
 - predictia hazardelor și soluțiile practice propuse pentru prevenire și combatere;
 - studiul resurselor naturale și gasirea unor modalități de valorificare compatibile cu protecția mediului;
- Cooperarea științifică internațională pentru rezolvarea problemelor ambientale;
- Conservarea mostenirii culturale prin adoptarea unei strategii noi, capabile să contribuie prin metodologii și tehnici geologice de căutare și corelare (Geo-arheologice).

Spiritul Florentei renascentiste, dominat de Leonardo da Vinci, Dante, Petrarca, Michelangelo, se constituie lent într-o provocare a celor ce astăzi încearcă să schiteze viitorul.

Si dacă Renasterea ne-a lăsat ceea ce stim cu totii, atunci de ce nu, și noi, să primim provocările EUROPEI și să le raspundem cu încredere și optimism față de Tânără generație pe care școala geologică o formează, iar comunitatea geologică o adoptă și o lansează. Teritoriul nostru, spațiul geologic românesc, ramâne plin de surpirze și tinde, deja, să devină o „arie europeană de cercetare”, gata să ofere probe aparaturii de vârf care așteaptă în laboratoarele occidentale. Tineri masteranzi sau doctoranzi din toate Universitățile românești lucrează în ele. Tineri geologi, specialisti în geologia petrolului, geologia carbunilor, a pietrelor și metalelor prețioase, a rocilor decorative, în probleme de protecția mediului și hidrogeologie sunt angajați de numeroase companii străine. Si atunci? Sa nu ne mai temem, să acceptăm competiția și să ne lansăm în noile proiecte europene.



BECOMING A EUROPEAN GEOLOGIST, “EURGEOL”; LOOKING TO THE FUTURE

GARY O'CONNOR, CECILIA SZENTESY
Rosia Montana Gold Corporation S.A.

The creation of the European Economic Union or more simply the EU (originally formed as the Coal and Steel Marketing Union) has created the single world's largest economic market encompassing more than 400 million people and growing. The EU is now the world's largest consumer of raw materials with consumption still growing at some 3-4% a year. The EU has managed to break down borders and create unprecedented wealth and prosperity as well as opportunity for a large percentage of the world's population. It has created markets and freedom of movement never previously imagined. Within this frame work the EU has also managed to bring with it high standards and levels of technological and professional compliance.

Within the EU technical and professional standards have been extended to cover the “geosciences” under the auspices of EU directives 89/48/EEC and 92/51/EEC. An EU wide professional body has been formed to ensure standards and promote across border recognition and movement of professionals. The professional association termed the European Federation of Geologists (EFG, www.eurogeologist.de) has been formed as the blanket association to cover and ensure cross-border compliance of the current 18 National member associations with some 275,000 members as well as 4 observer members, 3 stand-by members and 1 associate member associations (Unfortunately Romania is not mentioned by the EFG). The body has developed and regulates the professional position of “EurGeol”. The professional association and regulated role is closely linked to that of the “competent person” (“The qualified person concept”, EurGeol Dr. I. F. Fuentes, Feb. 2004) as used in other internationally recognized profession geological associations including those defined by the JORC code in Australasia and NI43-101 law in North America and in turn recognizes the professional standards and persons from these organizations. One of the newest and most successful National professional bodies, the Institute of Geologists of Ireland, was only formed in 1999 but already is the second largest membership in Europe with statutes and codes of conduct and ethics based on those of the Professional Association of Geologists of Canada as well as that of the EFG. There should be no surprises as to why the economy of Ireland is the fastest growing in the EU along with the number of “EurGeol’s. Irish professional geologists are recognized worldwide (www.igi.ie).

The EFG ensures that the standard of all National professional associations meets or betters those of the EFG which includes a code of ethics, a code of conduct, statutes, a standard means of sponsoring new members, vetting of new members including a requirement to meet certain experience levels (8 years) as well as present examples of professional competence (Professional Practice Report of 3000-4000 words, or at least 2 Academic papers). Selected independent committees first vet and others then interview candidates to ensure competence and professionalism. Also the EFG requires members and associations to have a system of monitoring and disciplining members and associations must show evidence for having enforced this. Additionally members are required to have a program in place for their own ongoing professional development. Associations should be independent, non-political and self governing. Plus members must be aware and conversant with health and safety policies, programs and issues in their profession. (The European Geologist Professional Title, EurGeol J.A. Clifford)



The Profession European Geologist or “EurGeol” must be dedicated to ensuring a common European policy in regards to the responsible use of the Earth’s natural resources, the avoidance of environmental pollution, land use and environmental protection. Emphasis is given to the free movement of geologists in Europe. Geologists are the experts when it comes to discovering the raw materials that underpin and sustain modern life, such as oil and gas, base and precious metal ores and construction material. They are the specialists in the earth’s structure and tectonics, including earthquake and volcanic predictions as well as other related phenomena (e.g. Tsumanis). Hydrogeologists and environmental geologists are the foremost experts on most water resource and environmentally linked issues.

Romania has the unique chance to professionalize the position and role of the geologist in society by quickly adapting EU directives in regards to the professional position of geologist. The Society of Geologists of Romania has the potential to form the platform for such a professional association by the adoption of the appropriate code of ethics, codes of conduct as well as statutes and ensuring the monitoring and enforcement of the codes on members. Romania, as one of the few truly natural resource endowed (minerals, oil, gas, water, construction materials) countries in Europe (Others include Sweden, Finland and Spain with possible member countries Bulgaria and Turkey, with Norway not in the EU) to take a lead role in sustaining the professional level of geologists in European society for the future.



SESSIONS'S PAPERS



THE QUARTZ-SANDY OCCURRENCES FROM ROMANIA – AN INVENTORY AND PROPERTIES FOR INDUSTRY

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Keywords: *quartz-sandy, grain-size, sphericity index, roundness index, mineralogical and petrographical analyses*

The aim of the study was the identification and the research works about quartz-sandy formations from Romania as the utility in glass industry and drilling operations for hydrocarbon wells, in conformity with national and international standards. In this order, we have selected 285 sandy occurrences from all geological units of Romania: the East Carpathians, the South Carpathians, the Apuseni Mountains, Transylvania Basin, Pannonic Basin, Moldavia Basin, Getic Basin, Romanian Plain, and Dobrogea. Inside of this units, the distribution of the quartz-sandy occurrences is very irregular, and their qualities high variable. The stratigraphic interval where quartz-sandy occurrences is from Eocene to Quaternary. The oldest formations are weak-consolidated, or affected by strongly decimented processes in their diagenetic evolution (as in the case of kliwa oligocene “sand”), and the youngest formations (Pliocene and Quaternary) include unconsolidated deposits, as mobile, so easy to exploration works.

The total thickness of sandy formations for each occurrence are between 0.50 – 10.15 meters, and the quality of sand, generally, is not very close to most of the requirements for drilling operations in the hydrocarbon wells. In the class grain-size intervals, the quartz content is 50 – 95 %, the sphericity index 0.2 – 0.8, and the roundness index 0.2 – 0.8.

In spite of the great number of the identified occurrences, only 35 of them have sandy formations with more than 90% quartz content, and also, the properties are close to requirements in the standards. The quartz-sandy occurrences selected by us in order to carried out detailed evaluations to reveal their properties for drilling operations in hydrocarbon wells are placed in the East Carpathians, the Apuseni Mountains, Pannonic Basin, Transilvania Basin, and the Moldavia Platform areas.

The analysis – grain-size, shape, mineralogical, chemical, and technological (leaching analysis, acid solubility, crush-resistance) – was carried out on 40 samples, and was revealed textural parameters c, md, m, so, ro, sf, mineralogical composition, also their technological quality of sand (in compliance with reference standards api 56/1986 and 58/1986).

By lithological point of view, the examined occurrences are siliciclastic deposits with different stages of consolidation, as following:

- Strong-cemented sandstones – as occurrences from Patarlagele area (valea Sibiciului, Colti, Alunis), and Suceava area (Plesa, Manastirea Humor);
- Decemented sandstones – as occurrences from Valeni area (Gura Vitioarei, Copaceni), Patârlagele area (valea Sibiciului, Crivineni, valea Rea), Suceava area (Paltinoasa);
- Weak cemented sand – as occurrences from Botosani (Hudesti, Miorcani), Faget area (Faget Scaune si Jupanesti), Aghires area, and Fagetul Ierii area.



The age of this sandy deposits is Eocene for Fagetul Ierii sands (Apuseni Mountains), Oligocene for Teleajen valley sands (Valeni area), Buzau valley sands (Patarlagele area), Humor valley sands (Humor area), Nadasu valley (Aghires quarry), Miocene for Moldavia platform sands, and Pliocene for Panonic basin sands.

In all cases, the petrotypes correspond to quartz-rocks (i.e. rocks with more than 90% quartz). In sequences, at outcrop scale, these types of rocks are associated, as interbeded, with fine grain-size rocks (i.e. siltites, shales and slates, marls), and even carbonate rocks (as in the case of the Manastirea Humor area). Arenite vs. other grain-size fractions ratio is variable between 9/1 – the maximum values (>75%) in cases of Humor area (Plesa, Paltinoasa), Patarlagele area, and Valeni area (Copaceni, Gura Vitioarei) – to 6/4 – the minimum values in the case of Faget Timisoara area (Faget Scaune quarry).

The qualities of the sand form the selected and analyzed occurrences are as following:

The grain-size analysis was shown, for occurrences in exploration technological processes, a high percent of very fine arenites (under 0.125 mm class), fine arenite (0.1 – 0.2 mm), and medium arenite (0.2 - 0.4 mm, and 0.3 – 0.6 mm). Coarse arenites occurs in Paltinoasa-Iliesti area, but in this case the deposits are cemented. Often, arenitic class (named “usefully”) is associated with silty-lutitic class and/or carbonatic (“poor” class). In the analyzed samples “u” vs. “p” ratio is 78.98 – 98.97 %, generally 88 – 92%. The maximum content of “u” (over 90%) is registered by samples from Aghiresi, Hudesti, and Miorcani areas. For each sort (e.g. preferred grain-size class) the obtaining randament is 30.82 – 90.39%, but generally 62 – 73% (for 0.1 – 0.2 mm sort), and 39 – 69 % (for 0.2 – 0.4 mm, and 0.3 – 0.6 mm sorts). The maximum randament is for samples collected from Hudesti, Miorcani, Aghires, Fagetul Ierii areas, and in part for samples collected from Valeni, and Plesa areas.

The grain-size classes and sorts required by API standards are:

0.6 – 1.8 for 16/30 sort; 0.4 – 0.8 for 20/40 sort; 0.3 – 0.6 for 30/50 sort.

From our analyzed samples, the most close are which are collected from Aghires, Miorcani, Faget Scaune, Hudesti and Paltinoasa areas.

Morphometrical analysis, evaluation of roundness index – ro, and sphericity index – sf (Krumbein & Sloss, 1956), for global sample and for different sorts, was shown a large variations. In this reason, for all samples analyzed by this study, indexes $ro = 0.1 - 1$, and $sf = 0.1 - 1$, so, the whole interval. A direct proportional relationship was detected on increase of ro (sf) as increase grain-size class of quartz grains. The grains with more than 0.6 ro index are variable between 30 – 60 % per sample, and maximum values was determined on samples from Hudesti, Miorcani areas, and in part from Valeni (Gura Vitioarei), Patarlagele (valea Rea) and Aghires areas.

The api required values for morphological parameters are:

- Ro = min. 0.6 for 16/30, 20/40, and 30/50 sorts;
- Sf = min. 0.6 for 16/30, 20/40, and 30/50 sorts.

The closest samples analyzed are collected from Hudesti, Miorcani, Faget Scaune, Gura Vitioarei and Aghires areas.

The mineralogical and petrographical analyses evidenced mineralogical assemblages domined by quartz (over than 90 %), then feldspars, micas (biotite and muscovite), glauconite, heavy minerals, calcite, also bioclast and lithic fragments. Quartz grains are both mono-, and polycrystalline, with or without right extinction, showing many inclusions (frequently solid, but some time as fluid state), oriented (corresponding systems) or unoriented microjoints, corroded



surfaces, overgrowths (in Kliwa-type sandstones).

Requirements by standards, for quartz content are min. 90% (API 56/1986), and min. 96% (API 58/1986). In our case, the closest samples are collected from Valeni, Patarlagele, Hudesti, Miorcani, Faget-Scaune, Jupanesti, Aghires areas, and in part from Humor area (Paltinoasa, and Plesa perimeters)

Our study show that from the all identified quartz-sandy occurrences, both in exploration technological processes as well as in outcrops state, the most close, in respect to API standards requirements (as international reference system – Hulliburton sorts), are only as the follow occurrences:

- in the East Carpathians – Valeni (Gura Vitioarei, Copaceni), and Patarlagele (Valea Rea) areas;
- Panonic Basin – Faget Scaune area;
- Transilvania Basin – Aghires quarry.

Concluding, only by technological processes, as secondary, the API standard requirements could be realizable.

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GEOCHEMISTRY OF THE MUSCOVITE FROM THE CONTU-NEGOVANU PEGMATITES (LOTRU-CIBIN MTS.)

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1. Introduction

Micas and especially muscovite are fair indicators of changes occurring in both tectonics and chemical characteristics of the different episodes of the pegmatitogenesis. Although very complex, this process may be synthetically represented by a series of evolving stages, which considered independently constitute distinct petrogenetic types of pegmatites. Muscovite is the third abundant mineral in the Contu-Negovanu pegmatites, after quartz and feldspars.

Previous research carried out in the Contu-Negovanu pegmatite field since 1968 revealed the presence in the area of two different types of pegmatites: (i) feldspar ± muscovite pegmatites (**FPs**) and (ii) albite–spodumene pegmatites or Li – rich pegmatites (**LPs**). For each of them, the mineralogical composition, as well as the bulk chemistry features and geochemical signatures suggest the affiliation to two distinct pegmatite classes, according to Cerný's (1992) systematic classification of pegmatites: (i) muscovite class and (ii) rare-element class, respectively. More specifically, the **FPs** are of ceramic and mica-bearing type, being barren or poorly mineralised (Li, Be, Ti, Nb, Ta, U, Th, Y, REE). The **LPs** are assigned to one of the five subdivision types of the rare-element class: the albite–spodumene type, characterised by minor to extensive mineralisation (Li, Rb, Cs, Be, Ga, Y, REE, Sn, Ti, U, Th, Hf, Nb, Ta), with typical substantial Li and variable rare-alkali fractionation (Cerný, 1992). Therefore, in terms of pegmatitogenetic evolution, the **FPs** from the Contu-Negovanu field represent a „primitive” stage (Gordyenko and Leonova, 1972), whereas the **LPs**, typical for this area, constitute a much evolved stage, marked by progressive differentiation and also by extensive albitization processes.

2. Geological setting

Contu-Negovanu pegmatite field is located in the Lotru-Cibin Mts., in the central-west part of the Southern Carpathians. The investigated area is situated on the slopes of Contu valley - a tributary to Sadu river - and up towards Negovanu peak. The pegmatite bodies are hosted by mesometamorphic rocks belonging to Sebes-Lotru Group, which is the most extensive petrographic entity of the Getic Realm. The Sebes-Lotru sequence represents a crustal fragment of Cadomian-Caledonian age, part of an active continental margin or an island arc near a continent. Possibly, around 430 M.a., this terrane collided with an ahead block, being consequently involved in a subduction process (Balintoni *et al.*, 2004). The lithology of the Sebes-Lotru Group is generally represented by a lower



migmatitic complex and upper kyanite–staurolite–bearing micaschists alternating with biotite paragneisses; metaultramafites are also present (Berza *et al.*, 1994). The entire region has polycyclic metamorphic features, the best known being those related to the last metamorphic event, developed in the amphibolite facies, within the kyanite–staurolite zone (Hărtopanu, 1978, 1986, 1988). The pegmatite bodies display an incomplete zoning, the best represented being the aplite and the intermediate zones. The latter shows no differentiation of the quartz core (Maieru *et al.*, 1968). The genesis of the Contu-Negovanu pegmatites is considered to be metamorphic, as they seem to have originated mainly, if not entirely, as a result of segregation within a fluid under preanatetic conditions (Sabau *et al.*, 1987).

3. Analytical methods

In X-ray fluorescence technique, bulk analyses were performed with a Philips PW 2400 X-ray spectrometer, using the analytical procedure called „oxiquant”. Seventy-two natural rocks and clays were used to determine the calibration curves of the pertinent elements.

EMPA were carried out with a JEOL JXA-8900 instrument, using an operating current of 20 nA and accelerating voltage of 20 kV. X-ray intensities of the alkalis, the minor (Ti and Mn) and the major elements were counted for 5s, 40s and 60s, respectively. In order to minimize losses of Na and K, the beam diameter was expanded to 10 mm. Components were standardised using natural minerals, glasses of natural rocks (Jochum *et al.*, 2000) and synthetic oxide compounds. The results were corrected using the ZAF procedure (Reed, 1996).

The ICP-MS investigations were performed with a Perkin Elmer/Sciex ELAN 6000 ICP-MS (quadrupole mass spectrometer). Measurements of element concentrations were performed using as internal standards Ru-Re (10 ng/ml) to minimize drift effects and two calibration solutions (high purity chemical reagents). A batch of 5-7 samples was bracketed by two calibrations procedures. Accuracy and precision of determinations were checked with certified reference materials (CRM) (Govindaraju, 1994; Dulski, 2000).

4. Muscovite characteristics. Samples.

The samples have been collected from both **FPs** and **LPs**, a special interest being given to the micaschist contact zone. The former are characterized by the typical mineral assemblage: Qtz + K-feld + Plg + Ms ± Bt ± Tur ± Grt and present also small amounts of staurolite, kyanite, apatite, ilmenite, rutile *etc.* The latter consist of: Ab (cleavelandite) + Sp + Qtz + Ms and contain also subordinate amounts of microcline, biotite, garnet, tourmaline, triphylline–lithiophyllite, heterosite–purpurite, amblygonite–montebrasite, tavorite, uraniferous magnocolumbite, huréaulite *etc.* (Sabau *et al.*, 1987; Hann, 1987; Murariu, 2001). The muscovites from the two pegmatite types are essentially different as far as physical features, chemical composition and trace element distribution (Li, Rb, Cs, Nb, Ta *etc.*) are concerned.

The **FPs** muscovite occurs as rather large „books” (up to 20 cm²) of white–yellowish colour. It is usually a result of biotite muscovitization and therefore, it preserves minute hematite inclusions on the cleavage sheets. This type of muscovite can be often found associated with the biotite from which it originated. In the **LPs**, muscovite is greenish, perfectly transparent, with fresh aspect and glassy–nacred luster. It is related to the albitization stages of pegmatitogenesis and is considered to be an indicator of the spodumene mineralization.

The present study is based on more than 70 muscovite analyses, among which 41 were investigated by EMPA, 8 samples by XRF and 8 samples by ICP-MS.



5. Muscovite geochemistry

5.1. Major elements

The chemical analyses of the muscovite samples and the unit formulae on the basis of 22 (O, OH, F) are given in Table 1.

In all the analysed samples, Si^{4+} has similar or slightly higher values than the ideal value (i.e., 6 apfu)(Fig.1). The samples with $\text{Si}^{4+}>6$ apfu may represent intermediate terms of the muscovite-aluminoceladonite series (Rieder, 1998), or muscovite end-member with micronic quartz inclusions (Buzgar, 2000). Nevertheless, the existence of some EMPA investigated samples with $\text{Si}^{4+}>6$ apfu shows that the samples represent intermediate terms of the muscovite-aluminoceladonite solid solution.

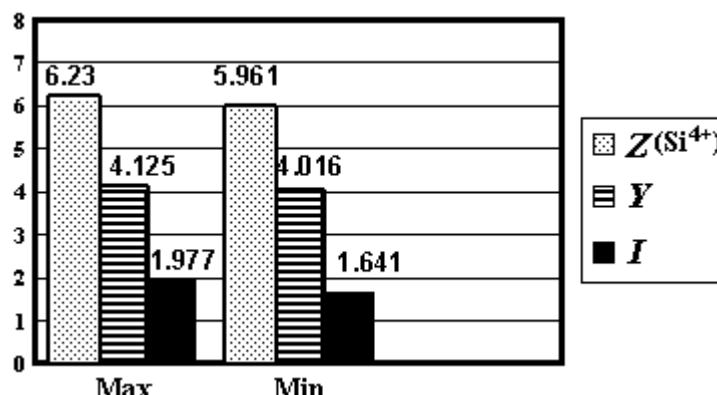


Fig. 1. Tetrahedral (only Si^{4+}), octahedral (Y) and 12-fold coordinated (I) site occupation in the Contu-Negovanu muscovites.

The octahedral site occupation is close to the ideal 4 apfu characteristic to muscovite and the higher values probably belong to the samples deviating towards the trioctahedral K-micas composition (Rieder, 1998). The exchange vectors ${}^{\text{VI}}\text{Al}_{2-1}(\text{Mg},\text{Fe})_3$ for the biotite substitution and ${}^{\text{VI}}\text{Al}_{1-1}{}^{\text{IV}}\text{Al}_1(\text{Mg},\text{Fe})\text{Si}$ for the Tschermark substitution indicate the fact that the deviation towards the trioctahedral composition is less effective in the LPs muscovites, whereas the Tschermark substitution is dominant (Fig. 2).

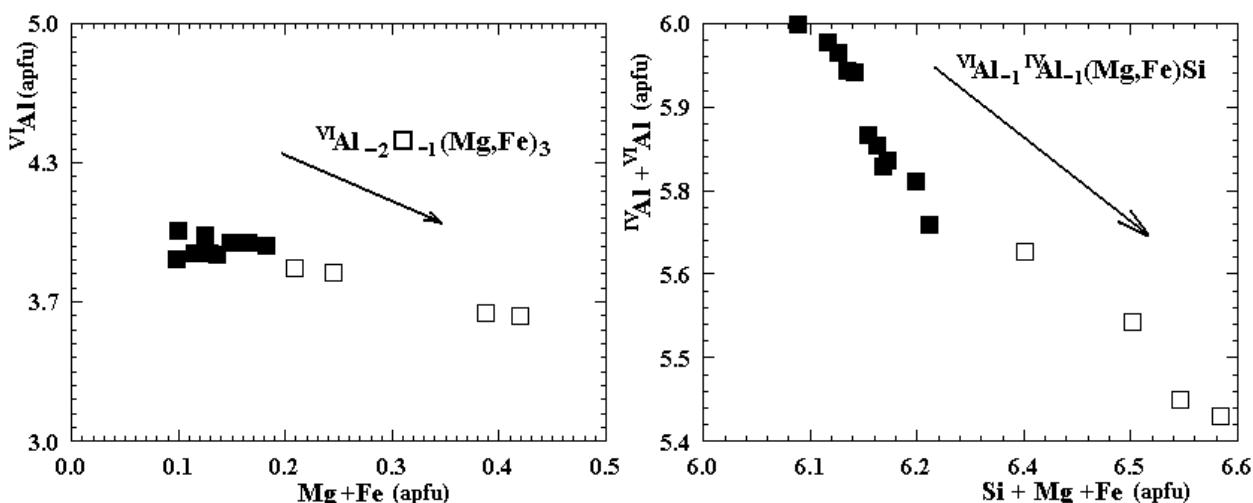


Fig. 2. ${}^{\text{VI}}\text{Al}$ vs. $(\text{Mg}+\text{Fe})$ and $({}^{\text{IV}}\text{Al}+{}^{\text{VI}}\text{Al})$ vs. $(\text{Si}+\text{Mg}+\text{Fe})$ diagrams and the corresponding exchange vectors for the muscovite-biotite and muscovite-aluminoceladonite series, respectively; filled squares: LPs muscovite; open squares: FPs muscovite.

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Table I. Chemical composition (wt %) of muscovites from Comtu-Negovaru pegmatites

Oxides (wt%)	Samples													
	Ms61	Ms531	Ms551	Ms512	Ms832	Ms841	Ms881	Ms88	Ms911	Ms941	Ms952	Ms95B ²	Ms962	Ms96B ²
SiO ₂	45.11	47.29	45.23	47.28	47.69	44.77	44.41	46.88	44.12	44.87	44.39	45.27	44.70	45.65
TiO ₂	0.608	0.02	0.017	0.52	0.509	0.026	0.037	0.002	0.068	0.654	0.02	0.073	0.105	0.064
Al ₂ O ₃	33.75	37.20	36.51	35.84	37.28	36.18	36.19	38.84	35.75	33.80	36.29	37.88	37.99	38.93
Fe ₂ O ₃	1.842*	0.84*	1.687*	-	0.861*	1.29*	-	1.305*	1.81*	1.184*	-	-	-	-
FeO	-	-	-	0.94*	0.818*	-	-	0.898*	-	-	1.201*	1.188*	1.153*	1.054*
MnO	0.008	0.03	0.048	0.002	0.011	0.011	0.041	0.024	0.039	0.012	0.026	0.021	0.038	0.019
MgO	1.133	0.07	0	0.73	0.624	0.14	0	0.005	0.003	0.979	0.028	0.075	0.237	0.197
CaO	0.024	0.06	0.02	0.008	0.024	0.032	0.014	0.026	0.093	0.03	0.012	0.026	0.072	0.078
Na ₂ O	1.05	1.09	0.58	0.729	1.658	1.13	0.52	0.361	0.68	0.75	0.79	0.445	0.396	0.631
K ₂ O	9.27	10.06	10.49	8.646	8.112	9.34	10.20	9.82	10.05	9.80	9.74	9.94	10.05	9.77
P ₂ O ₅	0.014	0.02	0.02	0.011	0.03	0.041	0.062	0.021	0.036	0.014	0.044	0.021	0	0.065
Li ²	0.008	0.048	0.071	-	-	0.024	0.034	-	0.042	0.005	0.029	-	-	-
Y	4.087	4.016	4.091	4.103	4.082	4.037	4.058	4.101	4.049	4.069	4.058	4.104	4.125	4.116
K	1.616	1.671	1.795	1.454	1.335	1.621	1.776	1.622	1.763	1.713	1.696	1.685	1.709	1.635
Na	0.278	0.275	0.150	0.186	0.414	0.298	0.137	0.090	0.181	0.199	0.209	0.114	0.102	0.160
Ca	0.004	0.008	0.003	0.001	0.003	0.004	0.002	0.003	0.013	0.004	0.001	0.003	0.010	0.00
Rb ³	0.003	0.022	0.017	-	-	0.021	0.037	-	0.020	0.002	0.003	-	-	-
Ba ³	0.019	0.000	0.000	-	-	0.000	0.000	-	0.000	0.023	0.000	-	-	-
I	1.920	1.976	1.965	1.641	1.752	1.944	1.952	1.715	1.977	1.941	1.908	1.802	1.821	1.789



This idea is consistent with the presence of variable aluminoceladonite amounts within the muscovite samples, as stated above. Still, the Tschermak substitution does not increase the octahedral site occupation over 4 apfu (Rieder, 1998) and this odd situation is strongly related to the interlayer site occupation. The closer the I site occupation is to the ideal value, the more the Y site occupation is approaching 4 apfu (Fig. 3). This relation is effective especially in the **LPs** muscovites, suggesting that in the I site there are also H_3O^+ and/or NH_4^+ cations. Ignoring them in the analytical procedure leads to a slight overestimate of the high molecular mass oxides with significant participation, such as Al_2O_3 (Fig. 3). The presence of H_3O^+ and/or NH_4^+ is more likely considering that the **LPs** are formed during the late stages of pegmatitogenesis.

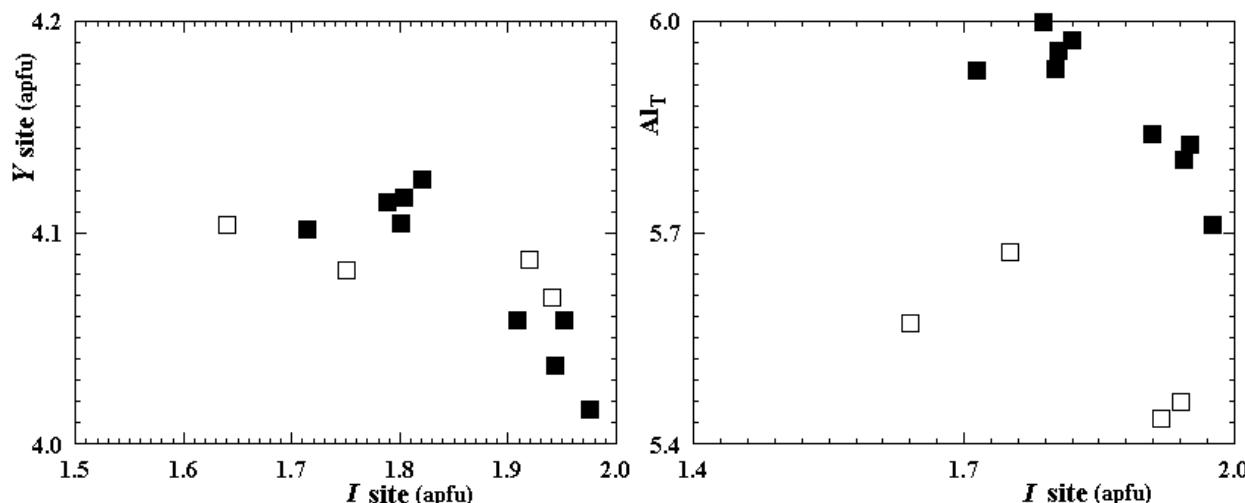


Fig. 3. Y and I site occupation (left) and negative correlation between Al_T and I site occupation in the Contu-Negovanu muscovites (same symbols as in Fig.1).

As previously mentioned, the two muscovite types belonging to **FPs** and **LPs** are formed during different stages, fact which is consistent with their different chemical composition. The **FPs** muscovite presents higher Mg, Fe and Ti contents and consequently lower ^{VI}Al content (Fig. 2 and 4). In fact, this muscovite type, formed by biotite substitution, preserves in its structure higher amounts of Mg, Fe and Ti. During the final stages of pegmatitogenesis, when **LPs** muscovite was formed, Mg and Fe concentrated mainly in the Fe-Mn phosphates and tourmaline.

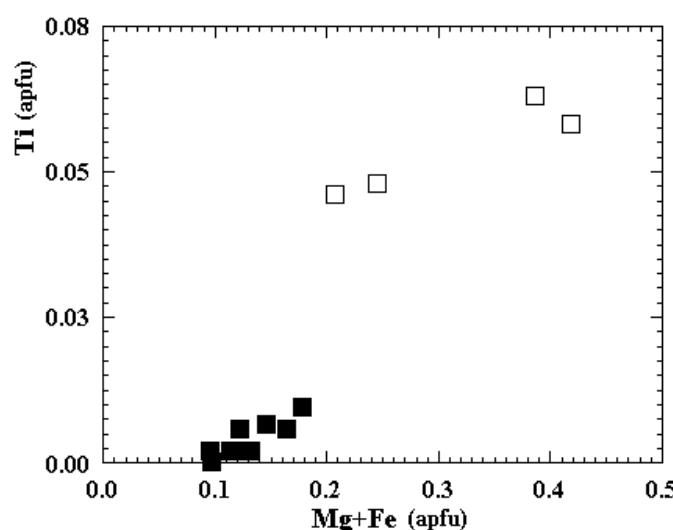


Fig. 4. Ti^{4+} vs. $(Mg+Fe)$ diagram for the Contu-Negovanu muscovites (same symbols as in Fig.1).



As for the normative composition, the two muscovite types are also different. The **FPs** muscovite is characterized by lower amounts of muscovite end-member and higher paragonite end-member (Fig. 5), which indicates an early formation, during a stage thought to have a higher X_{Na} (X_{Na} defined as $\text{Na}/(\text{Na}+\text{K})$).

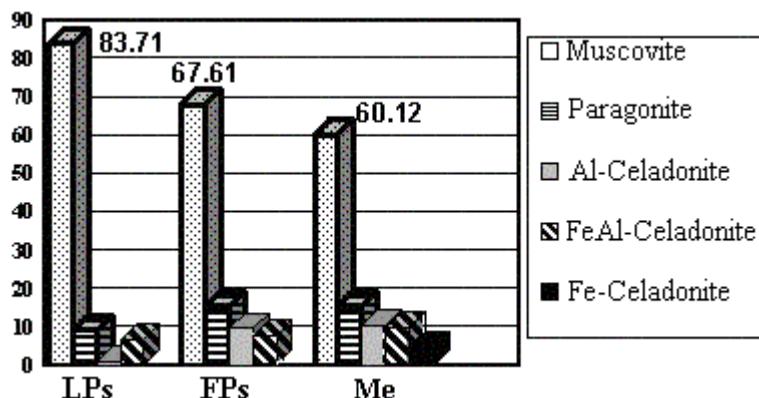


Fig. 5. Normative composition (average values) of the Contu-Negovanu muscovites; Me - muscovite of the metamorphic rocks (Murariu, 1991).

5.2. Minor and trace elements

The minor and trace elements contents of the muscovite samples are given in Table 2. Generally, the minor and trace elements contents of muscovite are indicators of these elements behaviour and their relationship with the peraluminous anatetic fluids and the subsequent mineral phases.

Lithium. Lithium is incompatible with the peralkaline fluids during the early stages of pegmatitogenesis and becomes compatible in the final ones. Thus, the Li compatibility generally increases in the late muscovites. In the Contu-Negovanu samples (Fig. 6), the Li content is lower in the **FPs** muscovites ($41\text{-}72 \mu\text{g}\cdot\text{g}^{-1}$ Li) and becomes significant in the **LPs** muscovites ($210\text{-}618 \mu\text{g}\cdot\text{g}^{-1}$ Li).

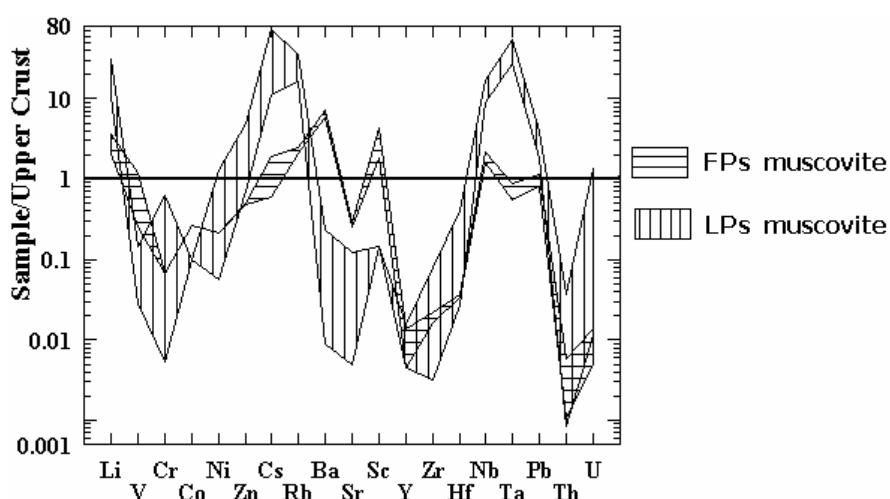


Fig. 6. Spider diagram of the Contu-Negovanu muscovites (upper crust normalization values from Taylor and McLennan, 1985; 1995).

Transition metals. The transition metals (Sc, V, Cr, Ni and Zn) display a strong geochemical affinity for Mg and Fe and therefore, they concentrate in the ferromagnesian silicates. The transition metals contents in the pegmatitic minerals are determined on one hand, by the nature



of the preanatectic protolith and on the other hand, by the compatibility of these elements with the anatectic restite (biotite, amphibole, garnet, zircon, sphene *etc.*). In a metapelitic protolith, the transition metals are compatible with the anatectic restite (biotite, garnet, zircon, allanite) and consequently they have a lower content in the pegmatitic fluid. Regarding the pegmatitic fluid – pegmatitic mineral system, the transition metals are incompatible with the peraluminous pegmatitic fluids and compatible only with the metaaluminous ones (*e.g.*, according to Mahood and Hildreth, 1983, the Sc, Cr and Co partition coefficients are greater than 1 for biotite). That is why in both pegmatite types, where the main K-mica is the muscovite (*i.e.*, the character of the pegmatitic fluid is peraluminous), the transition metals are incompatible and they present low and comparable concentrations (Fig. 6). Scandium, by its capacity to form complex combinations with the volatiles (Gramaccioli *et al.*, 2000), tends to concentrate in the evolved pegmatitic fluids.

Table 2. Minor and trace elements ($\mu\text{g}\cdot\text{g}^{-1}$) of muscovites from Contu-Negovanu pegmatites.

Elements ($\mu\text{g}\cdot\text{g}^{-1}$)	Samples							
	Ms-6	Ms-53	Ms-55	Ms-84	Ms-88	Ms-91	Ms-93	Ms-94
Li	72	423	618	210	290	358	41	253
Sc	25	1.949	0.0	0	0	0	58	0
V	127	3	0	5.96	11.02	11.74	27.04	15.43
Cr	5.68	53	0	4.58	0.07	0.46	0	3.45
Co	0	0	0	0	0	0	4.54	1.66
Ni	9.36	0	56	0	8.12	2.49	0	4.77
Zn	36	71	349	50	139	214	35	80
Rb	278	2455	1843	2254	3879	2125	229	3808
Sr	87	43	1.7	29	2.7	4.7	103	12.1
Y	0.3	0.344	0.0	0.1	0	0	0.1	0
Zr	4.2	15.21	0.6	1.5	0.7	0.9	3.1	0.8
Nb	25.8	106	117	133	201	172	19.1	182
Cs	8.9	142	81	127	305	51	2.8	335
Ba	3259	126	5	98	42	35	3921	36
La	0.089	0.144	0.015	0.14	0.022	0.02	0.278	0.056
Ce	0.051	0.292	0.035	0.27	0.042	0.04	0.173	0.101
Pr	0.005	0.034	0.003	0.034	0.004	0.005	0.042	0.01
Nd	0.043	0.186	0.351	0.164	0.015	0.14	0.203	0.093
Sm	0.03	0.036	0.018	0.032	0.005	0.016	0.059	0.021
Eu	0.368	0.017	0.002	0.066	0.007	0.003	0.47	0.016
Gd	0.008	0.034	0	0.025	0.003	0.002	0.022	0.007
Tb	0.003	0.008	0	0.003	0.001	0.001	0.003	0.001
Dy	0.029	0.008	0.008	0.016	0.002	0.007	0.024	0.009
Ho	0.008	0.012	0.001	0.003	0.001	0.001	0.003	0.001
Er	0.026	0.035	0.002	0.007	0.002	0.003	0.013	0.002
Tm	0.006	0.005	0	0.001	0.001	0.00	0.005	0
Yb	0.125	0.043	0.001	0.009	0.002	0.003	0.349	0.003
Lu	0.012	0.006	0	0.001	0.001	0.001	0.024	0
Hf	0.2	2.391	0.257	0.343	0.281	0.155	0.215	0.212
Ta	0.555	38	45	53	47	27.04	0.876	51
Pb	19.5	66	45	53	42	31	13.8	42
Th	0.011	0.399	0.009	0.105	0.042	0.032	0.062	0.318
U	0.039	3.868	0.032	0.576	0.233	0.535	0.014	0.13
$(\text{La}/\text{Yb})_{\text{N}}$	0.48	2.26	10.11	10.49	7.42	4.49	0.54	12.58
$(\text{Ce}/\text{Yb})_{\text{N}}$	0.11	1.76	9.05	7.76	5.43	3.45	0.80	8.71
La/Sm	1.87	2.52	0.52	2.75	2.77	0.79	2.96	1.68
$(\text{Gd}/\text{Yb})_{\text{N}}$	0.05	0.64	0.00	2.24	1.21	0.54	0.05	1.88
Eu/Eu^*	72.63	1.49	0.00	7.13	5.53	1.62	39.89	4.03



In the Contu-Negovanu muscovite samples, Sc concentration apparently has an opposite behavior, as the **FPs** muscovite shows higher contents than the **LPs** muscovite. This situation is due to the **FPs** muscovite formation, by biotite substitution, and not to a specific compatibility/incompatibility of scandium during the early pegmatitic stages.

Large-ion lithophile elements (Rb, Cs, Ba and Sr). LILE are known to display a specific geochemical affinity for K and Ca (only Sr). In the pegmatitic fluid – anatetic restite system, LILE present different degrees of incompatibility: Rb and Cs are highly incompatible, whereas Sr and Ba are less incompatible, depending on the K-feldspar (for Ba) and plagioclase (for Sr) participation in the anatetic process. Nevertheless, in the pegmatitic minerals it is more likely for Rb and Cs to present higher concentrations than Ba and Sr and thus, the Ba/Sr ratio to be highly over 1, as the plagioclase is less involved in the anatexis than the K-feldspar.

The LILE` incompatibility in the pegmatitic fluid – pegmatitic mineral system is highly depending on both the fluid and the crystallized phase evolution. Rb and Cs are incompatible with the pegmatitic fluids (Nash and Crecraft, 1985; Icenhower and London, 1995, 1996) and this behavior stands for their different concentration degrees in the muscovites resulting from different stages of pegmatitogenesis, the highest values being characteristic to the late ones. In the investigated samples, this geochemical behavior is also noticed, the **LPs** muscovites accommodating the highest Rb and Cs amounts ($1843\text{-}3879 \mu\text{g}\cdot\text{g}^{-1}$ Rb and $50\text{-}335 \mu\text{g}\cdot\text{g}^{-1}$ Cs) (Fig. 6). Ba and Sr have an opposite behavior, as they mainly concentrate in the early pegmatitic stages. In the Contu-Negovanu samples, the highest concentration are typically found in the **FPs** muscovites ($3259\text{-}3921 \mu\text{g}\cdot\text{g}^{-1}$ Ba and $87\text{-}103 \mu\text{g}\cdot\text{g}^{-1}$ Sr) and much lower values, falling in a larger range are characteristic to the **LPs** muscovites ($5\text{-}126 \mu\text{g}\cdot\text{g}^{-1}$ Ba and $1.7\text{-}43 \mu\text{g}\cdot\text{g}^{-1}$ Sr). As compared to Ba, the much lower Sr concentration in the **FPs** muscovites (Fig. 6) could be determined either by an early crystallization of Ca-rich plagioclase (for which Sr shows a high compatibility) or by the accommodation of Sr in the plagioclase of the anatetic restite.

High field strength elements (Y, Zr, Hf, Nb and Ta). HFSE are highly incompatible and therefore they usually concentrate in evolved magmas and pegmatitic fluids (Watson, 1979; Rollinson, 1983). In the pegmatitic fluids resulted by anatetic processes, the incompatibility of these elements must be regarded separately as it is highly variable. In the anatetic fluid – anatetic restite system, Nb and Ta are incompatible and Y, Zr and Hf are “compatible”, as they accumulate in the anatetic restite minerals (zircon, garnet). For this reason, in the anatetic pegmatitic fluids, Nb and Ta concentration is usually higher than that of Y, Zr and Hf. In the pegmatic fluid – pegmatic mineral system, Nb and Ta are highly incompatible and they concentrate in the last stage minerals, being consequently expected to concentrate in the late muscovite, *i.e.* the **LPs** muscovite. This fact is consistent with our analytical data ($106\text{-}201 \mu\text{g}\cdot\text{g}^{-1}$ Nb and $38\text{-}53 \mu\text{g}\cdot\text{g}^{-1}$ Ta) (Fig.6). Y, Zr and Hf are incompatible in the pegmatitic fluid – pegmatitic mineral system, but they also are incompatible in the muscovite structure and thus, they have a constant behavior within its crystal lattice and does not depend on its formation stage (Fig. 6).

Uranium and thorium. U and Th are also highly incompatible elements which accumulate in the final fluids and concentrate mainly in the phosphates. In muscovite, U and Th concentration is low, due to their diadochical incompatibility with the octahedral and interlayer sites’ cations. In the Contu-Negovanu muscovites, U and Th contents are low and variable, with significant, yet not generalized values in the **LPs** muscovite (Fig. 6).

Rare earth elements. REE are incompatible with the pegmatitic fluids and tend to accumulate in the final stage fluids (Shmakin, 1992), when they concentrate in the late minerals as phosphates. In the Contu-Negovanu muscovites, REE have low contents and the upper crust normalized values are much less than 1 (Fig. 7). LREE show no difference between **FPs** and **LPs**



muscovites, whereas Eu presents a highly positive anomaly in the **FPs** muscovite. This Eu anomaly is normal, considering the high Ba content in this type of muscovite. As the Ba content decreases (in the **LPs** muscovite) the amplitude of the Eu anomaly decreases altogether. HREE contents are also low, but with different concentration levels in the two muscovite types. The slight increase of the normalized values in the **FPs** muscovite reflects its formation from biotite which is a stronger HREE concentrator than LREE. In the **LPs** muscovite the upper crust normalized HREE values are rather similar to the LREE ones. Their large range could be due to phosphates formation, which are REE accumulators, producing pegmatitic fluid concentration variations either spatial and/or temporal (different REE diffusion rate).

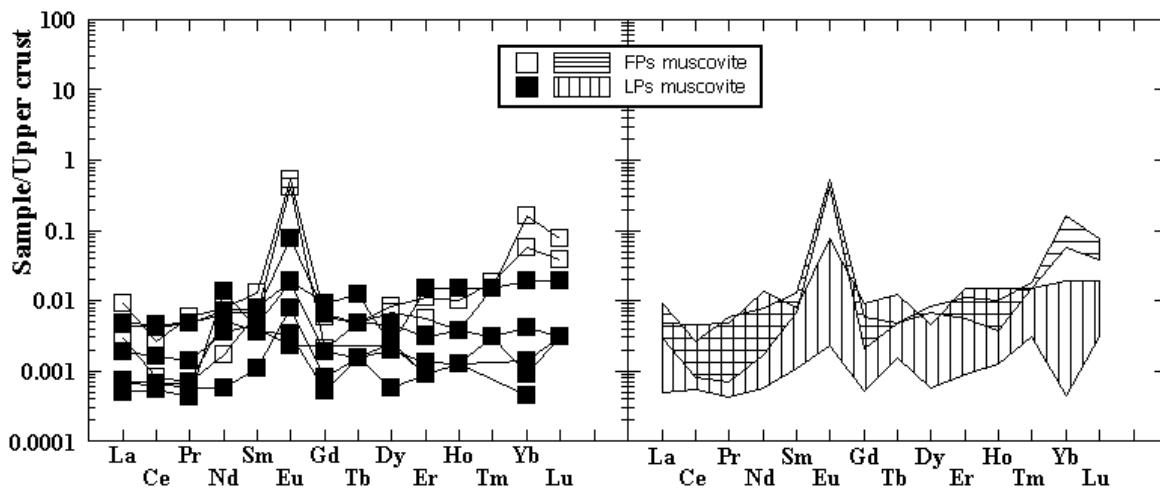


Fig. 8. REE diagram of the Contu-Negovanu muscovites (upper crust normalization values from Taylor and McLennan, 1985; 1995).

6. Conclusions

The muscovites from the two pegmatite types are essentially different as far as physical features, chemical composition and minor and trace element distribution (Li, Rb, Cs, Nb, Ta etc.) are concerned.

The **FPs** muscovite is colorless in thin sheets and white-yellowish in thicker blocks. It preserves minute hematite inclusions on the cleavage sheets and it may be often found associated with the biotite from which it originated. In the **LPs**, muscovite is greenish, perfectly transparent, with fresh aspect and glassy-nacred luster.

The bulk composition shows higher Al₂O₃ contents in the **FPs** muscovite than the **LPs** muscovite, whereas the (Mg, Fe)O contents have reversed proportions. Ti content, strongly related to the (Mg, Fe)O, is also higher in the **FPs** muscovite.

The normative composition using muscovite s.s., paragonite and celadonite end-members shows differences due to different (Mg, Fe)O and Na₂O contents which are related to the specific stage of the pegmatite formation. Therefore, the muscovite s.s. increases from the **FPs** muscovite to the **LPs** muscovite, while the paragonite and the celadonite are decreasing in the same direction.

The upper crust normalized values of the minor and trace elements are used as an objective tool for the muscovite stage formation, *i.e.* the pegmatitic fluid evolution degree, as well as for the preanatetic protolith nature and its melting degree. The elements which are generally considered to be incompatible behaved this way in the pegmatitic fluid – anatetic restite system only when they have normalized values over 1. In the Contu-Negovanu pegmatites, only Rb, Cs, Ba, Nb and Ta certainly preferred the anatetic melt. Among these, Rb, Cs, Nb and Ta behaved as incompatible elements also in the pegmatitic fluid – pegmatitic mineral, eventually concentrating in the **LPs**



muscovite. Ba had an opposite behavior, having a direct impact on the Eu positive anomaly amplitude, of maximum extension in the **FPs** muscovite. The low Sr, Y, Zr and Hf normalized values stand for a compatible behavior in the pegmatitic fluid – anatetic restite system, being preserved in the restite minerals. Among the elements which are generally considered to be mobile, only Li apparently behaved as an incompatible element both in the pegmatitic fluid – anatetic restite and pegmatitic fluid – pegmatitic mineral systems. The specific “apparently behaviour” terms are used because Li metapelitic protolith origin cannot be taken for certain.

The low U and Th normalized values in muscovite cannot provide information on their compatible/incompatible behavior in the pegmatitic fluid – anatetic restite, because these elements could be concentrated in the Fe-Mn phosphates from the Contu-Negovanu pegmatites.

REE normalized values are slightly different in the two muscovite types. The low values are normal for muscovite, as REE are strongly incompatible elements in the pegmatitic fluid – muscovite system. The Eu positive anomaly suggests a highly over 1 $\text{Eu}^{2+}/\text{Eu}^{3+}$ ratio in the early stages of pegmatitogenesis, when Eu was accommodated along with Ba in the **FPs** muscovite *I* site. During the late stages, this ratio decreases and subsequently the Eu anomaly decreases and eventually changes direction.

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ROMANIAN GEOLOGIST VISITING KRAKATAU – BRIEF SITE DESCRIPTION AND PETROLOGICAL ANALYSES

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Keywords: volcanic islands, landscapes, tectonic, petrography, geochemistry, Krakatau volcanic complex, Indonesia.

Krakatau volcanic complex (KVC) is located in the Sunda Strait, between Java and Sumatra islands in the western region of Indonesia. The KVC consists nowadays of Sertung Island, Panjang and Rakata islands – the last remnants of the explosive eruption on August 1883 and Anak Krakatau – the only active volcanic island. Towards the Sumatra Island at north, there are other two volcanic islands - Sebuku and Sebesi.

Tectonically, KVC lies at the intersection of two graben zones and of two N-S trending fracture zones between the Sumatra trench oblique subduction and the Java frontal subduction, at 140 km from the Java trench in SW. The Benioff zone is approximately at a depth of 120 km beneath it (Zen, 1983 in Zulkarnain et Priadi, 2003).

Part of this exotic place - Anak Krakatau (Child of Krakatau) Island - was visited by Laurentiu Ionescu – geologist at ROMGAZ - Medias, during September 2004. He kindly offered us photos and samples to be analyzed and discussed. Besides, monitoring satellite images of before and after December 26th 2004 devastating tsunamis have been acquired.

Petrographically, the island forming rocks vary from olivine basalts and olivine-pyroxene basalts as products of the early stage KVC, through pyroxene andesite to dacites – the latest magma composition reached at explosive eruption time (Zulkarnain et Priadi, 2003). Samples from Anak Krakatau (island born in 1927) show the same general petrographical evolution, the dacitic stage being reached in 54 years time (till 1981), followed by intensive eruption, not so explosive like the 1883 one. Although differentiation – crystal fractionation seem to be the main petrological processes, geophysical data revealed two different magma chambers and latest petrological investigation suggest magma mixing between basaltic and acidic magmas (Zulkarnain et Priadi, 2003) as a main process prior to eruption.

7 samples (5 porphyritic basaltic andesites with pyroxene and olivine, one of basaltic sand and one silicious scoria) from Anak Krakatau have been analyzed mineralogically and geochemically (for major, trace and rare earth elements). The basaltic sand belongs to a tholeiitic magma of an immature island arc volcanic stage, whereas the 5 basaltic andesites have formed in a mature stage arc volcanism. TE and REE behavior indicates a medium LILE and LREE slab enrichment and a normal mantle (lherzolite) phase entering partial melting during a more or less typical ocean-ocean collision phenomenon. High Ba/La ratios reveal mantle wedge enrichment by Ba-rich slab derived fluids, much of Ba being derived from subducted oceanic sediments (Hole et al., 1984 in Wilson, 1989).



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DETAILED GEOLOGICAL-GEOPHYSICAL REMOTE SENSING MAPPING OF THE BAIA MARE AREA, EAST CARPATHIANS, ROMANIA

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Key words: *remote sensing mapping, horizontal gravity gradient, horizontal airmagnetic gradient, Bouguer anomalies, tectonic sketch, volcanologic map, Oas-Gutâi Mts., Baia Mare region.*

The Baia Mare region (East Carpathians - Romania) is one of the famous ore deposit areas (gold, cooper, base metal) in Europe, with a two centuries mining history, which offered to science a huge mineralogical heritage. Geological mapping on the area began in the last decade of the 19th century. The actual maps (scale 1: 200,000 1: 50,000, 1: 25,000 on the entire area or 1: 10,000, 1: 5,000, 1: 2,000 on parts of it) performed by Geological Survey of Romania and IPEG Maramures using detailed prospecting, drilling and mining data, have a serious handicap because of widespread soil and vegetation covering. The advanced Remote Sensing (RS) processed a geological map using photo-geological interpretation – scale 1: 25,000 of a 2,500 sq. km area coupled with SPOT imagery interpretation, needed to be validated using equivalent methods of investigation (geological field mapping, geophysical and geochemical methods). Tectonic elements, volcanic edifices, lava flows, intrusives bodies and hydrothermal aureoles, Paleogene and Neogene sedimentary formations have been identified, classified and mapped. The accuracy of this map has been proved by overlapping the geological map obtained by field mapping, horizontal gravity gradient (HGG) and horizontal aeromagnetic gradient (HAG) maps.

Statistical observation regarding the accuracy of the map interpretation, indicate that:

- 80% of the volcanological and structural characteristics are confirmed by the structural sketch-map made on HGG and HAG maps;
- only 25% of geological limits and around 6% of volcanological elements from the field map are recognized by the RS geological maps;
- ore veins direction and their relative position detected by RS are accurately confirmed by geological mining data.
- the main structural elements (discordancies) like Dragos Voda Fault, Bogdan Voda Fault, North Gutâi Fault outlined by RS mapping are strictly confirmed by HGG or HAG sketch maps, as well as the Bouguer anomalies detailed inflexion analyses (the air magnetic elongate swarms of anomalies analyses);
- there is no discordant lines of the major Early Miocene napes are confirmed by HGG, HAG or RS mapping;
- there is no difference between intimate design structure of the major Early Miocene napes (Botiza, Wildflysch), but a certain difference between the mentioned ones outcropping in the area and the Petrova (Magura) one.

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As a result of a methodical photogeological interpretation the volcanological map of the Oas-Gutâi Mountains is presented.

The Remote Sensing mapping of the Baia Mare (Oas-Gutâi Mts.) proves its very efficient utilization as an overriding accurate approach. Its complex utilization join with field geological mapping, geophysical, geochemical or mining and drilling data is powerfully recommended.



NEW DATA ON ARDEALITE FROM THE TYPE LOCALITY (THE "DRY" CIOCLOVINA CAVE, SUREANU MOUNTAINS, ROMANIA)

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Keywords: ardealite, morphology, physical properties, crystal chemistry, unit cell parameters, infrared absorption data, "dry" Cioclovina Cave, Romania.

Ardealite, a rather rare hydrated calcium acid phosphate sulfate [ideally $\text{Ca}_2(\text{HPO}_4)(\text{SO}_4) \cdot 4\text{H}_2\text{O}$] was first described and named by Schadler (1932) from the bat-guano deposit in the "dry" Cioclovina Cave (Sureanu Mountains, Romania). Previous analyses performed by Halla (1931) on the type material furnished by Schadler gave rise to confusion concerning the discoverer of the mineral. During the study of brushite (Dumitras *et al.*, 2004), ardealite was abundantly found in material from the cave, which prompted to the re-investigation of this mineral species. The aim of this paper is to provide new optical, chemical, infrared absorption and X-ray data on ardealite from the type locality and to describe its associations and mutual relationships with other associated phases.

The information below is based on scanning electron microscopy coupled with energy-dispersive scans (SEM-EDS), optical study, wet-chemical analysis, inductively-coupled plasma atomic emission spectrometry, X-ray powder diffraction ($\text{Cu K}\alpha$, $\lambda = 1.54056 \text{ \AA}$) and Fourier-transform infrared absorption spectrometry. All the analytical facilities, procedures and experimental details are similar to those described by Marincea *et al.* (2002) and Dumitras *et al.* (2004).

The "dry" Cioclovina Cave, hereafter referred to as Cioclovina, is located in the southern part of the Sureanu Mountains (Southern Carpathians), at about 16 km east-southeast of Hateg, on Luncanilor Valley. It represents the upper fossil level of the Ponorici - Cioclovina cu Apa karst system (7890 m in length). The cave, known since 1873, was extensively exploited for phosphates because of the huge bat guano deposit inside: about 3200 wagons (or 23 000 tons) of this precious fertilizer were mined until the 1940's (Bleahu, 1976). The mined part of the cave consists in a nearby sub-horizontal gallery with some short divergent passages, measuring on their all about 900 m. The cave is developed in Tithonic - Neocomian algal micritic limestones with calcarenite levels.

The mineral association from Cioclovina is typical for a "dry" system of phosphate bearing cave deposits. About 30 mineral species have been identified until now. In addition to ardealite in the phosphate-bearing zones occur brushite, taranakite, tinsleyite, hydroxylapatite, crandallite, carbonate-hydroxylapatite, gypsum, bassanite, calcite, vaterite, aragonite, quartz, goethite, birnessite, hematite, romanechite, todorokite, kaolinite and illite. Onac *et al.* (2002) and Onac and White (2003) reported the presence of some other mineral species, such as berlinitite, burbankite, churchite, chlorellestadite, fogrite, paratacamite, collinsite and sampleite, but the occurrence of such exotic species is not enough substantiated.

Macroscopically, ardealite occurs as damp, off-white to pale yellow earthy masses of chalky appearance, whose luster is pearly but less pronounced than that of brushite. Under the optical microscope, the ardealite masses appear as extremely fine-grained spherule-coated surfaces that



hardly allow recognition of any single aggregate. Only SEM studies reveal the presence of minute micrometer-sized platy crystals, forming masses of felted appearance. The compact radial or bunched groups are generally built up of thin blades or individual crystals with the c^* axis pointing outwards from the center. Locally, they are composed by many “butterfly”-like interlocking or loosely packed aggregates of crystals, up to 100 μm in length. This kind of aggregate was synthesized by Rinaudo and Abbona (1988) and Rinaudo *et al.* (1994) and consists on two systems of interlaced laths, which radiate from a common center in opposite directions. The common center of a “butterfly”-like aggregate may be an unusual twin of ardealite with reentrant angles, with two individuals developed from the tips, the same as observed in the case of brushite by Abbona *et al.* (1993). Individual crystals, platy on (010), are typically 10-20 μm in length and 5-8 μm wide, but generally up to 1 μm thick.

No fluorescence has been observed under either short-wave (254 nm) or long-wave (366 nm) ultraviolet radiation. The mineral is, however, cathodoluminescent under Cu $K\alpha$ radiation. The indices of refraction measured in immersion for a number of ten platy grains taken from a representative sample (D 53 A) are $n_{\min} = \alpha = 1.530(2)$ and $n_{\max} = \beta = 1.537(2)$. Accepting that our sample has the same 2 V angle as the sample from La Guangola (Italy) analyzed by Balenzano *et al.* (1984), hence $2V_\alpha = 86^\circ$, the calculated value of γ is 1.543. The mean density of ten separate clusters taken from the same sample was measured by sink-float in methylene iodide diluted with toluene, at 20°C. The obtained value is $D = 2.335(3) \text{ g/cm}^3$, being in excellent agreement with the calculated densities of ardealite samples from Cioclovina, based on the chemical formulae (Table 1) and unit cell parameters (Table 2), accepting $Z = 4$ (Sakae *et al.* 1978): $D_x = 2.317 - 2.350 \text{ g/cm}^3$.

The chemical compositions obtained for some samples of ardealite from Cioclovina are summarized in Table 1.

Table 1. Bulk analyses of selected samples of ardealite from Cioclovina (in wt.%)

Sample	D37A ⁽¹⁾	D53A ⁽¹⁾	D31 ⁽²⁾	D35A ⁽²⁾	D35B ⁽²⁾	D37B ⁽²⁾	D53B ⁽²⁾
K ₂ O	0.03	0.06	0.08	0.05	0.06	0.05	0.07
Na ₂ O	0.06	0.00	0.05	0.06	0.07	0.06	0.04
CaO	32.18	32.39	32.45	32.45	32.46	32.46	32.45
MnO	0.02	0.01	0.00	0.01	0.00	0.01	0.01
MgO	0.02	0.02	0.01	0.02	0.01	0.01	0.02
P ₂ O ₅	19.80	19.10	19.13	19.56	19.62	20.45	19.26
SO ₃	23.80	24.80	24.92	24.44	24.36	23.43	24.78
H ₂ O ⁽³⁾	23.63	23.27	23.36	23.41	23.42	23.53	23.37
Total	99.54	99.65	100.00	100.00	100.00	100.00	100.00

Number of ions on the basis of 2 (S+P) and 8 (O) in the anhydrous part of the formula

K	0.002	0.004	0.006	0.004	0.004	0.004	0.005
Na	0.007	0.000	0.006	0.008	0.008	0.007	0.004
Ca	1.992	1.996	1.993	1.992	1.993	1.993	1.993
Mn	0.001	0.000	0.000	0.000	0.000	0.000	0.000
Mg	0.002	0.002	0.001	0.002	0.001	0.001	0.002
(HPO ₄) ²⁻	0.968	0.930	0.928	0.949	0.952	0.992	0.934
(SO ₄) ²⁻	1.032	1.070	1.072	1.051	1.048	1.008	1.066

Degree of hydration (water molecules pfu)

H ₂ O	4.069	3.998	4.000	4.000	4.000	4.000	4.000
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(1) wet-chemical analyses; (2) ICP-AES analyses; (3) in the ICP-AES analyses, as calculated for stoichiometry.



In disagreement with the previously published analytical results (e.g., Schadler, 1932), our findings show that in ardealite from Cioclovina the sulfate groups dominate over the protonated phosphate ones. Overall, the compositions fall within the ranges for the synthetic $\text{Ca}(\text{HPO}_4)_x(\text{SO}_4)_{1-x} \cdot 2\text{H}_2\text{O}$ ($0.15 \leq x \leq 0.70$) phase (Rinaudo and Abbona, 1988), having $0.46 < x < 0.50$. The cumulative substitution in the eight-fold coordinated sites, normally occupied by Ca, accounts for only 0.30 to 0.60 % of them. Other substitutions, i.e. (Ba, Sr, Y, REE)-for-Ca, are minor since the samples in Table 1 contain between 4.8 and 39 ppm Sr, between 1.4 and 20 ppm Ba, and up to 3.18 ppm La, 7.82 ppm Ce, 0.25 ppm Eu, 1.82 ppm Y and 0.07 ppm Yb.

The XRD patterns of ardealite samples from Cioclovina shown characteristic differences as compared with the pattern obtained by Sakae *et al* (1978) for the synthetic $\text{Ca}_2(\text{HPO}_4)(\text{SO}_4) \cdot 4\text{H}_2\text{O}$ (monoclinic, space group Cc), due to the presence of at least four additional lines at about 4.32, 4.15, 3.15 and 2.95 Å. Because all but few of the XRD lines could be, however, indexed using a monoclinic C-centered unit cell (space group Cc), the obtained patterns were indexed in this hypothesis. Although present, the supplementary reflections quoted before, which not index satisfactorily accepting the Cc structure, were not accounted for in the calculations of the unit cell parameters. Table 2 gives the unit cell parameters calculated for 24 representative samples of ardealite from Cioclovina.

Table 2. Unit-cell parameters of selected samples of ardealite from the “dry” Cioclovina Cave

Sample	a (Å)	b (Å)	c (Å)	b (°)	V (Å ³)	n ⁽¹⁾	N ⁽²⁾
2217 A	5.725(2)	30.993(2)	6.252(3)	117.08(2)	987.6(6)	3	78
2217 B	5.723(2)	31.113(8)	6.259(2)	117.23(2)	991.1(4)	10	87
D 2 B	5.721(2)	30.994(9)	6.253(2)	117.28(2)	983.8(4)	4	65
D 2 C	5.720(2)	31.018(1)	6.242(2)	117.27(2)	984.4(5)	6	48
D 7 A	5.726(1)	30.995(7)	6.243(2)	117.28(1)	984.7(3)	6	51
D 9 B	5.723(2)	30.999(1)	6.251(2)	117.20(2)	986.3(4)	5	50
D 22 B	5.709(3)	30.968(2)	6.252(5)	117.19(3)	983.2(8)	6	57
D 31 A	5.721(5)	30.990(2)	6.245(9)	117.23(1)	984.6(1)	4	73
D 35 A	5.720(2)	31.014(1)	6.255(3)	117.24(2)	986.6(5)	5	105
D 35 B	5.717(2)	31.019(1)	6.254(3)	117.18(2)	986.5(5)	5	103
D 35 C	5.725(4)	31.041(2)	6.238(4)	117.31(3)	985.1(7)	7	51
D 35 D	5.716(1)	30.988(7)	6.242(2)	117.23(1)	982.9(3)	7	58
D 37 A	5.721(2)	30.995(1)	6.248(3)	117.25(2)	984.8(5)	6	88
D 37 B	5.713(2)	31.031(1)	6.233(3)	117.09(2)	983.9(5)	9	95
D 45 A	5.719(2)	31.034(1)	6.247(3)	117.26(2)	985.7(5)	8	104
D 47 B	5.723(2)	30.995(7)	6.254(2)	117.25(1)	986.1(3)	5	94
D 53 A	5.736(2)	31.024(9)	6.263(2)	117.26(2)	990.8(4)	7	85
D 59 B	5.716(2)	31.021(1)	6.254(3)	117.22(2)	986.2(5)	4	73
D 59 C	5.718(2)	31.016(1)	6.250(2)	117.14(2)	986.4(5)	6	74
D 59 D	5.716(2)	31.013(9)	6.248(3)	117.23(2)	984.9(4)	7	96
D 59 E	5.724(2)	31.014(1)	6.242(3)	117.25(2)	985.0(5)	5	70
D 63 B	5.714(3)	30.973(1)	6.261(4)	117.09(3)	986.5(7)	5	68
D 63 C	5.716(2)	30.991(1)	6.258(3)	117.09(2)	986.9(5)	5	101
D 66 B	5.721(2)	30.977(8)	6.249(2)	117.27(1)	984.3(4)	5	60

(1) number of refining cycles; (2) number of reflections in the 2θ range 10-76° used for refinement.



The mean values, obtained as average of the 24 sets in Table 2, are $a = 5.720(5)$ Å, $b = 31.009(29)$ Å, $c = 6.250(7)$ Å and $\beta = 117.21(7)^\circ$. Both values for individual samples and the mean values are in good agreement with those reported by Sakae *et al.* (1978) for the synthetic $\text{Ca}_2(\text{HPO}_4)(\text{SO}_4) \cdot 4\text{H}_2\text{O}$ [$a = 5.721(5)$ Å, $b = 30.992(5)$ Å, $c = 6.250(4)$ Å and $\beta = 117.26(6)^\circ$].

The infrared absorption spectra obtained for four representative samples of ardealite from Cioclovina, which can be obtained by request from the first author, were recorded in the frequency range between 250 and 4000 cm⁻¹. The mean values obtained for the absorption bands are depicted in Table 3, which tried to offer the wavenumbers, characters and intensities of the infrared absorption bands, as well as attempts for the band assignments.

Table 3. Positions of and assignments of the infrared absorption bands recorded for selected samples of ardealite from Cioclovina ⁽¹⁾

Structural group	Vibrational mode	Wavenumber (cm ⁻¹) ⁽²⁾	Character, intensity ⁽³⁾
H_2O	ν_3 antisymmetric stretching	3552(28)	vs, shd
H_2O	ν_3' antisymmetric stretching	3396(15)	vs, b
H_2O	ν_1 symmetric stretching	3205(10)	vs, shd
$(\text{HPO}_4)^{2-}$	(P)O-H stretching	2958(5)	s, shd
$(\text{HPO}_4)^{2-}$	(P)O-H stretching	2240(8)	w, b
H_2O	ν_4 in-plane H-O-H bending	1718(5)	m, b ⁽⁴⁾
H_2O	ν_4' in-plane H-O-H bending	1640(18)	s, b ⁽⁵⁾
$(\text{HPO}_4)^{2-}$	P-O-H in-plane bending	1209(6)	m, shd
$(\text{HPO}_4)^{2-}, (\text{SO}_4)^{2-}$	ν_3 O-P(S)-O antisymmetric stretching	1142(2)	vs, b
$(\text{HPO}_4)^{2-}, (\text{SO}_4)^{2-}$	ν_3' O-P(S)-O antisymmetric stretching	1101(7)	vs, b (shd)
$(\text{HPO}_4)^{2-}, (\text{SO}_4)^{2-}$	ν_3'' O-P(S)-O antisymmetric stretching	1003(1)	vs, sh
$(\text{HPO}_4)^{2-}, (\text{SO}_4)^{2-}$	ν_1 O-P(S)-O symmetric stretching	956(8)	m, shd
$(\text{HPO}_4)^{2-}$	P-O(H) symmetric stretching	867(5)	s, sh
$(\text{HPO}_4)^{2-}$	P-O-H out-of-plane bending	821(3)	m, b (shd)
$\text{H}_2\text{O} + (\text{SO}_4)^{2-}$	ν_2 H-O-H + ν_4 O-S-O	671(1)	m, sh
$\text{H}_2\text{O} + (\text{SO}_4)^{2-}$	ν_2 H-O-H + ν_4' O-S-O	627(6)	m, sh
$(\text{HPO}_4)^{2-} + (\text{SO}_4)^{2-}$	ν_4 O-P-O + ν_4'' O-S-O	600(3)	s, b
$(\text{HPO}_4)^{2-}$	ν_4 O-P-O in-plane bending	559(3)	s, b (shd)
$(\text{HPO}_4)^{2-}$	ν_4'' O-P-O in-plane bending	526(4)	s, sh (shd)
$(\text{HPO}_4)^{2-}, (\text{SO}_4)^{2-}$	ν_2 O-P(S)-O out-of-plane bending	418(5)	w, shd
$[\text{CaO}_6(\text{H}_2\text{O})_2]^{10-}, (\text{SO}_4)^{2-}$	ν_2' O-S-O + lattice mode (+ H-O-H ?)	360(5)	m, shd
$[\text{CaO}_6(\text{H}_2\text{O})_2]^{10-}$	lattice mode (Ca-O)	304(3)	s, sh
$[\text{CaO}_6(\text{H}_2\text{O})_2]^{10-}$	lattice mode (Ca-O)	289(1)	s, sh (shd)
$[\text{CaO}_6(\text{H}_2\text{O})_2]^{10-}$	lattice mode (Ca-O)	280(2)	s, sh
$(\text{HPO}_4)^{2-}$	ν_2' O-P-O out-of-plane bending	265(3)	s, sh (shd)
$[\text{CaO}_6(\text{H}_2\text{O})_2]^{10-}$	lattice mode (Ca-O)	256(3)	s, sh (shd)

(1) assumptions made by analogy with brushite and taking into account the work of Aslanian and Stoilova (1982) and Rinaudo and Abbona (1988); (2) standard deviations into brackets; (3) s = strong; m = medium; w = weak; vs = very strong; sh = sharp; b = broad; shd = shoulder; (4) split into two bands; (5) broad complex, split into six distinct absorption bands.

The main remarks concerning the IR spectra are as follows: (1) There are only three bands clearly recognizable in the OH-stretching region between 3000 and 4000 cm⁻¹, although the



structure determination by Sakae et al. (1978) show that in synthetic $\text{Ca}_2(\text{HPO}_4)(\text{SO}_4) \cdot 4\text{H}_2\text{O}$ there are four hydrogen bonds implying the water molecules in the structure. Two of the bands due to the stretching vibrations of molecular water consequently overlap. (2) As well as in brushite, a fifth hydrogen bond seem to be established between OH groups pertaining to the protonated phosphate groups, which could explain the presence, in our spectra, of the shoulder at $\sim 2960 \text{ cm}^{-1}$. (3) The different structural positions of the four water molecules in the structure of ardealite explain the pronounced splitting of the bending motions of molecular water: the didentate bending at $\sim 1718 \text{ cm}^{-1}$ and the broad complex composed of six bands, centered at $\sim 1640 \text{ cm}^{-1}$. (4) As shown by Marincea et al. (2004) The bands due to the $(\text{SO}_4)^{2-}$ and $(\text{HPO}_4)^{2-}$ ν_1 and ν_3 fundamentals overlap, whereas ν_2 and ν_4 are presumably distinct. These bands are tentatively given in Table 3. (5) Even attempted, assumptions in the low-range wavenumber of the spectra are very hypothetical because there is difficult, if not impossible, to ascertain a band in the spectral range between 650 and 250 cm^{-1} , where lattice modes, phosphate bendings and P-O-H librations largely overlap.

As well as in other guano bearing caves (e.g. Pestera Mare de la Meresti – Marincea et al., 2004) ardealite from Cioclovina seems to be formed in a crystallization sequence from hydroxylapatite to brushite and finally ardealite. Both direct precipitation from an acid solution and formation by replacement of brushite may be proposed, but closer examination reveals that many brushite aggregates are partly replaced by ardealite. S-bearing veins fill fractures of the brushite aggregates, but because of their small sizes (usually $< 10 \mu\text{m}$) and of the volatility of the sample under the electron beam, there are some difficulties in the identification of the composing mineral species, that is only supposed to be ardealite. Ardealite is frequently present as overgrowths on brushite aggregates and in all cases textural relationships are conclusive for an early formation of brushite. It results that the hypothesis of formation of ardealite through step-by-step replacement of preexistent brushite may be favored at Cioclovina. This is consistent with progressive crystallization under open-system conditions within fractures affecting brushite masses, due to the incoming acidic sulfate-rich solutions.

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EVIDENCES OF SUBMARINE ERUPTION/EMPLACEMENT OF INTERMEDIATE VOLCANICS IN OAS-GUTAI MTS., EASTERN CARPATHIANS; CASE STUDIES

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Part of the inner volcanic arc of the Eastern Carpathians, Oas-Gutai Mts. experienced the volcanism imprinted by the Miocene tectonics developed in the Carpatho-Pannonian Region. The both types of volcanism, the felsic and the intermediate one, are still questionable about the subaerial versus submarine eruptions or emplacement of the volcanic products. The felsic volcanism developed mostly subaerial but a wide range of volcaniclastics developed in connection with the submarine emplacement. The complexity of the intermediate volcanism and the difficulties in attempting the structures reconstruction raised a lot of questions about the environment of evolution.

Studying the intermediate volcanics, some evidences of submarine emplacement of volcanics have been picked up. The used criteria reffer to lava features and morphology of lava units, types of associated volcaniclastics and relationships with the Miocene sedimentary deposits.

This paper aims to present two case studies from Oas and Gutai Mts. respectively, some extrusive domes, showing clear proves of the submarine deposition of their volcanic products.

In Oas Mts., Turulung dacitic dome records the submarine evolution of both lava and associated volcaniclastics. The quarry opened in Turulung dacites shows spectacular transitions from glassy lavas to hyaloclastites with enclosed lava lobes and different phreatomagmatic deposits, both primary and reworked. The quarry is surrounded by Miocene sedimentary deposits.

In Gutai Mts., Piatra Rosie dacitic dome shows many evidences of the submarine setting involvement. The massive glassy lavas pass to a thick sequence of reworked volcaniclastics of both non-explosive and explosive origin. Hyaloclastites and phreatomagmatic products had been reworked in submarine setting by different mass flows and slides or slumps. The sequence is covered by laminated mudstone.

The case studies suggest the submarine emplacement of the lava domes, from shallow to deep water, involving either the quench fragmentation of lavas or the phreatomagmatic disruption. Most of all, the submarine transport determines a wide variety of volcaniclastics, according to the degree of fluidization, all prone to deep water, below wave base setting. The same setting is assessed by the sedimentary deposits from the top of Piatra Rosie sequnce.

Similar deposits have been found in some other different parts of Oas-Gutai Mts. but the assessemnt of the submarine environment of eruption or emplacement is rather difficult because of the scarce outcrops, often hidden by vegetation, because of interfingered lavas and volcaniclastics belonging to different sources and the very complicated syn- and postvolcanic tectonics.



TARINA AND RODU: GOLD MINERALISATION HOSTED IN MAAR/DIATREME CONTACT ENVIRONMENTS

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Significant deposits of gold and silver have been exploited in the past from the contact zones of intrusive maar/diatreme complexes at Rosia Montana and Rodu-Frasin, but these deposits are often overlooked due to their proximity to the larger and more conspicuous deposits hosted within the dacite intrusives at the centre of these maar/diatreme complexes. The Tarina, Igre, and Jig zones (combined as the Tarina deposit for the purposes of this paper) are located on the northern periphery of the Rosia Montana intrusive maar/diatreme complex, close or coincident to the contact with the surrounding Cretaceous sedimentary sequence. The Rodu deposit is located in a similar environment on the western edge of the Rodu-Frasin maar/diatreme complex, a smaller maar/diatreme located 5 km to the southeast of Rosia Montana. Both intrusive complexes are located on dilational sites associated with a regional scale fault system that has a northwesterly trend and hosts numerous phases of igneous activity.

The Tarina deposit is characterised by a zone of mineralisation that is coincident with the contact between the vent breccia and the surrounding Cretaceous sediments and has a general orientation of 300°/50°, dipping moderately towards the centre of the maar/diatreme complex (southwest). Where this contact is exposed in underground workings it is faulted, mineralised and has been extensively mined. The mineralised zone can be up to 80 metres in width and contains gold within both the vent breccia and the underlying Cretaceous sediments. A series of northwesterly to north-northwesterly trending faults associated with the diatreme/maar bounding structure appear to be the principal controls on the location and distribution of gold mineralisation. This can clearly be seen by the fault and vein orientations stope out by the historical miners at Jig, Igre and Tarina. Gold mineralisation is also associated with dacite dykes and intrusions of polymictic diatreme breccia that have been intruded along fault structures and at major fault intersections. The gold mineralisation at Tarina is hosted within brecciated and silicified sedimentary rocks, strongly altered and fractured vent breccia and altered dacite and polymictic intrusives.

The Rodu deposit has been less thoroughly explored, but a significant amount of gold mineralisation has been identified in a zone of strongly altered vent breccia, usually located immediately above the contact with the Cretaceous sedimentary package. Two principal styles of gold mineralisation have been identified at Rodu; low to medium grade, large tonnage alteration style mineralisation and high grade low tonnage mineralisation hosted within narrow fault zones or quartz veins. A number of the high grade structurally hosted zones have been mined in the past. Unlike at Tarina, no significant gold mineralisation has been identified in the Cretaceous sediments to date, and no dacite dykes or intrusive polymict breccias have been intersected. However, one significant similarity between the Tarina and Rodu deposits is the coincidence of intense structural deformation associated with zones of major faulting intersecting the maar/diatreme contact. Late normal faulting caused by relaxation across the maar/diatreme has also been identified at both deposits. These faults are most likely the conduits for mineralising hydrothermal fluids.

The most probable reason why the gold mineralisation at Tarina is sometimes preferentially concentrated in the Cretaceous sediments is due to impermeable nature of the overlying clay-rich reworked vent breccia, which leads to the pooling and boiling of hydrothermal fluids below this



horizon. However, at Rodu, the vent breccia is not reworked or clay rich, and is therefore more permeable than the underlying Cretaceous shales. This causes the hydrothermal mineralising fluids introduced along fault structures to be concentrated in the vent breccia above the contact.

The latest resource calculation for Tarina has identified 41.5 million tonnes at 1.14 g/t Au and 3.9g/t Ag using a 0.6 g/t Au cut off, for a total of 1.52 million ounces of gold and 5.59 million ounces of silver. However, the calculation method used large blocks (40m by 40m by 10m) which has resulted in an exaggeration of the tonnage and a consequent reduction in the grade. At Rodu a recent resource update calculated a resource of 26 million tonnes at a grade of 0.97 g/t Au and 2.02 g/t Ag using a 0.6 g/t Au cut off, for a total of 812, 530 ounces of contained gold and 1.7 million ounces of silver. The potential exists to increase both these resources significantly.



THE RELATIONSHIP BETWEEN THE CRYSTAL-CHEMISTRY AND THE MORPHOLOGY OF STIBINITE FROM BAIA MARE AREA

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Introduction

The paper describes the crystal-chemical and morphological features of stibnite – Sb_2S_3 from Baia Mare area, and aims at establishing the causes behind its morphological variations. Stibnite crystals from various occurrences in Baia Mare area have been classified into three main morphological types, denominated after the ore deposits where such forms prevail, i.e.: the *Baia Sprie* type, the *Herja* type and the *Baiut* type. The Baia Sprie type (BS) consists of elongated, often acicular crystals of up to 10-12 cm in length. Crystals are strikingly striated along [001] axis, and often aggregate into fragile radiating or large, irregular groups. BS-type crystals are frequently associated with translucent flaky crystals of barite. No basal faces are obvious. The Herja type (H) is characterized by radiating or parallel groups of short, bladed crystals of up to 45 in length. Striations parallel to elongation are also present. Unlike the BS-type, the H-type shows terminal {111} forms. The Baiut type (B) is defined by thick, prismatic crystals of up to 8-10 cm in length. Crystals are striated along [001] and often form parallel or slightly radiant groups with few individuals. Transversal sections are isometric or slightly flattened after [010]. The same occurrence may include all three morphological types, either in separate or combined aggregates. However, a defined morphological type will always be predominant within a given occurrence. For example, at Baiut, the “autochthonous” B-type is often overgrown by H-type crystals, whereas at Herja, bladed H-type are intergrown with acicular BS-type crystals. A common feature of stibnite from all described occurrences is that they lack smooth (prism or pyramid) faces with true (semi-)metallic luster.

The morphological classification of stibnite may be further detailed based on several criteria which avoid the locality-related character of a certain crystal habit:

- *The “visual” ratio* between the contributions of the longitudinal vs. the transversal sections. Six morphological subtypes may thus be defined: short-columnar, long-columnar, short-prismatic, long-prismatic, short-acicular and long-acicular, respectively.
- *The aspect ratio*: defined as the ratio between the length and the diameter (L/D) of elongated (acicular, fibrous, prismatic etc.) crystals, and commonly used in industrial applications of minerals. The aspect ratio may be introduced as a quantitative parameter, in order to test its possible relevance in the morphological study of stibnite. A set of 16 samples, belonging to all crystal habit types and from all studied occurrences was considered, with 1 to 5 different crystals being investigated for each sample. Three values were calculated for the aspect ratio (A.R.) in each case: the measured, A.R.(m) (L/D), the maximum, A.R.(+), obtained as $(\text{measured } L \text{ value} + \text{error bar}) / (\text{measured } D \text{ value} - \text{error bar})$, and the minimum, A.R.(-), obtained as $(\text{measured } L \text{ value} - \text{error bar}) / (\text{measured } D \text{ value} + \text{error bar})$.



D value + error bar). Large variations of the three values calculated for each crystal (underlining the contribution of the errors), and also of the values obtained on distinct crystals belonging to the same habit type and even sample, were obtained. The values for each habit type were compared in order to establish certain “natural” broad limits that could be still used in an informative way. For acicular crystals, the lower limit of the A.R. values could be set around 5. For the short-acicular type a value of 15 may be suggested a possible upper limit. Prismatic crystals provide the most heterogeneous values of the A.R., with apparently no significant difference between the short-prismatic and the long-prismatic. The lower limit may be set to A.R=5, whereas for the short-prismatic type, the upper limit may be considered as 30. Columnar crystals – all originating from Baiut – have typical A.R. values lower than 5, and in the case of the short columnar subtype the characteristic values are less than 3.

- *The crystal aggregate habit:* the tri-dimensional display of individual crystals may be described by using the following categories: (b) bladed – quasi-compact aggregates of thin lath-like crystals; (p) (quasi) parallel orientation of crystals; (f) fan shaped – closely-packed divergent display (in the case of the studied samples, the crystals were always arranged in a quasi-parallel display of the {010} forms); (r) radial – crystals radiate from a centre without producing stellar forms; (s) stellate – spherical, radial aggregates radiating from a star-like point; (pc) parallel crust – quasi-parallel display of short acicular (fibrous) crystals of a combined (f+r)-type, forming a continuous crust; (sc) stellate crust – fibrous crystals in radial and “star” like aggregates (similar to s-type) forming a continuous crust; (i) interlace – intergrowth of similarly-sized aggregates radiating from 2-3 centres; additionally, there were samples consisting of loose, single crystals.

The possible causes determining this morphological variability were approached from two different angles: a) *intrinsic causes*: i.e., depending on the chemical composition and/or potential subtle structural differences between the three varieties; b) *extrinsic causes*: i.e. related to certain features of the mineral environment and of the crystal growth regime (mainly, temperature and saturation).

1. Chemical and X-ray data

The intrinsic factors were assessed by means of EMP chemical analysis, with special emphasis on Sb-Bi or Sb-As substitution, as well as by using powder X-ray diffraction.

A number of 28 chemical analyses were carried out on 6 samples which were considered to be representative for each of the three main morphological types. The results show chemical compositions very close ($\text{Sb}_{1.993}\text{S}_{2.930}$) to ideal stibnite, with no significant variations among the three morphological types.

X-ray diffraction analysis of the three morphological types of stibnite involved a detailed calculation of the unit-cell parameters. Measurements were carried out using Debye-Scheerer powder diffractometers of various builds. A total of 7 samples were measured in various analytical conditions. Unit-cell parameters calculated on the basis of X-ray powder data published by Pop and Costin (2003) were added to the data set. The results show no major differences between the three morphological types (Table 1).

In conclusion, the analyzed stibnite samples proved very similar in both chemical and structural aspects. All cases were of ideal stibnite, without major traces of chemical substitutions. Unit-cell data were very similar, and also, very close to the original data by Bayliss and Nowacki (1972). Any differences may be assigned to different analytical conditions. Thus, neither chemical nor structural parameters should be accepted as possible causes of stibnite morphological variability.



Table 1. Summary of unit-cell parameters data for stibnite samples of various origins and morphological types (italic lettering denotes standard deviations).

Sample	Morphological type	<i>l</i> (Å)	<i>a</i> (Å)	<i>b</i> (Å)	<i>b</i> (Å)	Volume (Å ³)
28~99*	Baiut	1.5419 <i>0.0021</i>	11.2996 <i>0.0044</i>	3.8317 <i>0.0012</i>	11.2369 <i>0.0042</i>	486.5250
136~32*	Baiut	1.5419 <i>0.0044</i>	11.3098 <i>0.0028</i>	3.8337 <i>0.0011</i>	11.2514 <i>0.0063</i>	487.8420
28~106*	Herja	1.5419 <i>0.0028</i>	11.3152 <i>0.0036</i>	3.8367 <i>0.0013</i>	11.2459 <i>0.0049</i>	488.2190
136~15*	Herja	1.5419 <i>0.0036</i>	11.3115 <i>0.0036</i>	3.8366 <i>0.0013</i>	11.2519 <i>0.0062</i>	488.3000
28~105*	Baia Sprie	1.5419 <i>0.003</i>	11.3160 <i>0.003</i>	3.8340 <i>0.0012</i>	11.2454 <i>0.0053</i>	487.8880
136~35*	Baia Sprie	1.5419 <i>0.0043</i>	11.3086 <i>0.0043</i>	3.8380 <i>0.002</i>	11.2204 <i>0.0073</i>	486.9950
1028*	Baia Sprie	1.5419 <i>0.0036</i>	11.3170 <i>0.0036</i>	3.8367 <i>0.0013</i>	11.2526 <i>0.0055</i>	488.5910
4	Baiut	1.5406 <i>0.0052</i>	11.2926 <i>0.0052</i>	3.8391 <i>0.0028</i>	11.2212 <i>0.0082</i>	486.4820
10	Herja	1.5406 <i>0.0055</i>	11.2902 <i>0.0055</i>	3.8319 <i>0.0024</i>	11.2432 <i>0.0059</i>	486.4100
12	Baia Sprie	1.5406 <i>0.0033</i>	11.2957 <i>0.0033</i>	3.8294 <i>0.0016</i>	11.2375 <i>0.0045</i>	486.0820
13	Baia Sprie	1.5406 <i>0.0083</i>	11.3110 <i>0.0083</i>	3.8416 <i>0.0039</i>	11.2306 <i>0.0111</i>	487.9930
14	Herja	1.7903 <i>0.0023</i>	11.3069 <i>0.0023</i>	3.8346 <i>0.0007</i>	11.2292 <i>0.0034</i>	486.8650
16	Baiut	1.7903 <i>0.0021</i>	11.3088 <i>0.0021</i>	3.8348 <i>0.0007</i>	11.2303 <i>0.0032</i>	487.0260
18	Baia Sprie	1.7903 <i>0.0022</i>	11.3045 <i>0.0007</i>	3.8341 <i>0.0007</i>	11.2245 <i>0.0034</i>	486.4960

* - Unit-cell parameters calculated on the basis of X-ray powder diffraction data published by Pop and Costin (2003).

2. Genetic interpretation of crystal morphology

Any interpretation of the extrinsic control factors of crystal growth which determined the differentiation of the three main morphological types of stibnite (i.e., temperature, supersaturation etc.) must be done in close correlation with the atomic structure of this mineral. The present interpretation is based in the classical theory of the *periodic bond chains* – PBC by Hartmann and Perdock (1955) and on the bonding energy of Sb₄S₆ to the crystalline edifice. The present study is an attempt to explain the mere absence of faces belonging to the [001] zone, by defining a single PBC vector leading to the exclusive development of S (stepped) and K (kink) faces.

Interatomic potentials of stibnite were calculated in order to assess the distribution of bonding energies across the crystal lattice and to define the PBCs. The calculation was carried out in *quasi ab initio* manner using the program GULP (Gale 1999, 2000) without numerical constraints concerning relevant physical properties (e.g., dielectric constant, elastic constant, phonon spectra etc.). Covalent bond polarity was modeled using a factor of nucleus eccentricity with respect to the electron shells. Bonding potentials were calculated using the Buckingham model, expressed by:



$$V_{ij} = A \exp\left(\frac{-r_{ij}}{B}\right) - \left(\frac{C}{r_{ij}^6}\right)$$

Where A , B and C correspond to the repulsion sphere, bond deformation factor and attraction sphere, respectively; r_{ij} is the interatomic distance. In order to describe the anisotropy of covalent bonds, a spring-type interaction was introduced between atomic nuclei and electron shells. The calculated A , B , C factors and the spring constants are given in the table below.

Table 2. Calculated A , B and C factors of the Buckingham potential and the spring constants for the interaction between the nucleus and the electron shell.

Atoms		Potential	A (Å)	B (Å)	C (Å)
1	2				
S	s Sb	Buckingham	0.569E+04	0.347	2.03
S	s S	Buckingham	0.120E+04	0.149	135
Sb	c Sb	Spring (c-s)	148		
S	c S	Spring (c-s)	23.2		

Given the actual state of knowledge and the lack of relevant physical constraints, the accuracy of this calculation may still be improved. However, for the purpose of this paper, the calculated Buckingham potentials may be regarded as acceptable. Due to the reversed proportionality between the bonding energy and the interatomic distance, we may conclude that this type of potential approximates rather well the proximal interactions within the stibnite lattice. The distribution of the calculated bonding energy (Table 3, Figure 1) fairly explains among others, the perfect cleavage along (010).

Table 3. Lengths and energies of interatomic bonds in stibnite

Atomic bonds	Bond lengths (Å)	Buckingham bonding energies (arbitrary units)
Sb1-S3	2.521626	3.9730
Sb1-S2	2.539451	3.7741
Sb1-S1	3.102776	0.7443
Sb1-S2'	3.167179	0.6183
Sb1-S3'	3.640816	0.1579
Sb2-S1	2.460352	4.7404
Sb2-S3	2.678020	2.5315
Sb2-S1'	2.860233	1.4974
Sb2-S2'	3.372670	0.3420

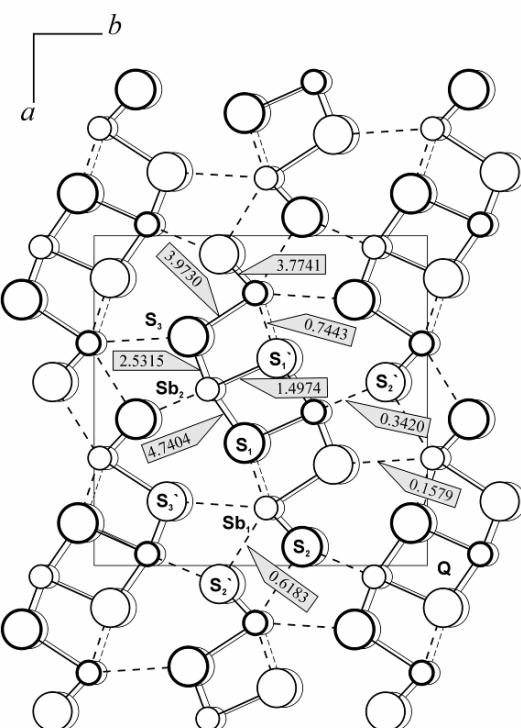


Figure 1. The crystal structure of stibnite (Bayliss and Nowacki, 1972) and the distribution of Buncigham bond energies.

Stibnite contains *a single type of PBC*, which is represented by infinite Sb_4S_6 rods along [001] (Figure 2). These PBCs contain the strongest bonds, but are weakly bonded to each other. The PBCs are thus compatible with the Hartmann-Perdock concept, due to: 1) the distribution of strongest bonds, 2) the early formation of Sb_4S_6 growth units in the hydrothermal solution and 3) the centricity of Sb_4S_6 chains which renders nil electric polarization.

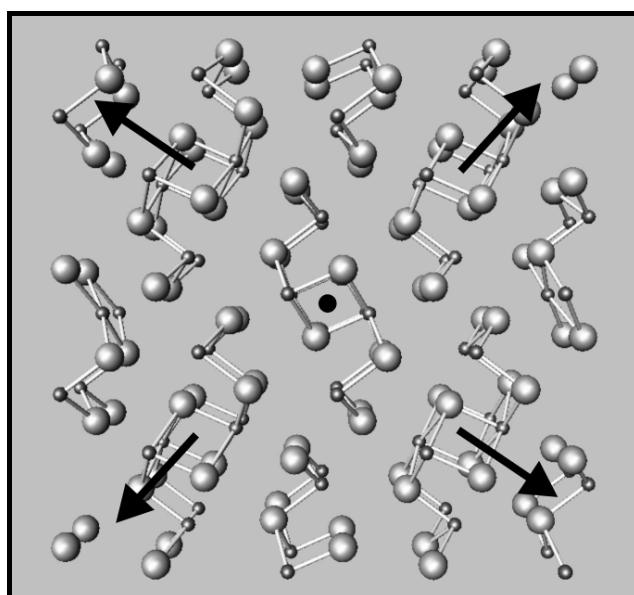


Figure 2. Equivalent PBCs parallel to [001] in the crystal structure of stibnite (symbolized by the black arrows and central dot).



The distribution of the inter-PBC bonding energies and their relative position to the main crystallographic axes is shown in Figure 3. The consequences deriving from the existence of a single PBC-type in stibnite may be described as follows:

- F-type (flat) faces are absent;
- S-type faces are prevailing in the [001] zone; crystals exhibit striations parallel with the elongation;
- K-type faces are subordinate, and they develop mainly as terminal pyramid forms {hkl}.

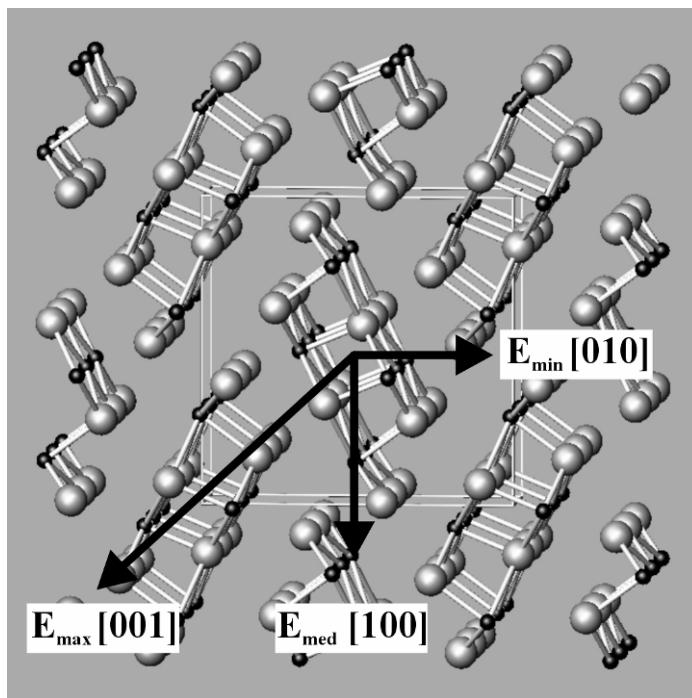


Figure 3. The distribution and ranking of inter-PBC bonding energy, with regard to the main crystallographic axes.

A characterization of the three main morphological types is possible based on the PBC concept and of the genetic factors. The aspects described below confirm the behavior of the morphological varieties of stibnite in the way predicted by the PBC theory.

The Baia Sprie type (Figure 4, 5)

Growth conditions:

- Low temperature
- High supersaturation
- Numerous crystallization centers

Morphological features:

- Prevalent growth parallel to PBC direction – acicular habit
- Almost exclusive occurrence of S-type faces in the [001] zone
- Lack of basal pynacoid or pyramid faces

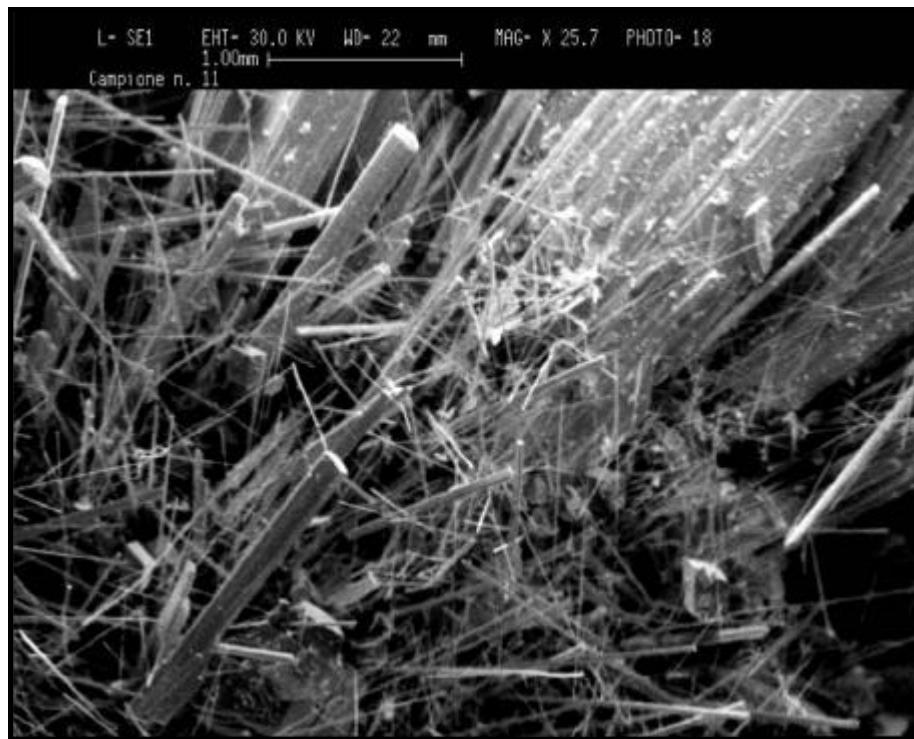


Figure 4. Typical aggregate morphology of BS-type, needle-like microcrystals, suggesting a large number of crystallization centers and a high supersaturation level. (SEM image).

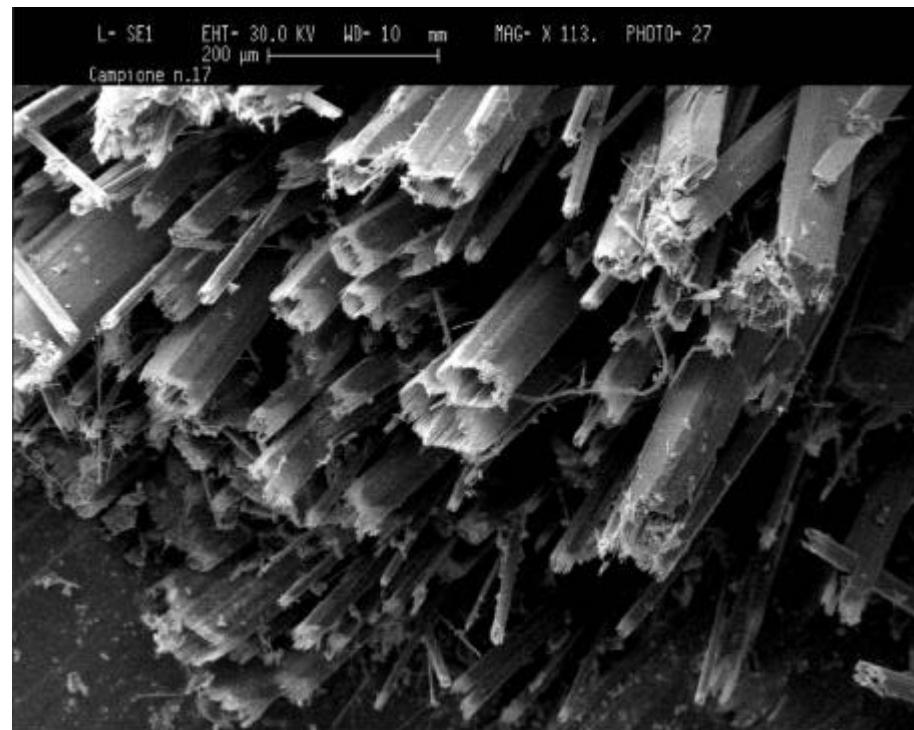


Figure 5. Absence of basal pynacoid faces in the case of BS-type morphology. Edging corresponds to PBCs developed after [001]. (SEM image).



The Herja-type (Figure 6)

Growth conditions:

- Comparatively higher temperature than in Baia Sprie
- Average supersaturation
- Less crystallization centers

Morphological features:

- Prevalent growth parallel to PBCs and parallel to average inter-PBC bonding energies [100] – lath-like habit flattened after [010]
- No growth units can be attached parallel to [010] – i.e., to the direction of minimum inter-PBC bonding units – as they are rapidly dissolved back
- Prevailing S-type faces in the [001] zone
- K-type terminal forms {hkl} may develop

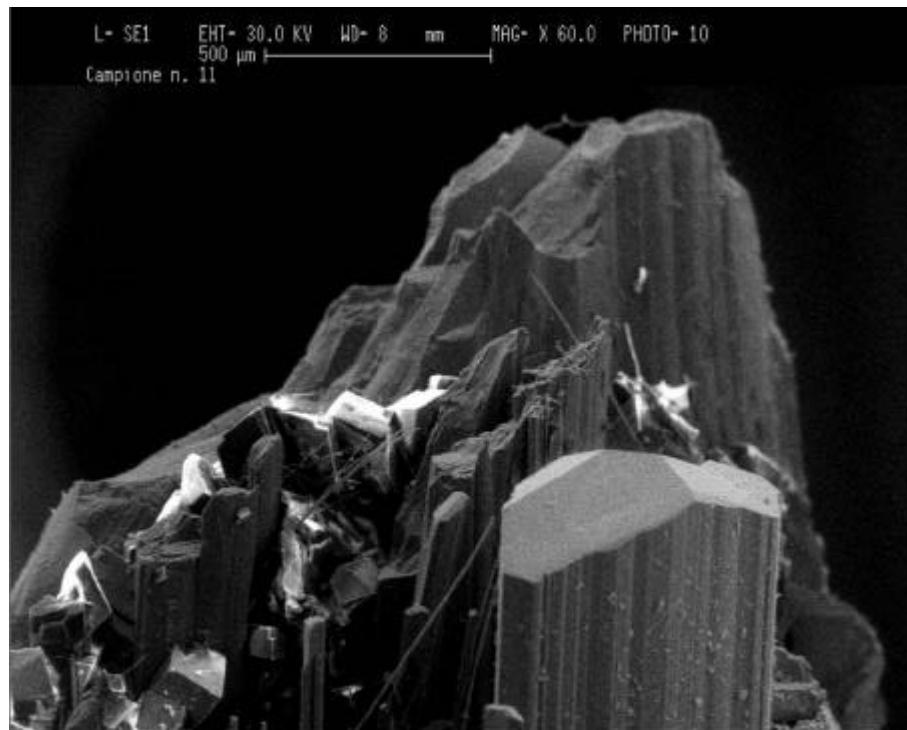


Figure 6. Prismatic microcrystal flattened along [010]. Longitudinal striations given by S-type faces and K-type terminal pyramid forms are visible. (SEM image).

The Baiut type (Figure 7)

Growth conditions:

- Highest forming temperature compared to the other occurrences
- Lowest supersaturation
- Lowest number of crystallization centers

Morphological features:

- Prevalent growth parallel to PBCs, but also parallel to the average and low inter-PBC bonding energies – columnar habit with quasi-isometric cross section
- Prevailing S-type faces in the [001] zone



- Relatively well developed k-type terminal forms {hkl}

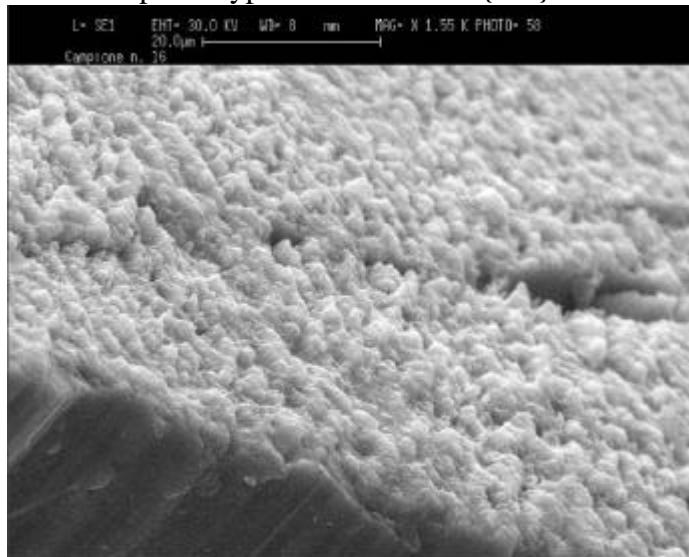


Figure 7. Detail of a K-type face on a B-type crystal. The rugged surface is given by staggered terminations of PBCs. (SEM image).

3. Discussion

The morphology of stibnite crystals from various occurrences in the Baia Mare area results from the interaction between external genetic factors (temperature, supersaturation) and unidirectional PBCs parallel to [001] axis. The unidirectional PBCs favor the development of S-surfaces parallel to the [001] zone axis (i.e., of characteristic striation) and subordinately, of the terminal K-type faces. No F-type forms could be observed. The degree of supersaturation and the number of crystallization centers are progressively higher in the following order: Baiut \Rightarrow Herja \Rightarrow Baia Sprie. With regard to the growth of crystal faces related to the single PBC type present in stibnite, the most reliable criteria for characterizing external morphology is the growth anisotropy related to [001], [100] and [010]. The absolute size of crystals and the “visual” ratio (e.g. “short-acicular”, “long-acicular”) may thus fail to be a reliable criterion to separate relevant morphological types. The genetic conditions suggested by each of the three morphological types are prevalent within the corresponding ore-deposit, but they should not be described as unique on the deposit’s scale. BS-type crystals are often found as overgrowths on H or B-type crystals. Similarly H-type crystals may overlap B-type columns, thus denoting dynamic thermal and supersaturation conditions. The relative continuity of temperature and supersaturation conditions could result in the formation of transient types such as B-like crystals with incipient flattening along [010] which is characteristic to H-type morphology.

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**POTENÚIALUL AURIFER AL MUNÚILOR METALIFERI**NICOLAE LUDUTM AN

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Având în vedere interesul deosebit fa \leftrightarrow | de exploatarea zăcămintelor aurifere, interes datorat cererii tot mai mari, pe pia \leftrightarrow a mondială, a acestui metal, care î Π i găsește utilizarea în tehnologii de vârf ale lumii contemporane, precum și con \leftrightarrow inutul noilor concep \leftrightarrow ii privind dezvoltarea durabilă, se impune informarea, nu numai a specialiștilor ci și a opiniei publice, cu privire la rezervele de aur cunoscute, pentru a se diminua impresia că exploatarea intensivă a unui zăcământ aurifer ar putea duce la epuizarea resurselor acestui metal și compromiterea vie \leftrightarrow ii socio-economice a viitoarelor genera \leftrightarrow ii.

Plecând de la aceste considerente, credem că este utilă o sinteză, nu neapărat exhaustivă, ci mai degrabă informativă, asupra rezervelor aurifere din Mun \leftrightarrow ii Metaliferi, pentru ca cel puțin o parte din suspiciunile legate de epizarea rezervelor de aur să fie diminuate sau eliminate.

Din punct de vedere al regiunii metalogenetice, Mun \leftrightarrow ii Metaliferi fac parte din **Provincia metalogenetică a Mun \widehat{U} ilor Apuseni, Subprovincia asociată vulcanismului neogen**.

Abordând acestă problematică pe districte metalogenetice, situa \leftrightarrow ia rezervelor de aur se prezintă, după cum urmează:

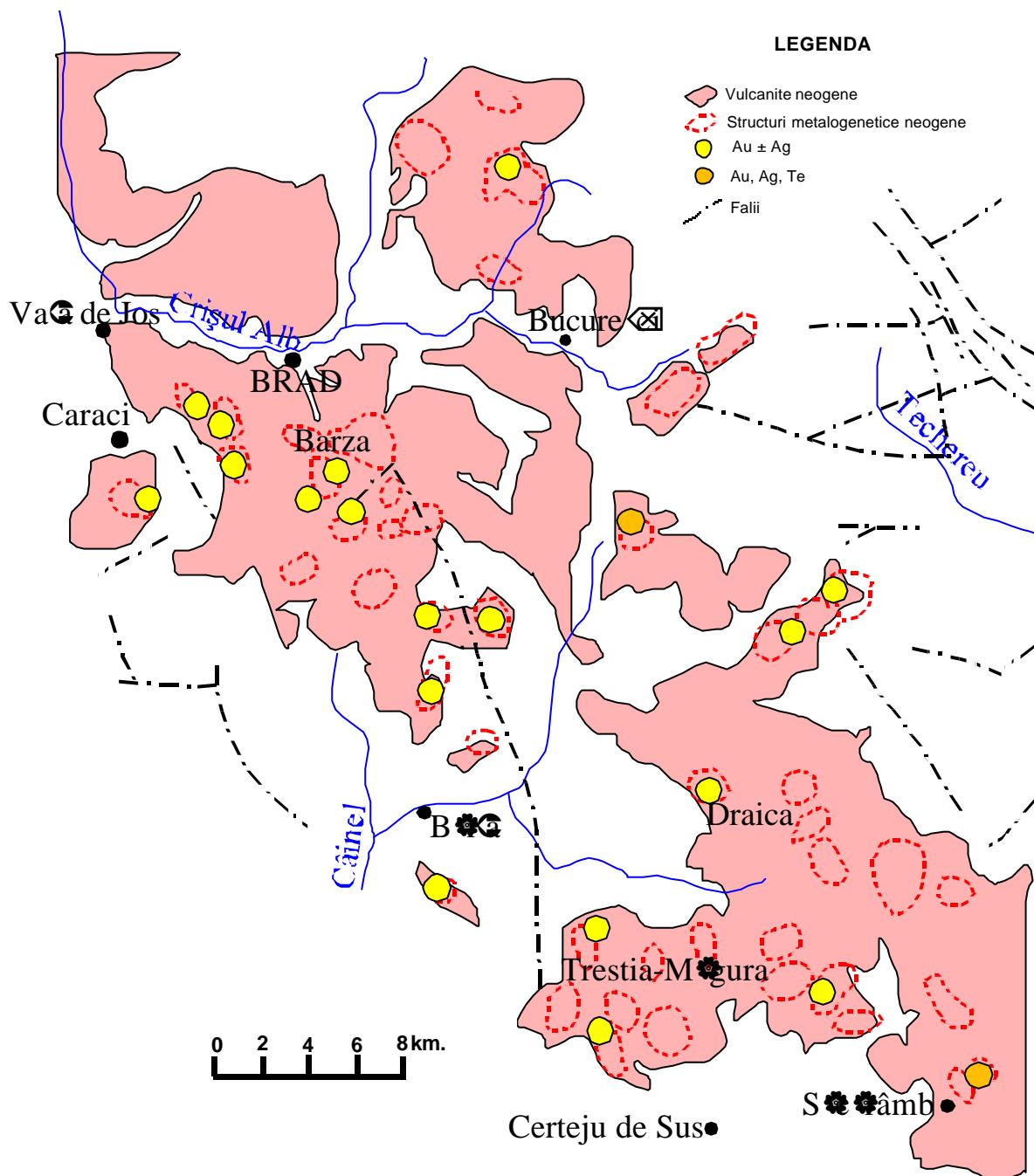
1. Districtul metalogenetic Brad-Săcărâmb (fig. 1) reprezintă unitatea metalogenetică cu cea mai mare extindere din cadrul subprovinciei, fiind caracterizată, îndeosebi, prin mineralizări auro-argentifere, asociate aparatelor vulcanice neogene. Toate câmpurile metalogenetice ale acestui district cantonează mineralizări auro-argentifere, în cea mai mare parte fiind vorba de zăcăminte exploatare sau în curs de exploatare. Câmpurile metalogenetice ale districtului au fost separate în două grupuri [Gh.C.Popescu, 1986], respectiv câmpurile de pe latura sud-vestică a districtului și câmpurile de pe latura nord-estică a districtului.

Grupul câmpurilor de pe latura sud-vestică a districtului include:

Câmpul metalogenetic Caraci, în cadrul căruia aurul este impregnat în umplutura caolinoasă a unui sistem filonian complex, majoritatea filoanelor fiind deja exploatare, însă au fost puse în evidență și stockuri, care ar putea constitui obiectul unor explorații de perspectivă. Aici a fost semnalată, destul de rar, și prezența telururilor.

Câmpul metalogenetic Barza ce cantonează mai multe grupe de filoane aurifere, legate de structura complexă a stratovulcanului Barza care, altături de filoanele polimetale subordonate și coloanele de tip "porphyry copper", formează un adevarat nod metalogenetic, respectiv "nodul metalogenetic Barza". În categoria grupurilor de filoane, pot fi amintite: grupul Ruda-Barza (cunoscut din epoca romană), grupurile Valea Morii Veche și Valea Morii Nouă, Brădișor, Antoniu și Dealul Feții-Valea Talpelor, majoritatea filoanelor aurifere fiind exploatare, însă structura complexă a stratovulcanului Barza ascunde încă rezerve importante de metale.

Nodul metalogenetic Câinel-Bîța cuprinde structurile mineralizate din bazinul văii Bîța, legate de sistemul de fracturi de pe limita sud-vestică a zonei centrale a bazinului Brad-Săcărâmb. Toate cele trei câmpuri metalogenetice ale nodului, respectiv Câinel, Bîța și Draica, cantonează importante sisteme filoniene aurifere, majoritatea exploatare, perspectiva de extindere fiind însă



foarte optimist.

Nodul metalogenetic Trestia-Măgura-Hondol, care cuprinde câmpul metalogenetic Troița-Măgura, în care sunt incluse mai multe grupuri filoniene (Trestia, Magdalena, Troița) și grupul de filonian Măgura, în care predomină filoanele auro-argentifere, în subsidiar fiind prezente și mineralizațiile polimetalice.

Câmpul metaogenetic Sloicrâmb a devenit cunoscut datorită prezenței, în parageneza mineralologică, a telurului nativ și a telururilor de aur și argint, fiind unul dintre cele mai importante câmpuri metalogenetice aurifere, cu rezerve încă nenumărate.

Grupul câmpurilor metalogenetice de pe latura nord-estică a districtului se caracterizează prin prezența mineralizațiilor polimetalice-auro-argentifere. Unitățile metalogenetice care sunt



cuprinse în acest grup *Structurile cu mineralizări Merișoare*, *Mineralizările din structura Cetății*, *Câmpul metalogenetic Vilișoara*, *Structura vulcanică Cordurea-Cerburea*, *Câmpul metalogenetic București-Rovina*), au fost mai puțin exploatare, multe din structurile mineralizate fiind în diferite stadii de crecere, astfel că potențialul auro-argentiger al acestui grup de câmpuri metalogenetice este departe de a putea fi pronozat.

2. Districtul metalogenetic Almaș-Stânișoara cuprinde zăcămintele plimetalice și auro-argentifere cantonate în lungul unor aliniamente tectono-vulcanice cu orientare dominantă NV-SE (fig. 2).

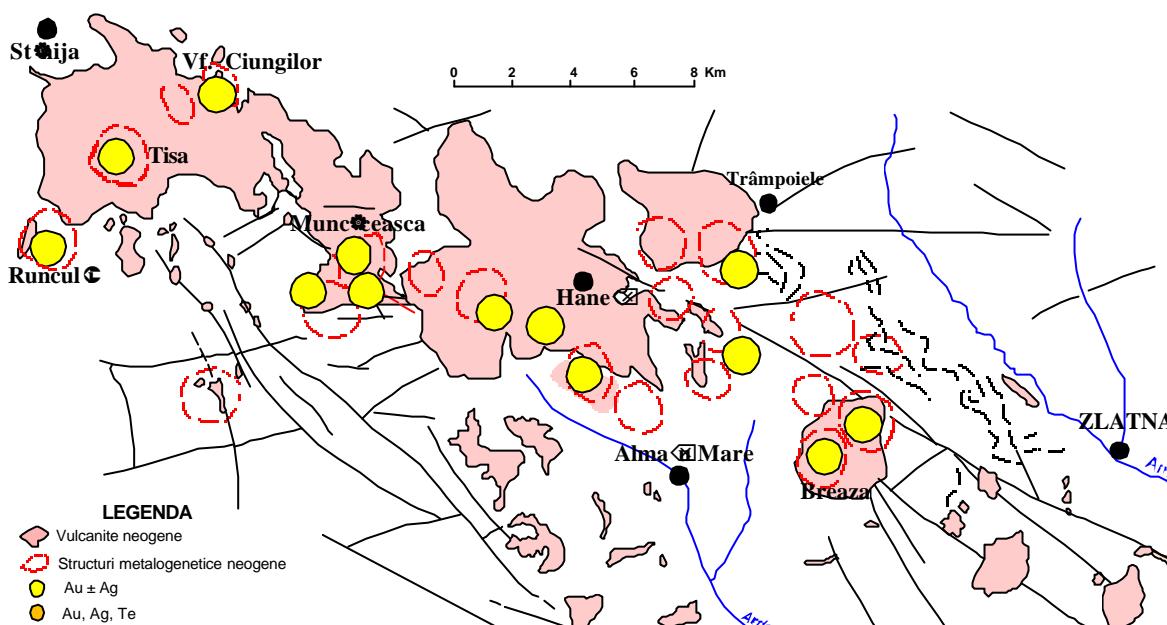


Fig. 2. Districtul metalogenetic Zlatna-Stânișoara
(după: Map of mineral resources, 1983)

Mineralizările auro-argentifere din cadrul acestui district nu mai au amploarea celor din districtul Brad-Sărămbău, însă în gradul de cercetare al multora dintre ele se află doar în faza de prospecție grea, prin intermediul lucrărilor miniere subterane (galerii de coastă).

Unitățile metalogenetice ale districtului sunt constituite din aliniamente metalogenetice [GH.C.Popescu, 1986], mineralizările fiind caracterizate printr-o zonalitate a paragenezelor pe verticală, zonalitate dată de asocierea următoarelor elemente metalice: Au nativ și telururi; pirite aurifere; Au și Pb, Zn și Cu; Pb, Zn, Cu și Au; Cu, Au; Mo, W și Cu.

Aliniamentul metalogenetic Hanești-Breaza cuprinde două câmpuri filoniene (*Breaza* și *Hanești*), în care mineraizările aurifere apar sub formă de cuiburi de aur nativ sau impregnat în masa cuarțului care formează umplutura filoanelor.

Aliniamentul metalogenetic Prepeștera-Trămpoiele cuprinde două câmpuri filoniene, respectiv *Fața Băii Larga*, caracterizate prin faptul că în partea superioară, mineralizările sunt impregnări preponderente piroitoase și caracter aurifer, iar spre adâncime, este din ce în ce mai bogată în Pb și Zn, menținându-se însă caracterul aurifer.

Aliniamentul metalogenetic Tihuța-Umanu-Baba-Almaș se dezvoltă paralel cu aliniamentul Hanești-Breaza, înspre sud-vest și cuprinde trei câmpuri metalogenetice: *Ușoara Umanu*, *Baba-Băbuța* și *Almaș*, exploatarea unora dintre zăcămintele acestor câmpuri fiind începută cu mult timp în urmă. Mineralizările auro-argentifere sunt amplasate, de regulă, la contactul dintre corpurile



andezitice, ca o particularitate, în cazul câmpului Almaș, printr-un proces intens de metasomatoză s-au format lentile de minereu polimetalic, cu conținut ridicat de aur, denumit “comb”.

Aliniamentul metalogenetic Neagra-Dealul Ungurului se caracterizează printr-o metalogenie de tip auro-argentifer, controlată spațial atât de fracturile cu orientare NV-SE cât și de structurile vulcanice amplasate pe traseul aliniamentului. În cadrul principalelor câmpuri metalogenetice (*Munccea Est, Munccea Vest și Stănița*), aurul nativ s-a depus în ultima fază de mineralizare, în porțiunile brecificate filoanele fiind mai intens mineralizate cu galenă, blendă, altăit și aur nativ.

Districtul metalogenetic Zlatna-Stănița își apare într-o cîmpurile metalogenetice **Mormântu-Vârful Ciungilor** (de interes economic minor) și cel din bazinul văii **Tisa** și **Dealul Runculești**, în filoanele acestuia, în parte exploatare, fiind prezent și aurul nativ.

3. Districtul metalogenetic Roșia Montană-Bucium-Baia de Arieș este legat de două mari concentrații de edificii vulcanice și centre de erupție, care au generat o intensă metalogenie auro-argentiferă de tip “porphyry copper” (fig. 3). Districtul prezintă caracteristicile unui veritabil aliniament metaogenetic, în cuprinsul căruia se individualizează trei noduri metalogenetice.

Nodul metalogenetic Roșia Montană-Roșia Poieni constituie una din cele mai importante concentrări metalogenetice ale Apusenilor și cuprinde două structuri mineralizate, respectiv *Roșia Montană*, cu parageneze auro-argentiferă și *Roșia Poieni*, cu parageneze “porphyry copper”.

Câmpul metalogenetic Roșia Montană cuprinde acumulările cunoscute și exploatare încă din antichitate, mineralizarea fiind prezentă sub formă de volburi și filoane, asociate și controlate spațial de structura vulcanică Roșia Montană. Cîmpurile metalogenetice se dezvoltă în cadrul a două structuri, respectiv “grupul Cetate” și “grupul Cârnic”, care au format, și formează și la ora actuală, obiectul unor ample lucrări de cercetare și exploatare.

Nodul metalogenetic Constanța-Arama-Corabia, în ceea ce privește mineralizația auro-argentifere, cuprinde mai multe grupuri de filoane, respectiv: *grupul filonian Arama*, constituit dintr-un filon central, cu mai multe ramificații; *grupul filonian Corabia*, caracterizat prin zone foarte bogate în aur, cantonate la intersecția filoanelor, în zonele denumite “cruci” și *grupul filonian Boteni*, în care aurul apare fin dispersat sau sub formă de cuiburi.

Nodul metalogenetic Baia de Arieș conținează o puternică metalogenie polimetalică-auriferă, repartizată în două câmpuri metalogenetice:

Câmpul metalogenetic Afinitățile, ce cuprinde mineralizația repartizată în două stockuri, aurul fiind prezent subordonat, în asociere cu sulfuri, sulfosulfuri, sau argentit, majoritatea filoanelor fiind deja exploatare.

Câmpul metalogenetic Ambru, care prezintă o parageneză cu caracter polimetalic, asociată unor corpuri metasomatici, mai puțin importantă în ceea ce privește paragenezele auro-argentifere.

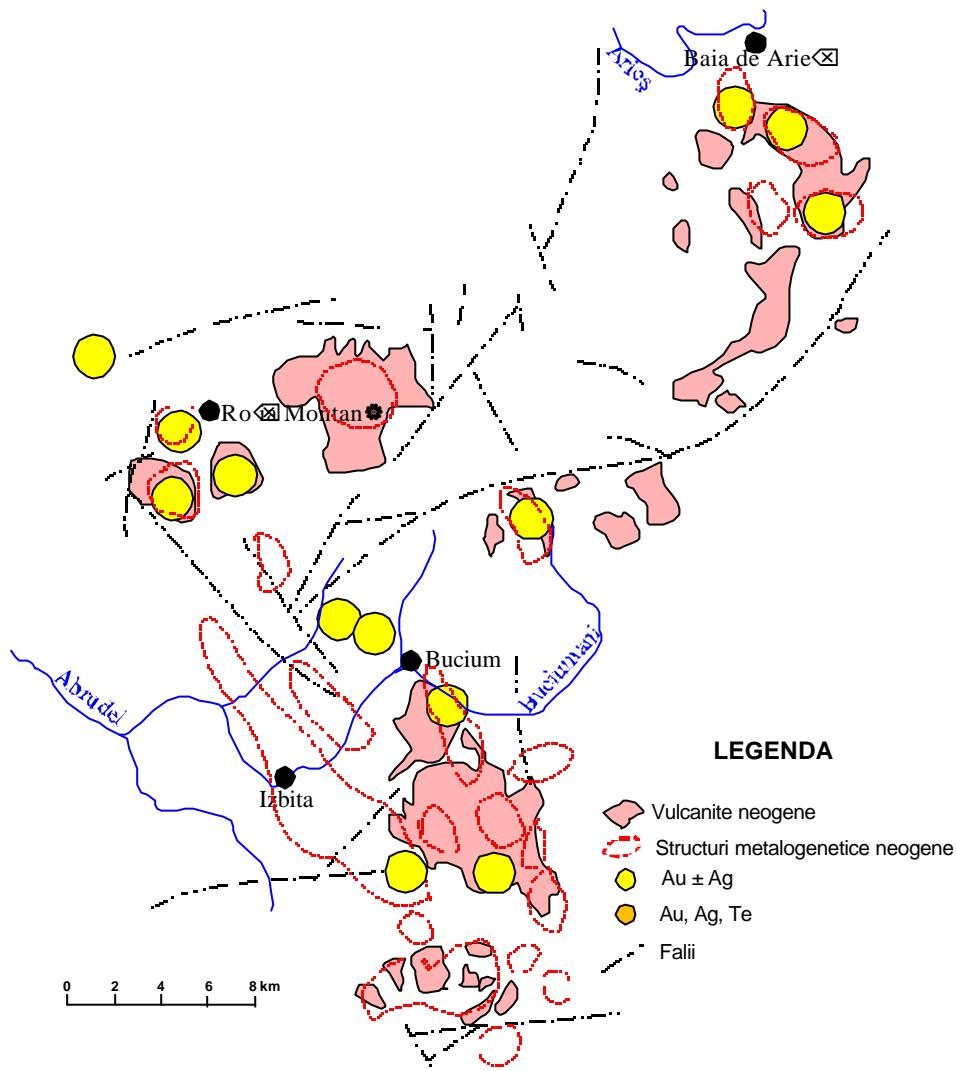


Fig. 3. Districtul metalogenetic Roșia Montan-Bucium-Baia de Arieș (după: Map of a mineral resources, 1983)

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THE STRUCTURE OF WORLD-WIDE POWER AND MINERAL COMMODITIES

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Abstract

Annually, in the world-wide level, huge quantities of mineral and energy substances are extracted for economic activities. In the 2000s 5 billion tones of metallic ores had been exploited and used as raw materials for recovery almost 0.7 billion tones of metal by industrial preparing treatment and metallurgic processing. It had been extracted and obtained as concentrates a lot of nonmetallic ores and industrial minerals (magnesite, chalk, phosphates, potash salts, native sulphur, unrefined salt, gypsum, natural abrasives, kaolin, bentonite, talc, baryte and witherite, ilmenite, zirconium, fuller's earth, andalusite, kyanite, sillimanite, borate minerals, arsenic trioxide, natural graphite, mica, etc) totalized 0.6 billion tones, but the exploited quantity is bigger. On a large scale (6.0 billion tones) it had been also proceeded the building stones exploitation. Fossil flues had been exploited in huge quantities (9.2 billion tones coal equivalent): hard coal (3.9 billion tones), lignite (0.8 billion tones), crude petroleum (3.4 billion tones), natural gasoline (0.03 billion tones), natural gas (0.003 billion tones coal equivalent) and peat for fuel (0.017 billion tones). Natural gas, lignite, peat and some of the hard coal (for energy purposes) had been used as raw power products and as power commodities by processing.

In the 2000s the world production of mineral and power commodities had been exceeded 21 billion tones.

Introduction

The mineral and energy commodities are solid or fluid minerals or fossil fuels (named also mineral fuels) extracted from their naturally occurring ore deposits by exploitation activity. They stand only crude mineral and energy substances if are not prepare for increasing quality (1). Only few are used as raw materials in industry or for domestic purposes.

After exploitation from deposits other minerals, rocks and fossil fuels are converted through a number of physical and chemical processes, into the pure or almost pure state, to separate the product or metal bearing ore, of interest, from the other non-economic or undesirable minerals, or gangue. Some of these products become final products (e.g. granulometric sorts of rocks and coal, nonmetallic ore concentrates and industrial minerals) being used as raw materials, since others (e.g. ore concentrates, coking coals, liquid hydrocarbons without water and solid impurities) are industrially converted into metals, coke, gases, petroleum, motor fuel, diesel oil – from coal; liquid petroleum gases, petroleum and diesel oil, bitumen, coke, oils, paraffin hydrocarbons, etc., from liquid hydrocarbons) or (by combination) other new products (organic substances from hydrocarbons).

We try to present a global, unique image of commodities resulted from mineral and energy substances by exploitation and dressing.



The mineral commodities production

1. The ore production

Metallic ore is considered now a mineral or an aggregate of minerals which can be mined at a profit and can be processed to produce metals desired by society. Nonmetallic or industrial ore is used for its properties or its mineral compounds (including metallic minerals), but finally no metal is extracted.

A. The metal quantities of ores and metallic concentrates

The world-wide statistics for metallic ores productions excluded (excepting bauxite and sometimes iron-bearing ore) tonnage of ore but not metal quantity or metal recovery content. The exploitation activity of ores in deposits (including different levels of same deposit) is made knowing that there is a variety of metal contents; so that ore extended product specifications are not important without metal-bearing estimations. Additionally, some companies or countries use to not report the metal contents of exploited ores, but the content of concentrates intended for treatment for metal recovery.

In the year 2000 some of metals (included in exploited ores and concentrates intended for treatment for metal recovery) are obtained in million of tons: iron (617.479), aluminum (24.025), copper (13.282), manganese (11.008), zinc (8.333), chromium (4.764), lead (3.143) and nickel (1.212). Other 12 metals (tin, molybdenum, antimony, vanadium, tungsten, uranium, cobalt, niobium and tantalum, silver, mercury and gold totalized together up to 0.707 million tones.

The percentage production of the main metals contained in ores and concentrates in the year 2000, is presented in figure no 1.

B. The nonmetallic ores and industrial minerals production

Nonmetallic ores and industrial minerals (both terms are almost a common significance) are much more numerous than metallic ores. From geological point of view some of them (clay, gravel, crushed stone, dolomite, etc.) are rocks or they could be considered rocks (salt, gypsum, bentonite, calcareous stone, etc.) but are not directly used as building materials and are well known as nonmetallic ores (1).

Other nonmetallic ores are some metal compounds (baryte, witherite, salt, gypsum, etc.) but are used for their properties and not for metal extraction. There are also ores with double character (magnesite, bauxite, limonite, illmenite, etc.) (1), both metallic (for metal recovery) and nonmetallic (used for their properties).

Only the most important nonmetallic ores and industrial minerals (from the exploitation and consumption point of view) are world-wide classified like in figure no 2.

2. The building stones production

This is about rocks used as monumental and building stones, different from others with nonmetallic ore significance. In the world-wide statistics (3, 4) these rocks are not genetically presented (marble, travertine, alabaster, granite, porphyry basalt, sandstone, etc.), but are classified from use criteria.

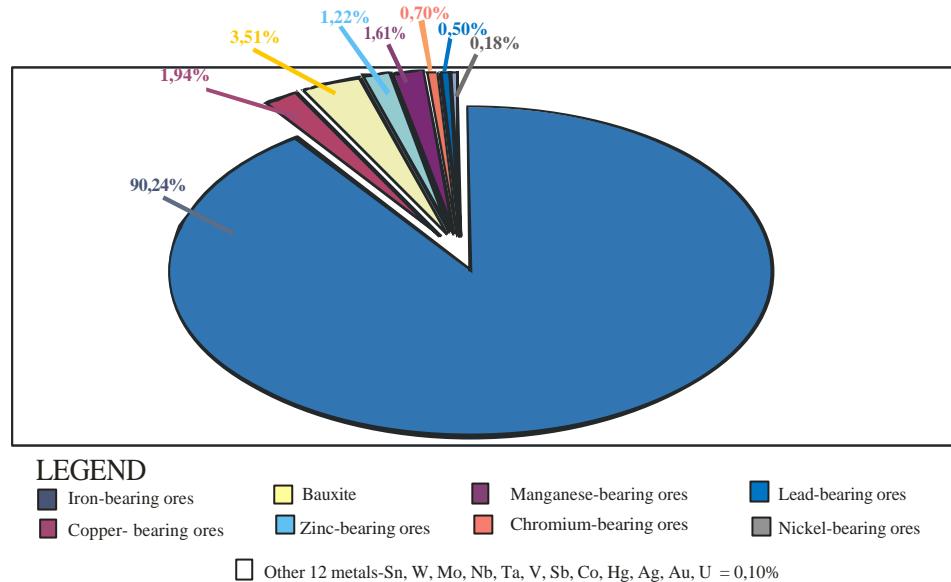


Figure 2 - The world percentage in metal content of proper ores and all other ores and concentrates intended for treatment for metal recovery in the 2000s

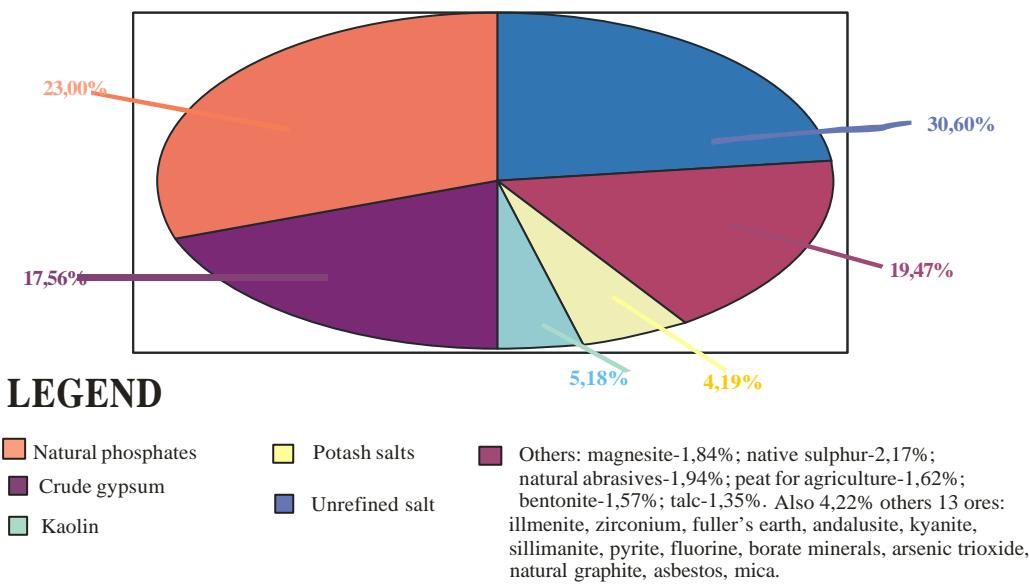


Figure 2 - The world percentage repartition of nonmetallic ore and industrial minerals productions in the 2000s

The most important productions (3.282 billion tones) are mentioned for gravel and crushed stone group, including pebbles and crushed or broken stone, gravel, macadam and tarred macadam of a kind commonly used for concrete aggregates, for road metal lining or for railway or the ballast. Flint and shingle are included (3). Silica and quartz sand (1.389 billion tones) are commercially-extracted sand used in building, in glass industry, for cleaning metals, etc. The limestone flux and calcareous stone (1.067 billion tones) group includes limestone flux and limestone and calcareous rocks, commonly used for the manufacture of lime or cement, excluding building or monumental stone. Those materials in powdered form for soil improvement are included. Dolomite and chalk are excluded (3).



Granite, porphyry, basalt, sandstone and other hard indigenous rocks (99 mil. tones) is a monumental and building stones group, including such stone not further worked than roughly split, roughly squared or squared by sawing, such as granite, porphyry, basalt, sandstone and other hard rocks.

In the group of marble, travertine, ecaussine and other similar hard calcareous monumental and building stone and alabaster (24 mil. metric tones) are included rocks provided that their apparent specific gravity is 2.5 or more, presented in the mass or in form of blocks, slabs or sheets. Stones identifiable as mosaic cubes or as paving flagstones are excluded.

The percentage level of world-wide building rock productions in the year 2000, classified by use criteria groups, is presented in figure 3.

For specifying the entire tonnage production it is necessary to multiplicate the value of production (in m³) with 2.5 (apparent specific gravity). It could be assumed in order to decrease, the important contribution of gravel and crushed stones, silica and quartz sand, limestone flux and calcareous stones groups.

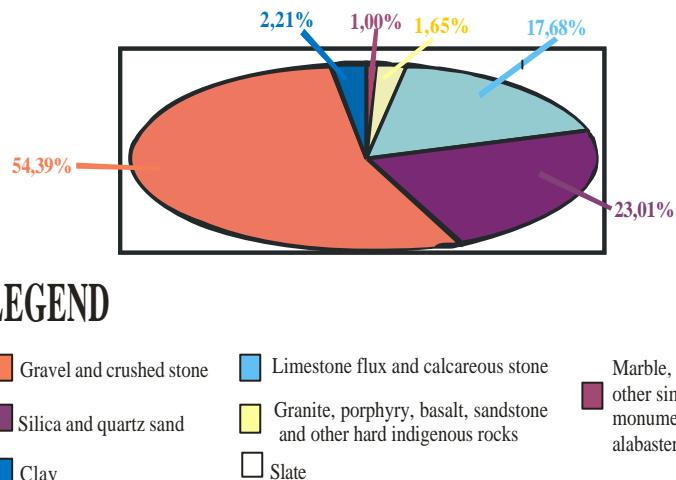


Figure 3 - The percentage level of world-wide building rock productions in the 2000s, by use criteria groups.

The power commodities production

The main raw power products obtained by exploration are fossil fuels: fluid hydrocarbons (crude petroleum, natural gasoline, natural gas) and coals (anthracite, hard coal, brown coal, peat for fuel).

As the demand for electric power increased, prospecting for coal was accelerated, obtaining for it the biggest quantity for all fuels (4.716 billion tones). In the world-wide statistics (2, 3, 4) coal are not presented as four genetic types, but as hard coal (3.867 billion tones) and lignite (0.849 billion tones). Hard coal is a coal with high degree of coalification and with a gross calorific value above 5000 kcal/kg on an ash-free but moist basis, and with a reflectance index of vitrinite of 0.5 and above. Lignite is coal a low degree of coalification which has retained the cellular structure of the vegetable matter from which it was formed and its gross calorific value is less than 5000 kcal/kg on an ash-free but moist basis, and with a reflectance index of vitrinite less than 0.5. In world statistics peat is not itself classed as coal, having a special position. It is mentioned that peat is used also in agriculture, not only for fuel. In the year 2000 about 18 mil. tones peat for fuel was obtained.

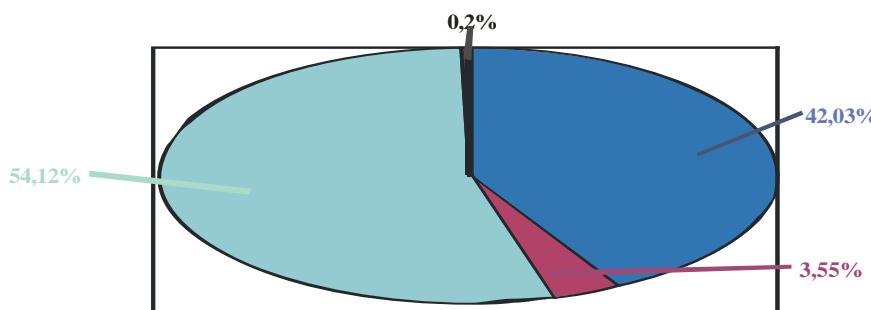


Crude petroleum (3426 mil. tones) is a mineral oil consisting of a mixture of hydrocarbons of natural origin, yellow to black in color, with variable specific gravity and viscosity, including crude mineral oils extracted from bituminous minerals (shale, bituminous sand, etc.).

Natural gasoline (29 mil. tones) is a light liquid hydrocarbon, extracted from wet natural gas, often in association with crude petroleum. It is used as petroleum refinery and petrochemical plant input and is also used directly for blending with motor spirit without further processing.

Natural gas (10316 pet joules) is a mixture of hydrocarbon compounds and small quantities of non-hydrocarbons existing in the gaseous phase, or in solution with oil in natural underground reservoirs at reservoir conditions. It may be sub-classified (3) into associated, dissolved or non-associated gas. Methane recovered from coal mine sand sewage gas are also included as well as natural gas liquefied for transportation.

To estimate the energy contribution of each fuel, all energy commodities was converted into coal equivalent (its gross calorific value is 7000 kcal/kg). Percentage distribution of world fossil fuels productions, as coal equivalent tones, may be observed in figure no 4. It could be mentioned that crude petroleum has the most important contribution at the world-wide production, because of its calorific power (higher with 1.454), despite that it id produced in smaller quantities like hard coal.



LEGEND

■ Crude petroleum	■ Hard coal	□ Peat for fuel-0,06%
■ Lignite and brown coal	■ Natural gasolene	□ Natural gas-0,04%

Fig. 4 The percentage distribution of world fossil fuels productions

Conclusions

By an intensive and less known world-wide exploitation activity, in the 2000s it had been obtain approximately (world statistics do not mention all metals) 684 million tones metal from ores with enough important contents to be economically recovered by preparation techniques and metallurgic processing.

Concerning to the medium exploitation content of each metal, it had been resulted that our discussed metals are included in almost 5 billion tones exploited ores. If to these metallic ores are added 0.613 million tones nonmetallic ores and industrial minerals (quantities of ore used for extracting are bigger but unknown), and also 6.035 billion tones building stones, results almost 11,600 billion tones of mineral commodities.

The world mineral and power commodities production in the 2000s is impressive: 21 billion tones. It is enough to satisfy the world economy necessities for raw materials and raw power.



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EPICLASTIC ORIGIN OF THE GREENSCHISTS OF THE TULGHES GROUP (LOWER ORDOVICIAN) IN THE EASTHERN CARPATHIANS

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Keywords: Tulghes Group; Ordovician epimetamorphic pile; Caledonian metamorphism; stratigraphic formations, metabasites; greenschists, basic metaepiclastites; impurified greenschists; non-impurified greenschists; back-arc basin.

The Tulghes Group (Tg) is a Caledonian epimetamorphic pile (> 6000 m): Lower Ordovician, regionally metamorphosed (under the conditions of the greenschist facies – Barrovian conditions – Kräutner et al, 1975) at the Arenigian / Llanvirnian boundary (Muresan, 2000 a – cf. K/Ar age of 272 m. y., determined by Mînzatu et al, 1975 – and cf. ordovician Chitinozoa, determined by Vaida, 1999).

I. GENERAL LITHOSTRATIGRAPHY AND TECTONICS OF THE TULGHES PILE. (5,000-6,000 m)

Tg is made up of four lithostratigraphic formations (Bercia et al., 1976; Voda, 1980; Kräutner et al, 1992): **Tg1** (> 500 m) – quartzitic; **Tg2** (> 500 m) – graphitous schists, black quartzites (metalydites) whith syngenetic manganese ore; rarely, intercalations of limestones; **Tg3** (2000-2500 m) – terrigenous schists with thick intercalations of rhyolitic metavolcanics (metatuffs, metalavas and epiclastics) and syngenetic ores Kuroko type (Kräutner, 1965, 1984, 1989; Muresan, 1996, 2002), rarely, greenschists and metabasites; **Tg4** (> 2500 m) – terrigenous schists with intercalations of quartz-feldspar blastodetrital rocks, greenschists and metabasites, rarely, rhyolitic metavolcanics, black quarzites and limestones. In the metamorphic socle of Bucovinian Nappe (sensu Sandulescu, 1975, 1984), the metamorphics of the Tulghes Group constitute some charriage nappes (Model of Kräutner & Bindea, 1995; Model of Muresan, 1996, 2002). In our conception, the succession (from bottom to top) and lithostratigraphic content of these tectonic units (post-metamorphic shear nappes – ante-mesorectaceous) is as follows: (1) **Capu Corbului Nappe (CCN)** (constituted by Tg 1 and Tg 2); (2) **Sumuleu Nappe (SuN)** (formed by Tg4); (3) **Sadocut Nappe (SaN)** (constituted by Tg3); (4) **Belcina Nappe (BeN)** (whith Tg 4 - petrographically similar with the term corresponding to the Sumuleu Nappe); (5) **Balan Nappe – BaN** (constituted by Tg 3 and Tg 4 – according to Kräutner & Bindea, 1995).

II. LITHOSTRATIGRAPHIC POSITION OF THE GS.

In the Tg (Tg3 and, specially, Tg4) one can find, at some levels, greenschists (GS) horizons usually together with (or in the proximity of) metabasites (MBas), where the metagabbros (MGb) are clearly predominant (the metadolerites – MD_O are rare). The MGb and MD_O (predominantly sills, with the metric-decametric thickness and with kilometrical lengths) are represented, particularly in some parts of the Tg4, in **Sumuleu Nappe (SuN)**, **Belcina Nappe (BeN)** and **Balan Nappe (BaN)**.

(a) Most MBas (metagabbros) are situated in BeN (Muresan, 1996, 2004; the name of MBas and GS levels from: G. Muresan, 1969; H.G. Kräutner et al., 1978, 1983, 1990): the **Sipos, Varsaraia** and **Gherpotoc Metabasites**, usually together with (or in the proximity of) the greenschists horizons (**the Sipos Greenschists Horizon; the Varsaraia Greenschists Horizon; the Gherpotoc Green Schists Horizon**).



(b) In BaN (after Kräutner & Bindea, 1995), in Tg4, one can find the **Arama Oltului Metabasites** (metagabbros) associated to **Arama Oltului Greenschists**; in BaN, the metabasites are also present in Tg3 (**the Sedloca Metabasites** – metagabbros). (c) We also mention the presence of the **Isipoaia Metabasites** (with ancient doleritic structures – Kräutner et al., 1986) and of the **Gârbele Greenschist** in lower part of Tg3 of the Sadocut Nappe. The frequent association of MBAs with GS horizons, as well as some geochemical differences (Kräutner et al., 1978, 1986, 1990; Bindea et al., 1993) between MBAs intruded at different levels of the Tg metamorphics enable us to suppose that the MBAs are younger and younger as they are located at higher and higher levels in Tg pile.

III. MAIN MINERALOGICAL, PETROGRAPHICAL AND CHEMICAL FEATURES

A. Mineralogical data.

GS (also called "basic metatuffs" or "diabasic metatuffs") have the characteristic paragenesis chlorite + albite, to which minerals of the epidote group, actinote, calcite, locally biotite and, in small amounts, titanite, magnetite and hematite are frequently associated. The possible presence of paragonite is revealed by the occurrence of GS in the muscovite + paragonite - epidote - spessartine field, on the ACF diagram (Turner, Verhoogen, 1967). Some GS also contain stilpnomelane (e.g. Tg 4 in BeN, well drilling 533 of Pângarati region), postkinematically grown (Variscan ?). This mineralogical composition of GS resembles metamorphically some primary deposits with a basic chemistry. The greater amounts of quartz, sericite or biotite (indicated by higher percentages of SiO₂, Al₂O₃, K₂O) point to clayey-sandy supplies in the primary material, which generated GS; excessive amounts of calcite can also be found in GS. These impurified rocks, previously described as "basic metatuffs" prevail quantitatively in comparison with the "non-impurified" ones. The chemical composition of the latter points out the appurtenance to basalts (G. Muresan, 1969; Muresan, Muresan in Bercia et al., 1971; Kräutner et al., 1986; Muresan, 2000 b). In case of the impurified GS, this appurtenance to the basic rocks is more "diminished", still recognisable. The high contents of Na₂O (3.5 - 4 % even more) in some GS are considered to be due to some halmyrolytic redistribution in the initial deposits.

We consider that the generic term "**greenschists**" (eventually with the specification "**non-impurified**" and "**impurified**") should be used instead of previously used terms such as "tuffogene greenschist", "basic metatuff", "basic metatuffite", "diabase greenschist" etc.

B. The most widespread GS are, as follows:

- (1) **non-impurified GS:** (a) chlorite-albite schists (chlorite 45-55 %; albite 20-30 %; quartz 8-10 %; calcite 1-7 %; magnetite 0-2 %; sporadically: epidote and apatite); (b) chlorite-calcite-albite schists (chlorite 25-45 %; calcite 15-30 %; albite 8-20 %; quartz 5-10 %; magnetite 0,5-2 %; sporadically: epidote, apatite); (c) chlorite-albite-epidote schists (chlorite 30-45 %; epidote 18-25 %; albite 5-30 %; quartz 5-10 %; calcite 0-10 %; sporadically: actinote, magnetite, apatite, rutile, sericite); (d) chlorite-actinote-epidote schists (chlorite 25-35 %; actinote 20-35 %; epidote 20-25 %; quartz 5-10 %; albite 3-8 %; sporadically: calcite, biotite, magnetite, apatite).
- (2) **impurified GS:** (a) albite-chlorite-titanite schists (albite 25-30%; chlorite 20-30 %; quartz 10-15 %; titanite 5-10 %; magnetite 5-10 %; sporadically: epidote, sericite, apatite); (b) chlorite-albite-sericite-quartz schists (chlorite 15-30 %; sericite 5-20 %; quartz 10-20 %; albite 10-20 %; calcite 0-5 %; magnetite 1-3 %; sporadically: titanite, rutile, zircon, apatite); (c) chlorite-calcite-albite-sericite schists (chlorite 20-40 %; calcite 15-30 %; quartz 5-10 %; albite 10-20 %; sericite 5-15 %; magnetite 0-3 %; sporadically: epidote, apatite); (d) chlorite-sericite-albite-epidote schists (chlorite 23-37 %; sericite 10-20 %; albite 5-17 %; epidote 5-15 %; quartz 5-10 %).



%; magnetite 1-3%; sporadically: calcite, titanite, apatite); (e) chlorite-sericite-epidote bearing biotite schists (chlorite 20-35 %; sericite 20-25 %; epidote 10-15 %; biotite 2-15 %; quartz 5-10 %; albite 5-10 %; magnetite 0-3 %; sporadically: titanite, apatite); (f) chlorite-quartz-albite bearing biotite schists (chlorite 20-40 %; quartz 10-15 %; albite 5-10 %; biotite 1-10 %; sporadically: calcite, epidote, magnetite).

C. Metamorphism of the GS.

Unlike MGB (which present an incomplete synmetamorphic, textural, structural and mineralogical adaptation (Muresan, 1998, 2000 a, 2000 c), in the case of GS there is a complete textural, structural and mineralogical adaptation to the thermo-baric conditions of the greenschists facies. This can be explained by the fine-grained texture (therefore a large reaction surface) of the primary material as well as by the higher water amounts (which favoured the metamorphic reactions and the formation of the penetrative schistosity) existing within it. The obvious geochemical of most MB with common basic intrusive rocks (gabbro, gabbrodiorite) as similarity of GS with basalts rocks proves that their regional epimetamorphism occurred in practically isochemical conditions (excepting water and CO₂ balance).

D. Chemical composition of the GS

(51 analysis – after G. Muresan, 1969; M. Muresan et al., 1972; C. Caruntu et. al., 1972; N. Petrescu, 1976; H.G. Kräutner et al., 1978, 1983, 1990; G. Bindea et al., 1993; Al. Voda et al., 1996): **SiO₂** = 32.98 – 49.86 %; **TiO₂** = 1.21 – 3.38 %; **Al₂O₃** = 12.73 – 17.66 %; **Fe₂O₃** = 2.00 – 10.51 %; **FeO** = 4.18 – 10.26 %; **MnO** = 0.11 – 0.46 %; **MgO** = 4.18 – 9.69 %; **CaO** = 4.22 – 11.95 %; **Na₂O** = 1.10 – 5.47 %; **K₂O** = 0.10 – 1.51 %; **P₂O₅** = 0.17 – 0.44 %; **CO₂** = 0.80 – 9.75 %. **N.B.** TiO₂ > 0,8 % is specifically for basic rocks (P. Lapadue-Hargue, (1958)).

IV. SYNDEPOSITIONAL EVOLUTION OF THE PRIMARY MATERIAL OF THE GS.

During the accumulation, these basic deposits were subjected almost permanently to the carbon dioxide action (mostly of a volcanic origin), existing in the sea water, which extracted calcium from plagioclase, the latter being altered into albite. The calcium bicarbonate thus formed was either removed from the material, thus the present GS with albite and poor in calcium being formed, or it redeposited as calcite (most frequently). The latter case is rendered evident by chemical analyses, petrochemical parameters and corresponding diagrams, which indicate the initial provenance of calcite (included in calcite, present in GS) even from the primary basic material. It is true that if one removes from analyses CaO corresponding to CO₂ abberant appurtenances are obtained; on the contrary, if the whole amount of CaO is considered, the analyses indicate a basaltic character (gabbroid) for GS except those in which calcite deposited in excessive amounts.

The prevalence and the large amounts of GS with calcite are explained by the frequent action of CO₂ over the supplies of basic material. The calcite deposition, beside other processes, contributed to the diagenesis of the basic deposits. It is supposed (by analogy with the actual halmyrolysis processes) that the CO₂ action as well as that of other halmyrolytic processes could affect other components of the primary basic material, as well (e.g. pyroxenes, amphiboles, particles of volcanic glass, even albite formed as a result of the CO₂ action), thus sedimentary chlorites, clayey minerals, zeolites etc. could be formed. Finally, deposits with a complex mineralogical composition, enriched in water (by its inclusion in hydroxilate neominerals), have resulted. It is possible that prior to the Caledonian regional metamorphism, the primary deposits of Tg had been affected by an incipient burial metamorphism (anchimetamorphism), determined by their great thickness (at least 5000-6000 m).

V. GENETIC CONSIDERATIONS.

Up till now the researchers (inclusively us) who studied GS of the Tg accepted without any objection their basic tuffogene origin. Because very rare basic tuffs were found among the products



of the actual basaltic volcanism (from marine and oceanic basin domains) we agree, without many reserves, with the exclusive tuffogene origin of the GS. In our opinion, it is possible that the primary material of these rocks was mostly of epiclastic origin, formed of fine particles resulting from the erosion, by waves, of the basaltic volcanoes, the contemporaneous cases of rapid destruction (weeks or months) of volcanic islands being quite numerous. The epiclastic material that resulted could be transported at great distances (tens or hundreds of kilometres) by the marine currents as shown by the oceanographic researches on the nature and repartition of the different types of actual submarine deposits. Thus, the local great thickenings of some GS horizons cannot represent themselves apparatus of some paleovolcanoes. Unlike it, these GS agglomerations could indicate that the provenance (namely the basaltic volcanoes, apparatus, influenced by the waves movement) of the epiclastic material was not far away. Each GS horizon was penecontemporaneous with the formation of a generation of basaltic volcanoes as, during their emergence, they had been immediately destroyed by waves, thus the basic epiclastic material being formed. This situation explains the MB location in some GS horizons or nearly them; the formation of epiclastics could possibly be concomitant with the basic intrusive activity, which could accompany the basaltic volcanism. Thus, the formation and reiteration, in different moments, of the "greenschist-metabasite" binominal, materialized in the Tg pile as GS horizons, spatially associated with MB bodies, can be explained.

If one admits the preponderent epiclastic origin of GS, then one can coherently explain: **(a)** general basic chemistry of GS; **(b)** frequent impurifications with terrigene material; **(c)** frequent high amounts of titaniferous minerals (titanite, ilmenite, rarely rutile), the relatively high analytical percentages of TiO₂ (3-4 %), observed in many GS, which, in our opinion, can represent gravitational concentrations of heavy minerals with titanium (melanocrates of the titaniferous augite type inclusive), added to the existing titaniferous minerals within a basic magmatogene material; **(c)** frequent intercalation of the terrigene rocks (blastodetrital rocks, quartz schists, sericite-chlorite schist etc) in the GS horizons; **(d)** frequent and intimate alternation of non-impurified and impurified GS with terrigene material; principle of uniformity, mentioned above (scarcity of basic tuffs within the actual basaltic volcanism); **(e)** frequent relic blastodetrital structure of GS (round feldspath and quartz granules in the chlorite matrix).

VI. PRIMARY GEOLOGICAL CONTEXT

In our opinion, although GS are mostly formed as a result of the regional metamorphosis of epiclastics, they can still be affiliated to the basic magmatogene rocks due to the basaltic origin of the prevailing primary elements. GS and MBs being situated in the pile of the actual Tg, it is significant to define the environment in which the primary deposits of this pile accumulated. We can state that the Tg deposits originate in the marine medium, as proved by: **(1)** the presence of rocks coming from the marine medium, such as the metalydites (the actual black quartzites); **(2)** intercalation, at certain lithostratigraphic levels, of syngenetic manganiferous ores deposits (in Tg2) and of syngenetic ores Kuroko type; **(3)** presence of marine palynomorphs; **(3)** known great thickness (5000-6000 m) of the actual TG pile; **(4)** remarkable facial constant of the Tg1, Tg2, Tg3, Tg 4 Formations observed on hundreds of km² within the Crystaline Mesozoic Zone. This situation, correlated with the bimodal character of the pre-metamorphic magmatism (revealed also by the coexistence in the TG pile of the rhyolitic and basic magmatogene products), excludes the possibility of the formation of this pile in an oceanic medium or under conditions of island arc. One can infer that the sedimentary medium was offered by a marine back-arc basin. The great thickness of the known pile of Tg can be explained by the uninterrupted accumulation of the primary deposits under conditions of quasicontinuous subsidence of the marine basin basement, of the above-mentioned type.



VII. EPICLASTIC ORIGIN OF THE GRENSCHISTS FROM OTHER PALEOZOIC PILES FROM THE CARPATIANS.

Based on the data from literature, we consider that we can generalize the major conclusions of the present paper for all greenschists from the epimetamorphic piles from the Carpathians. Thus, in many caledonian and sudetic epimetamorphic piles (greenschists facies), from the Southern Carpathians, Eastern Carpathians and Apuseni Mountains, are often located the greenschists (GS) levels. Following are some representative epimetamorphic piles with GS: (a) **Caledonian epimetamorphic piles: in the Southern Carpathians**: Locva Group – Ordovician (Maier, 1974; Maier, Visarion, 1976; Iancu, 1986), Batrana Crystaline (sensu Muresan, 2003) – Lower Ordovician – Poiana Rusca Massif (Krautner et al., 1969, 1973); **in the Apuseni Mountains**: Biharia-Muncel Group (Dimitrescu, 1958, 1966, 1976; 1994; Visarion, Dimitrescu, 1971; Visarion, unpublished data; Solomon et al., 1984); (b) **Sudetic epimetamorphic piles: in the Southern Carpathians**: Lescovita Group (Maier, 1974; Maier, Visarion, 1976) – Devonian; Poiana Rusca Crystalline s.s. (sensu Muresan, 2000 a, 2003) – Devonian (Krautner et al., 1969, 1973; Muresan, 1973, 1998); **in the Eastern Carpathians**: Repedea Group (Silurian) and Cimpoiasa Group (Devonian-Lower Carboniferous) – Rodna Mts. (Krautner, Krautner, 1970; Iliescu Krautner, 1975, 1976; Krautner, 1987); **in the Apuseni Mountains**: Paiuseni Group (Devonian – Lower Carboniferous) – Highis-Drocea Mts. (Savu, 1962, 1965, 1986; Giusca, 1962, 1979; Giusca et al., 1964 ; Visarion, in Istocescu, 1971).

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L'ALLOTOCHTONIE TECTONIQUE ALPINE DES UNITÉS DE LA DOBROGEA ET LEUR STRUCTURE INTERNE (ROUMANIE)

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Dobrogea (Db) est constituée, au jour, par trois unités tectoniques majeures: **Db Méridionale (DbM)**, **Db Centrale (DbC)** et **Db Septentrionale (DbS)**; plus au nord, il y a la quatrième unité, la **Dépression Prédobrogéenne (DPD)**, aujourd’hui complètement couverte par le dépôts tertiaires et quaternaires. Tous les chercheurs sont d'accord qu'entre ces unités il y a des contacts (plans) tectoniques, mais concernant la nature et la localisation des ces contacts il y a des opinions différentes. DbM et DbC reviennent au Plateforme Moésienne; DbS englobe principalement l'Orogen Nord Dobrogéen (OND), qui (cf. Sandulescu, 1980, 1984) s'a formé pendant les mouvements cimmériens (anciens et neufs – intra-néocomiens), ou même pendant le mouvements du Crétacé inférieur (les mouvements austro-alpins).

I. L'ALLOCHTONIE TECTONIQUE DE LA DOBROGEA MÉRIDIONALE (DbM)

A. La constitution et la structure de la Dobrogea Méridionale (DbM).

Sous une couverture sédimentaire presque horizontale (Jurassique moyen et supérieure, Crétacé supérieur, Tertiaire), développée à surface tant dans DbM que dans la partie méridionale de Dobrogea Centrale (DbC), les forages, effectués dans la partie septentrionale de la DbM, dans la région Palazu Mare-Cocosu, ont mis en évidence trois entités métamorphiques: (1) **Cristallin de Iasi-Palazu (CIP)** – antéprotérozoïque (prekarélien – archéen ?); (2) **Groupe Palazu Mare (GPM)** – protérozoïque inférieur (karélien); (3) **Groupe Cocosu (GCo)** – protérozoïque supérieur (briovérien).

(1) **Le Cristallin de Iasi-Palazu (CIP)** renferme des gnaisses migmatiques, gneisses granitiques paragneisses, micaschistes et pégmatites (Ianovici, Giusca, 1961; Giusca et al., 1967). Nous avons adopté la notion CIP pour deux raisons: la constitution pétrographique est mixte, CIP englobant tant des roches terrigènes (des paragneiss et des micaschistes), tant des granitoïdes métamorphisés (granites gneissiques); enfin, pour souligner la similitude pétrographique avec l'ensemble rencontré dans le soubassement de la Plateforme de Moldavie (Plateforme est-européenne) par les forages (par exemple, dans la région Iasi-Todireni (Ianovici, Giusca, 1961; Giusca et al., 1967). Tout cet ensemble métamorphique, présent tant dans le socle de la Moldavie, que dans la DbM, a un âge prékarélien (Giusca et al., 1967), étant tout semblable avec la formation gneissique prekarelienne du Bouclier Ukrainien. Les âges K/Ar de 1777-1620 m.a., déterminés dans les roches du CIP, ont été interprétés (Giusca et al., 1967) comme des âges régénérés pendant le métamorphisme régional protérozoïque inférieur du Groupe Palazu Mare (GPM), disposé sur le CIP.

(2) **Le Groupe Palazu Mare (GPM)** englobe une pile (500-800 m d'épaisseur connue dans les forages; bien entendu qu'en réalité, l'épaisseur primaire de la pile du GPM est sensiblement plus grande, parce que sa partie supérieure a été écartée par l'érosion protérozoïque, avant de déposition du Groupe Cocosu (d'âge briovérien). GPM renferme une suite quasi-rythmique (parfois finement rythmique) des roches quartzeuses à amphiboles et magnétite, amphibolites ± magnétite ± cummingtonite, roches carbonatées (des roches calcaires, ankérites, dolomites), quartzites –



métasilicrolites, micaschistes, représentant une formation férugineuse, considérée (par Giusca et al., 1967 – selon sa pétrographie et les données K / Ar) un équivalent de la Série de Krivoi Rog, d'âge karelien (protérozoïque inférieur). La présence du andalusite et du sillimanite dans les micaschistes et l'absence du cordiérite montrent que le degré du métamorphisme régional s'a déroulé dans le faciès des amphibolites, autour de 600° C et 4-5 Kb (Giusca, 1977). Selon Maier & Mihu (1976; et dans Visarion et al., 1979), l'orientation générale de la stratification des roches du GPM est ONO-ESE / 35-40° SO. Tous les chercheurs antérieurs ont considéré que le GPM est disposé transgressivement sur le CIP. Tenant compte que: (a) à la base du CIP, il n'y a pas des roches détritogènes (métaarcoses), connues, en échange, à la base de la pile de la Formation de Krivoi Rog d'Ukraine; (b) entre les séparations lithologiques du CPM et la limite avec le CIP il y a un angle aigu (autour de 10°), fait observable dans les coupes géologiques (exécutées par: Maier, Mihu, 1976; Maier, Mihu dans Visarion et al. 1979); (c) la croissance visible vers le sud de l'épaisseur de la pile du GPM (observé par Maier, Mihu, 1976), nous considérons que le CPM est décollé tectoniquement (un court charriage?) vis-à-vis de CIP, événement synchrone avec le plissement synmétamorphique protérozoïque du GPM. Cette décollement a écarté les termes basales du GPM. Mais ces observations n'oblitèrent pas le fait que GPM s'a formé primairement dans un bassin sédimentaire, installé sur un ancien socle métamorphique, représenté par le CIP. Pour nous, la nature pétrographique du GPM, l'épaisseur de sa pile et son intense métamorphisme régional sont des arguments pour affirmer que la pile primaire du GPM s'a formée dans un milieu marin, probablement dans un bassin du type « back-arc » active, avec une profondeur modérée, mais affecté par des mouvements de subsidence. La présence des roches calcaires dans la pile du GPM, montre que les dépôts primaires se sont formés au-dessus de la limite d'existence du CO₃Ca. L'enrichement en fer (la présence, presque dans toutes les roches du GPM, des minéraux férisères (magnétite, ankérite, cummingtonite etc) accuse des apports hydrothermaux sous-marins rythmiques, qui se sont déposés concomitamment avec les roches primaires du GPM. C'est-à-dire, les concentrations férisères de Palazu Mare ont une génèse primaire hydrothermale-sédimentaire. Jusqu'à nos jours, il n'y a pas des preuves directes concernant l'existence du Groupe Palazu-Mare dans le soubassement métamorphique de la Plateforme de la Moldavie; les rares forages d'ici (Iasi, Todireni, Batrânesti) ont rencontré seulement de roches prékaréliennes du Cristallin Iasi-Palazu (Ianovici, Giusca, 1961; Giusca et al., 1967). En dépit de ces résultats, nous croyons qu'il est possible que des séquences du Groupe Palazu Mare ont été, peut-être, ménagées par l'érosion, hypothèse basée sur la présence d'une zone à plusieurs anomalies magnétiques positives, développées dans une assez large région, entre Iasi et la frontière septentrionale de la Roumanie (Airinei St. et al., 1983).

(3) **Le Groupe Cocosu (GCo)**, intercepté par les forages, effectués entre Constanta et Mangalia, représentent une pile épimétamorphique (la zone du chlorite), disposé transversalement sur le GPM. Le GCo renferme deux entités lithostratigraphiques (Giusca et al., 1967; Mirauta, 1969; Maier, Mihu, 1976 et dans Visarion et al., 1979): (a) **La Formation inférieure spilitique** (> 350 m d'épaisseur) – ayant, à la base, des roches blastodétritique, suivies par des métalaves spilitiques et des métapyroclastites basiques (selon notre opinion, des métäepiclastites basiques, tenant compte que les volcans basiques ont seulement des activités effusives, n'ayant pas des pyroclastites véritables); (b) **La Formation supérieure détritique** a une épaisseur connue > 300 m tenant compte que les parties supérieures du GCo ont été érodées, avant de la déposition du Cambrien, présent à la base de la couverture de la Plateforme Moéssienne. Cette formation renferme une pile térrigènes, formée par une alternance de métagrés arkosiennes, métamicroconglomérats, métaconglomérats and métasiltites). Les métaconglomérats contiennent fréquemment des gallets de quartzites à magnétite du type CPM. Basé sur l'âge maxima K / Ar (547 m.a. – Giusca et al., 1967), qui estime l'âge du métamorphisme du GCo, très ressemblant avec l'âge K / Ar (Giusca et al., 1967) du métamorphisme du Groupe des Schistes verts (développé dans la Dobrogea Centrale),



Mirauta (1969) a soutenu qu'il est possible équivaler les deux entités du Protérozoïque supérieur (qui se présentent sous des facies sedimentaires différents), qui correspondent au Briovérien du Massif Bohémien (Kräutner et al., 1988). Tenant compte que le GCo est disposé transgressivement sur un socle métamorphique (représente par le GPM) et que, dans sa moitié inférieure, le GCo renferme de roches magmatogènes basiques et des épiclastites basiques, nous pouvons affirmer que cette entité s'est formée primairement dans un bassin du type «back arc», installé sur un socle métamorphique – représenté par le GPM (c'est-à-dire sur une croûte continentale) et tout près d'un arc insulaire (d'où proviennent les roches basiques spilitisées et les épiclastites basiques). Dans la région Cocosu, le GCo est couvert (données de forages) par les dépôts jurassique (Bathonien, Callovien, Oxfordien, Kimmeridgien). Dans la partie méridionale de la DbM et à l'est de Danube, le GCo est couvert transgressivement par la couverture paleozoïque de la Plateforme Moésienne, qui commence avec Cambrien fossilifer – avec des trilobites – Mutiu, 1991, fide Ionesi, 1994).

B. Le contact tectonique de la Dobrogea Méridionale (DbM) avec la Dobrogea Centrale (DbC).

Dans la région Palazu Mare, il y a quelques forages qui, traversant le Cristallin de Iasi-Palazu (CIP), ont rencontré en profondeur les roches faiblement métamorphisées du Groupe de Schistes Verts (GSV) – largement développés dans la Dobrogea Centrale, montrant que le CIP chevauche tectoniquement la DbC. Tenant compte que le Groupe Cocosu (du DbM) et le GSV (du DbC), ont la même âge, mais se présentant sous des faciès différents – des faciès isochrones hétéropiques (fait remarqué par Mirauta, 1969), nous avons présumé (Muresan; 1971) que ce contact représente un charriage. Nous le dénommons le **Charriage Palazu (CP)**, intra-moésien, orienté NO-SE (données géophysiques et de forages). Dans une zone logée immédiatement au-dessus de ce plan de charriage, les roches du CIP sont rétromorphisées et milonitisées (Maier, Mihu, 1976; Visarion et al., 1979), fait qui relève l'ampleur du transport tectonique le long de plan du ce contact tectonique (la présence des roches dynamo-rétromorphosées à la base des entités mésométamorphiques, charriées sur les épimétamorphites, a été constatée dans les Carpates – par exemple, dans la Zone cristallino-mésozoïque des Carpates Orientales, à la base de la Nappe de Rarau, les mésométamorphites du Groupe Bretila sont fortement dynamo-rétromorphosés – cf. Muresan, 1967, 1984, 2002). Auparavant, la Faille Capidava-Ovidiu (FCO), située plus au nord de ce charriage (à surface, FCO affecte la couverture du Jurassique moyen et supérieur – ayant le flanc méridional fortement baissé – données de forages) a été considérée par tous les chercheurs – inclusivement nous – comme la limite tectonique entre la DbM et la DbC. Puisque, en profondeur, au nord et au sud de FCO, il y a des roches du Goupe des Schistes Verts (données de forages), spécifiques pour la DbC, devient évidemment que le contact tectonique réel entre la DbM et la DbC est représenté par le Charriage Palazu. Bien sur que ce charriage est plus récent que le faible métamorphisme régional fini-protérozoïque du GSV. A l'ouest de Dobrogea, les données des forages et les données géophysiques montrent que ce charriage affecte tant le soubassement que la couverture paléozoïque de la Plateforme Moésienne (Cambrien-Ordovicien-Silurien-Dévonien-Carbonifère- Permien) et du Trias. Nous mentionnons que Sandulescu & Visarion (1979) et Sandulescu (1984) ont prolongé vers NO ce contact tectonique (comme une faille) en dehors de Dobrogea. D'autre part, le charriage en discussion est «soudé» par les dépôts du Jurassique moyen (Bathonien et Callovien) et supérieur (Oxfordien et Kimmeridgien) – Draganescu, dans Draganescu et al., 1978 (la Feuille géologique Dorobantu, à l'échelle 1: 50.000). Cette situation montre que le Charriage Palazu est formée après le Trias et avant du Jurassique moyen, c'est-à-dire synchronement avec les mouvements cimmériens ante-bathoninnes, très bien exprimés dans la Dobrogea Septentrionale. Pendant le processus de charriage, l'ensemble formé par les métamorphites du CIP, du GPM et du GCo a été charrié sur le Groupe des Schistes Verts de la Dobrogea Centrale. Ainsi, dans notre conception, s'est formé, par cissaillement, la **Nappe de Palazu**.



(NP), une unité tectonique de socle, qui, en effet, constitue la Dobrogea Meridionale. Etant charriée, en dehors de Dobrogea, sur les entités lithostratigraphiques soumises à l'érosion, la NP est une nappe épiglyptique.

II. L'ALLOTOCHTONIE DE LA DOBROGEA CENTRALE (DbC)

A. La constitution et la structure de la Dobrogea Centrale (DbC).

Coformément à notre conception, qui sera exposée dans cet ouvrage, Dobrogea Centrale renferme, à jour, seulement la pile térrigène flyschoïde (Mirauta, 1964, 1965, 1969; Jipa, 1970; Anastasiu, Jipa, 1984) du Groupe des Schistes Verts (GSV), attribué au Briovérien supérieur (Glowacki, Karnkowski, 1963; Mirauta, 1965, 1969; Kräutner et al., 1988), selon les comparaisons lithostratigraphiques régionales (avec le Briovérien supérieur de la Bretagne, du Massif Bohémien et des Montagnes Swientokrzyskie), les données palynologiques (Glowacki, Karnkowski, 1963; Iliescu, Mutihac, 1965) et l'âge maxima K / Ar de 573 m.a. (Giusca et al., 1967), la dernière concernant l'âge du son métamorphisme régional. À l'est de Danube, par exemple dans le forage profond de Bordeiu Verde, les roches du GSV sont couvertes transgressivement par les dépôts du et du Cambrien, qui se trouve à la base de la couverture paléozoïque de la Plateforme Moesienne (Mutiu, fife Ionesi, 1994). Le GSV englobe des roches faiblement métamorphisées, représentées, principalement, par: des métagraywackes, quartzites, roches métapélitique et des phyllites à chlorite. Dans la partie supérieure du GSV, il y a aussi des métaconglomérats (Mirauta, 1969). Puisque, le GSV est délimité tectoniquement tant vers le Sud, que vers le Nord, aujourd'hui, nous ne connaissons pas l'épaisseur primaire de la cette entité; pour la pile accessible à l'observations, Mirauta (1969) a estimé une épaisseur de 5000 m. Selon le même auteur (1964, 1965, 1969), la pile du GSV renferme quatre formations lithostratigraphiques, qui se développent, de bas en haut, successivement, du Nord vers le Sud; à surface, Mirauta a observé que le degré du métamorphisme du GSV se diminue vers le Sud. Concernant ce sujet, nous croyons que cette assertion est amendable, tenant compte que, vers le Sud, les formations lithostratigraphiques du SGV sont de plus en plus neuves, normalement de plus en plus moins métamorphisées; mais, en profondeur, le degré du métamorphisme peut-être est plus grande, semblable avec celui constaté dans la partie septentrale du GSV. La structure d'ensemble du GSV est caractérisée par un large plissement synmétamorphique (Mirauta, 1964, 1965, 1969), qui, en ensemble, constitue une structure descendante vers le Sud, fait qui explique l'apparition, dans cette direction, des termes de plus en plus neufs. Les roches du GSV sont affectées par de failles, principalement quasi-directionnelles, parmi lesquelles, les plus importantes sont la Faille Capidava-Ovidiu et la Faille Ostrov-Sinoe, les deux étant orientées NO-SE, avec des forts pendages vers le Sud et avec le flanc septentrional élevé; c'est-à-dire, ces sont des «failles normales». (a) **La Faille Capidava-Ovidiu – FCO** (mise en évidence par Ciocârdel & Patrulius, 1950) affecte, à surface, la couverture du Jurassique moyen et supérieur et, dans la profondeur, les roches du GSV (données de forages). (b) **La Faille Ostrov-Sinoe – FOS** (mise en évidence par Mirauta, 1969), sépare au nord les roches du SGV qui ont les directions structurelles orientées NO-SE, contrastant avec les directions E-O, existantes au Sud de FOS. Les roches, logées dans le flanc septentrional de la FOS, représentent les termes les plus profonds du GSV (Mirauta, 1964, 1969) et ont, parfois, des intercalations métépiclastiques basiques (auparavant, considérés comme des roches tuffogènes basique – Peters, 1867; Motas, 1913; Bujor, 1936; Cosma et al., 1962). Probablement, en liaison avec des processus hydrothermales-sédimentaires, dans les roches du ce compartiment, se sont formées fréquemment des disséminations finement grenues de magnétite (qui provoquent des anomalies magnétiques positives (mises en évidence par Pesky, 1968), fait inexistant dans les autres aires de développement du GSV. Selon notre opinion, la concentration de sulfures près de Ceamurlia (au SO de Movila Goala), représentée par des imprégnerations, nids lentilliformes, petites lentilles, faites en pyrite



(prédominamment), chalcopyrite, magnétite, quartz, chlorite – est liée par une activité hydrothermale-sédimentaire sous-marine, synchrone avec la déposition des épiclastites basiques (mentionnées) et des sédiments primaires du GSV. De même, au Nord de FOS, les roches du GSV représentent des véritables schistes épimétamorphiques (dans la Colline Dalâcă, est cité même le chloritoïde – Motas, 1913). L'épaisseur très grande de la pile du GSV s'explique, premièrement, par une subsidence quasi-continue qui a affecté le bassin marin (Anastasiu, Jipa, 1984), où s'a formée cette entité. En même temps, les apports des sédiments étaient abondants, situation explicable: les sources continentaux ont eu des reliefs (d'âge protérozoïque) à haute altitude. Le transport des matériaux dans le milieu continental a été, sûrement, court et rapides, les arguments principaux étant la présence des métagraywackes, l'hétérogenité des lithoclastes et des minéraux détritiques présentes dans ces roches et les formes non-arrondies des ces éléments (Mirauta, 1964, 1965, 1969; Jipa, 1970; Jipa & Anastasiu (1984). Dans le milieu marin, le transport s'a fait par l'intermédiaire des courants de turbidité (Jipa, 1970; Anastasiu, Jipa, 1984), ainsi se gardant les formes irrégulières des éléments mentionnés. Anastasiu & Jipa (1984) ont démontré que les sources continentales des matériaux sédimentaires se trouvaient au Sud (comme ont admis: Mirauta, 1969; Jipa, 1970), à l'Ouest et au Nord du bassin en question. Tenant compte de composition des lithoclastes et des minéraux détritiques présentes dans les métagraywackes du GSV, on peut apprécier que les premières deux sources étaient représentées dominamment par les roches du Cristallin de Iasi-Palazu et les métamorphites du Groupe Palazu Mare. Aujourd'hui, nous pouvons faire quelques comparaisons nouvelles plus exactes entre le GSV et le GCo: (a) ainsi, la pile du GCo peut être mise en parallèle principalement avec les parties inférieures du GSV, développées, principalement, entre la Faille Peceneaga-Camena et la Faille Ostrov-Sinoe, qui ont des rares intercalations de épiclastites basiques, en échange, très bien développées dans la moitié inférieure du GCo (la Formation inférieure spilitique), qui renferme, principalement, des métalaves spilitiques et des météepiclastites basiques). (b) Le soubassement du GCo est connu directement (données de forages); étant représenté par le Groupe Palazu Mare (GPM), disposé, à son tour, sur le Cristallin de Iasi-Palazu (CIP); cette situation représente pour nous une indication forte (à côté de présence, dans les métagraywackes du GSV, des gallets du type CIP et du type GPM) que le soubassement du Groupe des Schistes Verts (GSV) a été représenté par les métamorphites des deux entités; c'est-à-dire que tant le GCo, que le GSV se sont formés dans le même bassin marin, installé sur un socle métamorphique d'une plaque continentale. Il s'agit d'un bassin du type back-arc, dont sa partie méridionale, où s'a formé la pile du GCo, était très près d'un arc insulaire (d'où proviennent les roches magmatogènes basiques du GCo). Puisque, les roches du GSV se prolongent vers NO de Dobrogea, dans le soubassement de la Plateforme Moésienne (données de forages) et puis sous les Carpates Orientales, réapparaissant dans les Montagnes Swientokrzyskie, on peut apprécier que le bassin primaire du GSV était fortement allongé, selon une direction générale NNO-SSE.

B. Le contact tectonique de la Dobrogea Centrale avec la Dobrogea Septentrionale.

Nous avons démontré (Muresan, 1971, 1972) que l'ensemble du GVS est charrié sur les mésométamorphites du Groupe Altîn Tepe (AT – Protérozoïque moyen), ainsi constituant la Nappe Istrienne. Les deux entités, vers le Nord, sont limitées par la Faille Peceneaga-Camena (FPC), orientée NO-SE et fortement inclinée vers le Sud – faille normale. Aujourd'hui, on peut préciser que la FPC s'a formée dans un intervalle post-jurassique supérieur (après les dépôts de l'Oxfordien, de la région Cârjelari – mises en évidence par Gradinaru, 1981) – ante-cénomanien, parce que le Cénomanien, le Turonien et le Coniacien du Bassin Babadag (étudiés par: Mirauta & Mirauta, 1964; Szász, dans Szász et al., 1981) couvrent le tronçon du SE de la FPC (dans Dealul Ienicerilor). En échange, la FPC est partiellement rejouée (Muresan, 1971, 1972) après la déposition des Cénomanien-Turonien-Coniacien, le long d'un tronçon (assez court), développé au Nord Fântâna Mare (Baspunar). Puisque, dans la Dobrogea septentrionale, les dernières mouvements (cf.



Sandulescu, 1980, 1984) se sont produit, peut-être, soit intra-néocomien (les mouvements cimmériens neufs), ou pendant le Crétacé inférieur (les mouvements austro-alpin ou même autrichiens) et que la FPC interrompt les structures alpines de la Dobrogea septentrionale, nous supposons que cette fracture sa formée dans l'étape finale des ces mouvements, c'est-à-dire pendant le Crétacé inférieur. Par conséquence, le charriage de la Nappe Istrienne (c'est-à-dire du GSV) étant plus vieux que le segment ancien de la FPC, il en résulte qu'il est jurassique; cette assertion est soutenue par les âges K /Ar de 193-202 m.a. (c'est-à-dire pendant du Jurassique inférieur), obtenus (Semenenko et al., 1969) sur les épreuves de micaschistes voisins avec le plan de charriage istrien. Encore de sa découverte (Pascu, 1909; Mrazec, 1910; Mrazec, Pascu, 1912; Macovei, 1912), la Faille Peceneaga-Camena (FPC) a été considérée (inclusivement par nous) comme un contact tectonique tranchant entre DbS et DbC; puis, la FPC a été appréciée comme un plan de charriage, le long de DbC, qui chevauche la DbS (Mrazec, 1912; Preda, 1959, 1964; Muresan, 1971; Olteanu, 1975). En ce qui concerne le contact entre la DbC et la DbS, notre opinion d'aujourd'hui est bien différente et sera exposée et argumentée dans cette partie du notre article. Ainsi, dans notre conception, le Groupe Altîn Tepe représentent un élément tectonique de l'Orogen Nord-Dobrogéen (récemment, nous avons exposée cette idée sommairement – Muresan, 2004), logé au sud de Faille Peceneaga-Camena. En ce cas, les relations tectoniques primaires réelles entre la Dobrogea Centrale et celle Septentrionale sont exprimées par le charriage du Groupe des Schistes Verts sur l'AT (c'est-à-dire le charriage de la Plateforme moésienne sur la Dobrogea du Nord), formé antérieurement vis-à-vis de Faille Peceneaga-Camena. Par conséquence, nous attachons le Groupe Altîn Tepe à la Dobrogea Septentrionale (DbS), c'est-à-dire à la Nappe de Macin, la plus interne unité à métamorphites de la DbS. Nos arguments principaux seront exposés au-dessous: (1) Dans la Nappe de Macin, la plus interne unité tectonique majeure de la Dobrogea Septentrionale, il y a deux entités mésométamorphiques protérozoïques (le Groupe Orliga, le Groupe Megina), semblable avec le Groupe Altîn Tepe, tant pétrographique (les trois entités ont des roches terrigènes à des intercalations d'amphibolites magmatogènes basiques), que le degré du métamorphisme (les trois groupes sont mésometamorphisés régionalement au niveau du faciès des amphibolites à almandine); (2) Au ONO de localité Altîn Tepe, près de Sommet Sacar Bair II, il y a une apparition (des blocs) faite en rhyolites (Vlad Roșca, information orale) du même type avec les roches rhyolitiques mésozoïques, logées au nord de Faille Peceneaga-Camena (FPC). Jusqu'aujourd'hui, ces roches eruptives étaient connues seulement dans la Dobrogea Septentrionale, située au nord de FPC. C'est-à-dire, au sud de FPC, il y a des éléments nord-dobrogéens.

III. L'ALLOTOCHTONIE DE LA DOBROGEA SEPTENTRIONALE

A. La structure d'ensemble de la Dobrogea Septentrionale (DbS) et son contact avec la Dépression Prédobrogéenne.

Dans la DbS se développe l'Orogen Nord Dobrogéen (OND), alpin, la plus complexe unité tectonique majeure de la Dobrogea, comme résultat d'une très longue évolution, (autour d'un milliard d'années), commençant avec le Protérozoïque moyen, jusqu'à dans le Néocomien, même dans le Crétacé inférieur (selon Sandulescu, 1980, 1984). Aujourd'hui, la tectonique alpine est la plus visible, étant exprimée par une structure en nappes de charriage, orientées NO-SE, qui sont interrompues vers le Sud par la Faille Peceneaga-Camena. Nous mentionnons que toutes les nappes suivantes ont été mises en évidence par Mirauta (fide. Patrulius et al., 1973, 1974). (1) **Nappe de Macin (NMa)**, la plus interne unité tectonique majeure du l'OND, est formée, principalement, par des métamorphites protérozoïques (parmi lesquelles les Groupes AltînTepe, Megina et Orliga) et paléozoïques, des formations sédimentaires paléozoïques (parfois faiblement métamorphisées), des granitoïdes (varisques et anté-varisques) et, moins répandus, par des dépôts sédimentaires du Trias moyen et du Jurassique supérieur, tout l'ensemble supportant, vers SE, la couverture post-tectonique du Crétacé supérieur (du Bassin Babadag). La structure interne de la NMa est



compliquée par quelques écailles (ou nappes ?), parmi lesquelles, d'Orliga et de Cârjelari. Par sa constitution, la NMa représente une nappe de socle, épiglyptique, formée par cissaillement. (2) **Nappe de Consul (NCo)** renferme des dépôts du Trias moyen, percés par de basaltes et rhyolites triassiques (c'est-à-dire, comme résultat d'un magmatisme bimodal). Si les dépôts détritiques bréchieux évidemment polygénés, non-métamorphisés (la Formation de Mihai Bravu), connus à l'O de Mihai Bravu, revient au Paléozoïque supérieur (comment ont opiné Patrulius et al. (1974), il est possible que cette entité soit logée dans une fenêtre tectonique, soit représentent un mélange tectonique associée avec la Nappe de Consul. Vers NO, la NCo supporte la charriage de la NMa, le long de Ligne Luncavita-Consul-Babadag. Vers NO et vers SE, la NCo est dépassée par la NMa. (3) **Nappe de Niculitel (NNi)**, qui a un contour d'érosion fortement festonné – Fig. 1 (probablement due aux quelques failles transversales), renferme des dépôts du Trias moyen et des roches basaltiques triassiques; les roches dévonniennes, connues dans quelques affleurements à l'ouest d'Isaccea, sont logées, peut-être, dans une fenêtre tectonique (Sandulescu, 1984). La NNi est couverte tectoniquement par la Nappe de Consul. (4) **Nappe de Tulcea (NTu)**, qui supporte la NNi, englobe, au jour, quelques apparitions de roches pré-mésozoïques (paléozoïques, faiblement métamorphisées et des granitoïdes) et de roches basiques et acides triassiques; un grand volume revient aux dépôts triassiques et du Jurassique supérieur, inclusivement ceux du Malm (Gradinaru, 1974; 1984; Ionesi, 1994); vers la partie orientale de la NTu, en profondeur (données de forages), dans la zone du shelf de la Mer Noire, Catuneanu & Maftei (dans Ionesi, 1994) ont décrit, aussi, des dépôts triassiques et du Jurassique (inclusivement ceux du Malm) et des roches magmatogènes basiques (basaltes et andésites basaltoïdes) du Jurassique supérieur (données d'âge K / Ar), qui montrent que, dans la NTu, il y a des épreuves pour la continuation de l'activité effusive jusqu'à la fin du Jurassique supérieur. Nous avançons l'hypothèse qu'il est possible que la Nappe de Tulcea n'a pas un soubassement pré-mésozoïque; en ce cas, les dépôts de la NTu sont charriés sur les roches pré-mésozoïques du ce soubassement. À son tour, ce soubassement, que nous le dénommons la «**Nappe Soutulcéenne**» (NSTu) – unité de socle, est charrié, soit directement sur les formations de la Dépression Prédobrogéenne, soit sur une autre nappe inconnue aujourd'hui (qui s'interpose entre la Nappe Soutulcéenne et la Dépression Prédobrogéenne). Notre hypothèse est stimulée par les résultats obtenus par les méthodes de l'induction électromagnétique (Stanica & Stanica, 1985, Stanica et al., 1985; 1989, 1993, 1999) et par les méthodes géothermiques (Veliciu, 1987) qui montrent l'existence, en profondeur, l'allotoctonie tectonique de l'Orogen Nord Dobrogéen vis-à-vis de Dépression Prédobrogéenne (DPD). La Faille Sfântu Gheorghe (faille normale) lève fortement la DPD, ainsi que le plan de charriage situé à la base de Dobrogea Septentrionale est fortement levé (Fig. 3). Finalement, nous considérons que, par leurs contenus, la NCo, la NNi et la NTu (la dernière, si notre hypothèse est correcte), étant dépourvues de métamorphites et de granitoïdes, représentent des nappes de couverture, formées, selon nous, par obduction (la présence des produits du magmatisme bimodal dans ces nappes est due au caractère du rift intracontinental (intra-plaque, considéré d'âge triassique – Vlad, 1978; Savu et al., 1980; Sandulescu, 1984; Balintoni, 1997) sur une continent (représenté au moins par le soubassement de la NTu, englobée aujourd'hui, selon nous, dans la Nappe Soutulcéenne et, probablement, dans les autres unités inférieures inconnues vis-à-vis de cette dernière unité). L'autre marge continentale du ce rift a été, probablement, représentée par les formations anté-mésozoïques englobées dans l'actuelle Nappe de Macin (NMa). Nous croyons que la période d'existence du ce rift intraplaque, a dépassée le Trias, jusqu'à la fin du Jurassique supérieur (ou pendant le Néocomien), une épreuve étant la présence des dépôts sédimentaires et des magmatites basiques du Jurassique supérieur dans la Nappe de Tulcea. Sandulescu (1984) a considéré que la NTu s'est formée après le Jurassique, probablement pendant la Phase cimmérienne neuve (intranéocomienne) ou même pendant les mouvements plus neuves (pré-albien). Dans notre conception, pendant la fermeture (fini jurassique ou même intranéocomienne) du rift triassique-jurassique, se sont formés, par obduction, les NCo, NNi, NTu; les



NMa et NSTu (unités de socle, formées par cissilement) ont la même âge. Nous considérons que les nappes d'obduction (NCo, NNi, NTu) constituent «Le système des nappes egissiennes» (Aegissus, l'ancien nom antique de Tulcea), ou «Egissinides». Les études stratigraphiques détaillées récentes des dépôts mésozoïques des ces nappes (Gradinaru, 1974, 1984; Szász et al., 1981; Baltres, dans Seghedi et al., 1991; Baltres et al., 1992; Ionesi, 1994; Catuneanu & Maftei, dans Ionesi, 1994) ont montré qu'il y a des différences faciales notables entre la plupart des entités mésozoïques isochrones. Cette situation démontre que ces unités sont des véritables nappes de charriage, ayant à leurs bases des plans de charriages importants.

B. Traits des Groupes mésométamorphiques protérozoïques Altîn Tepe, Megina et Orliga.

(1) **Le Groupe Altîn Tepe (AT)**, étant logé au sud de Faille Peceneaga-Camena, est la plus méridionale entité métamorphique de la Nappe de Macin, composante tectoniques de l'Orogen Nord Dobrogéen L'AT renferme une pile (du Protérozoïque moyen) prédominamment terrigène (à des intercalations des amphibolites), dont l'épaisseur connue dépasse 2000 m (Muresan, 1969, 1971, 1972). L'omniprésence des paléorythmes sédimentaires (du type arénite-siltite) dans toutes les séquences terrigènes, confère à la pile du AT un caractère primaire de flysch (Muresan, 1971, 1972); un rythme (autour d'un mètre), est formé, surtout de paragneiss plagioclasique-micaschiste ou, parfois, de quartzite-micaschiste. L'AT est divisé en quatre formations lithostratigraphiques (Muresan, 1971, 1972): Les Formations AT1 (> 500 m), AT3 (850-1250 m+) et AT4 (> 350 m) sont, principalement, terrigènes (paragneiss plagioclasiques, micaschistes, quartzites) et ont un caractère flyschoïde. Dans l'AT4, est logé le gisement de sulfures à magnétite Altîn Tepe, de nature hydrothermal-sédimentaire (Muresan, 1969, 1972). La Formation AT2 (350-550 m) englobe, principalement, des paraamphibolites (des météaplastites basiques – Muresan, 2004) et des orthoamphibolites; selon nous (Muresan, 1972, 2004), les dernières constituent des anciens sills de gabbros (les épaisseurs maxima: 50 m). L'étude géochimique des ortoamphibolites et des roches terrigènes de l'AT nous a montré (Muresan, 2004) que la pile du AT s'a formée dans un bassin du type back-arc, près d'un arc insulaire, d'où proviennent les roches magmatogènes basiques logées dans l'AT. La pile de l'AT a été mésométamorphisée régionalement (dans le subfaciès staurotide-quartz du faciès des amphibolites à almandine – Giusca et al., 1967), pendant le Protérozoïque – cf. aux âges K/Ar de 696-711 m.a. (Codarcea Dessila et al., 1966; Giusca et al., 1967). Puisque le rétromorphisme est répandu dans les roches de l'AT (Berbeleac et al.; 1985; Nedelcu & Gridan, dans Nedelcu et al., 1987, 1988), c'est très possible que ces âges sont fortement rajeunis. Dans les métamorphites de l'AT, Nedelcu & Gridan (dans Nedelcu et al., 1987, 1988) ont mise en évidence quelques générations de déformations (surtout des clivages), qui accompagnent les transformations rétromorphes mentionnées.

(2) **Le Groupe Megina (Me)** est développé dans la Nappe de Macin, au nord de la Faille Peceneaga-Camena. À l'ouest de Danube, sous les dépôts tertiaires; la Nappe de Macin a été rencontrée par les forages executés à Frumusita-Branesti, qui ont mis en évidence des amphibolites attribués au Groupe Megina (Muresan et al., 1971); ultérieurement, ces roches ont été reconnues dans les forages de Calmatui, Piscu, Independenta (Seghedi, dans Rosca et al., 1994). Dans la partie centrale de la NMa, le Groupe Megina englobe, selon Seghedi (1980; Seghedi, dans Seghedi et al., 1980; Seghedi, dans Krautner et al., 1988; Seghedi, 1998); trois entités lithologiques (entre lesquelles il n'y a pas un ordre lithostratigraphiques): la Formation des Amfibolites de Roman Bair (considérés, par Seghedi, comme des tholeiites du type MORB – données géochimiques), la Formation de Boldea (quartzites muscoviteux, gneiss à oligoclase, schistes à porphyroblastes d'albite – c'est-à-dire, un ensemble terrigen-volcanogen acide), la Formation des Micaschistes de Saia Culak (à grenat et staurolite). Puisque, Seghedi décrit l'alternance étroite des Micaschistes de Saia Culak avec les roches de la Formation de Boldea (par exemple, dans le forage



70.301 Balabancea), nous pouvons conclure qu'il s'agit des séquences flyschoïdes, aussi présentes dans la pile de l'AT. Le degré du métamorphisme primaire du Me correspond au faciés des amphibolites à almandine (grenat et staurotide). Seghedi a démontré que le Me est polymétamorphique et polydéformational, situation constatée aussi dans les métamorphites de l'AT (Berbeleac, Stefan, 1984; Berbeleac et al., 1985; Nedelcu & Gridan, dans Nedelcu et al., 1987, 1988).

(3) **Le Groupe Orliga (Or)** est répandu dans la partie du NO de la Dobrogea Septentrionale, où constitue l'Écaille Orliga (ou la Nappe d'Orliga). L'Or (séparé par Mirauta & Mirauta, 1962; Mirauta, 1966; Patrulius et al., 1973) a été divisé par Seghedi (1980) en cinq entités lithologiques (sans un ordre lithostratigraphique): les Gneiss de Sararie, amphibolites (et, subordonnement, périclases), micaschistes interstratifiés avec des paragneiss (selon nous, probablement, il s'agit de séquences flyschoïdes), micaschistes, quartzites à muscovite. Dans quelques roches de l'Or, il y a garnet, kyanite et, plus rare, staurotide et sillimanite. Tous les âges K / Ar concernant l'Or sont rajeunis. Les seules indications sont les palynomorphes déterminées par Olaru – 2001, parmi lesquelles: *Satka undosa* (Janskauskas) Janskauskas, *Satka elongata* Janskauskas, *Eomyctopsis robusta* (Schopf) Knoll & Golubic, *Flabelliforma compacta* German, *Polytrichoides lineatus* (German) Jankauskas, *Leiotrichoites typicus* (German) Jankauskas. Ces formes indiquent l'âge protérozoïque de cette entité. Finalement, nous sommes d'accord avec l'âge protérozoïque moyen, accordé pour la pile primaire et pour le métamorphisme régional de l'Or (Kräutner, Savu, 1978).

Conclusions.

Dans la Nappe de Macin de la Dobrogea septentrionale, il y a trois entités mésométamorphiques protérozoïques, dont leurs piles renferme, principalement, des roches terrigènes et des amphibolites: (1) **le Groupe Altîn Tepe (AT)**; (2) **le Groupe Megina (Me)**; (3) **Le Groupe Orliga (Or)**. Tenant compte des traits de ces groupes (exposés plus haut), nous sommes d'accord avec l'équivalence des ces trois entités (AT, Me, Or), faite antérieurement par Ianovici & Giusca (1961), Giusca et al. (1967), Seghedi & Oaie (1993), Seghedi (1998), mais nous considérons que toutes ces entités occupaient initialement des positions lithostratigraphiques différentes dans la pile métamorphique primaire commune, que nous la dénommons "**Le Supergroupe Orliga-Megina-Altîn Tepe.**" (**ORMEAT**). Ainsi, nous considérons que dans la pile primaire de ce supergroupe, le Groupe Orliga a eu une position lithostratigraphique inférieure vis-à-vis de les autres deux entités (comment a opiné Sandulescu, 1984), parce que les amphibolites sont assez rares (marquant le début de l'activité du magmatisme basique) et le degré de métamorphisme est plus élevé (grenat, staurotide, kyanite, sillimanite – Seghedi, 1975; Seghedi, dans Kräutner et al., 1988). Sans avoir des arguments décisifs, nous supposons que le Groupe Megina s'a situé dans le deuxième niveau (ayant le plus grand volume des amphibolites, qui correspond au maximum de l'activité du magmatisme basique). Enfin, la troisième position a été occupée par l'AT, pour le raison que les amphibolites sont plus rares en comparaison avec la pile du Me, ça signifiant la diminution de l'activité du magmatisme basique. Tenant compte de: (a) l'épaisseur grande de la pile commune (seulement la pile connue de l'AT dépasse 2000 m); (b) la prédominance des roches terrigènes dans les trois entités; (c) la présence des roches volcanogènes acides, dans le Me; (d) l'omniprésence des métagraywackes, dans la pile de l'AT; (e) la présence des séquences flyschoïdes dans les trois groupes; (f) pour la pile de l'AT, a été démontré (sur la base des données géochimiques) que s'a formée dans un bassin du type back-arc, installé sur un plaque continentale active, voisin avec un arc insulaire (Muresan, 2004), nous pouvons conclure que la pile de l'ORMEAT s-a formée dans un tel bassin marin du type back-arc voisin avec un arc insulaire (d'où proviennent le roches basiques (les actuelles amphibolites)). Après le métamorphisme régional protérozoïque, la pile mésométamorphiques primaire du l'ORMEAT a été démembrée



tectoniquement pendant les mouvements calédoniens et/ou varisques et, bien entendu, alpines, ainsi que, aujourd’hui, nous ne connaissons pas les séquences lithostratigraphiques situées entre les piles des groupes examinés.

L'ALLOCHTONIE TECTONIQUE ET LA STRUCTURE DES UNITES DE LA DOBROGEA

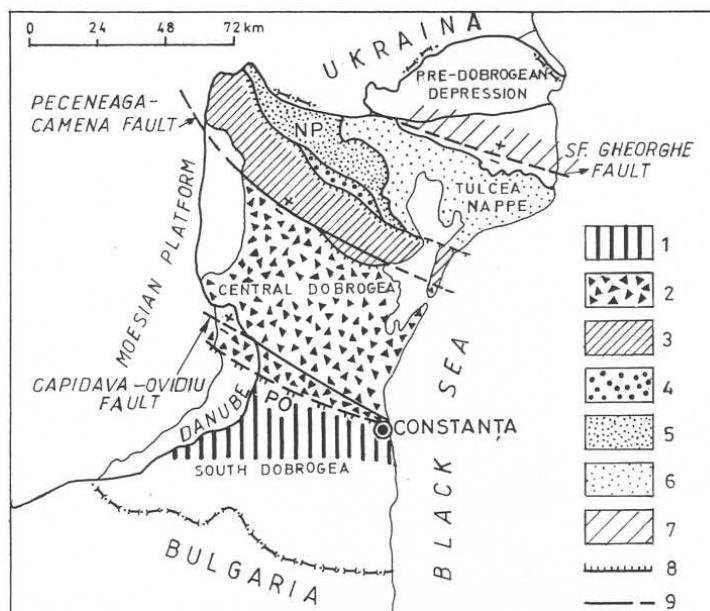


Fig. 1. Esquisse tectonique de la Dobrogea. 1, Dobrogea Méridionale; 2, Dobrogea Centrale; 3, 4, 5, 6, Dobrogea Septentrionale (3, Nappe de Măcin; 4, Nappe de Consul; 5, Nappe de Niculițel; 6, Nappe de Tulcea); 7, Dépression prédobrogéenne; 8, Charriage alpin; 9, Faille.

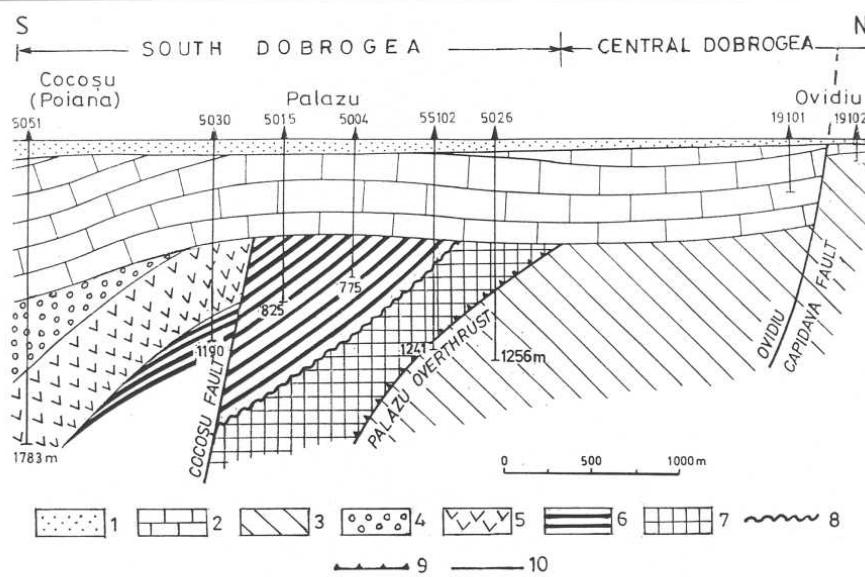


Fig. 2. Coupe géologique dans la zone de contact entre Dobrogea Meridionale et Dobrogea Centrale. (lithostratigraphie des métamorphites, selon Maier & Mihu, 1976; Mirăuță, 1965, 1969; Kräutner et al., 1988; Mésozoïque, selon Drăgănescu, dans Drăgănescu et al., 1978). 1, Crétacé supérieur et Tertiaire; 2, Bathonien, Callovien, Oxfordien, Kimmeridgien; 3, Groupe des Schistes Verts (Briovérien); 4, 5, Groupe Cocoșu – Briovérien (4, Formation supérieure détritique; 5, Formation inférieure spilitique); 6, Groupe Palazu Mare – GPM (Protérozoïque inférieur); 7, Cristallin de Iași-Palazu (Pré-



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karélien); 8, contact de décollement synchrone avec le plissement synmétamorphique protérozoïque du GPM (court charriage?); 9, Charriage alpin Palazu; 10, Faille.

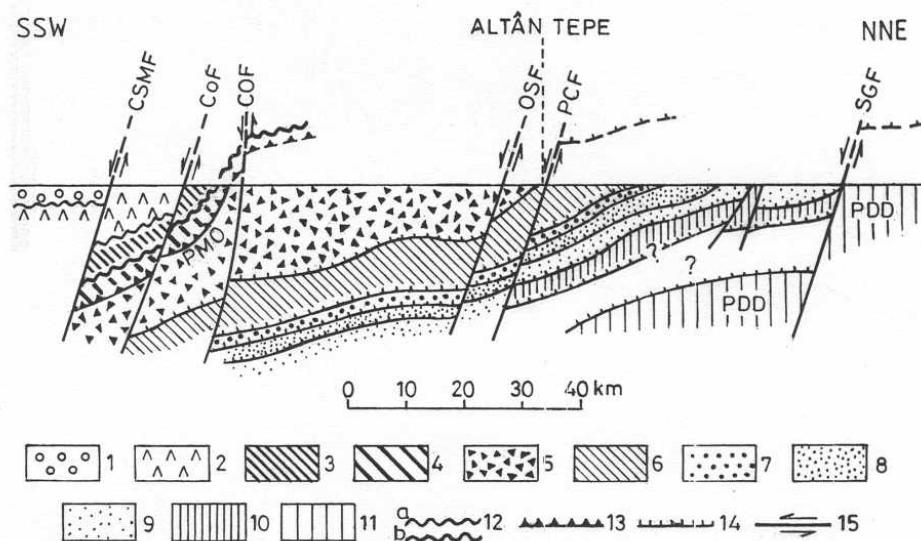


Fig. 3. Coupe géologique schématique transversale par la Dobrogea – les épaisseurs des nappes sont exagérées; la couverture mesozoïque de la Plateforme Moesienne n'est pas représentée. 1, Couverture paléozoïque de la Plateforme Moesienne; 2, 3, 4, Dobrogea Méridionale (2, Groupe Cocoșu – Briovérien; 3, Groupe Palazu Mare (Protérozoïque inférieur – équivalent de la Série de Krivoi Rog, d'âge karélien); 4, Cristallin de Iași-Palazu – CIP (pré-karélien); 5, Groupe des Schistes Verts (Dobrogea Centrale - DbC); 6, 7, 8, 9, 10, Dobrogea Septentrionale – DbS (6, Nappe de Măcin; 7, Nappe de Consul; 8, Nappe de Niculitel; 9, Nappe de Tulcea); 10, Unité soustulcénienne à formations paléozoïques et granitoïdes); 11, Dépôts mésozoïques (anté-crétacées supérieures) et paléozoïques de la Dépression Prédobrogéenne; 12, Contact de décollement synchrone avec le plissement synmétamorphique du Groupe Palazu Mare (court charriage?) vis-à-vis de CIP; 13, Charriage alpin Palazu de la Nappe Palazu sur le Groupe des Schistes Verts (Dobrogea Centrale – DbC); 14, Charriages alpins (entre DbC et DbS et des nappes de l'Orogen Nord Dobrogéen); 15, Faille; Abréviations: CSMF, Faille Cochirleni-Sud Medgidia; CoF, Faille Cocoșu; COF, Faille Capidava-Ovidiu; OSF, Faille Ostrov-Sinoe; PCF, Faille Peceneaga-Camena; SGF, Faille Sfântu Gheorghe; PDD, Dépression Prédobrogéenne



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**UNELE TRASATURI DE ANSAMBLU
ALE MINERALIZATIILOR METAMORFOZATE DIN ROMÂNIA**

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Keywords: Regionally metamorphosed mineralizations; hydrothermal-sedimentary mineralisations; Krivoi Rog type, Kuroko type, Scandinavian Type, Mississippi Valley type; Lahn-Dill type; Ghelar-Teliuc type; hydrothermal metamorphosed mineralisations; relict stratification; lithostratigraphic control; petrographic control; “hard” and “soft” minerals; metamorphic structural elements.

Abstract.**Some general traits of regionally metamorphosed mineralizations (RMM) in Romania.**

Such mineralizations are located in various predominantly terrigenous metamorphosed piles in the Carpathians (especially Eastern and Southern Carpathians) and in Dobrudja. In these piles also basic and / or acid magmatogene metamorphosed rocks are present. (1) Most RMM are syngenetic (concordant), having a hydrothermal-sedimentary origin (HSRMM). They were formed in: (a) the Early Proterozoic (Palazu Mare ferriferous HSRMM of the Krivoi Rog type, in Southern Dobrudja); (b) the Middle Proterozoic (HSRMM of sulphides with Altîn Tepe magnetite, mainly of Scandinavian type, but with transition to the Lahn-Dill type, in Dobrudja); (c) Middle Proterozoic (Gusev-Valea Blaznei lead-and-zinc bearing, HSRMM, Mississippi Valley type, in the Rodna Mountains of the Eastern Carpathians), (d) the Early Ordovician (Iacobeni-type manganese-bearing HSRMM; (d) in the Early Ordovician (sulphide HSRMM at Lesul Ursului-Fundul Moldovei, Kuroko type; both types in the Bistrita Mountains, Eastern Carpathians); (e) Silurian (HSRMM of Izvorul Cepii sulphides, Scandinavian type, in the Rodnei Mountains, Eastern Carpathians); (f) the Middle and Late Devonian (ferriferous HSRMM, in carbonatic facies of the Teliuc and Ghelar type and in the HSRMM Grenzenstein oxides facies of the Lahn-Dill type, both types in the Poiana Rusca Massif, Southern Carpathians). (2) In a few cases, the RMM are produced by the metamorphosis of some hydrothermal concentrations (for instance, the RMM of Muncelul Mic, located in the metarhyolites intrusive bodies of the Late Carboniferous of the Poiana Rusca Massif). (3) In all HSRMM one can find relict stratifications, recrystallized quartz metagels, primary facies variations (transitions from the ore to the surrounding rocks), as well as an obvious lithostratigraphic control. (4) The present configuration of the relations among metal-bearing minerals does not reflect the succession of the formation of minerals, but the latter's behavior (depending on their degree of tectonic competence) during the process of metamorphism (“soft” minerals of the chalcopyrite, galena, blende, etc. and “hard” minerals – cf. Ramberg, 1963).

Lucrarea de fata reprezinta o prima tentativa de sintetizare a celor mai importante caracteristici ale mineralizatiilor metamorfozate regional (MMR), aflate în tara noastră, realizata atât pe baza experientei autorului cât și, mai ales, pe baza numeroaselor date gitologice cuprinse în literatura românească, referitoare la astfel de mineralizatii. Progrese foarte importante ale studiilor gitologice s-au înregistrat începând din 1957, odata cu demararea studiilor complexe asupra zacamintelor singenitice ferifere din Paleozoiul epimetamorfic al Unitatii Epimetamorfice din Masivul Poiana Rusca (situat în partea nord-vestică a Carpaților Meridionali), concomitent cu descifrarea



litostratigrafiei acestuia și a evoluției magmatismului premetamorfic (ale cărui produse sunt înglobate în stiva acestui Paleozoic). Rezultatele obținute au reliefat importanța deosebită pentru determinarea corectă a genezei MMR, pe care o are corelarea datelor de zacamânt propriu-zise cu contextul geologic regional. Introducerea în literatura gitologică românăscă, de către Dimitrescu (1961), a conceptului de «**zacaminte vulcanogen-sedimentare**», s-a dovedit a fi fructuoasa pentru cercetarea MMR, concept ulterior dezvoltat și aplicat la studiul mineralizațiilor singenetic, la început, al celor ferifere din Poiana Rusca și apoi extins la cele de sulfuri din cristalinul Carpaților Orientali și al altor unități carpatici și dobrogene. În stadiul actual, se poate afirma că marea majoritate a MMR din România au fost încadrate genetic, deosebindu-se două tipuri principale: **MMR singenetic** (stratiforme) și **MMR hidrotermale** (filoniene). MMR sunt răspândite mai ales în metamorfitele din Carpații Orientali și Carpații Meridionali și, subordonat, din Dobrogea. Pe teritoriul românesc, MMR sau format în diferite perioade, începând din Proterozoicul inferior (Carelian), până în Paleozoic, ultimele MMR luând nastere în Carboniferul inferior. În lucrare, nu ne vom referi la MMR lichid-magmatice metamorfozate, localizate în unele metabazite și metaultrabazite din Carpați.

I. ENTITATILE METAMORFICE PRINCIPALE CU MMR.

A. Grupul Palazu Mare.

În Dobrogea Meridională, în regiunea Palazu Mare-Cocosu, sub o acoperire sedimentară (Jurasic mediu și superior; Cretacic superior, Tertiар), forajele au evidențiat metamorfitele **Grupului Palazu Mare (GPM)**, analogat (Ianovici, Giusca, 1961; Giusca et al., 1967, 1976) cu formatiunea feruginoasă a Seriei de Krivoi Rog, de vîrstă careliană (Proterozoic inferior). GPM stă peste un soclu gnaistic precarelian – arhaic? (Ianovici, Giusca, 1961; Giusca et al., 1967) și este acoperit transgresiv de epimetamorfitele brioveriene ale Grupului Cocosu. **GPM** (500-800 m grosime cunoscută) include o suita quasi-ritmica de roci cuartoase cu amfiboli și magnetit, amfibolite ± magnetit ± cummingtonit, roci carbonatice (calcare, ankerite, dolomite), cuartite – metageluri, micasisturi, tot ansamblul fiind metamorfozat în faciesul amfibolitelor, la aproximativ 600°C și la 4-5 Kb (Giusca, 1977). Îmbogătirea în fier (rezidență, în aproape toate rocele ale GPM, a mineralelor ferifere (magnetit, ankerit, cummingtonit etc) acuza aporturi hidrotermale submarine ritmice, care sau depus concomitent cu rocele primare ale GPM. În concluzie, concentratiile ferifere de la Palazu Mare au geneza primară hidrotermal-sedimentară.

B. Grupul Altîn Tepe.

În Dobrogea Centrală, imediat la sud de Falia Peceneaga-Camena, de sub sariajul (Muresan, 1971) Grupului Sisturilor Verzi, aflorează mezometamorfitele **Grupului Altîn Tepe – GAT** (Proterozoic mediu), care, în concepția noastră (Muresan, 2004, 2005), reprezintă cea mai sudică apariție a Pânzei de Macin, care aparține Orogenului Nord Dobrogean. GAT cuprinde o stiva terigenă flisoidală, cu intercalări de amfibolite, divizată în patru formări litostatigrifice (Muresan, 1969, 1971, 2004): AT1, AT2, AT3 și AT4. În AT4, este localizat MMR Altîn Tepe de sulfuri și magnetit, hidrotermal-sedimentar, de tip scandinav, cu trecere spre tipul Lahn-Dill.

C. Grupul Rebra.

În Carpații Orientali, în Zona Cristalino-Mezozoică, se dezvoltă, alături de alte entități metamorfice, stiva mezometamorfică a **Grupului Rebra** (Proterozoic mediu), divizată litostatigrafic în trei formări litostatigrifice (Kräutner, 1968, 1980; Bercia et al., 1976): Rb1 (terigenă), Rb2 (carbonatică, în opinia noastră, corespunzând unei paleoplatforme carbonatice) și Rb3 (terigenă-amfibolitică). În Rb2, sunt gazduite MMR Blazna-Guset de sulfuri (în principal,



galena si blenda), hidrotermal-sedimentatare, de tip Mississippi Valey (Udubasa, 1970, 1996; Udubasa et al., 1981).

D. Formatiunea de Razoare.

În Muntii Preluca (Insula Cristalina Preluca), în rocele terigene ale Formatiunii de Razoare, din Cristalinul de Preluca (analog Grupului Rebra), se localizeaza MMR Masca Razoare, manganifera, hidrotermal-sedimentara (Gherasi, Sandu, 1957; Savul et al., 1958; Giusca, 1962; Balan, 1976; Udubasa et al., 1996).

E. Grupul Fagaras.

În Muntii Fagaras, **Grupul Fagaras** (Proterozoic mediu) – reprezentat prin micasisturi, amfibolite si calcare, analogat cu Grupul Rebra (Kräutner, 1980), gazduieste (Dimitrescu, 1967; Schuster, Hîrtopanu, 1973) MMR Porumbacu-Arpaș de sulfuri (in principal, galena si blenda), hidrotermal-sedimentatare, de tip Mississippi Valey (Borcos et al., 1984).

F. Grupul Cibin.

În partea sud-estica a Masivului Poiana Rusca (între Silvas si Lunca Cernii), se dezvolta **Grupul Cibin (GCib)** – atribuit Proterozoicului superior (Kräutner, 1986) care a fost separat anterior sub numele de «Seria de Dabâca» (Maier et al., 1970, 1975). GCib a fost ulterior echivalat de Kräutner (1965, 1968, 1980) cu Seria de Sibisel si încadrat în Grupul Cibin. GrCib este metamorfozat regional pâna la nivelul almandinului (rocile retromorfozate sunt frecvente) si cuprinde trei formatiuni: vulcano-sedimentara bazica, carbonatica si blastodetritica. În prima formatiune, se localizeaza MMR Boita-Hateg de sulfuri, hidrotermal-sedimentara (Kräutner, 1965 a), care fiind asociata intim cu calcarele formatiunii mentionate, o atribuim tipului Mississippi Valey.

G. Grupul Repedea.

În Muntii Rodnei, pe un soclu metamorfozat (Grupul Bretila – Proterozoic mediu) se dispune transgresiv Grupul Repedea (Silurian), separat de Kräutner & Kräutner (1970) si datat palinologic de Iliescu et al. (1976). Grupul este constituit preponderent din sisturi terigene epimetamorfice (mai ales sisturi sericito-cloritoase), în care se intercaleaza calcare si MMR Izvorul Cepii de sulfuri, hidrotermal-sedimentara; partea bazala a Grupului Repedea contine sisturi verzi.

H. Grupul Rusaia.

În Muntii Rodnei, pe un soclu mezometamorfic (reprezentat prin Grupul Bretila), se dispune stiva terigenena epimetamorfica a Grupului Rusaia – GRu (Silurian - studiat de Fl. Kräutner, 1970; datat palinologic de Iliescu et al., 1976). În partea bazala a stivei, se intercaleaza, MMR ferifere tip Lahn-Dill Rusaia. GrRu a fost echivalat (Kräutner, 1987) cu Grupul Repedea.

I. Grupul Tulghes.

In Carpatii Orientali, în Zona Cristalino-mezozoica, Grupul Tulghes (GTg) are multe si însemnante MMR. GTg cuprinde o stiva groasa (> 5000 m grosime), formata în decursul Ordovicianului inferior, epimetamorfozata în timpul miscarilor caledoniene de la sfârsitul acestuia (Muresan, 2000 a). S-au deosebit patru formatiuni litostratigrafice (Bercia et al., 1976; Kräutner et al., 1983, 1992; Voda, 1980), Tg1, Tg2, Tg3, Tg4. Dintre acestea, Tg2 (o entitate constituita mai ales din cuartite negre – metalidite, carora li se asociaza, subordonat, sisturi sericito-grafitoase) contine MMR manganifere, hidrotermal-sedimentare **tip Iacobeni** (Savul, Ianovici, 1957; Balan, 1976; Hîrtopanu, 2004), iar la Holdita-Brosteni, MMR hidrotermal-sedimentare de baritina, rodocrozit, blenda si pirita (Voda, Voda, 1982), mineralizatie, care, prin caracterele ei, este unica la noi, fapt pentru care propunem sa fie desemnata ca un tip aparte: **«tipul Holdita-Brosteni»**. In Tg3,



(in principal, o formatiune cu metavulcanite riolitice – metatufuri si metalave, cu secvente de sisturi terigene) se localizeaza MMR de sulfuri **tip Kuroko** (Kräutner, 1965, 1970, 1984, 1989; Kräutner, Popa, 1973; et al., 1992; Muresan, Muresan, 1977 a; Zincenco, 1998; Muresan, 2002 a, 2002 b).. În Tg4, se cunosc mici MMR ferifere ± manganifere de tip **Lahn-Dill** (Muresan et al., 1972; Muresan, 2002 a), asociate cu sisturi verzi. De asemenea, pe alocuri, în epimetamorfitele Tg4, sunt intruse porfiroide riolitice (de Mândra), în care, uneori se localizeaza MMR hidrotermale filoniene, plumbo-zincifere, **tip Paltin** (Muresan, Muresan, 1977 b).

J. Cristalinul de Poiana Rusca s. s.

În Unitatea Epimetamorfica a Masivului Poiana Rusca, situat în partea de NW a Carpatilor Meridionali, peste un soclu metamorfic caledonian (Muresan, 2003), se dispune, discordant și transgresiv, **Cristalinul de Poiana Rusca s. s.**, care cuprinde stiva epimetamorfica varistica (sudeta), a Devonianului și Carboniferului inferior (Kräutner et al., 1973), subdivizată în trei entități litostratigrafice majore (Kräutner et al., 1969, 1973): **Grupul Govajdia** (la partea inferioară, detritogen; grafitos, la cea superioară) – atribuit Devonianului inferior; **Grupul Ghelar** (o stiva groasa, de peste 5000 m, predominant terigenă, cu frecvențe intercalatii de sisturi verzi și cu puternice secvente carbonatice) – Devonian mediu și superior; **Grupul Pades** (acum **Grupul Roscani** – cf. Muresan, 2003) – care, la partea inferioară, înglobează masivele carbonatice de Hunedoara și de Luncani (care pot reprezenta o paleoplatformă carbonatică) coreponde Devonianului superior și Carboniferului inferior. În **Grupul Ghelar**, se află MMR hidrotermal-sedimentare ferifere (Kräutner, 1964 a, 1969, 1974, 1977, 1996; 1999), oxidice – **de tip Lahn-Dill** și carbonatice – **de tip Teliuc și Ghelar**. În **Grupul Roscani**, în metamorfitele Carboniferului inferior, de la partea superioară a acestuia, se află MMR Vetel de sulfuri, hidrotermal-sedimentara (Gurau, 1980) – **tip Kuroko**. În corupurile de porfiroide riolitice, intruse în epimetamorfitele Grupului Roscani, se localizează MMR Muncelul Mic de sulfuri (blenda și galena), hidrotermal (filonian), de sulfuri (Kräutner, 1963, 1964, 1974, 1995, 1996).

II. TIPURI GENETICE DE MMR.

(A) **MMR singenetic** prezente în România, sunt toate, în esenta, de natură hidrotermal-sedimentară, deosebindu-se mai multe tipuri. (1) **MMR hidrotermal-sedimentare de sulfuri tip Kuroko** (ex. MMR Lesul Ursului și Fundul Moldovei, localizate în Formatuna Tg3 a Grupului Tulghes). (2) **MMR hidrotermal-sedimentare de sulfuri de tip scandinav** (acest tip a fost introdus în literatura gitologică românească de către Kräutner, în Borcos et al., 1984). (a) MMR Altîn Tepe (în principal, constituie din trei lentile de minereu masiv de pirita cu calcopirita) este situat în mezometamorfitele de la partea superioară cunoscută a Grupului Altîn Tepe (Proterozoic mediu – Dobrogea). Acum, noi considerăm că acest zacamant este, în principal, de tip scandinav (fiind intercalat într-o stiva terigenă, care are intercalatii de roci magmatogene bazice – amfibolite), însă care, în zonele periferice din partea de NW a lentilelor mineralizate, trece gradat la un minereu magnetic cu hematit de tip Lahn-Dill; Trecerea primă la faciesul oxidic ferifer exprima faptul că sulful a fost în cantități insuficiente pentru a lega fierul sub forma de pirita; este interesant că aceasta situație s-a repetat pentru fiecare letila de minereu masiv. (b) MMR Izvorul Cepii (Radulescu et al., 1963 – Arh. I:G.R.), constituie, în principal din pirita, blenda și galena, este localizat în sisturile sericito-cloritoase ale formatiunii cu acelas nume, care revine Grupului Repedea (Silurian); întrucât la baza acestui grup, se dezvoltă sisturi verzi, atribuim aceasta mineralizare tipului scandinav. (3) **MMR hidrotermal-sedimentare de sulfuri de tip Mississippi Valley**, de exemplu, MMR Valea Blaznei-Guset de sulfuri (în principal, blenda, galena, pirita) (Udubasa, 1970, 1996; Udubasa et al., 1981), asociate cu platformă carbonatică (calcare și dolomite) a Formatunii Rb2, din Grupul Rebra (Proterozoic mediu – Carpatii Orientali). Atribuim acestui tip și MMR pirotoasa cu blenda și galena de la Boita-Hateg, localizată în metamorfitele Grupului Cibin (Proterozoic superior), unde se asociază intim cu calcarele și rocile amfibolice ale



acestuia (Kräutner, 1965). Aceluiasi tip îi revin si dolomitele cu blenda si galena (descrise de Kräutner, 1964), localizate la partea superioara a Devonianului inferior (Grupul Govajdia) din Cristalinul de Poiana Rusca s. s., ca si mineralizatia concordanta de baritina, blenda, galena si fluorina, de la Românesti (considerata de noi anterior a fi o mineralizatie filoniana hidrotermala metamorfozata în faza sudeta – Muresan, 1973), localizata între rocile carbonatice (situate in culcus) si cuartitele (metasilikolite – situate în coperis) Devonianului superior epimetamorfic, din Unitatea epimetamorfica a Masivului Poiana Rusca. (4) **MMR ferifere tip Krivoi Rog** (MMR. Palazu Mare – Proterozoic inferior – Dobrogea sudica). (5) **MMR ferifere oxidice de tip Lahn-Dill** (de exemplu: MMR Rusaia – Silurian, din Zona Cristalno-Mezozoica a Carpatilor Orientali; MMR Grenzenstein – Devonian mediu – Masivul Poiana Rusca). (6) **MMR ferifere carbonatice de tip Teliuc-Ghelar** – Devonian mediu – Masivul Poiana Rusca (un nou tip genetic, introdus în literatura de Kräutner, 1977). (7) **MMR manganifere hidrotermal-sedimentare** (ex. MMR tip Iacobeni – situata în Formatiunea Tg2 a Grupului Tulghes).

(B) **MMR hidrotermale de sulfuri**, mult mai putin raspândite, sunt reprezentate mai ales prin filoane, localizate în porfiroide riolitice, cea mai importanta fiind MMR Muncelul Mic (galena, blenda), situata in pofiroidele riolitice intruse în epimetamorfitele Carboniferului inferior din Poiana Rusca; în Porfiroidele riolitice de Mândra (intruse premetamorfic, în sisturile terigene ale Formatiunii Tg4), sunt localizate MMR filoniene de sulfuri (galena, blenda), de tip Paltin (Muresan, Muresan, 1977).

III. TRASATURI PRINCIPALE SINDEPOZITIONALE ALE MMR.

(A) Alcatuirea mineralogica primara.

Pe baza datelor din literatura, se poate aprecia ca, în ansamblu, metamorfismele regionale, de intensitati si vîrste diferite, care au afectat atât stivele primare cât si mineralizatiile metalifere asociate, au decurs practic în conditii izochimice, exceptând H₂O si partial CO₂. In aceasta situatie, plecând de la compozitia actuala a MMR, putem sa facem unele presupuneri privind alcatuirea primara (premetamorfica) de ansamblu a acestora. (1) **MMR singenetice, de tip Kuroko, de tip Scandinav, de tip Mississippi Valey si MMR hidrotermale** au o alcatuire care nu s-a modificat substantial prin metamorfozare regionala, deoarece sulful si majoritatea metalelor (Cu, Pb, Zn etc) din sulfuri, nu pot intra în reteaua silicatilor metamorfici. Astfel, **MMR de sulfuri, tipul Kuroko si tipul scandinav** au avut si înaintea metamorfismului o alcatuire apropiata de cea actuala, formata predominant din pirita (10-85 % din minereu), al doilea mineral metalifer fiind calcopirita (3-10 %, de obicei sub 2 %); blenda si galena sunt adesea sporadice si apar in cantitati mici (de obicei, 0-1 %). Cuartul este omniprezent si apare în cantitati mari (15-70 %). În schimb, **MMR de sulfuri, tipul Mississippi Valey si MMR hidrotermale** se remarcă prin cantitati mereu mai mari de blenda si galena, care adesea depasesc pe cele ale piritei; calcopirita este adesea absenta. **Mobilizarile hidrotermal-metamorfice ale sulfurilor** (mai ales, calcopirita si, subordonat galena sau blenda) si ale cuartului sunt de cele mai multe ori locale, materializate prin filoane si filonase, cu grosimi milimetrice-centimetrice (mai rar, decimetrice) si cu lungimi decimetrice-metrice, localizate mai ales în interiorul corpurilor de MMR sau în imediata apropiere a acestora. **Aspectele fin sau marunt granulare ale sulfurilor ale acestor MMR, ca si prezenta cuartitelor mozaicate**, asociate acestora, arata ca materialul primar era în principal geliform, uneori impurificat de aporturi argiloase (actualele stratulete si lamine de filosilicati) si de cele detritogene (cuart, feldspat, titanit etc). (2) **MMR de tip Lahn-Dill**, formate mai ales din oxizi de fier (hematit, magnetit) si cuart mozaicat, au fost anterior formate in principal tot din oxizi de fier, de tipul hematit (?), goethit, hidrogoethit, sugerând provenienta acestor MMR din geluri silicioase feruginoase, de origine hidrotermal-sedimentara. Prezenta relativ frecventa a spessartinului în aceste MMR, arata ca manganul a fost prezent, probabil sub forma de oxizi si / sau hidroxizi, care, în decursul epimetamorfismului regional, în prezenta cuartului si a fractiunii argiloase, au trecut în granat



manganifer (de exemplu, cuartitele hematitice cu magnetit si spessartin din Dealul Gherman, la nord de Raul Bicaz (Muresan, 2002) – din Formatiunea Tg4 a Grupului Tulghes. **(3) MMR de tip Krivoi Rog**, formate atât din oxizi de fier (magnetit, subordonat, hematit) cât și din silicati bogati în fier (de exemplu, cummingtonit) au avut o alcatuire de oxizi de fier (de tipul hematit (?), goethit, hidrogoethit s.a.), eventual, asociati cu oolite si clorite ferifere sedimentare. **(4) MMR de tip Teliuc-Ghelar**, constituite din carbonati feriferi (siderite, ankerite etc), au avut o constitutie premetamorfica similara. **(5) MMR singenetice manganifere (tip Iacobeni)** au actualmente o alcatuire mineralogica extrem de complexa, caracterizata prin asocierea rodocrozitului cu zeci sau chiar sute de alte minerale de mangan (Hîrtopanu, 2004), ansamblu format prin recristalizarea metamorfica a unor geluri manganifere carbonatice-silicioase-argiloase.

(B) Conservarea în ansamblu a formelor primare depozitionale.

Desi au fost afectate de deformarile sinmetamorfice, MMR reflecta, în esenta, forme initiale de zacamânt, putând fi stratiforme sau filoniene. **(1) Formele de zacamânt ale MMR hidrotermal-sedimentare** sunt variate, dar toate exprima, mai ales, prin caracterul lor concordant fata de rocele-gazda, singeneza acestor mineralizatii. Datorita genezei lor, aceste MMR nu constituie niciodata strate cu extindere directionala regionala, formarea acestora având loc în zonele de venire a hidrotermelor, care erau dispuse de-a lungul unor aliniamente directionale (corespunzând unor fracturi directionale din fundamental bazinelor marine), de-a lungul carora erau distribuite discontinu. De asemenea, dimensiunile directionale limitate ale corpurilor de minereu au fost cauzate si de faptul ca sarcina minerala a hidrotermelor era relativ instabila, în contact cu electrolitii din apa marii. **(a) Formele de zacamânt ale MMR de tip Kuroko, de tip scandinav si de tip Mississippi Valey** cele mai frecvente sunt: **(a1) stratele lenticulare**, de obicei, cu lungimi de ordinul sutelor de metri (rar, peste 1000 m – cazul celor trei coruri de minereu masiv de la Altîn Tepe), cu latimi de zeci de metri (rareori, peste 100-150 m) si cu grosimi maxime metrice (rareori, peste 10 m); **(a2) lentilele stratiforme**, cu lungimi sub 100 m, latimi de 20-30 m si grosimi submetrice (rar, peste 1-2 m); **(a3) lentile, cuiburi si diseminari stratiforme**. Cel mai adesea, se remarcă în același zacamânt coexistenta mai multor forme de acumulare, rareori existând forme independente, cum este cazul celor trei coruri de minereu compact de la Altîn Tepe. **(2) Formele de zacamânt ale MMR de tip Lahn-Dill si cele manganifere hidrotermal-sedimentare** au, de obicei, dimensiuni simitor mai mici, predominând lentilele stratiforme si, mai ales, lentilele mici. **(3) Diseminarile stratiforme de magnetit ± hematit**, sunt de asemenea de natura hidrotermal-sedimentara, cum sunt cele din sisturile verzi ale Devonianului mediu de la Iazuri (din Unitatea Epimetamorfica a Masivului Poiana Rusca) si cele de la Sunatori (Valea Bistritei – în Muntii Bistritei) din paraamfibolitele Formatiunii Rb3 a Grupului Rebra (Proterozoic mediu), din Carpatii Orientali. **(4) În formatiunea ferifera de tip Krivoi Rog**, reprezentata prin **Grupul Palazu**, din Dobrogea Meridionala, mineralele ferifere de aici (magnetit, subordonat hematit si ankerit) nu au format cuartite feruginoase, de tipul itabiritelor careliene, constituind astfel numai diseminar, localizate în aproape toate rocile grupului. **(5) Forma de zacamânt a MMR ferifere, de tip Teliuc si Ghelar**. Aceste MMR, din Devonianul mediu epimetamorfic al Cristalinului de Poiana Rusca s. s. au forma unor lentile puternic bombate (dupa datele lui Kräutner, 1964) – în sectiune transversala, grosimea maxima, de 100-200 m, este de același ordin cu latimea, de 400-500 m; aceste lentile se efileaza rapid, până la disparitie.

C. Succesiuni de formare în cadrul corpurilor mineralizate.

În cazul MMR singenetice (stratiforme), de fapt, hidrotermal-sedimentare, există o succesiune de formare în cadrul fiecarui corp mineralizat, astfel ca partile inferioare ale corpului stratiform au fost sedimentate înaintea celor situate stratigrafic mai sus. Aceasta situatie reflectă în ansamblu depunerea succesiva a sarcinilor minerale din solutiile hidrotermale ajunse pe fundul marii. Aceasta succesiune de depunere a fost influentata de variatia PH-ului apei marine, din



vecinatatea zonelor de debordare a hidrotermelor, dar și de faptul că sulfurile precipita mai rapid decât silicea. Deci, în procesul de depunere, se produce un proces de diferențiere fizico-chimic. O analiză detaliată a acelor aspecte depozitionale a fost facuta de Kräutner (1965), în cazul MMR Boita-Hateg, de sulfuri, de natură hidrotermal-sedimentară. Asemenea reconstituiri, ca și modele depozitionale, au mai fost facute, în cazul MMR ferifere din Cristalinul de Poiana Rusca s. s. (Kräutner, 1996), în cel al MMR de tip Kuroko, localizate în GTg din Carpatii Orientali (Kräutner, 1984, 1989) și în cel al MMR Altîn Tepe (Muresan, 1972).

D. Metagelurile din MMR hidrotermal-sedimentare.

Structurile geliforme primare sunt exprimate actualmente prin structura mozaicată a cuartitelor, adesea asociate intim cu mineralele metalifere. Se poate afirma că starea geliformă a fost dominată în acumularile initiale, rata depunerilor detritice și a celor argiloase fiind mult mai mică și deci depasita net cantitativ de aportul mineral al hidrotermelor în apa marii.

E. Texturi relicte, stratificatii.

În cazul MMR singenetic, s-au pastrat texturi relicte (rubanari, stratificatii), treceri direcționale (variații faciale) de la minereul compact, la minereul diseminat și apoi la sisturi.

F. Conservarea unor zone de paleoalteratie hidrotermala ale rocelor înconjuratoare (în special a porfiroidelor riolitice – sericitizari, foste argilizari, albitizari, silicifieri, calcitizari), în cazul unor MMR hidrotermale (exemplu: MMR filoniene de la Muncelul Mic – Poiana Rusca (Kräutner, 1963); MMR filoniene tip Paltin – localizate în Pofiroidele de Mândra – Carpatii Orientali (Muresan, Muresan, 1977 b)).

G. Contextul geologic actual.

(1) **MMR singenetic** au fost localizate în stive initial sedimentare, în care sunt prezente elemente magamatogene bazice și / sau acide, metamorfozate regional împreună cu entitatea în care află și reprezentate prin metatufuri, metaepiclastite bazice (sisturi verzi, respectiv paraamfibolite) și acide, corpuri de metagabbrouri și / sau de metariolite (actualmente pofiroide). Astfel, uneori, MMR sunt localizate direct în vulcanoclastite acide sau în imediata apropiere a acestora; de exemplu, în Zona Cristalino-Mezozoica a Carpaților Orientali, MMR de tip Kuroko, din Grupul Tulghes (Ordovician inferior), sunt strâns asociate cu vulcanoclastite riolitice sau sunt localizate direct în acestea. (Kräutner, 1965, 1977, 1984; Muresan, Muresan, 1977; Muresan, 2002a, 2002 b; Zincenco, 1998), reflectând contextul geologic primar. Mentionăm, de asemenea, că, în Devonianul mediu și superior din Poiana Rusca (Carpații Meridionali), se remarcă asocierea strânsă cu sisturile verzi a MMR ferifere (atât în faciesul de Lahn-Dill, cât și în faciesul carbonatic, de tip Teliuc-Ghelar), situație evidențiată de câteva decenii (Dimitrescu, 1961; Kräutner, 1964, 1965, 1969; Maier et al., 1964, 1969; Kräutner, et al., 1973; Muresan, 1968, 1973).

(2) **MMR hidrotermale** sunt localizate practic exclusiv în dyke-uri de pofiroide riolitice, cum sunt, de pilda, filoanele plumbo-zincifere de la Muncelul Mic (Kräutner, 1963), intruse în epimetamorfitele terigene ale Carboniferului inferior din Poiana Rusca, precum și MMR plumbo-zincifere din regiunea Tulghes-Paltin (Muresan, Muresan, 1977), localizate în corpul Porfiroidelor de Mândra (metariolite), intrus în epimetamorfitele terigene ale Formațiunii Tg4 din Grupul Tulghes (Ordovician inferior), din Carpații Orientali.

H. Controlul litostratigrafic

Este evident, în cazul MMR singenetic, de exemplu: (1) **MMR de tip Kuroko** (Kräutner, 1965, Kräutner, Popa, 1973; Kräutner, Bindea, 1995; Bercia et al., 1976; Muresan, 2002 b), din Grupul Tulghes – Ordovician inferior (Carpații Orientali), sunt situate la mai multe nivele litostratigrafice apropriate în cadrul Formațiunii Tg3. (2) **MMR manganifere hidrotermal-**



sedimentare, cele mai importante din Grupul Tulghes, sunt gazduite în Formatiunea Tg2 a acestuia. (3) **MMR ferifere** (atât în faciesul de Lahn-Dill, cât și în faciesul carbonatic, de tip Teliuc-Ghelar, din Grupul Ghelar, sunt localizate la anumite nivele în Grupul Ghelar (Devonian mediu și superior – Poiana Rusca – Kräutner, 1964, 1965, 1969; Maier et al., 1964, 1969; Kräutner, et al., 1973, Muresan, 1968, 1973). I. Controlul petrografic magmatic este vizibil în cazul MMR hidrotermale; de exemplu, filoanele plumbo-zincifere de tip Muncelul Mic (Kräutner, 1963), sunt practic localizate numai în porfiroidele riolitice, intruse în Grupul Roscani – Carbonifer inferior (din Poiana Rusca).

IV. TRASATURI METAMORFICE ALE MMR.

A. Concordanta dintre gradul de metamorfism al MMR și cel al entitatilor metamorfice-gazda. De exemplu, în minereul de la Altîn Tepe, apare o recristalizare metamorfica prograda, caracterizata prin aparitia almandinului, biotitului și muscovitului în minereu (Muresan, 1969).

B. Concordanta dintre caracterele structurale sinmetamorfice ale MMR și cele ale entitatilor metamorfice-gazda. (plunjuri axiale, structuri plicative, liniatii, sistuoziatii etc). Un exemplu clar în aceasta privinta este cel al MMR Altîn Tepe (Muresan, 1969), alcătuit din trei lentile de minereu compact, puternic alungite, a căror axa lungă se afunda cu circa 30-35° spre SE, situatie concordanta cu plunjul axial al structurii B, constatata în mezometamorfitele grupului. Microcutele, de tip drag-folds, liniatii, sistuoziatii din minereu, au aceiasi orientare spatiala ca a elementelor corespunzatoare prezente în metamorfitele acestei entitati. În portiunile mai bogate în silicati ale diferitelor MMR, se constata frecvent transpunerea situoziatii de stratificatie dupa sisteme de clivaje, cel mai adesea clivajele axiale, rezultând sistuoziitatea clivajului axial, care poate oblitera parțial (uneori, total) sistuoziitatea de stratificatie (de exemplu, MMR Balan, tip Kuroko – din Grupul Tulghes – și MMR ferifere, tip Lahn-Dill, din Devonianul Unitatii Epimetamorfice din Masivul Poiana Rusca).

C. Concordanta dintre transformarile mineralogice retrograde ale MMR și cele ale entitatii metamorfice-gazda. De exemplu, fenomene de transformari prograde în MMR de sulfuri Altîn Tepe (Berbeleac et al. 1984, 1985), au fost constatate și în mezometamorfitele proterozoice ale Grupului Altîn Tepe (Nedelcu & Gridan, în Nedelcu et al., 1987, 1988).

D. Actualele relații mutuale dintre mineralele metalifere componente ale MMR (atât singenetică cât și hidrotermală) nu reflectă ordinea initială de formare a acestora. Astfel, în timpul blastezei sincrone cu metamorfismul regional, se evidențiază mineralele «**dure**» făta de cele «**moi**», deosebite de Ramberg (1963), între acestea existând și minerale cu comportament intermediu. Din descrierile MMR de sulfuri din România, rezulta că primele sunt reprezentate, mai ales, prin pirita și magnetit, care prin competența lor și prin puterea de idioblastica, au frecvent fete cristalografice (forme idiomorfe sau hipidiomorfe), în raport cu mineralele moi. Acestea din urmă, mai numeroase (calcopirita, galena, blenda, tetraedritul, burnonitul, galeno-bismutina, smaltina, molibdenitul s. a) nu prezintă fete cristalografice, fiind mereu allotriomorfe. Se observă că sulfurile moi au fost frecvent mobilizate, atât pe cale mecanică (adică presate pe fisurile mineralelor metalifere dure și pe ale cuartului), cât și pe cale hidrotermal-metamorfică, fiind adesea asociate filonaselor de cuart, cantonate în minereu, sau în imediata apropierea a acestuia. Dintre sulfurile cu comportament intermediu, cităm mispichelul și pirotina. Astfel, indiferent că este vorba de MMR singenetică sau MMR hidrotermală, mineralele dure prezintă aparentă unei generații mai vechi, urmată (tot aparent) de cele moi, situație, care, în trecut, a condus la interpretări genetice eronate. Mentionăm că contextul mineralologic poate influența modul de prezentare al mineralelor dure: (a) astfel, pirita, când apare în zonele de minereu diseminat, are frecvent forme idiomorfe sau hipidiomorfe; (b) în zonele cu minereu compact pirozit, pirita are de obicei două moduri de



prezentare: granule mici (în medie, sub 0,2 mm) și porfiroblaste (1-5 mm); granulele mici de pirita constituie o masa mozaicată, în care apar porfiroblaste sau grupe de porfiroblaste de pirita, situație care da un aspect «porfiric» minereului.

E. Indicatii asupra metamorfozarii regionale a minereurilor luate în considerare sunt date și de structura intimă a acestora: (1) **Umbrele de presiune (pressure shadow)**, reprezentate de indivizi alungiti de cuart sau de lamele de clorite, dispuse quasi-perpendicular pe fetele cristalografice opuse ale piritei sau ale magnetitului, respectiv orientate paralel cu alungirea paralela cu sistozitatea minereului (conform principiului lui Riecke). (2) **Includeri poikiloblastice**, cum sunt cele ale piritei și magnetitului în granati (în cazul MMR mezometamorfozate). (3) **Extinctia ondulatorie a quartului** (mai rar a baritinei, feldspatilor, calcitului). (4) **Microcutarea submilimetrica frecventa**, mai ales a minereului cu filosilicati. (5) **Cimentarea mineralelor metalifere** (în special, a piritei, magnetitului și, mai rar, a altor sulfuri) cu minerale întâlnite frecvent în roci metamorfice (clorit, sericit, muscovit, granat, zoizit, epidot, sfen, albite s. a.)

V. CONTEXTUL GEOTECTONIC PRIMAR AL UNOR MMR.

Contextul geotectonic primar al stivelor primare în care se află MMR, îl putem presupune pe baza celui actual, prin analiza stivelor acum metamorfozate și pe baza datelor geochemice.

A. Stivele transgresive pe socluri metamorfozate anterior (deci, repauzând pe o crusta continentală) și continând intercalatii de roci magmatogene bazice și / sau acide, pot fi apreciate ca fiind formate în bazin sedimentare marine, de tip back-arc, învecinate cu arcuri insulare, de unde au provenit rocele magmatogene (bazice și / sau acide), intercalate în aceste stive sedimentare, ulterior metamorfozate regionale. (1) **MMR Palazu Mare ferifera**, de tip Krivoi Rog, este localizată în Grupul Palazu (Proterozoic Inferior), dispus peste soclul metamorfic vechi pre-carelian (arhaic?), cunoscut în Scutul Ucrainian. Pentru noi, natura petrografică a GPM, grosimea stivei sale (care initial a fost mai mare, o parte fiind erodată înaintea depunerii Grupului Cocosu) și metamorfismul ei regional intens, sunt argumente pentru a afirma că GPM s-a format initial într-un mediu marin, probabil într-un bazin de tip «back-arc» activ, cu o adâncime moderată, însă afectat de miscări de subsidență.. Prezenta calcarelor în stiva GPM, arată că depozitele primare s-au format deasupra limitei de existență a CO₃Ca. (2) **MMR de sulfuri Valea Cepei**, din Munții Rodnei, este localizată în Grupul Repedea – GRe (Silurian – GRe a fost separat de Kräutner (1968, 1987), Kräutner & Kräutner (1970) și a fost datat palinologic de Iliescu et al., (1976), care sta transgresiv peste soclul mezometamorfic (Proterozoic mediu), constituit din Grupul Bretila. Aceasta situație, corelată cu caracterul predominant terigen al stivei GRe, în care se intercalează MMR menționată, ne determină să admitem că formarea acestei entități a avut loc într-un bazin marin instalat pe o margine continentală activă, în apropierea unui arc vulcanic insular (de unde a provenit materialul epiclastic bazic al actualelor sisturi verzi, din baza stivei GRe). (3) **MMR ferifere tip Lahn-Dill Rusia** sunt intercalate în baza detritogenă a stivei terigene epimetamorfice a Grupului Rusia – GRu (Silurian – studiat de Fl. Kräutner, 1970; datat palinologic de Iliescu et al., 1976), transgresiv pe soclul metamorfozat, format din mezometamorfitele Grupului Bretila. GRu a fost echivalat (Kräutner, 1987) cu Grupul Repedea. În aceasta situație, se impune concluzia că GRu s-a format în aceleasi conditii ca și GRe, dar mai departe de un arc vulcanic insular (nu are intercalatii de roci magamatogene bazice). (4) **MMR ferifere (în facies carbonatic și respectiv oxidic)**, din Cristalinul de Poiana Rusca s. s., se localizează în Devonianul epimetamorfic, dispus transgresiv peste un soclu metamorfic caledonian (Muresan, 2003). (a) Tinând seama de aceasta situație; (b) de grosimea foarte mare a stivei Cristalinului de Poiana Rusca s. s. (> 10.000 m); (c) de caracterul predominant terigen-carbonatic al succesiunii; (d) de intercalarea în Devonianul mediu și superior a rocilor magamatogene bazice (sisturi verzi și, subordonat, metagabbrouri); (e) de caracterul bimodal al magmatismului, ale carui produse se află în actuala stiva a Cristalinului de Poiana Rusca s. s.



(metagabbrouri si metariolite), am admis (Muresan, 1998, 2004 – sub tipar) formarea succesiunii primare într-un bazin marin de tip back-arc.

B. Stive cu subasment necunoscut, dar despre care există informații geochimice (elemente majore și minore). În aceasta categorie, se inscrie Grupul Altîn Tepe – GAT (Proterozoic mediu) – din Dobrogea, ale carui mezometamorfite aflorează imediat la sud de Falia Peceneaga-Camena, de sub sariajul (Muresan, 1971) Grupului Sisturilor Verzi. GAT cuprinde o stiva terigenă flisoidă, cu intercalatii de amfibolite (Muresan, 1969, 1971, 2004). În partea superioară a acesteia, este localizat MMR Altîn Tepe de sulfuri și magnetit, hidrotermal-sedimentar. Studiul geochimic (bazat pe elemente majore și minore) al ortoamfibolitelor (metagabbrouri), corelat cu caracteristicile stivei, au aratat formarea acesteia într-un bazin marin de tip back-arc, imediat învecinat cu un arc vulcanic insular (de unde se insinuau, în succesiune, magmele bazice – actualele amfibolite).

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VARISCITE (AlPO₄·2H₂O) FROM CIOCLOVINA CAVE (SUREANU MOUNTAINS, ROMANIA): A TALE OF A MISSING PHOSPHATE

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Recent investigations on a phosphatized sediment sequence in the Cioclovina Cave led to the identification of a second occurrence in Romania (first time in the cave environment) of variscite, AlPO₄·2H₂O.

The mineral exists as dull-white, tiny crusts and veinlets within the thick argillaceous material accumulated on the cave floor.

Under scanning electron microscope (SEM) variscite appears as subhedral to euhedral micron-size crystals. The {111} pseudo-octahedral form is rather common. Variscite was further characterized by means of X-ray diffraction, thermal, vibrational FT-IR and FT-Raman spectroscopy, and by SEM energy-dispersive spectrometry (EDS). The calculated orthorhombic cell parameters are $a = 9.823(4)$, $b = 8.562(9)$, $c = 9.620(5)$ Å, and $V = 809.167(6)$ Å³. The ED spectrum of variscite shows well-resolved Al and P lines confirming thus the presence of the major elements in our compound.

The formation of variscite is attributed to the reaction between the phosphate-rich leachates derived from guano and the underlying clay sediments.



PRELIMINARY GEOCHEMICAL DATA ON THE PAULIS AND GALSA VARISCAN GRANITOIDS (HIGHIS MTS.)

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A major proportion of the Pre-Neogene basement of the Apuseni Mts. (Romania) and the Pannonian Basin (Hungary) is built up by the Tisia Composite Terrane Alpine Megatectonic Unit. The crystalline mass of the Tisia Composite Terrane is characterised by granitoid ranges and anticline wings of middle and high grade metamorphites (Pál Molnár et al., 2001 A,B, 2002). The largest basement exposure within the Tisia Composite Terrane is represented by the Apuseni Mts. The Apuseni Mts. are partially built up by two Alpine overthrust units (Codru and Biharia Nappe Systems), carrying Variscan granitoid intrusions (Pana, 1998). These granitoids were mainly characterized by petrographical and geochronological studies (Pana, 1998) while their relations to the Pannonian Basin granites are less studied.

The study presents results of major element and trace element geochemistry, performed on granitoids of the Codru and Biharia Nappe Systems, exposed in the Highis Mts. The final aim of the research is to reveal correlations between the granitoids of the Apuseni Mts. and other Variscan granitoids of the South Hungarian Basement.

Samples were taken from the vicinity of settlements Galsa (Siria granitoids - Codru Nappe System) and Paulis (Highis granitoids - Biharia Nappe System). The studied granitoids are relatively alkali-rich, fractionated from moderate to high extent, with calc-alkaline and slightly peraluminous features. Significant differences can be observed in the tectogenetic moment of the granitoids: the Galsa granitoids are continental collision granites (CCG) while the Paulis granitoids are post-orogenic granites (POG). Based on trace element geochemistry both granitoid groups show syn-collisional (syn-COLG) tectonic setting with an S-type character, Paulis granitoids are plotted into fractionated granitoid type field (FG). Both granitoid groups have Nb negative anomaly and are enriched in Rb, Ba and K as compared with primitive mantle and chondrites. The Paulis granitoids are depleted in Sr as compared with Galsa granitoids which show also a low contents in Y and Yb too. These results show that Paulis granitoids are formed in a later stage of fractionation process.

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**PRELIMINARY MICROTHERMOMETRIC DATA RELATED TO A COMPLEX MAGMATIC – HYDROTHERMAL SYSTEM FROM SANTIMBRU-BAI (SOUTH - HARGHITA MOUNTAINS, ROMANIA)**IOAN PINTEA¹, ATTILA A. LACZKÓ²¹*Geological Institute of Romania, P.O. Box 181, 3400, Cluj-Napoca, 1, Romania ipintea@personal.ro*²*SC Geolex SA, 530154 Miercurea Ciuc, str. Harghita 70/B, jud. HR, Romania laczkoati@yahoo.com*

Five types of fluid inclusions were documented in the coarse quartz veins assemblages (*Fig. 1a*) from the hornblende and biotite-bearing quartz andesite shallow body occurred at Santimbru-Bai in the Luci-Lazu area from South Harghita Mountains. Based upon phase's relations at room temperature they were classified as follows:

1. Monophase and biphases glass inclusions;
2. Vapor rich ($H_2O + CO_2$?);
3. Multiphases ($L + V + solids$);
4. Aqueous biphases ($L+V$) which also showed time to time a third phase separated as a rim around the vapor bubble, probably formed by CO_2 - rich liquid;
5. Gaseous (V) or liquid (L) monophase.

Types 1 and 2 were founded mainly as primary and secondary assemblages in primary quartz coarsened grains and recrystallized prismatic quartz crystals, respectively. Types 3, 4, and 5 occurred mainly as recrystallised planes in both quartz subtypes. It is emphasized, based on petrographic evidences, that glass inclusions are representative for a silicate melt segregated as separate fluid in the magmatic stage, the polyphases inclusions are samples of a second fluid phase evolved from the silicate melt rich in volatiles and the biphases and monophase inclusions are the latest fluids related to the Santimbru-Bai magmatic - hydrothermal system.

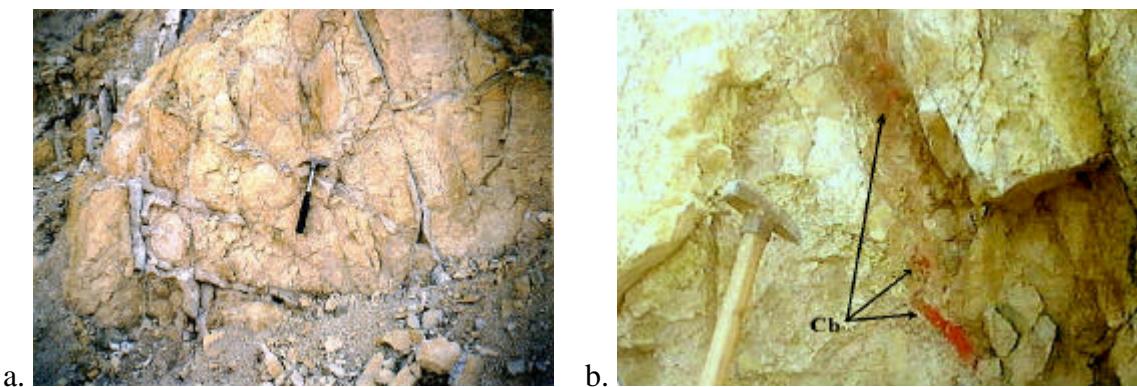


Fig. 1. a. Veins of coarse grain quartz and dravite in a magmatic network system in quartz andesite from Santimbru -Bai; **b.** Reddish cinnabar precipitated on the same altered andesites (*from LACZKÓ, 2004 - unpublished data*).

Preliminary microthermometric experiments showed high homogenization temperature for the glass inclusions near of 900 - 1000°C. The complex melt inclusions founded mainly in the coarse quartz



grains indicated heterogeneous trapping by vapor disappearance above 640°C when the cavities contained only liquid and solid opaque or undisolved transparent solid phases. In the multiphase inclusions with soluble solid grains (halite ?) the melting temperature in the presence of a complex mixture formed by other solids and vapor, ranged between 300 - 320°C (ws = 38 - 40 wt % NaCl). On further heating a coexistence of silicate melt + vapor and opaque was noted around 1000 - 1010°C.

The secondary multiphases inclusions from the prismatic quartz crystals showed melting temperatures of the solid phase (halite ?) around 120°C (ws = 28 wt % NaCl equivalent) and homogenization temperature ranged between 145 and 200°C. The primary biphasic aqueous inclusions were homogenized in the liquid phase by vapor bubble disappearance at 310°C. No microthermometric data about CO₂ are reported in this study the work is being in progress. The mineralogy, petrography of the host rocks and the microthermometric characteristics of the fluid phases presented in this study allowed us to compile a working metallogenetic model associated to the Sântimbru-Bai magmatic - hydrothermal system. The quartz veins network centred on the hornblende and biotite quartz andesite shallow structure were formed in the magmatic stage during magma emplacement near surface conditions. A fluid volatile rich phase was separated together with a B - bearing silicate melt (e.g. the quartz - dravite paragenesis presented by LACZKÓ, 2003) during the transition from magmatic to hydrothermal stages. The last important event took place at low temperatures by evolution of a rich aqueous and vapor phases which overprinted the overall magmatic structure, and complex hydrothermal transformations as sericitisation, silicification, tourmalinization, argilization were noted (TANASESCU, 1978; STANCIU, 1996; LACZKÓ et al., 2003). At this stage a special cinnabar parageneses (*Fig.1b*) were precipitated together with pyrite, marcasite, melniovite, magnetite, hematite, stibnite and orpiment (LACZKÓ et al., 2004).

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ALMAS-STANIJA: EPITHERMAL MINERALIZATIONS INTERRELATED WITH AN UNDERLYING PORPHYRY COPPER SYSTEM (SOUTHERN APUSENI MOUNTAINS, ROMANIA)

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Abstract

The Almas-Stanija(A-S) ore field from "Golden Quadrilateral" is located in the Southern Apuseni Mountains of western Romania, in the Zlatna-Stanija neogne volcanic structure. A-S structure comprises two interconnected magmatic-hydrothermal systems: the low sulphidation type of polymetallic gold + silver veins like a transitional epithermal system, which underlying a porphyry system situated in the depth of the geological structure. The studies show the evolution of an early magmatic pattern of alterations (propylitic and potassic-biotitic alteration) to a composite pattern alteration (adularia, intermediate argillic, sericitic and silica alteration), resulting by interaction with ground water.

In A-S area, the vein systems near surface are contained within a large alteration envelope that includes the adjacent Almas-Neagra, Muncaceasca East, Muncaceasca West and Popa-Stanija deposits. The host rocks have undergone propylitic- potassic -intermediary argillitic -sericitic -silicification alteration, increasing in intensity to mineralise veins. Mineralization is hosted in veins, fracture zones, stockworks and breccias, predominantly within igneous rocks, sometimes within the Cretaceous sediments and rarely in ophiolites. The veins have strike lengths of around 3-500 metres, and have been traced for approximately 500 metres down dip. Mineralization shows typical zonation patterns, from gold and silver-rich near surface, to more base metal-rich, then copper-rich ores with depth.

The porphyry copper-gold system is situated approximately 400 meters below the present surface in Muncaceasca West area. The porphyry mineralization is hosted in the microdiorite intrusion that contains sulphides (pyrrhotite+chalcocite+pyrite+molybdenite+ magnetite) stockworking.

Keywords: *porphyry copper, low sulphidation, Neogene andesites, South Apuseni Mountains*

Introduction

The Almas-Stanija ore field from "Golden Quadrilateral" is located in the Southern Apuseni Mountains of western Romania, in the Zlatna-Stanija neogne volcanic structure, exposed as a linear shape with 30 x 4 km dimensions, adjacent to the Zlatna-Stanija sedimentary basin. Here there are the precious metal (Au, Ag), base metal (Pb, Zn) and porphyry copper mineralization, appearing in steeply quartz-carbonate veins, large metasomatic pyrite lenses, hydrothermal breccias and stockworks, which are hosted by Miocene andesitic stocks and lava flows, ophiolitic basement and Cretaceous sedimentary rocks. The mineralization consists predominantly of sulfides (pyrite, chalcocite, sphalerite, and galena), sulfosalts of As, Sb, and gold. The A-S ore field comprises the following ore deposits: Almas, Muncacesca Est, Muncacesca West, Stanija, which were explored and have been mined to some degree over a long period of time.



Geological Setting

The South Apuseni Mountains district represents an internal part of the Carpatho-Pannonian Cenozoic calc-alkaline belt, which is one of the Europe's most important porphyry Cu-Au provinces. The district covers an area of about 900 km² and includes numerous porphyry-types (Bostinescu, 1984). Mineralization is associated with calc-alkaline Neogene igneous rocks that were intruded through pre-Mesozoic low-grade metamorphics, Mesozoic rocks (Ianovici, et al., 1976). The Neogene magmatic activity and related hydrothermal mineralization are controlled by strike-slip fault system forming pull-apart basins (Rosu et al., 2000; Drew&Berger 2001)

The geology of the A-S ore field consists of a middle Jurassic ophiolites basement and a variety of overlying Cretaceous sediments, including flysch deposits conglomerates, sandstones, carbonaceous shales, siltstones and calcareous sediments. The sediments were deposited in marine and non-marine environments; some of them have been juxtaposed by intense structural activity, including thrusting, during the Cretaceous.

Neogene andesitic sequences have resulted from the calc-alkali volcanism during the middle to late Miocene. The Neogene andesitic intrusions (stocks) are located in Fericeaua Hill, Neagra Hill, Ungurului Hill, Magura Hill and are associated by lava flows in Negrii Hill and Ludului Hill and by associated breccia bodies (Bradisor, Villanel, Magura, Fericeaua).

These andesites contain green hornblende with varying amounts of biotite, quartz and resorbed pyroxene.

Volcanic centres were aligned along the series of northwest and west-northwest trending fault zones and constituted the point of convergence of subsequent hydrothermal activity and associated mineralization. The volcanic centers trending are: Neagra Hill-Ungurului Hill; Fericeaua-Acra Hill.

In the depth, Fericeaua Hill developed a potassic altered microdiorite stock and associated with a porphyry system, which will evolve. At the shallow levels on an extended zone there are present hydrothermal systems, as a result of the cooling intrusions and active groundwater convection cells.

Alteration and Mineralization

Geological field data joining with drill data and mining data reveal the A-S structure comprises two interconnected magmatic-hydrothermal systems: the low sulphidation type of polymetallic gold + silver veins like a transitional epithermal system, which underlies a porphyry system situated in the depth of the geological structure. Drilling and underground data point out that epithermal mineralization, as veins, mineralised-breccia bodies cross-cut the upper part of porphyry system. In this manner, the low sulphidation type deposits are younger than the porphyry system.

The circulation of the reduced near neutral fluids generates the potassic alteration. In the low sulphidation veins system environment, the K-feldspar is produced mainly by loss of pressure and carbon dioxide.

At the shallow the mineralization style is gold and silver rich ores, more silver and base metal rich mineralization styles. In depth typical porphyry copper systems with associated potassic and phyllitic alteration styles are developed.

Mineralization is hosted in veins, fracture zones, stockworks and breccias, predominantly within igneous rocks, sometimes within the Cretaceous sediments and rarely in ophiolites.

In A-S area, the vein systems near surface are contained within a large alteration envelope that includes the adjacent Almas-Neagra, Muncaceasca East, Muncaceasca West and Popa-Stanija



deposits. The field mapping and microscopic analysis, it was identified a zoning of alteration selvages (Popa et al, 1990, 1995). The host rocks have undergone propylitic- potassic –intermediary argillitic -sericitic -silicification alteration, increasing in intensity to mineralise veins.

The Muncaceasca West deposit is located in the west-central part of the area. The Muncaceasca West deposit consists of a series of steeply dipping, northwest trending veins that occur above a porphyry copper-gold system approximately 400 metres below the present surface.

The porphyry mineralization is hosted in a porphyritic andesitic intrusion (microdiorite stock) that contains abundant quartz vein and sulphide stockworking.

Geological data obtained by logging DDH301 (1300m depth), localised in Fericeaua Hill, add new information about porphyry system, porphyry mineralization and typical alteration and epithermal mineralization (Popa et al, 1995, Popa & Popa, 2004).

The lithological column consists of Cretaceous and ophiolites rocks in the upper part. The Neogene intrusion (microdioritic stock) was intercepted to various depth levels and in the depths.

Lithological column of DDH 301 reveal a vertical alteration zoning: in the depth, there is a porphyry copper system (who is known also by mining works (level +545 m), related with the microdiorite stock, and upward an epithermal system is developed in sedimentary rocks and Neogene andesites (25-Corabia vein and Rosia system vein). In the depths, in the interval m468-1141 the drill has intercepted discontinuous intervals with potassic (biotitisation), phyllitic zones and sericitic zones and surrounded externally by propylitic zone.

The deeper porphyry system exhibits classic alteration types normally associated with this style of mineralisation, including:

-potassic **zone** of porphyry copper system has the assemblages: biotite ± K-feldspar+ +quartz ± rutile ± pennine + sulphides (pyrrhotite+chalcopyrite+pyrite+molybdenite+ magnetite).

- potassic zone is surrounded by **phyllitic zone**, characterised by the assemblages: sericite +carbonate ± chlorite ± biotite ± quartz ± rutile+sulphides± magnetite

-a distal propylitic assemblage (chlorite + pyrite + epidote + carbonate) is a diffuse and widespread fringe around the margins of the deposits.

The porphyry mineralization is hosted in the microdiorite intrusion that contains sulphide stockworking.

The typical porphyry assemblages were establishing:

1. for the interval m 400-700: [quartz + biotite + alkali feldspar + tourmaline+ (hematite?) + pyrite + pyrothine + chalcopyrite +Au, Ag + sphalerite + molibdenite + galena + tetrahedrite];
2. for the interval m700-1300: [quartz + alkali felspar + biotite/chlorite + tourmaline + rutile + anatase +magnetite + pyrothite + pyrite+chalcopyrite + (native Au) + molibdenite + marmatitic sphalerite + galena + galeno-bismuthine(?)].

Quartz-calcite-sulphides veins cross the porphyry system.

Telescoped system generally display later formed mineralization typical of higher crustal levels overprinting deeper earlier formed mineralisation. Precious metal mineralization occurs in veins and stockwork systems near surface. Gold grades apparently decrease with depth, with a corresponding increase in copper values as the porphyry system is approached.

The veins, breccias and stockwork zones (25-Corabia Vein with the branches: Haber, Scarii, Spoiala and Robotin; hydrometasomatic pyrite lenses, stockworks, the Rosia group with the veins:



18, 19, Vâna Mare, Fântâna, Tulnic, Iolanda) are hosted in an andesite sub-volcanic intrusion and in the surrounding Cretaceous sediments and volcaniclastics. The quartz-calcite veins can be traced for over 500 meters along strike and up to 500 meters down dip. The veins are generally less than one meter in thickness, but can attain widths of over two meters (Cioflica et al., 1962). The veins and associated stockwork and breccia zones contain abundant sulphides.

For 25-Corabia vein the assemblage is pyrite, chalcopyrite, marcasite, arsenopyrite, galena, sphalerite, sylvanite, petzite, tetrahedrite, gold (Ghitulescu & Socolescu, 1941].

For Rosia system the upper part of mineralized column the assemblage is: pyrite, chalcopyrite, marcasite, arsenopyrite, galena, sphalerite, sylvanite, petzite, tetrahedrite, plumbomolybdate, and gold, in quartz, calcite and argillic minerals gangue and in the depth predominant assemblage is: sphalerite, galena, and chalcopyrite.

The convective circulation of the hydrothermal solutions expanded a halo of overlay potassic-intermediate argillisation-sericitic alterations around the veins. The conglomerates and ophiolites are host of the veins and undergo a process of potassic alteration through quartz-adularia fissures, or by substitution of cement and compound elements with quartz and adularia /illite and pyrite impregnation. Silicification involves an increase in proportion of quartz in altered rocks and associated with polymetallic-gold veins.

The Muncaceasca East deposit is situated immediately east of Muncaceasca West. The vein group (1 vein, 2 vein, 3 vein, 4 vein and B vein) consists of a number of quartz-carbonate-sulphide veins, with associated stockwork and breccia zones, hosted in an andesitic sub-volcanic intrusion and associated extrusive rocks.

The andesitic host rocks contain a number of northwest trending quartz-carbonate veins that dip steeply to the southwest. The veins have strike lengths of around 500 meters, and have been traced for approximately 500 meters down dip. The veins are generally thin, usually in the order of 0.5 meters, but can attain widths of up to 4 meters. Most veins are associated with stockwork and breccia zones that are preferentially developed in the footwall of each vein. Quartz-carbonate veins contain abundant sulphides, including pyrite, marcasite, galena; sphalerite and chalcopyrite. The andesites are strongly altered, and are contained within the same alteration envelope that encompasses the Muncaceasca West and Popa-Stanija group of deposits. Sericitic alteration is the most abundant, widespread and significantly.

The Popa-Stanija area is located, immediately north of Muncaceasca East and Muncaceasca West. It consists of a number of epithermal quartz-carbonate vein and stockwork systems that are hosted in a porphyritic andesitic sub-volcanic intrusion (Ungurului Hill) and associated extrusive rocks, and to a lesser degree, in Cretaceous sediments. The area consists of a number of related vein and stockwork systems, known from west to east as Malita-Colt group, Stanija group (with the veins: Vilanel, Ana, Ludovica, Lazar, Villanel, Teluros, Fortuna, Aurel, Gratiela Emanuel, Butac, Sfânta Treime, Iulius, Wolhirth, Sever, Viorel, Elvira, Ieronim, Andronic), Piua, Bradisor stockwork, Barburisca stockwork and Magura (breccia-vein Magura, 17 vein, 650 vein, 500 vein, Sorina vein). Veins and fractures contain abundant sulphides, including pyrite, marcasite, galena, sphalerite and chalcopyrite. Mineralization shows typical zonation patterns, from gold and silver-rich near surface, to more base metal-rich, then copper-rich ores with depth. Tellurides and native gold are common at shallow levels (Ghitulescu & Socolescu, 1941). The vein systems generally have overall thicknesses of between 20 and 40 metres. Individual veins are steeply dipping, and have a northwest strike direction. Strike lengths in the order of 500 metres are reported. The vein systems are hosted in strongly altered andesites that are contained within the large alteration halo that encircles the mineralization at Muncaceasca East and Muncaceasca West. Potassic alteration of andesites and hydrothermal breccias consists of varying proportion of quartz,



adularia, chlorite, neoformation biotite, tourmaline, rutile, and iron hydroxides (Popa et al. 2004). The ground mass and the fenocrysts are transformed and replaced by neominerals association. Sericitic alteration it is in the vicinity of the veins and is the most abundant, widespread and significantly (table 1). The sericitic alteration is observed in andesites, conglomerates, sandstones and breccia bodies and it is characterized by the neomineral association, with areas of significant silicification.

Magura vein is located, immediately east of Popa Stanija system and has a genetic link with Magura andesitic stock and intrusive breccias, which have penetrated the stock. It consists of a number of epithermal quartz-carbonate breccia-vein systems that are hosted in a porphyritic andesitic sub-volcanic intrusion and associated extrusive rocks, and to a lesser degree, in Cretaceous sediments and ophiolitic rocks. The breccia-vein systems generally have overall thicknesses of between 8 and 30 metres. The breccia- vein can be traced for over 400 metres along north-west strike and up to 120 metres down dip (3 mining levels). Breccia-vein and fractures contain abundant sulphides, including pyrite, sphalerite, galena, and rarely chalcopyrite, and sporadically tetrahedrite, marcasite, and gold. Gangue mineral comprise quartz, calcite and clay minerals

The Almas-Neagra vein system is located immediately east of Muncaceasca East. It consists of two groups of mineralised vein systems, known as the southern vein group, and the central vein group. The southern vein group (2- Concordia vein) is hosted in a quartz andesite sub-volcanic intrusion (Neagra stock).The central group (Dascaleasca, Gheorghe, Ovidiu, Alexandru, Horatiu, Tulia, Elena and also her ramifications) is hosted in the contact zone between the andesitic intrusion and Law Cretaceous sediments. The veins attain thicknesses of up to 4 meters, and have been traced along strike for up to 600 meters. They are sometimes associated with large zones of mineralized breccias.

The Almas-Neagra vein systems are strongly mineralized in precious metals, and contain abundant sulphides, including pyrite, arsenopyrite, sphalerite, chalcopyrite, galena, gold, tetrahedrite and jamesonite. Gold is associated with galena, sphalerite and arsenopyrite, and with quartz and calcite. Characteristic for these veins was association gold–galena (Petrulian, 1943). The gangue of mineralisation is generally composed of quartz, calcite and clays. The vein system is hosted in strongly altered andesites that are contained within the large alteration halo that encircles the mineralization at Popa Stanija and Muncaceasca East. Altered wallrock comprise the intermediate argillic alteration which it is like an envelope overlays between potassic and sericitic alteration (Popa et al, 1990). It conserves the porphyry texture of the andesite, the fenocrysts of the plagioclase and the mafics have a uniform aspect and are substituted by neominerals association.

Conclusions

In A-S structure, a vertical and lateral zonation of mineralization is present varying with depth and distance from up flow zones. The geological data reveal a horizontal and vertical alteration zoning which surrounding the vein systems, increasing in intensity to mineralize veins.

Mineralization consists of epithermal gold- base metal mineralization and porphyry copper mineralization, associated with intense hydrothermal activity.

Hydrothermal systems were formed as a response to the cooling intrusions and active groundwater convection cells.

In agreement with Hunt (1991), the porphyry copper deposit and associated epithermal veins belong to a longer, more complex and dynamic hydrothermal process, in which the gradually changing conditions create multiple overprinting events.



A model of polyphasic overprinting events is more suitable for this case (Cioflica & Lupulescu, 1998), similarly as the conceptual model elaborated by Corbett & Leach (1996).

Studies in active porphyry systems indicate that hydrothermal alteration results from a prolonged history of activity, ranging from: initial emplacement of the intrusive, to a sequence of events involving evolution of fluids from the cooling melt followed by influxes of meteoric waters at progressively lower temperature (Corbett & Leach, 1996). Initial emplacement of melts at high crustal levels and associated cooling and crystallisation is accompanied by the formation of zoned alteration assemblages formed in response to the transfer of heat from the melt into host lithologies. The alteration mineralogy grades from an inner potassic zone outward to progressively cooler propylitic alteration assemblages.

The convective circulation of the hydrothermal solutions expanded a halo of overlay potassico-intermediate argillisation-sericitic alterations around the veins

The epithermal mineralization is hosted in veins, fracture zones, stockworks and breccias, predominantly within volcanic rocks, sometimes within the Cretaceous sediments and rarely in ophiolitic rocks. Precious metal mineralization occurs in veins and stockwork systems near surface.

The porphyry mineralization is hosted in a porphyritic andesitic intrusion (microdiorite stock) that contains abundant quartz veinlets and sulphide stockworking. Pyrrhotite, chalcopyrite, pyrite, magnetite, molybdenite assemblage is the dominant in porphyry copper system.

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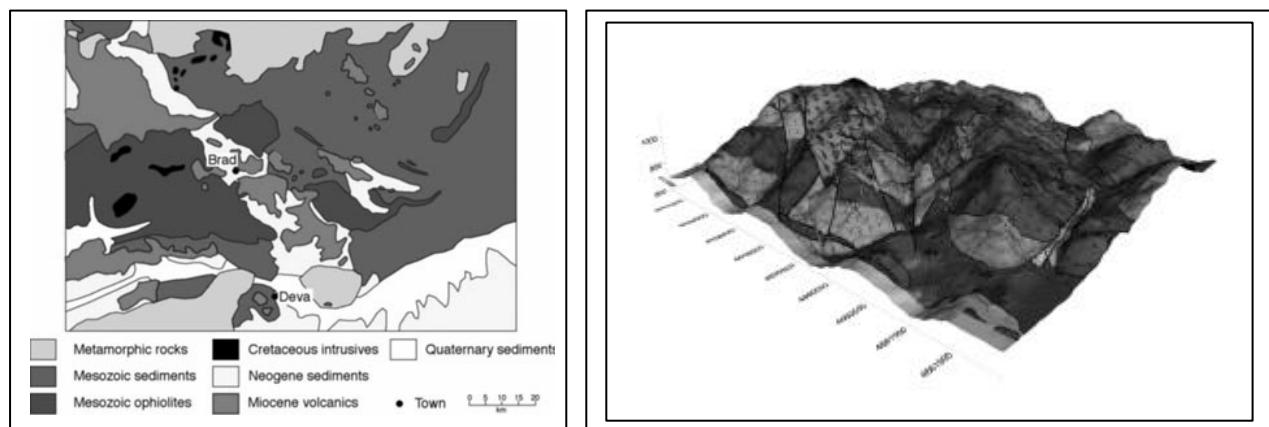


Figure 1. 3D view of Almas-Stanija Geological Map

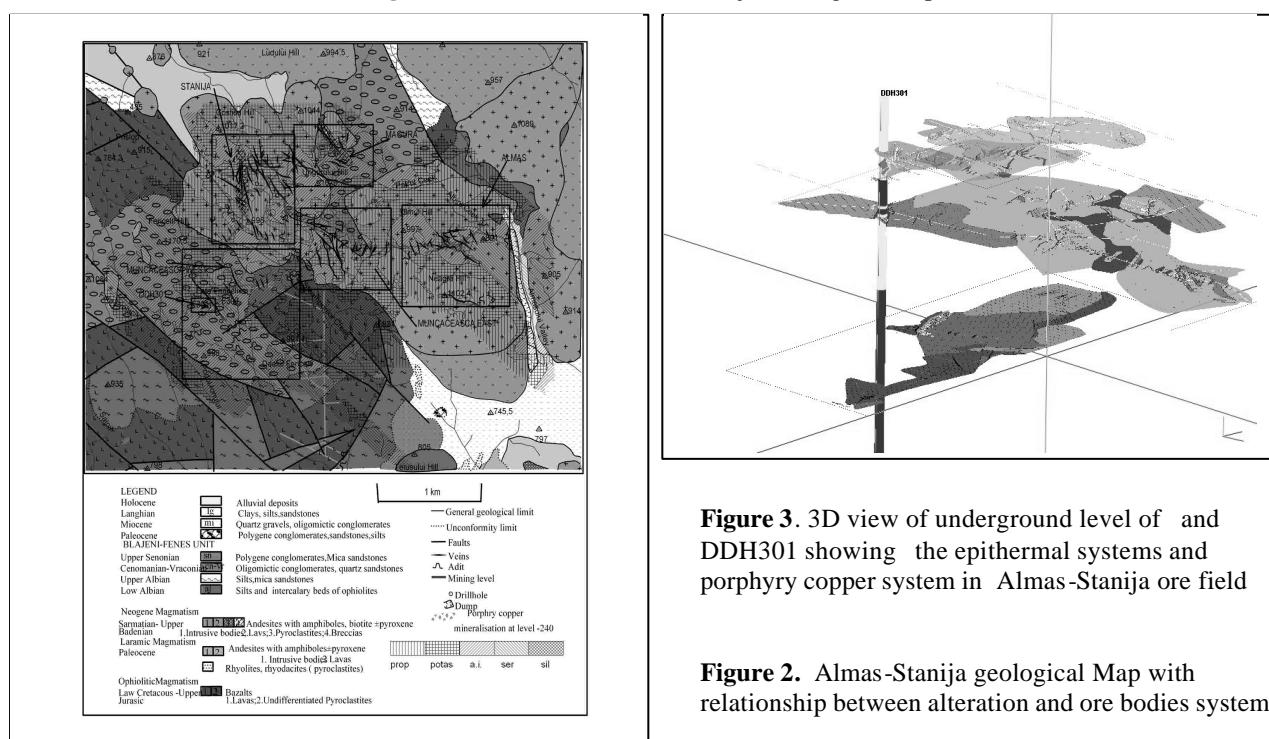


Figure 3. 3D view of underground level of and DDH301 showing the epithermal systems and porphyry copper system in Almas-Stanija ore field

Figure 2. Almas-Stanija geological Map with relationship between alteration and ore bodies systems


Table 1. Zone alteration adjoining sulfide ore bodies in Almas-Stanija orefield

The ore deposits		Alteration zones						Type of vein and mineral content			
		Host rock			Vein, breccia,stocks						
Muncaceasca West	Vein system	Congl.	Intermediate argillic zone illite -montmorillonite – chlorite –calcite –pyrite - iron hydroxides			Potassic (adularia)	Sericitte	Silicified	Quartz-carbonate veins, breccias with pyrite, marcasite, galena, sphalerite, and chalcopyrite, gold		
		Sandstone	Phyllitic zone (quartz-clays-sericite-pyrite)			Potassic (biotitisation) (potassium feldspar-biotite-magnetite-pyrite)	Stockwork Pyrrhotite+chalco pyrite+pyrite+molybdenite+magnetite				
Muncaceasca East	Porphyry	Microdiorite	Propyllitisation chlorite-pyrite-epidote-carbonate	Andesites			Propillitic fringe		Intermediate Argilisation	Pervasively sericitised zone	Quartz-carbonate veins including pyrite, marcasite, galena, sphalerite and chalcopyrite, gold
Popa-Stanija	The Almas-Neagra vein system	Andesite	Breccia	Propillitic fringe			Intermediate argilisation	Potassic-adularia	Sericitic Zone: sericite- illite-quartz- calcite ± tourmaline-muscovite- pyrite		Quartz-carbonate veins, breccias Pyrite, marcasite, galena, sphalerite, chalcopyrite tellurides, native gold
		Andesite	Propillitic fringe	Andesite			Argilisation	Sericitic zone Silicification		Quartz-carbonate veins, breccias Pyrite, arsenopyrite, sphalerite, chalcopyrite, galena, gold, tetrahedrite and jamesonite.	



MAGMATISMUL CALCOALCALIN SI METALOGENEZA PORPHYRY COPPER; IMPLICATII PRIVIND GEODINAMICA MUNTILOR METALIFERI

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Abstract

The Neogene magmatism and metallogenesis of the Apuseni Mountains is subject of numerous works referring to tectonic reconstruction of the Carpathian-Pannonian region. As many authors suggest from the beginning of '70s, the Carpathians consist of a series of structural zones closely comparable to the zones of island arc regions (subduction models); the Transylvanian Basin may be interpreted as an interarc basin. Recently, according to paleomagnetic data, it had been established that the Apuseni Mountains underwent very fast clockwise rotation since the Middle Miocene to at around 12 Ma: so, the geological evolution of the Apuseni Mountains is reviewed with respect to the interplay of geodynamics and geochronology evolution (liniamentary-basinal models). No one of these models insist on the metallogenetical characteristics. But the mineralogy, metallogenetical paragenesis and hydrothermal alteration characteristics of porphyry copper deposits are similar to those of SE of Asia. So the metallogenesis of the Apuseni Mountains could be better explained by the subduction models of their geological evolution.

Magmatismul si metalogeneza neogena din Mt. Metaliferi au constituit în ultima vreme obiectul unor referiri mai restrânsse sau mai ample, în numeroase lucrari privitoare la evolutia cruste terestre din spatiul Mt. Apuseni sau al ariilor carpatica si/sau pannonica (Ianovici et al., 1976, Balintoni & Vlad, 1998, Huismans, 2001, Ciobanu et al., 2004). În mod evident acestea au vizat în primul rând caracteristicile petrologice si doar cu totul subordonat pe cele metalogenetice.

Există mai bine de o duzina de modele elaborate în spiritul tectonicii placilor privitoare la România/spatiul carpatic, si implicit la Mt. Apuseni de Sud.

Metalogenia modernă a adus în prim-planul filosofiei explorării geologice corelatia de tip determinist în triada subductie? magmatism calcoalcalin? metalogeneza porphyry copper. Cu alte cuvinte, existența primilor doi termeni ai acestei triade presupune cu necesitate și pe cea a ultimului termen. Pe acest rationament s-au bazat proiectele de explorare de incontestabil succes în toate secolele de subductie cu magmatite calcoalcaline, inclusiv în România (ex. descoperirile structurilor porphyry copper din Mt. Metaliferi din anii '70 sunt o marturie în acest sens).

Primele modele geodinamice referitoare la Mt. Carpati au fost de tip subductionist, atât modelul Radulescu-Sandulescu (1973) cât și variantele modelului Bleahu-Boccaletti și colab. (1973), suportul lor argumentativ fiind de natură petrologică și geochemicală.

Ulterior, după evidențierea rotației în sensul acelor de ceasornic a Mt. Apuseni, și apoi a determinărilor de vîrste radiogene ale magmatitelor din spatiul Metaliferilor, s-au elaborat și alte modele, între care cele cu caracter "liniamentarist-bazinal", care sunt astăzi cele mai citate (Seghedi, 2004, Ciobanu et al., 2004).

În ceea ce privește componenta metalogenetică a argumentării, aceasta lipsește sau este nesemnificativa. În plus ultimele modele nici nu abordează problemele de metalogeneza,



considerându-le implicate; astfel, lucrările care fac referire la metalogeneza, fie nu fac referire la modelele geodinamice, fie se raportează fără discernământ, la unul sau altul dintre modele, de regula la cele mai recente.

Dupa parerea noastră, caracteristicile metalogenetice ale Mt. Metaliferi, definite prin existența binecunoscutei metalogeneze epitermale auro-argentifere și a metalogenezei porphyry copper descoperite în ultimele decenii, pot fi explicate mai convingător din perspectiva modelelor subductioniste, care pledează pentru o evoluție geodinamică în regim de arc insular (Boccaletti et al., 1973, Bleahu et al., 1973).

Prin toate caracteristicile sale – alcătuirea mineralogică, parageneza metalogenetică, structura haloului de alterație hidrotermală - metalogeneza de tip porphyry copper din Mt. Metaliferi prezintă foarte multe similarități cu metalogeneza porphyry copper din Asia de Sud-Est (mai ales din Filipine și partea estică a Arhipelagului Indonezian).

Evident că veridicitatea ipotezelor geodinamice, ca și oricărora ipoteze, trebuie apreciată în masură în care este explicată cvasi-totalitatea fenomenelor geologice din crustă sau din portiuni ale acesteia. Chiar dacă un model a fost elaborat pentru a explica o anumită situație geologică, el trebuie să elucidă și aspecte geologice care ies din sfera destinației sale. De exemplu, abordarea problematicii magmatitelor și a metalogenezei neogene din Apuseni de Sud, trebuie să fie explicată în corelație cu existența magmatitelor și a metalogenezei neogene din Carpații Orientali, sau cu prezența tufurilor din spațiul avanfosei neogene carpatici și Bazinul Transilvaniei.

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ANALIZA PETROGRAFICA A DEPOZITELOR PALEOGENE DIN NORD-ESTUL DEPRESIUNII GETICE - PETROTIPURI, RECONSTITUIRI DE ARII SURSA SI EVOLUTII DIAGENETICE

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La nivelul Rupelianului depozitele oligocene din nord-estul Depresiunii Getice sunt reprezentate prin *Conglomeratele de Cheia si Gresia de Corbi*.

Ambele formatiuni nu au continuitate laterală. *Conglomeratele de Cheia*, afloreaza pe Valea Cheia, Valea Olanesti si Valea Muiereasca. *Gresia de Corbi* afloreaza pe vaile Râul Doamnei si Vâlsan în localitatile Corbi, Bradulet, Bradet si, este predominant arenitica. Caracterul lor polimictic le confera calitatea unor petrotipuri optime pentru reconstituiri de arii sursa si evolutii diagenetice.

Faciesurile lor depositionale sunt diferite si sunt dominate de entitati petrografice acumulate prin curgeri gravitationale (debris flow), din curgeri turbulente canalizate si, respectiv, curenti turbiditici de densitate mare în sisteme marine de apa adâncă.

Examenul petrografic al *Conglomeratelor de Cheia* a pus în evidenta urmatoarele petrotipuri: *ortoconglomerate polimictice*, subordonat *paraconglomerate polimictice si gresii litice*, iar în faciesurile fine, *argile calcareoase*.

A putut fi precizata natura litoclastelor : metamorfite (70-80%) si sediclaste (20-30%). Litoclastele metamorfice sunt reprezentate de *gnaise ortoclazice*, *gnaise cu granati*, *micasisturi cu granati*, *micasisturi cu muscovit si biotit*. Pe lângă acestea au mai fost întâlnite *amfibolite cu hornblenda si plagioclaz*, *eclogite retromorfozate cu omfacit granat si disten*, de regula simplectitizat.

Litoclastele de origine filoniana, de tipul *pegmatitelor* prezinta cristale de dimensiuni mari de Q, Fk si Fp. Rocile carbonatice metamorfice sunt reprezentate de un *calcar cristalin*.

Sediclastele aparțin calcarelor si gresiilor litice.

Petrotipurile carbonatice sunt variate si reprezentate prin : *pelsparit, intrasparit, biosparit, micrit*. De regula, aspectul este brecios (calcirudite polimictice). In gresiile litice pe lângă quart (Q) si feldspati (F- pâna la 50%) contin si litocaste sedimentare carbonatice (*pelsparite*) cu o pronuntata amprenta diagenetica: cimentari carbonatice, recristalizari ale matricei micritice, procese de metasomatoza carbonatica pe feldspati si Q, supracresteri de quart autigen pe claste de quart alogen. Cantitatea mare de feldspati, releva o sedimentare rapida, fara ca influenta climatica sa fi avut timp sa provoace argilizarea feldspatilor.

Din punct de vedere al provenientei si semnificatiilor paleoclimatice asociatiile descrise indica o sursa majoritar metamorfica si secundar plutonica, într-un climat arid. Rata de sedimentare rapida este coroborata cu crearea unui spatiu disponibil de sedimentare în bazin, activat tectonic (prin subsidenta si/sau înaltarea ariei sursa) sau eustatic.



Compozitia clastelor din *Conglomeratele de Cheia* arata multe similitudini cu petrotipurile formatiunilor cristaline apartinând Muntilor Căpătâni (cu o componitie predominant gnaisica) si cu cele sedimentare cretatici si eocene de pe rama acestui bazin.

Examenul petrografic al *Gresiei de Corbi* a evidențiat urmatoarele petrotipuri: *gresii subcuartoase, gresii subfeldspatice, gresii litice*, si, subordonat *paraconglomerate oligomicetice si ortoconglomerate polimictice*. Litoclastele, majoritare în cadrul faciesurilor ruditice, sunt *gnaise oculare, quartite, micasisturi, milonite* precum si litoclaste sedimentare: *orthoconglomerate polimictice, gresii litice, calcare, marne* (argile).

Studiul granoclastelor (feldspati si quart), prin analize morfometrice si prin examenul produselor secundare a reliefat doua categorii: granoclaste proaspete, cu grad mai mare de angulozitate si granoclaste rotunjite, partial pseudomorfozate prin minerale argiloase si calcit (deci, evident, reciclate). Examenul optic a stabilit provenienta lor metamorfica si pegmatitica .

Pentru *Gresia de Corbi*, trasaturile mineralogice mentionate atesta apartenența lor la metamorfitele din sudul Muntilor Fagaras si, respectiv, la formatiunile sedimentare din nordul acestuia.

Testând si calitatea de colector pentru fluide a acestor siliciclastite (*Conglomeratele de Cheia si Gresia de Corbi*) prin analiza produselor diagenetice am putut constata evolutii aproximativ similare : procese de cimentare locala cu ciment târziu, carbonatic, poikilitic , o porozitate secundara considerabila si lipsa unor efecte ale îngroparii foarte adânci. In aceste conditii nu a fost intersectata *fereastra de petrol*, iar porozitatea primara s-a putut mentine si ea, ridicata, imprimând corpurilor depozitionale calitatea de roca magazin.

In aceste conditii, atât *Conglomeratele de Cheia* cât si *Gresia de Corbi*, prin geometria corpurilor cât si prin caracteristicile diagenetice – porozitate efectiva mare si procese poronecrotice care au afectat local depozitele în ultima etapa diagenetica – formeaza coruri cu bune calitati de rezervor de hidrocarburi.



POZITIA STRATIGRAFICA A STRATELOR DE TIP PUCIOASA DIN NORD-ESTUL DEPRESIUNII GETICE

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Stratele de tip Pucioasa, de vîrstă oligocena, (Chattian), au grosimea maxima de 700 m. Depozitele sunt predominant lutitice și siltice cu intercalatii arenitice și ruditice, în baza. Afloreaza în NE-ul Depresiunii Getice, între Valea Otasau, în vest, și valea Râul Doamnei, în est. Perimetrelle cercetate sunt localizate în zona vestică, pe valele Cheia, Olanesti și Muiereasca, iar în est, pe valea Vâlsan și Râul Doamnei.

Pozita biostratigrafica și nomenclatura acestei unitati litologice a ridicat controverse, legate atât de vîrstă, cât și de relata sa cu unitatile adiacente: Conglomeratele de Cheia și Gresia de Corbi în baza formatiunii studiate, respectiv, Gresia de Muiereasca și Conglomeratele de Matau în partea ei superioara și vîrstă miocena.

Conform lui Popescu B. et al. (1976) și Bombita et al. (1980) la nivelul Oligocenului Gresia de Corbi, din valea Vâlsanului trece lateral în Marnele de Jiblea (în Valea Argesului) și au drept echivalent stratigrafic, în valea Olanesti, Conglomeratele de Cheia. Revenind cu un studiu mai recent, Rusu A. et al . (1996), limiteaza extinderea Conglomeratelor de Cheia la vest de Valea Muiereasca și defineste, la est de aceasta, unitatea Conglomeratelor de Frâncești (200 m grosime), mai noi decât Conglomeratele de Cheia.

Lucrarea își propune studierea în detaliu a litofaciesurilor și secventelor de facies ale *Stratelor de tip Pucioasa* și corelarea lor, de la V la E, pe criterii sedimentologice.

Prin metoda analizei secventiale, pe criterii granulometrice și structurale au fost separate 13 faciesuri depozitionale: 1-rudite grosiere masive; 2-rudite medii și fine masive; 3-rudite fine laminat paralel; 4-arenite masive; 5-arenite normal granoclasate; 6- arenite laminat paralel; 7-arenite cu laminatie oblic tabulara; 8-arenite cu ondulatii de curent; 9-cupluri arenit/lutit, 10-arenite lenticulare, 11-cupluri silturi și lutite cu laminatii paralele; 12-lutite cu laminatii paralele și, 13-facies particular megalutitic deformational.

Caracteristicile, granulometrice, morfometrice, structurale precum și natura limitelor dintre unitatile faciesale sugereaza procese gravitationale și procese tractive. Faciesul *megalutitic* a fost generat prin procese deformaționale de tip *slump* evoluat în procese de tip *mud flow*, iar faciesurile *foarte grosiere* reprezinta curgeri de fragmente necoezive (*debris flow*). Faciesul *rudit laminat paralel* este interpretat ca un covor de tractiune, faciesurile *arenitice* au fost generate de curenti tractivi în regim de curgere superior. *Arenitele* din cuplurile arenit lutit și arenitele lenticulare au fost formate prin procese tractive în regim inferior, iar faciesurile *fine* au fost generate prin depunerea din suspensii.

Asociatiile faciesale ale Stratelor de tip Pucioasa, arata caracteristici diferite între zona vestica și cea estica. Astfel, în zona estica au fost separate în functie de caracterisicile litologice 3 asociatii: asociatia tip **I** arenito-lutitica, cu intercalatii « disodilice », asociatia de tip **II** arenito-



ruditica si asociatia de tip **III** silto-lutitica. În zona estica sunt prezente în baza, asociatia de tip I peste care se suprapune asociatia de tip III.

In privinta modelului stratigrafic se constata ca din punct de vedere litologic formatiunea Stratelor de tip Pucioasa nu este omogena, în partea vestica interpunându-se o secventa conglomeratica , de numai 10 m grosime. Aceasta secventa, mai noua decât Conglomeratele de Cheia ar putea corespunde formatiunii Conglomeratelor de Frâncesti de pe valea Muiereasca, (definita astfel de Rusu et al., 1996, dar apreciata ca fiind mult mai groasa). Aceste nivele conglomeratice sunt corelabile cu nivelele mai subtiri din Valea Cheia. Din punct de vedere depositional, consideram aceste conglomerate ca pe niste canale amalgamate, cu efilari laterale. Lipsa unitatii ruditice , în partea estica, poate fi datorata unor accidente tectonice sau lipsei unei surse corespunzatoare.

Modelul de facies preliminar pentru asociatiile I si II corespunde unui sistem depositional de apa putin adâncă, self proximal, iar modelul aferent asociatiei de tip III, unui self distal si taluz.

Modelul stratigrafic propus urmeaza a fi verificat cu metode biostratigrafice.



ANALIZA SECVENTELOR DE FACIES DIN UNITATEA LITOSTRATIGRAFICA A MARNELOR DE OLANEȚI DIN VALEA VALSAN, DEPRESIUNEA GETICA

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A fost analizată unitatea litostratigrafică a Marnelor de Olanesti de vîrstă Eocen superior, din Depresiunea Getica, în scopul stabilirii faciesurilor depozitionale, ierarhiei secventelor și definirii arhitecturii corpurilor sedimentare.

Formatiunea are o continuitate laterală bună, întâlnindu-se din Valea Otasau, în vest, până în valea Râul Doamnei, în est. Perimetru cercetat se află pe Valea Vălsan și affluentul său, Pârâul Mierlei. Grosimea formatiunii pe valea Vălsanului este de 400 m. Depozitele sunt predominant lutitice și siltice, cu intercalatii arenitice și ruditice.

Pe criterii granulometrice și structurale au fost separate 10 faciesuri depozitionale: 1-pararudite cu matrice lutitică și stratificare paralela difuză; 2-pararudite grosiere masive cu matrice grezoață; 3-pararudite medii masive cu elemente de tipul galetilor moi; 4-rudite medii și fine, masive cu tendință de granoclasare normală; 5-ortorudite fine laminat paralel cu tendință de granoclasare normală, 6-arenite masive; 7-arenite cu laminatii paralele; 8-arenite cu ondulatii de curent; 9-cupluri silt/lutit cu intercalatii arenitice, și, 10-lutite cu laminatii paralele.

Caracteristicile granulometrice, morfometrice, structurale precum și natura limitelor dintre unitatile depozitionale, sugerează procese gravitaționale și procese tractive. Faciesul 1 este format prin procese de tipul curgerilor laminare, coeze, faciesul 2 sugerează curgeri de fragmente necoeze, *ortoruditele* au fost generate prin procese tractive de fund, *faciesul rudit laminat paralel* este interpretat ca un covor de tractiune, *faciesurile arenitice* au fost generate de curenti tractivi în regim de curgere superior, iar *faciesurile fine* au fost generate din suspensie.

Au fost separate 3 asociatii de facies în funcție de caracterisicile litologice: Asociatia A, în baza, de aprozimativ 45 m grosime, reprezentata printr-o secventa masiva debritica, în care sunt intercalate asociatii de canal de tip FUS (*Marnele tiloide* –Jipa, 1980); Asociatia B, cu 2 aparitii (75m si 175 m) reprezentata de macrosecvente alcatuite din microsecvente FUS si ThNU (45 cm), formate din cupluri de gresii decimetrice si nivele lutitice cu intercalatii siltice; Asociatia C (85 m) reprezentata de silturi si lutite laminat paralel cu aspect omogen.

Modelul de facies preliminar pentru asociatia tip A apartine unui sistem depozitional marin, de panta, iar asociatiile B si C apartin domeniului marin de apa putin adanca, self proximal si distal.



SCIENTOLOGY ON THE STRATIGRAPHICAL INFORMATION REFERRED TO THE DOBROGEA TERRITORY

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Scientology is an interactive method of managing scientific information of public interest gathered in data banks.

This modern method of scientific documentation has been applied since 80^s (20th century) worldwide. In Romania, due to a harmful dialectical trinity of self-sufficiency, inertia and routine typical to an “anxious mentality”, scientology has not found yet a practical applicability.

This paper intends to evidence the theoretical and practical performances of this kind of managing scientific information of public interest.

In order to benefit from the opportunities offered by scientology as a method of management of scientific information, it is necessary first to make a performing data bank on a certain domain (Stratigraphy – in this case) or on a certain territory (Dobrogea).

Information of public interest are obtained by analytical computerized listing of all the publications of the Geological Institute of Romania, of all the yearbooks of the Bucharest, Cluj and Iasi Universities, as well as of the publications edited by the Romanian Academy and by some museums of natural sciences from Romania.

The analytical listing of these serials was performed by Excell computing program compatible Access from Microsoft Office.

The file has three functional modules:

- A. Bibliographical module;
- B. Scientific module;
- C. Organizing module.

To the classical Bibliographical module which comprises name of authors, title, year of publishing, collection, publishing house, town, country and number of pages, I needed to add some extra information like: number of authors, sex and nationality of the first author, editing language of the article, type of document, and ISSN - ISBN number of the publication. Thus I promote the theoretical and practical bases of the *information sociology*, term upon I claim paternity.

The scientific module is complex and responds to the necessity of classifying according to all the selection criteria I could imagine for all natural sciences, which constitutes, besides geology, my topic of interest. The information concerning the listed titles of articles is distributed orderly in the scientific module column. In order to preserve information, which could not be distinctly classified and which I call *residual information*, I imagined two columns for key words.

The organizing module is essential for the geological bibliography of Romania because it links the virtual information obtained from Internet and its location in a public or private library.

Finally I emphasize once again the big difficulty I deal with in reaching my professional goal. It is about the great dispersal of information in tens of thousands of publications that even the big library of the Romanian Academy does not owe totally and to which unfortunately I shall have no access without your help.



GEOCHEMICAL SAMPLING IN AMPOIULUI VALLEY- ALMASULUI VALLEY AREA, SOUTH APUSENI MTS., ROMANIA

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During the last years, the international community acknowledged the fact that, in order to sustain economic development and to accelerate scientific and technological progress, constant efforts in the direction of ensuring the protection of the environment, as well as a rational utilization of mineral resources became imperative to undertake.

The interdisciplinary projects achieved in common by UNESCO and UNAPPT have lead to the conclusion that, when planning activities using natural resources, it is mandatory to take into account the geological and geochemical aspects with decisive impact upon the lithosphere.

The present study, "Geochemical sampling in Ampoiului Valley-Almasului Valley perimeter", treats about a significant aspect of influence upon the environmental factors (soil, water, plants, alluvial deposits) of the activities associated with polymetallic ores exploitation and processing in Zlatna area.

The geochemical sampling carried on in Ampoiului Valley-Almasului Valley perimeter revealed some aspects associated to pollution problems in this sector.

Samples from the fine fraction of stream sediments, soil, water and vegetals were systematically collected.

All those samples were analyzed to determine the contents of Cu, Pb, Zn, Ag, Mo, Sn, Bi, As, Cd and Mn.

Mono elementary maps at 1: 50,000 were elaborated for all the analyzed metals and sampled environments. The main aspects resulted from the analytical data interpretation are the following ones:

- a massive pollution of the environmental factors (soil, water, vegetals), especially developed along the main water courses and their banks, is present in the sector;
- the principal pollution sources are the metallurgical plant, the ore preparation station and the areas in which polymetallic and gold and silver ores are exploited;
- the dissipation of the noxious compounds is achieved mainly by the mean of atmospheric currents and running waters.

We mention that, in order to determine the influence of the environmental factors pollution, the majority of the European countries have already elaborated national geochemical maps atlases, showing the distribution in surface of different heavy metals. The principal objectives of these programs were the following ones:

- to determine of the background levels of the metallic elements in sampled environmental factors (soil, water, vegetals, alluvial deposits);
- to identify interesting mineralizations of metals scarcely studied in the past (Sn, W, T.R.);
- to obtain useful data for statistic epidemiological surveys for human, animal or vegetal degenerative disease.



THE PRESENT DAY MINERALOGY OF ROSIA MONTANA ORE DEPOSIT, SOUTH APUSENI MOUNTAINS, ROMANIA

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Abstract

A complete list of clearly identified mineral species and several uncertain minerals known in Rosia Montana ore deposit and the host rocks is provided. The complete references that include also the identification techniques where available are highlighted for all these minerals. The mineral list comprises the rock forming minerals (principal, accessory, and evaporitic), as well as the minerals related to hydrothermal activity i.e. alteration minerals, ore minerals, gangue minerals and the secondary minerals. The sources for this study start with Zepharovich's volumes (1853, 1873, 1893) and his cited former authors and end with mineralogical contributions presented in 2004 at the 32nd International Geological Congress, Florence, Italy (August 20-28), and the IGCP 486 project workshop held in Alba Iulia – Rosia Montana, Romania (August 31st –September 6th).

Key words : *Rosia Montana, mineralogy*

Introduction

Since Zepharovich's mineralogical index more and more new data and a lot of new minerals have been identified at Rosia Montana, South Apuseni Mountains, Romania. The up to date mineralogical record of this area will be presented herein.

Known as a major Au-Ag epithermal ore deposit in Europe, Rosia Montana is worldwide known for its amazing free gold samples, spread today all over the world. Leaving apart the free gold specimens, there are also many other minerals that have been recently discovered in this ore deposit and consequently Rosia Montana is once again in the mineralogical actuality. At least the recently confirmation of Petruțian's (1934) Ag-minerals association by Tamas (2002), the secondary minerals reported by Onac et al. (2003), and the tellurides identified by Tamas et al. (2004) and confirmed by Ciobanu et al., (2004) have been made due to the development of the Rosia Montana mining project promoted by Rosia Montana Gold Corporation and Gabriel Resources.

Minerals so far identified at Rosia Montana

Certainly the free gold (electrum), the silver minerals and the tellurides or generally speaking the ore minerals are the most striking minerals species from the Rosia Montana ore deposit, but there are also many other minerals identified in this area. For simplicity sake we can separate several groups of minerals:

1) rock forming minerals (1.1 principal, 1.2 accessory and 1.3 evaporitic);

2) minerals related to hydrothermal activity (2.1 alteration minerals, 2.2 ore minerals, 2.3 gangue minerals);

3) secondary minerals.

The methods of identification are to be noticed, too. For the older references the recognition of the minerals has been in most of the cases limited to macroscopic identification. Instead of this



method the accuracy of identification can be estimated taking into account the minute crystallographic descriptions made by Goldschmidt (1913-1923) for the free gold from Rosia Montana, for example. Step by step the identification of the minerals greatly benefited by various analytical techniques like optical microscopy (transmitted and reflected light), X-ray diffraction, thermal analyses, electron microscopy (SEM and TEM), and electron microprobe analyses.

1. Rock forming minerals

These minerals are the basic components of the volcanic, sedimentary and metamorphic rocks. The below mentioned minerals have been identified from the volcanic and sedimentary rocks found in Rosia Montana area as well as from various fragments found in volcanoclastics as well as within breccia pipes bodies.

1.1 Principal minerals

The petrogenetic minerals have been described by various authors that conducted mineralogical investigations in Rosia Montana area, precisely mineralogical descriptions of Cetate dacite and Rotunda andesite. Among them we can mention Cozubas et al. (1986), Mărza et al., (1990), Mărza and Ghergari, (1992), Tamas (1995), Mărza et al., (1997), Tamas (2002). The following minerals have been mentioned:

BIOTITE

FELDSPARS (POTASSIC and PLAGIOCLASE).

Cioflica et al. (1973) reported an An_{30} plagioclase from Cetate dacite, while Cozubas et al., (1986) offer a larger mineralogical composition of plagioclase (An_{30-50}).

HORNBLENDE has been identified in Cetate dacite (Cozubas et al., 1986) but it is heavily replaced, as well as PYROXENE remnants found towards Gauri sector (Mărza and Ghergari, 1992). A variety of hornblende, namely OXYHORNBLRENDE has been reported by Tamas (1995, 2002) in Rotunda andesite using Fedorov method.

Magmatic QUARTZ rich in primary glassy inclusions is another characteristic mineral from Rosia Montana. This hexagonal bipyramidal quartz may reach up to 3cm.

Mărza et al., (1990) described a series of metamorphic minerals identified by microscopic investigations from the metamorphic rock fragments found in different breccia pipe bodies from Rosia Montana. These authors noticed the following minerals: ANDALUSITE, APATITE, GARNETS, ILMENITE, MAGNETITE, MUSCOVITE, RUTILE, TITANITE, TURMALINE, ZIRCON.

As concerns the garnets, Ciobanu et al. (2004) put into evidence the following mineral species: ALMANDINE, PYROPE, and SPESSARTINE.

1.2 Accessory minerals

Within the volcanic rocks (Cetate dacite and Rotunda andesite) APATITE, ORTHITE, ZIRCON have been identified by the means of optic microscopy (Cozubas et al., 1986, Tamas, 2002). ILMENITE and ANATAS are reported by Radulescu and Dimitrescu (1966).

1.3 Evaporitic minerals

DOLOMITE

This mineral was mentioned by Cadere (1926, fide Radulescu and Dimitrescu), with uncertain origin according to these latter authors.



GYPSUM

Borcos and Mantea (1968) described a gypsum horizon within the so-called Gray Marls horizon, located in the eastern part of the Rosia Montana basin.

2. Minerals related to hydrothermal activity

The hydrothermal activity from Rosia Montana is responsible for the pervasive alterations and the major ore body genesis. Due to this activity, a great number of minerals have been formed during the active period of the hydrothermal activity.

2.1 Alteration minerals

Potassic, phyllitic, argilic (intermediate and advanced) and silification alteration zones have been pointed out during the time by various authors by the means of optic microscopy, X-ray diffractions, thermal analyses, electron microscopy. The alteration minerals identified in the alteration zones (Cozubas et al., 1986, Mărza and Ghergari, 1992, Mărza et al., 1997, Tamas, 2002) are summarized below:

ADULARIA
BASSANITE
CHLORITE
HALLOYSITE
ILLITE
ILLITE/MONTMORILLONITE
ILLITE/CHLORITE
KAOLINITE
KUTNOHORITE
MONTMORILLONITE
PYROPHYLLITE.

NACRITE was also mentioned by Stoicovici (1937, fide Radulescu and Dimitrescu, 1966) from the alluvial deposits close to Salistei hill, Rosia Montana.

2.2 Ore minerals

Using for the first time in Romania the reflected microscopy for opaque minerals, Petruțian (1934) made a first systematic study of the ore minerals from Rosia Montana ore deposit. The following ore minerals have been identified by the above cited author:

ALABANDITE
ARGENTITE
ARSENOPYRITE
CHALCOPYRITE
GALENA
GOLD
MARCASSITE
PEARCEITE
POLYBASITE
PROUSTITE
PYRITE
SPHALERITE
TETRAHEDRITE.



Radulescu and Dimitrescu (1966) and Udubasa et al., (1992) mentioned other metallic minerals from Rosia Montana, like:

BERTHIERITE

COPPER (in fact from Rosia Poieni)

FREE SILVER

Ag-rich TETRAHEDRITE.

Tamas (2002) using electron microprobe analyses identified additional metallic minerals as follows:

ELECTRUM

FREE GOLD

TENNANTITE

STEPHANITE, which has been also reported by Helke, (1938) in microscopic associations with sphalerite.

The uncertain occurrence of ARGYRODITE¹ at Rosia Montana as emphasized by Tamas (2002) was recently confirmed by Tamas et al., (2004), Ciobanu et al., (2004), Bailly et al., (2005, in print) by the means of microprobe results.

Located in the South Apuseni Mountains a well known telluride rich province, Rosia Montana was for a long time considered a telluride free ore deposit. The occurrence of ALTAITE, HESSITE, and SYLVANITE was announced by Tamas et al. (2004) and confirmed by Ciobanu et al. (2004), which identified also PETZITE and CERVELLEITE, as well as PYROLUSITE at microscopic scale.

2.3 Gangue minerals

HELVITE. This mineral has been described by Helke (1938) from the Rosata Tare stockwork.

The Mn-rich gangue minerals (Benea et al., 2000) associated with the telluride assemblages consist of RHODOCHROSITE and RHODONITE. The Mn-carbonate was cited before by Petruian (1934), Ghitulescu and Socolescu (1941), while the Mn-silicate was described for the first time by Helke (1938).

CALCITE, and hydrothermal QUARTZ were reported by Petruian (1934) while BARITE by Radulescu and Dimitrescu (1966). ADULARIA is also a common mineral within the filling of the veins or the breccias (Mârza and Ghergari, 1992, Tamas (2002). Quartz amethyst may be also found in important amounts in several veins from Cârnica massif.

SIDERITE has been identified as fragments within the volcanoclastic sequences by Mârza et al., (1990). Schuster (1983) mentioned also the presence of FLUORITE at Rosia Montana, recently confirmed by field investigations (drillings).

3. Secondary minerals

Old references concerning secondary minerals are provided by Radulescu and Dimitrescu, but Onac et al. (2003) described recently nine secondary minerals found in abandoned underground developments in Cârnica massif. Secondary GYPSUM crystals up to 2cm long are very frequent mainly in old usually unvisited adits. The secondary minerals reported by Onac et al. (2003) have been identified by the means of XRD, energy-dispersive spectrometry, electron microprobe

¹ The microprobe results had shown a constant Te content of about 20wt% for the argyrodite from Rosia Montana.



analyses, optical and scanning electron microscopy. These new minerals for Rosia Montana are the followings:

ALUNOGEN
APJOHNITE
DIETRICHITE
HALOTRICHITE
JOKOKUITE
KALINITE
MELANTERITE
PICKERINGITE
ROZENITE.

Ackner (1885, fide Radulescu and Dimitrescu, 1966) mentioned also ALUMINITE as deposited from the mine waters underground, but no further references confirmed this secondary mineral.

ALUNITE has been mentioned by several authors during the XIX-th centuries as filling of the veinlets associated with secondary gypsum found in andesites from Chicera hill (Radulescu and Dimitrescu, 1966, Schuster, 1983). Zepharovich (1873) mentioned in the same location also native SULPHUR as crystals in quartziferous andesites.

Zepharovich (1859) reported the occurrence of EPSOMITE at Rosia Montana.

Radulescu and Dimitrescu mentioned also EVANSITE but these authors agreed that this is an uncertain identification.

MELANTERITE was also described from Rosia Montana by Maklári (1943, fide Radulescu and Dimitrescu, 1966).

VIVIANITE is cited by Radulescu and Dimitrescu (1966) and Udubasa et al. (1990) according to several old references, even associated with native gold.

GOETHITE may occurs as the sole mineral filling of narrow veins structures (up to 3cm width.) in Cârmic massif (Tamas, Ghergari, and Minut, unpubl. data).

Conclusions

The present state of the art concerning the mineral record from Rosia Montana area reveals the abundance of minerals in this ore deposit. Instead of the continuity of the mining confirmed by recent mining archeological investigations carried out by Cauuet et al., (2003) to be older than 2000 years, and the mineralogical investigations made for more than 150 years, Rosia Montana continues to provide new and new minerals. Our synthesis counted 81 minerals in Rosia Montana ore deposit.

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A BRIEF HISTORY OF GOLD MINING

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Gold was probably the first metal known to the early hominids, possibly as long ago as 6000 BC. Used initially in jewelry, it gradually became a major symbol of wealth and power, standard of value and money. In modern times, it has come to be used in industry such as dentistry, computers, electronic circuits, and even in the aerospace industry.

The earliest gold jewelry known in the world (about 4000 BC) was found in 1972 near the present city of Varna (Bulgaria).

By 3000 BC, gold rings were used as a method of payment by the Egyptians. A coin weighing 11.3 grams of gold (called *Shekel*), starts to be used as a standard unit of measure throughout the Middle East by 1500 BC, the gold becoming to be recognized as standard medium of exchange for international trade. The first gold bullion market was set-up about 1100 AD in Venice, trading over 5 tons of gold annually. The gold price established in London Bullion Market became to be used as reference price for international transactions since 1919. Since the first evaluation in US\$ (19.30 US\$/oz in 1792) the gold price increased about twenty times, reaching its historical maximum in 1980 (850 US\$/oz).

The principal source of gold in antiquity was undoubtedly stream placers. However there is evidence that in certain gold belts the alluvial deposits, auriferous gossans, and the near surface parts of friable (oxidized) veins were mined.

The earliest artifacts suggesting an extensive alluvial mining activity, dated about 3500 BC, were found in the ancient Mesopotamia. The archeological artifacts suggest that the alluvial mining started by 3000 BC in India, 1765 BC in China, 1400 BC in Europe, 1200 BC in Peru, 1122 BC in Corea and 900 BC in Indonesia. It seems that the earliest artifact sustaining the alluvial mining in Romania is the wood pan found in Capus (Bronze Age).

Early gold mining by the Egyptians is thought to have produced no more than 1 tone annually. Perhaps five- ten times more was produced during the time of the Roman Empire (mainly from Spain, Portugal, France, Romania and Africa).

In the Dark and Middle ages (500-1400 AD) production probably fell back to less than one ton.

From the middle of the 15th century, the Gold Coast of West Africa, now known as Ghana, became the most important source of gold of the world, providing 5-8 tons a year.

Between 1493 and 1600, the Romania's gold production represented about 19% of the gold production of the world. By the middle of the 19th century the percentage decreased to about 3.5%.

The first gold district discovered in South America after the Spanish conquest was El Oro (Mexic). During its mining history, that started in the 16th century (1521) and took place for about 400 years, about 150 tons of gold were produced. By 1550 alluvial placers were found elsewhere in Mexic (Veta Madre-Guanajuato), Columbia and Brazil. The famous Minas Gerais (Brazil) that was the leading gold producing mine from one century, with a total estimated production during his mining history of 1,300 tons of gold, was discovered in 1693.

The first gold discovery in North America has been made in 1803 in North Carolina (Reed Gold Mine). It should be noted that this marked the beginning of so called “gold rush” from the 19th century that affected the entire world.



The gold rush practically started in Russia. In East Siberia 57 alluvial deposits have been found between 1823-1842, Russia becoming the world major gold supplier. By 1847 Russian output accounted for 30-35 tons of the world total of about 75 tons.

The crucial turning point in the history of the gold mining industry came with the discovery of the first gold nugget at Sutter's Mill in the Sacramento Valley of California in 1848. The so-called Californian gold rush, extended further on in Nevada, involved about 500,000 people between 1849 and 1864, the annual gold production varying from 77 to 93 tons. The well-known gold districts Mother Lode and Grass Valley were found in 1850 and Comstock Lode (Nevada) deposit was found in 1859.

The discovery of payable placer gold in Australia was made first in 1851 near Bathurst (New South Wales). During so-called Victorian gold rush large alluvial placer and bedrock gold deposits were found at Bendigo and Ballarat (1851), Gympie (1867), Charters Towers (1870), Mount Morgan (1886), Kimberley (1886), Coolgardie (1892) and Kalgoorlie (1893). Following these discoveries the Australia gold production rose to 119 tons in 1903 (equal to the Australian gold production in 1988!).

In South Africa gold has first been found in Eastern Transvaal in 1873. The most important discovery from the gold mining history was made in 1886 when Witwatersrand deposit was found. It is estimated that Witwatersrand has been the source of close to 40% of all the gold ever produced in the world. It should be noted that in 110 years the mining reached about 3,500 m depth. Since the discovery of Witwatersrand, South Africa became the leading gold producer of the world. The most productive year was 1970 when over 1,000 tons of gold were mined, representing more than three-quarters of western world output.

The most famous gold rush started in 1896 in the Klondike Gold Rush. Of the estimated 100,000 men who set out for Alaska, only about 4,000 found any gold. However 75 tons of gold were produced over the next three years. It is estimated that about 375 tons of gold have been recovered by mining the rich alluvial deposits on the Klondike River and its tributary creeks, which include Bonanza, Eldorado, Last Chance, Bear and Gold Bottom creeks during the mining history of the area.

Elsewhere in Canada hard-rock gold fields were found in the early of the 20th century: Abitibi and Larder Lake in 1906, Porcupine in 1909, Swastika in 1910, Kirkland Lake in 1911, Matachewan in 1916, Rouyn in 1924, Red Lake in 1925, Yellowknife in 1933 etc.

The discovery of the Carlin deposit from Nevada in 1960 represented another crucial turning point in the history of the gold mining industry. Relatively low grade disseminated type deposits with very large ore reserves, became to be exploited in giant open-pits with impressive mining outputs (exceeding several millions, in some cases several tens of millions, of tones of ore a year).

Giant deposits were discovered between 1980-1990 in Papua-New Guinea (Lihir in 1983, Grasberg and Mishima in 1989 and Porgera in 1990). Giant deposits were also found in Peru (Yanacocha in 1993, Pierina in 1998 etc.) and Argentina (Bajo De La Alumbrera and Cerro Vanguardia in 1998, La Coipa in 1999 etc.).

It has been estimated that during the human history about 196,000 tons of gold have been mined, from which about 19.5% only in the last 20 years. This is not an unexpected figure as long as the world gold production rose from 1,200 tons a year to over 2,500 tones a year during this period. Such productions were allowed especially by developing some giant mining operations (especially in open pit). The output of the ten top gold producing mines represents about 18.5% of the world production.



During the history of gold mining several turning points took place in the gold metallurgy.

Concentration of gold by washing away the lighter river sands from alluvial sediments with water, leaving behind the dense gold particles which could then be further concentrated by melting was the first method for gold ore processing known by man. As the archeological artifacts suggest, the Egyptians knew sluicing technology by about 3500 BC. Egyptian wall relief's from 3100 BC shows gold in various stages of refining and mechanical working.

Later mining of lode or vein deposits required crushing prior to gold extraction. The artifacts found at Bir Umm Fawakhir suggest that the Egyptians used the stone made rotary mil by about 1300 BC.

The technique of amalgamation, allowing to improve the recovery of gold, was discovered by the Romans about 500 BC.

Further on, it took more than two millenniums till major changes in the gold metallurgy happened.

Miller's process of refining impure gold with chlorine gas, patented in 1867, and Wohlwill's Electro-refining process, introduced in 1878, made possible routinely to achieve higher purity than had been allowed by fire refining.

In 1886 the cyanidation process was patented, allowing recovering gold values that had escaped both gravity concentration and amalgamation. The first cyanidation plant from the world started to operate in 1889 at Marta Mine (New Zealand).

Froth flotation, discovered in 1903, allowed the processing of some more complex ores.

With the introduction of heap cyanide leaching in early 1970, in conjunction with high mining outputs rate, it became possible the economic processing of low-grade ores.

Some “unconventional” processing methods were tested in the last decades (thiourea and thiosulphate leaching, bioleaching etc.) but the application at the industrial scale is still very limited. Over 80% of the gold produced annually in the world is recovered by different methods of cyanidation.

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**GEODINAMICA MASIVELOR MIOCENE DE SARE
DIN AREALUL EXTRACARPATIC SI BAZINUL TRANSILVANIEI
SI EFECTELE INTERVENTIEI UMANE ASUPRA ACESTORA**

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Abstract

The Transylvanian Basin and the Outer-Carpathian area contain Miocene salt formations with numerous blocks of gem salt. Their presence is to be linked with two past distinct salt-generating phases: aquitanian and badenian. In the Transylvanian Basin the age of salt formations is exclusively badenian. The miocene evaporated salt formed in an intermittent lagoon environment and is punctuated with loess material. In the Outer-Carpathian area the aquitanian formations include also a potassium complex between Ozana valley and Putna valley. The often erratically placed blocks of salt are frequently affected by the phenomena of diapirism. This phenomena is related only with tectonic phases, in which, local conditions can be met that allow for a plastic flowing of salt. As a principle, there are no recognised cases of recently active diapirism. On the other hand, salt formations are subjected to a specific process of erosion, by dissolution. In turn, this process causes higher than average influx of minerals in hydrographic networks, and the apparition of a very large number of salty springs, aquatic levels or plots of land called “saraturi” (salt-marshes), lacking vegetation or with halophile vegetation. The dissolution is also responsible for a natural carst phenomena called ‘halocarst’ or saline carst with characteristic forms of endo and exocarst. The dry exploitation of salt, as well as that by underground dissolution with probes leads to non-typical carst formations, characterised as antropocarst. The abandonment of numerous old sites of salt exploitation (known from prehistoric times) had lead to uncontrolled dissolution of salt formations and to salt mines caving-in. Out of this, resulted a series of salt lakes that represent merely a dynamic (temporary) equilibrium that can be followed up by future mine-cavings. Amongst dramatic results of human intervention on salt formations must be mentioned, aside from recent collapsing at Ocnele Mari, the morphologic alterations of diapiric areas of Ocna Mures, the recent collapse at Gura Slanicului (Tg. Ocna) and Ocna Dej (south), as well as successive collapses in the salt mine of Slanic Prahova. Further, amongst areas with potentially unfavourable evolutions in the future must be mentioned, aside from the collapsing cone of Teica, the alignment Telega-Doftana-Campina, and the locations at Ocna Mures, Ocna Dej, Ocna Sibiului, Turda, Praida and Gura Slanicului (Tg. Ocna). In connection with these areas, detailed geodynamic studies would be necessary in the future. These studies should then include a complete history of the exploitation sites, as well as a preventive monitoring of potentially damaging evolutions within these sites, and especially in what concerns the surrounding and adjacent areas to these sites.

Keywords: sare (gema), masiv de sare, facies salifer, etapa salinogena, serie evaporitica, diapirism, solubilizare, izvoare sarate, saraturi, vegetatie halofila, halocarst, carst antroposalin, ocna, salina, Aquitanian, Badenian, Miocen



1. Introducere

In aceasta lucrare seincearca o sumara trecere in revista a problematicii specifice unui domeniu relativ particular al geologiei, legat de existenta pe teritoriul tarii a unor masive (corpuri) de sare cantonate stratigrafic in Miocen. Marea solubilitate a clorurii de sodiu si comportamentul plastic al sarii in conditii deosebite de stres sunt principalele cauze ce determina pe de o parte un mod specific (particular) de erodare (prin solubilizare), cu consecinte diverse, iar pe de alta parte structuri aparte, de tip diapir, foarte caracteristice. Interventia umana, sub diverse forme, asupra unor arii cu astfel de corpuri (masive) de sare a avut si are consecinte dintre cele mai variate, iar cunoasterea acestora se impune ca necesara mai ales datorita unor efecte dramatice asupra unor zone locuite.

Avand in vedere si disputele de natura stiintifica ce au vizat varsta formatiunilor cu sare din Romania si geneza zacamintelor de sare au fost facute unele precizari in acest sens in primele capitole ale acestui studiu. O atentie aparte trebuie de asemenea acordata fenomenului de solubilizare a corpurilor de sare, reflectat de numarul foarte mare de izvoare sarate cunoscute, respectiv carstului salin (endo si exosalin), atat natural cat si antropic, acesta din urma fiind corespunzator diferitelor moduri de extragere a sarii din zacaminte, fie pe cale uscata, fie prin solubilizare subterana cu sonde.

In fine, studiul inregistreaza si unele efecte dramatice datorate interventiei umane asupra zacamintelor de sare din arealul extracarpatic si din Bazinul Transilvaniei si are in vedere si unele arii cu posibile evolutii nefavorabile in viitorul apropiat, arii a caror studiere si monitorizare sunt inevitabile.

2. Varsta formatiunilor miocene cu sare din Romania

In aria carpatica formatiunile miocene cu sare pot fi raportate la doua etape salinogene distincte, prima revenind Miocenului bazal, respectiv Aquitanianului, iar cea de a doua fiind corespunzatoare Badenianului. La nivelul Burdigalianului in aria carpatica nu se dezvolta un facies salifer care sa fi ajuns pana la depunerea clorurii de sodiu (a halitului) astfel incat acesta este reprezentat doar prin gipsurile care se dezvolta la nivelul orizontului cenusiu al fostului "Helvetic" (Gipsurile de Perchiu si Gipsul de Stufu, in Subcarpati). De amintit aici ca in ultima perioada de timp, din considerente de natura paleontologica in primul rand, s-a manifestat tendinta, nejustificata, de a atribui Burdigalianului formatiunea salifera inferioara care este totusi aquitaniana.

Formatiunile miocene cu sare de varsta miocen inferioara (aquitania) se dezvolta doar in arealul extracarpatic in timp ce depozitele badeniene cu sare sunt prezente atat in Subcarpati cat si in Bazinul Transilvaniei.

In aria subcarpatica faciesul salifer al Aquitanianului se dispune peste depozitele detritice placate la nivelul Stratelor de Gura Soimului sau a echivalentelor acestora si include in baza "gipsurile inferioare" (numite local Gipsuri de Sarata sau Gipsul de Feschi) si se incheie cu nivelul Gipsului de Varnita. Peste "gipsurile inferioare" (miocene) se dispun masivele de sare aquitaniene care au la partea lor superioara, intre valea Ozanei si valea Putnei, un complex potasifer ce cuprinde argile sarate cu lentile de saruri de potasiu langbeinit-kainitice. Acestea suporta argile sarate peste care se dispune nivelul gipsului de Varnita, urmat la randul sau de argile dispuse sub depozitele burdigaliene in facies predominant grezos. Practic, totalitatea depozitelor corespunzatoare faciesului salifer al Aquitanianului, amintite mai sus, pot fi raportate "Formatiunii salifere inferioare", de varsta miocen inferioara (aquitania).

In fine, faciesul salifer al Badenianului, cu termenul halitic bine reprezentat, debuteaza cu un slab nivel gipsifer (sau dolomitic) plasat sub nivelul tufurilor dacitice badeniene, nivel ce suporta un



pachet de marne. Peste tufuri (Tuful de Dej) se dispune “formatiunea salifera superioara” numita si “Formatiunea evaporitica” ce include, mai ales in Subcarpati, pe langa masivele badeniene de sare si “brecia sarii”. In fine, faciesul salifer al Badenianului se incheie cu un nou nivel gipsifer, superior “Formatiunii evaporitice”, plasat sub sisturile cu radiolari.

3. Geneza zacamintelor miocene de sare

Sarea miocen inferioara (aquitana) si sarea badeniana sunt legate de existenta la aceste nivele stratigrafice a unor depozite evaporitice lagunare de climat arid ce reflecta momentele cu temperatura cea mai ridicata a unor faze calde ale unui ciclu climatic cu perioada de 4,1 Ma, numit “ciclul valah” (Ticleanu M. si al., Praga, 1998). Deci fazele calde ale acestui ciclu au determinat, direct, etapele salinogene cunoscute ale Miocenului, respectiv aquitaniana, burdigaliana si badeniana.

Prezenta sarurilor de potasiu la nivelul etapei salinogene aquitaniene reflecta o faza calda a ciclului valah cu temperaturi medii anuale mai ridicate decat ale fazei calde badeniene, deoarece aceste saruri apar doar in situatia derularii complete a seriei evaporitice, derulare ce depinde in primul rand de temperatura.

Momentele mai reci, respectiv mai umede, ale acestor faze calde sunt markate de prezenta unor nivele cu grosimi mai mici de argile sau marne (frecvent impregnate cu sare) astfel incat masivele (corpurile) de sare, reprezentand de regula lentile cu grosimi de cateva sute de metri, se prezinta cu o textura rubanata tipica, care reprezinta stratificatia primara a respectivelor depozite lagunare evaporitice.

Etapa salinogena a Burdigalianului, ce a dus la aparitia gipsurilor “formatiunii cenusii” “helvetiene”, corespunde unei faze calde ce nu a permis aparitia termenului halitic, deci cu atat mai putin a termenilor finali ai seriei evaporitice. Aceasta nu caracterizeaza nici etapa salinogena badeniana, desi rareori se poate banui ca strict local si pentru perioade foarte scurte de timp au fost totusi intrunite conditiile pentru depunerea acestor saruri, atat in aria subcarpatica, dar si in Bazinul Transilvaniei.

4. Diapirismul sarii – structuri halocinetice si halotectonice

Avand in vedere geodinamica formatiunilor miocene cu sare putem distinge foarte clar doua directii distincte de dezvoltare a acestei problematici: prima are in vedere structurile (tectonice) specifice, proprii formatiunilor cu sare, numite, impropriu (Trusheim F., 1960), “structuri halogene”, iar cea de a doua vizeaza evolutia hidrodinamica a acestor formatiuni in cadrul edificiului structural al ariei subcarpatice sau al Bazinului Transilvaniei.

Structurile “halogene” ar include structuri halocinetice si structuri halotectonice. Primele s-ar datora miscarii sarii sub impulsul fortelor isostatice, iar structurile halotectonice s-ar datora fortelor tectonice, in special a celor tangentiale, in prezenta sarii. Aceasta structurare a problematicii privitoare la structurile “halogene” acorda de fapt un rol secundar sarii (respectiv masivelor de sare) in amorsarea si dezvoltarea acestor structuri de tip special. Clasic, conceptul de diapirism presupune o actiune dinamica a sarii cu totul deosebita, bazata pe curgerea plastica a acesteia, in conditiile in care limita de elasticitate la forfecare este depasita, concept ce a condus si la sintagma “zona cutelor diapire”. Fata de aceasta imagine curenta, la care s-a ajuns cu timpul prin exagerarea rolului pe care l-a avut sarea in realizarea unor astfel de edificii structurale, ne placem in situatia de a considera ca de fapt sarea a jucat un rol secundar si ca nasterea unor structuri specifice s-a datorat in primul rand fortelor tectonice tangentiale (orizontale) care s-au manifesta in cursul fazelor tectonice ulterioare depunerii formatiunilor miocene cu sare. Practic fenomenele propriu-zise de diapirism s-au



manifestat doar strict legat de fazele tectonice in timpul carora local si pentru scurt timp au fost atinse conditiile de presiune si temperatura care au condus la curgerea (deplasarea) plastica a sarii.

De notat in acest sens ca sarea miocen inferioara este intalnita frecvent sub forma de lame sau stalpi diapir si foarte rar sub forma de perna, pe cand sarea mai noua, badeniana, apare mai ales sub forma de perna sau de strat si mai rar sub forma de lame sau stalpi diapiri. In plus, privitor la fenomenele de diapirism activ (recent), ne exprimam, cel putin pentru moment, rezerva.

De amintit aici si tipurile de structuri “halogene” cunoscute la nivelul formatiunilor miocene cu sare din Romania, respectiv tip “strat”, “perna” (halocinetica), stulp diapir, lama diapira, lama (lacrima) de rabotaj (halotectonice).

5. Erodarea specifica a masivelor de sare – solubilizarea

Interactiunea dintre formatiunile miocene cu sare si apa, de natura hidrodinamica, reprezinta latura cea mai importanta si mai interesanta a evolutiei dinamice a acestor formatiuni. Aceasta se datoreaza marii solubilitati a clorurii de sodiu in apele dulci precum si capacitatii retelelor hidrografice de a transporta mari cantitati de saruri in solutie.

Intensul proces de dizolvare a sarii este reflectat in mod deosebit de evident de numarul foarte mare de izvoare sarate ce apar in intregul areal extracarpatic, dar si in Bazinul Transilvaniei.

Daca apele meteorice sau ale raurilor isi fac drum in interiorul masivelor de sare incep sa se dezvolte forme carstice (halocarstice), cursurile subterane fiind tradate de dezvoltarea la suprafata a unor forme exocarstice (doline, avene, uvale, polii). Se dezvolta astfel pesteri in sare care au toate caracteristicile carstului tipic (clasic) dezvoltat pe roci calcaroase. Fenomenele halocarstice naturale pot conduce la fenomene de prabusire care se materializeaza prin aparitia la suprafata terenului a unor conuri de surpare sau a unor depresiuni alungite ce reflecta prin pozitia lor directia de dezvoltare a cursurilor subterane. In arealul extracarpatic pot fi amintite aici ca exemple aliniamentul cu doline de prabusire (nord) Broaste-Lacul Biserici (Campina), aria cu doline din zona Podu Vadului-Valea Campinita si, eventual, zona vaili Mordana (Catina) de pe Bisca Chiojdului.

Acoperisul masivelor de sare, chiar dezvoltate la zi, poate fi cu timpul protejat de o asa numita “manta a sarii”, formata din material insolubil (argilos) ce poate contine si elemente de roci mai dure. Uneori aceste depozite reziduale au grosimi mari si asigura o buna protectie astfel incat pe suprafata lor pot aparea lacuri cu apa dulce.

Eroziunea in aria formatiunilor miocene cu sare este deseori accelerata de faptul ca vegetatia nu se poate mentine pe suprafata acestora datorita cantitatilor mari de clorura de sodiu, realitate ce conduce la dezvoltarea unor zone lipsite de vegetatie, numite “saraturi”, tradate si de prezenta unor specii vegetale halofile. In lipsa vegetatiei apele meteorice spala cu mult mai usor aceste depozite, fapt ce poate conduce la aparitia unor fenomene de alunecare. In situatia in care argilele reziduale au o anumita grosime este posibila si instalarea unui covor vegetal protector ce duce la atenuarea fenomenelor de dizolvare.

5a. Izvoarele sarate (sursele clorosodice naturale)

In arealul extracarpatic si in cuprinsul Bazinului Transilvaniei sunt cunoscute numeroase izvoare sarate cu debite si concentratii foarte variate. Multe din aceste surse sarate ating concentratii foarte mari (peste 300g NaCl/l) si sunt deseori amenajate de localnici care utilizeaza in mod frecvent in alimentatie apa acestora. Debitul acestor izvoare este foarte variabil de la sursa la sursa, dar si de la un anotimp la altul. Frecvent debitul multor izvoare se plaseaza in jurul valorii de 1l/minut, dar uneori se poate ajunge la cca 5-10l/minut in arile in care corpurile de sare sunt



afectate de un sistem carstic activ.

In arealul extracarpatic aria in care apar izvoare sarate incepe in nord de la valea Sucevei si se continua fara intreruperi semnificative pana in zona vailor Ialomitei, in zona localitatii Glodeni, Laculete, Vilcana Bai. O alta zona in care apar izvoare sarate este legata de zacamintele de sare din Oltenia, incepand de la Ocnele Mari si pana in zona anticlinalului Magura Slatioarei. Manifestari saline nesemnificative apar si in zona diapirului Slatioarele de langa Pitesti.

In Bazinul Transilvaniei cele mai importante surse clorosodice naturale sunt cunoscute mai ales in partea de nord a acestuia, dar si pe rama estica a bazinei, pana in zona Homoroadelor. Multe din aceste izvoare sarate sunt foarte bine amenajate de locuitorii satelor din apropiere in special cu casute de lemn, iar alimentarea cu apa sarata se face uneori sub controlul autoritatilor locale (la Sinmarghita, Mintiu, Taure, Mercheasa, Rupea, Lueta, Meresti, Corund, Petreni, Martinis).

Prezenta izvoarelor sarate este in mod curent marcata de gradul ridicat de mineralizare al paraielor dinspre aval, de eflorescente saline albe ce apar in apropierea lor in perioadele calde ale anului, de disparitia vegetatiei in zonele imediat invecinate, de aparitia unor plante halofile (specii de Salicornia, Aster, Statice, Artemisia etc), de spumele albe ce apar in aval de punctele de emergenta sau de-a lungul raurilor incarcate cu clorura de sodiu, sau de lipsa inghetului in perioadele cu temperaturi foarte scazute din timpul iernii.

Izvoarele sarate sau baltile sarate sunt cunoscute de localnici mai ales in Moldova si in partea de nord a Transilvaniei sub numele de "slatine". In rest, sunt numite in Vrancea "saramuri" (cu varianta "salamuri") sau, mai rar (Muntenia) "moare" cu varianta "muratoare", cunoscuta ici-colo si in Ardeal.

5b. "Poluarea salina naturala"

Dizolvarea clorurii de sodiu a masivelor de sare sau a depozitelor ce contin sare sub forma de ciment sau sub forma de impregnare conduce inevitabil la cresterea gradului de mineralizare a raurilor si paraielor ce strabat aria de dezvoltare a formatiunilor miocene cu sare. Practic procesul de mineralizare are ca punct de plecare numeroasele izvoare sarate existente, la care se adauga spalarea directa a corpurilor de sare aflate la suprafata, fie de catre apele meteorice, fie de catre apele retelelor hidrografice. La elementele primare ale acestor retele, in apropiere de izvoare, se pot atinge concentratii mari, cuprinse intre 10 si 50-60g/l NaCl. Colectarea lor in rauri cu debite mai mari conduce la ridicarea substantuala a gradului de mineralizare a acestora astfel incat se ajunge ca in regiuni foarte joase, situate pe cursul inferior al acestora, sa se atinga valori de 1-5g/l continut de sare, ceea ce reprezinta destul de mult pentru rauri cu debite mari.

Fenomenul de mineralizare a apelor ce vin in contact cu depozitele cu sare, precum si a apelor subterane si a solurilor, numit impropriu "poluare salina naturala", reprezinta cauza principală a acumularii de sariuri într-o zonă extinsă situată în partea de nord-est a Campiei Române, zonă ce cuprinde numeroase sariuri, marcate și de o vegetație halofila, dar și câteva lacuri sariate. Aici se adună cursurile mai multor rauri caracterizate prin diferite grade de mineralizare între acestea remarcându-se în mod special Râmnicul Sarat, Slanicul de Buzău, Ialomita, Putna, Ramna și Siretul. Între afluentii acestor rauri se detasează Cricovul Sarat, Tazlaul Sarat, Oituzul, Slanicul (ca affluent al Trotusului) și Sarata (de Ialomita).

In aria cu sariuri amintita mai sus se remarcă, in afara mineralizarii retelei hidrografice, prezenta a numeroase lacuri sariate, a diferitelor tipuri de soluri sariate, precum si a unor ape subterane (freatice sau de adancime) mineralizate. Aceasta poate conduce deseori la o accentuare a gradului de mineralizare a solurilor din aceasta zonă, pe cale antropica, respectiv prin utilizarea unor



ape mineralizate la irigarea culturilor, fenomen cunoscut sub numele de “salinizare secundara a solurilor”.

In Bazinul Transilvaniei se remarcă salinitatea ridicată a Tarnavei Mici (datorită spalarii masivelor de sare de la Corund, Praid și Sovata), a Somesului și a Muresului. După Stoica și Gherasie (1981) cantitatile de sare dizolvate și transportate de raurile Tarnava Mica (100.000t/an), Slanicul de Buzau (cca. 500.000t/an) și Râmnicele Sarat (cca. 400.000t/an) este de cca. 1 milion de tone de sare pe an.

6. Carstul natural endo si exosalin

In zonele cu larga dezvoltare a masivelor de sare aquitaniene și badeniene din țara sunt practic nelipsite fenomenele exocarstice foarte cunoscute, fapt ce tradează dezvoltarea unor forme endocarstice în interiorul acestor masive. Prezența lapiezurilor pe sare, a dolinelor, avenelor, uvalelor și poliilor nu este de loc surprinzătoare mai ales în cazul corpurilor de sare de dimensiuni mari. Acolo unde a fost posibil au fost cercetate și unele cavități dezvoltate în interiorul unor masive de sare, fapt ce a scos în evidență dezvoltarea tuturor formelor endocarstice tipice peșterilor din zonele calcaroase. Au fost întâlnite astfel stalactite de sare, anemolite, stalagmite, curgeri parietale, conopide, puf de sare și cruste de sare.

Cercetarea unor peșteri în sare a pus în evidență de asemenea parcurgerea tuturor stadiilor de evoluție carstică, de la trasee subterane sub presiune până la trasee subterane cu nivel liber, cu cavități mari. Existenta uvalelor, poliilor, dolinelor și arenelor naturale la suprafața terenului, deasupra masivelor de sare, reflectă și atingerea unor stadii mai avansate de evoluție carstică în corpurile de sare, corespunzătoare unor prabusiri de-a lungul traseelor unor goluri. Uneori linia unor astfel de zone de prabusire marchează practic la suprafața directia generală de dezvoltare a carstului subteran.

In Subcarpați pot fi amintite aici zonele carstice pe sare Stupina-Fetele Targului (Tg. Ocna), cele din zona zacamantului N Valea Sării (Vrancea), de la Minza-Lopatari (platoul halocarstic Meledic), cele din bazinul superior al văii Jgheabului (în apropiere de Bisoca), zona Sarile-Bisoca, cele din bazinul văii Matita, de la Slanic Prahova, Slatioarele (Pitesti) și cele de la Ocnele Mari (Valcea). Între acestea se evidențiază net aria halocarstică din bazinul Slanicului de Buzau (zona carstică Meledic), arie în care a fost explorată cea mai lungă peșteră în sare din lume (peștera 6S), cu o lungime de 3.234m. Platoul carstului sălin de aici cuprinde numeroase doline și uvale ce adăpostesc lacuri cu apă dulce. În anul 1985 erau înregistrate aici 26 de peșteri în sare, dintre care 16 active (date după Ica Giurgiu).

Între localitățile Jgheab și Plavat sunt cunoscute 15 cavități în sare, nu prea dezvoltate, asociate cu avene și doline. În Sarile (Bisoca) este foarte greu de disociat carstul natural de cel antropic. În zona zacamintelor de sare de la Muchia Ocnei, Slăvu și Valea Dulce sunt cunoscute mai multe forme exohalocarstice între care un sorb pe valea Zimbroaia, o polie la Slăvu, un sir de doline la nord de satul Valea Dulce. Se banuieste tot aici ca zacamantul de sare Surani Vest este afectat de un sistem endocarstic.

La Slanic Prahova Draganescu (1995) evidenția pe Valea Verde lapiezuri, doline de prabusire, ponoare și peșteri în sare. Tot aici, la Baia Baciu, V. Sencu (1968) semnală și forme naturale pe sare, alături de cele antropic. La nord de Valea Sării (Vrancea), pe Paraul Sarat, există un poron (sorb) insotit mai spre aval (1,3 km) de un izbuc cu ape sărate. În schimb carstul pe sare de la Tg. Ocna pare a fi mai de graba amorsat pe cale antropică.

In Bazinul Transilvaniei sunt bine cunoscute fenomenele carstice pe sare din aria zacamintelor de la Sovata și Praid, alături de care pot fi amintite forme exosaline la Jabelnița și pe Valea Patei (la SE de Cluj). Acestea li se pot adăuga pseudopolile din aria zacamintelor de sare de la Mercheașa, Uila, Brancovenesti, Blajenii de Jos, Gadalin, Dezmir, Cojocna, Sic și Ocnisoara. La



Chiuza chiar poate fi citata o prabusire (surpare) datand din anul 1959, localizata intr-o arie cu relief depresionar tipic. De asemenea intre Sic si Gherla, langa Sacalaia se afla un lac ca pare sa fi rezultat in urma unei vechi prabusiri aparute in aria unui corp de sare.

7. Carstul antroposalin

Exploatarea masivelor de sare, indiferent de metoda folosita, conduce inevitabil la crearea unor mari goluri subterane, respectiv a unui sistem de goluri de dimensiuni foarte variate si de forme foarte diverse. Practic toate activitatile umane de exploatare a zacamintelor (masivelor) de sare reprezinta tot o forma de eroziune atipica, soldata cu dezvoltarea unui sistem endocarstic, de natura antropica, ce poate fi numit carst antroposalin sau halocarst antropic. Acesta poate fi de natura exocaristica (exohalocarst antropic) sau de natura endocaristica (endohalocarst antropic).

In general exploatarile vechi sau foarte vechi erau reprezentate in mod obisnuit prin palnii in corpul de sare, corespunzatoare unor forme de tip exocarstic. Exploatarile vechi, ocne de tip clopot, dezvoltate cu timpul ca goluri subterane de tip endocarstic, erau legate de suprafata prin puturi. Surparea acestor puturi dupa abandonarea ocnelor a condus frecvent la aparitia unor importante forme halocarstice (de tip exocarstic) care s-au umplut apoi cu apa sarata. In alte cazuri, surparea puturilor, fara prabusirea cerului de bolta la ocnelor a dus la formarea unor avene ce reprezinta receptorii unui sistem carstic dezvoltat in interiorul corpului de sare asa cum se intampla in mod obisnuit in apropiere de Tg. Ocna.

Metodele moderne de exploatare pe cale uscata a sarii au condus la aparitia unor sisteme carstice antroposaline a caror evolutie este diferita in functie de metoda utilizata, de conditiile hidrogeologice specifice fiecarui zacamant in parte, de gradul de carstificare naturala anterior exploatarii etc. Intre aceste metode se numara exploatarea cu camere lungi ogivale si trapezoidale, cu camere mici si pilieri dreptunghiulari sau cu camere mici si pilieri patrati. De amintit aici si varianta exploatarii sarii in cariere care a condus in mod obisnuit la aparitia unor forme negative de relief. In fine, exploatarea prin solubilizarea subterana cu sonde conduce tot la aparitia unor goluri subterane in masivele de sare, sistemul endocarstic rezultat fiind mult diferit de cel corespunzator metodelor de exploatare amintite mai sus.

Oricum insa, indiferent de natura sa genetica particulara, carstul antroposalin si in special cel endocarstic conduce prin simpla sa prezenta la schimbarea echilibrului dinamic initial al corpului de sare in care este dezvoltat, fapt ce poate sa conduca si la pericolarea stabilitatii terenului situat in acoperisul sistemului carstic respectiv. Evident, principalul rol destabilizator este jucat de apele de infiltratie, dulci, ce maresc prin dizolvare goulurile initiale si a caror activitate este lipsita de orice control in cazul vechilor lucrari miniere complet abandonate.

8. Efectele naturale ale prezentei masivelor de sare in cuprinsul depozitelor mio-pliocene

Inainte de a evalua efectele interventiei umane asupra ariei cu masive (corpuri) de sare se impune ca foarte necesara cunoasterea tuturor efectelor naturale ale prezentei acestor masive in aria de dezvoltare a depozitelor mio-pliocene din arealul extracarpatic si din Bazinul Transilvaniei. Fara aceasta cunoastere si fara cercetarea detaliata a istoricului exploatarilor ce au afectat un zacamant sau altul de sare nu este posibila aprecierea realista si corecta a consecintelor interventiei umane asupra corpurilor (masivelor) de sare.

Printre cele mai importante efecte naturale datorate simplei prezente a corpurilor de sare, la zi sau in adancime, se numara: mineralizarea retelelor hidrografice ce strabat ariile cu formatiuni miocene cu sare sau ce isi au originea in aceste arii, mineralizarea complexelor acvifere ale depozitelor miocene si pliocene, formarea unor saraturi cu eflorescente saline lipsite de vegetatie



sau cu vegetatie halofila si dezvoltarea unui relief depresionar, respectiv a unor forme de carst exosalin, in aria unor masive (corpuri, zacaminte) de sare.

Nu vom insista in acest studiu asupra acestor efecte naturale, mai ales ca in capitolele anterioare au fost deja atinse anumite aspecte privitoare la acest subiect. Mentionam aici faptul ca deseori este foarte dificil a se stabili daca in anumite zone unele realitatii (in special forme de relief negative) sunt generate pe cale absolut naturala sau la originea lor se afla foarte vechi interventii umane asupra corpurilor de sare. In orice caz insa, interventia humana asupra arillor markate de prezenta unor masive de sare se concretizeaza, fara exceptie, prin accelerarea fenomenelor naturale de erodare specifica a masivelor respective, fapt ce poate conduce deseori la consecinte nedorite iar uneori chiar dramatice.

Cunoasterea efectelor naturale mentionate mai sus poate conduce la interventii umane care sa duca la limitarea efectelor negative ale prezentei corpurilor de sare. Intre acestea putem aminti aici diminuarea fenomenului de “poluare salina naturala” prin exploatarea, in scopuri alimentare, din timpuri imemoriale, a izvoarelor sarate, atat in Bazinul Transilvaniei cat si in arealul extracarpatic. In acest sens putem afirma ca daca s-ar fi trecut la exploatarea sistematica, respectiv industriala, a acestor surse clorosodice s-ar fi ajuns la o limitare drastica a cresterii gradului de mineralizare a unor retele hidrografice, fapt ce ar fi atras dupa sine o reducere a suprafetelor ocupate de saraturi, respectiv la o reducere a gradului de mineralizare a complexelor acvifere “poluate” cu clorura de sodiu, atat in partea de NE a Campiei Romane cat si in unele zone din Transilvania.

Tot aici ar putea fi mentionat si un alt efect pozitiv al interventiei umane reprezentat de valorificarea terapeutica si turistica a apelor sarate, a namilorilor saraturilor sau a spatiilor vechilor saline (ocne) parazite, bine ilustrat atat in Ardeal cat si in Subcarpati.

9. Efecte dramatice ale interventiei umane asupra unor masive de sare.

In unele situatii interventia humana asupra arillor de dezvoltare a unor masive de sare se concretizeaza prin aparitia unor zone de prabusiri datorate unei mari dezvoltari a golurilor artificiale in subteran. Aceste prabusiri sunt legate de exploatari vechi sau foarte vechi, fie de exploatari mai noi, dar intensive sau hiperintensive.

In cazurile in care in prabusiri sunt antrenate zone locuite sau cai de comunicatie se poate ajunge la efecte absolut dramatice, mai ales in situatia in care nu pot fi prevazute in timp util. Din pacate insa, in ultimele zeci de ani, in aria unor exploatari de sare au aparut zone tipice de prabusire, markate de dezvoltarea in teren a unor conuri de surpare cu diametre de ordinul sutelor de metri, in care au fost antrenate uneori si zone locuite.

Zona vechilor exploatari de sare incepe in arealul extracarpatic cu zacamintele de la Tg. Ocna si Grozesti si se incheie in apropiere de valea Ialomitei, in zona zacamintelor de la Ochiuri si Ocnita. Zacamantul de la Ocnele Mari se afla astfel izolat mult spre vest. Pot fi enumerate aici exploatările vechi de la Grozesti, de la nord de Tg. Ocna, de la Valea Sarrii (la nord si sud de raul Putna), de la Reghiu si Andreiasu, de la Sarile-Bisoca, Rusavat (Sarea lui Buzau), Aricesti, Ghitioara (Aninis), Sararu, Teisani, Slanic Prahova, Telega, Baicoi, Ochiuri (Gura Ocnitei). In plus se cunosc arii mai restranse cu exploatari locale de mica ampoare in zona Vrancei, la est de valea Buzaului si la Predeal Sarari. In toate aceste zone au existat mai multe tipuri de organizare a exploatarii deosebindu-se ocne satesti, boieresti si domnesti, carora li s-au adaugat in toate timpurile si mici exploatari locale neautorizate, numite pe alocuri “titoace”

In apropiere de Tg. Ocna au functionat in trecut multe ocne distincte in forma de clopot, abandonate treptat. Lacul de prabusire de la Gura Slanicului s-a format prin surparea tavanului unei



astfel de ocne abandonate cu cca. 5 secole in urma. Lacul a aparut in anul 1978, in cateva zile formandu-se un lac sarat de cca 10.000m patrati si cu o adancime de 57m. In prabusire au fost antrenate cateva locuinte, iar caile de comunicatie din zona au fost afectate. Noi surpari au dus la extinderea lacului si la scaderea adancimii sale, dar si la abandonarea unor obiective din vecinatate. La Valea Sarii (nord) exploatarea s-a facut cu galerii si a fost insotita de accidente datorate surparii acestora. La Aricesti, in anul 1897, vechile ocne de sare erau deja surpate. Mina Doftana, inchisa intre anii 1900-1901 era deja inundata in anul 1911. La 1890 in aceasta mina a avut loc prabusirea unui bloc enorm de sare. Din anul 1894 au aparut grave probleme hidrogeologice ce au condus la abandonarea exploatarii.

La Slanic Prahova mina “Sistematica” a fost inchisa in anul 1875 datorita infiltratiilor de ape, fapt ce a condus la o serie de prabusiri ce s-a incheiat cu surparea in 1901 a ultimului put de aici. La mina Victoria, deschisa in 1971, a avut loc, in anul 1992, prabusirea etajelor III, IV si V din cele zece existente. Tot aici pot fi notate si prabusirile legate de exploatarea in cariera a Muntelui de Sare, la Grotă Miresei, prabusiri succesive ce au dus la aparitia lacurilor din zona.

In aria zacamantului de sare Podu Vadului, la NW de Campina, a aparut, la inceputul deceniului al 9-lea al secolului trecut, un con de surpare cu un diametru mai mic de 50m care a afectat in parte vechia sosea Campina-Breaza.

In apropiere de Ocnele Mari, la Teica, in septembrie 2001 s-a format prin prabusire un con de surpare cu axa de cca. 300m si care a antrenat si un numar insemnat de gospodarii. Spre sfarsitul anului 2004 alte prabusiri au dus la cresterea in dimensiuni a conului. Aparitia acestuia este strict legata de exploatarea hiperintesiva cu sonde a partii vestice a zacamantului de la Ocnele Mari, respectiv de existenta in aceasta zona a unui important gol subteran (golul “Socon”).

Ariile cele mai afectate de interventia umana asupra zacamintelor de sare din cuprinsul Bazinului Transilvaniei sunt legate de exploatarile de sare vechi si actuale de la Ocna Mures, Ocna Dej, Ocna Sibiului, Turda si Sovata, carora li se adauga cateva informatii privitoare la vechile exploatari de la Sic si de la Albestii Bistritei.

In 1913 apele Muresului inundau galeriile exploatarii de la Ocna Mures, atacau pilierii de sustinere si duceau la aparitia la suprafata a unor uriase palnii (conuri) de surpare, fiind antrenate in acestea si zone locuite ale localitatii Uioara. Exploatarea defectuoasa inceputa aici in 1921 (dizolvare statica) a condus la abandonarea minelor, reprofilate ulterior pentru exploatarea de sare bulgari.

Inundarea in 1978 a minei 1 Mai a intrerupt definitiv in aria acestui diapir exploatarea pe cale uscata. S-a continuat astfel aici doar exploatarea prin solubilizare subterana cu sonde, inceputa in anul 1952, dar care deja duse la abandonarea minei Nicolae, tip de exploatare ce poate avea consecinte neprevazute pentru aria diapirului de sare de la Ocna Mures.

La Ocna Dej, aria nordica a zacamantului, afectata de alunecari de teren, este marcata de prezenta unor lacuri sarate rezultate din prabusirea vechilor lucrari miniere. Avand in vedere intregul corp de sare de la Ocna Dej putem aminti aici surparea minei Stefan, aparitia lacului Mina Mare, aria de surpare a minei Iosif si zona de surpare a minei Ciciri. In ianuarie 1988 in zona minelor Ciciri-23 August s-a format un important con de surpare, destul de activ.

La Ocna Sibiului aria cea mai afectata de exploatare este legata de zona lacurilor Horea, Closca si Crisan care au aparut prin inundarea a sase vechi ocne. In 1890 a avut loc inundarea ocnei “Fodina Maior”, abandonata in 1817. Fenomene de instabilitate au fost semnalate in zona lacurilor Randunica (Sf. Ioan) si Brancoveanu si au dus la prabusirea unui put al salinei Ignatiu si la formarea lacului Gura Minei. Amenintarea apelor de infiltratie ducea la 1770 si 1775 la abandonarea ocnelor “Josef” si “Francisc”.



La Turda prabusirile vechilor ocne ce se aflau la obarsia Vaii Sarata au condus la formarea actualelor lacuri din aria numita in trecut Ocnele Turzii (lacurile Durgau, Carolina, Sulfuros, Rotund si Lacul Ocnei). In zona Bailor Turda vechile exploatari sunt marcate de mai multe lacuri intre care se remarcă Lacul Tarzan si Lacul fara Fund.

Fenomenele de instabilitate in aria zacamantului de la Sovata (intre care si alunecari recente) isi au sursa in vechile exploatari, datand inca din secolele II-III ale erei crestine. Formarea, prin prabusire, a Lacului Ursu, in 1878, nu poate reprezenta astfel doar un simplu fenomen halocarstic natural, ci un efect tardiv al unei interventii umane asupra diapirului de la Sovata.

In Prajd inundarea Ocnei Noi, deschisa in 1809, a dus la abandonarea acesteia in doar doi ani de la inaugurare. In 1939 incercarile de exploatare prin dinamitare, in mina Elisabeta, au dus la prabusirea tavanului camerei de exploatare si la aparitia unei mari doline la suprafata. In apropierea acestei localitati este semnalata si existenta unui sat scufundat, Beizid, insa in prezent nu dispunem de informatii suplimentare concrete privind acest fenomen. De notat aici si datele existente privitoare la unele accidente de munca, cu urmari dramatice, privitoare la exploatarile de la Sic si de la Albestii Bistritei.

10. Arii cu posibile evolutii nefavorabile in viitor

Dintre toate situatiile mentionate la capitolul anterior, probleme deosebite pentru viitorul apropiat sau mai departe pot aparea la Ocnele Mari, pe aliniamentul Telega-Campina, la Podul Vadului, la Ocna Dej, Ocna Mures, urmate mai apoi de aria zacamintelor Ocna Sibiului, Gura Slanicului, Slanic Prahova si Turda.

La Ocnele Mari (Valcea) evolutia conului de surpare este si in prezent absolut ingrijoratoare, iar prabusirea controlata (daca se va realiza) a tavanului marelui gol subteran "Socon" nu ar fi decat inceputul unor schimbari morfologice drastice in aceasta zona si cu urmari de lunga durata si greu de estimat pentru intreaga arie a zacamantului.

Pe aliniamentul Telega-Doftana-Campina aria cea mai activa este legata de zona fostei mine de la Doftana, dar posibile fenomene de instabilitate pot afecta traseul carstului salin in zona orasului Campina sau in aria fostelor exploatari de la Telega. Lacul de la Doftana are tendinta de a se extinde, prin surpari lente, spre nord-vest, punand in pericol o zona locuita, dar in viitor poate afecta soseaua Campina-Telega, dar si pe cea spre Brebu. La Telega este pusa in pericol zona locuita din apropierea ultimului lac (dinspre aval) din versantul stang al vallii Telega.

Evolutii neasteptate pot surveni si in aria zacamantului de sare de la Podu Vadului, afectat de un endocarst salin practic necunoscut, fapt ce ar periclitata importantele cai de comunicatie ce leaga orasele Ploesti si Brasov. In aria zacamantului de la Ocna Mures, practic compromisa, exploatarea intesiva cu sonde poate avea de asemenea urmari neasteptate. In principal insa se prefigureaza o extindere a lacului format prin prabusirile din zona vechilor ocne de sare.

La Ocna Dej se poate presupune extinderea fenomenelor de instabilitate in partea sudica a zacamantului, afectata si de dizolvari naturale.

In aria diapirului de la Ocna Sibiului dizolvarea necontrolata din zona tuturor lucrarilor vechi de exploatare va conduce la extinderea suprafetelor lacurilor sarate, in special prin unirea mai multor astfel de bazine, cea mai expusa modificarilor fiind zona lacurilor Horia, Closca si Crisan dar si zona lacului subteran al statiunii.

La Turda fenomenele de instabilitate (mai ales alunecari) vor afecta in primul rand zona locuita a orasului aflata la obarsia Vaii Sarate, in amont de zona lacurilor formate prin prabusirea vechilor ocne.



Evolutii neasteptate pot insa marca si aria zacamantului de la Tg. Ocna, mai ales in zonele in care exploatarea cu sone (Gura Slanicului) va conduce la modificari inevitabile ale starii de echilibru dinamic a zacamantului de sare.

11. Concluzii

In arealul extracarpatic si in Bazinul Transilvaniei se dezvolta la zi sau in adancime formatiuni miocene, aquitaniene si badeniene, cu sare ce contin numeroase masive (corpuri) distincte de sare gema (halit). In Bazinul Transilvaniei varsta acestor formatiuni este exclusiv badeniana. Prezenta acestora este legata de derularea a doua etape salinogene distincte, una miocen inferioara (aquitaniana) si alta miocen medie (badeniana), corespunzatoare unor faze calde ale unui ciclu climatic cu perioada de 4,1 Ma (ciclul climatic valah).

Sarea formatiunilor miocene, de natura evaporitica, s-a depus intr-un mediu lagunar intermitent, alternantele de sare alba, curata, si sare impurificata cu minerale argiloase reflectand o succesiune de faze aride si faze umede ale unor cicluri cu perioade mici de timp (nu mai mici decat cele anuale).

Formatiunile aquitaniene cuprind, intre valea Ozanei si valea Putnei, un complex de saruri de K si Mg, fapt ce reflecta derularea completa a seriei evaporitice in etapa salinogena a Miocenului inferior. Deseori masivele (corpurile) de sare se dispun discordant in raport cu formatiunile din jur, uneori cu mult mai noi, datorita fenomenului de diapirism bazat pe mobilizarea plastica a sarii. Se disting insa structuri halocinetice si structuri halotectonice, la ultimul tip rolul principal revenind miscarilor tectonice, iar nu sarii. In opinia noastra formarea structurilor tipic diapire este strict legata doar de fazele tectonice, cand, local si pentru scurt timp, se ating conditiile de temperatura si presiune necesare curgerii plastice a sarii. Din acest motiv consideram ca este putin probabila existenta in prezent a unor cazuri de diapirism activ (recent), poate doar cu exceptia structurii de la Praida.

In schimb aceste masive de sare sunt supuse in permanenta unui proces specific de eroziune, prin solubilizare, fapt reflectat de existenta unui numar foarte mare de izvoare sarate, cu debite si sarcini minerale foarte variate, atat in Bazinul Transilvaniei cat si in arealul extracarpatic. Drept urmare se constata o mineralizare a retelelor hidrografice ce strabat formatiunile cu sare, fapt ce are drept consecinta mineralizarea, inspre aval, a acviferelor si a solurilor, ajungandu-se pana la saraturi fara vegetatie sau cu vegetatie halofila. Acest fenomen, cunoscut sub numele impropriu de “poluare salina naturala”, are cele mai clare efecte in aria cu saraturi si lacuri sarate din partea de NE a Campiei Romane.

Solubilizarea masivelor de sare conduce si la aparitia unor forme carstice naturale cu totul asemanatoare celor dezvoltate pe masivele calcaroase, deosebindu-se forme exohalocarstice si endohalocarstice. Cel mai bun exemplu in acest sens este platoul halocarstic Meledic (din bazinul Slanicului de Buzau) ce cuprinde si cea mai lunga peștera in sare din lume.

In acelasi timp, exploatarea pe cale uscata a sarii, dar si cea prin solubilizare subterana cu sonde, conduce la aparitia unor forme carstice aparte apreciate a reprezinta un carst antroposalin. Abandonarea a numeroase vechi lucrari de exploatare a sarii (de la ocne pana la saline moderne) a dus la o dizolvare necontrolata a masivelor de sare respective si la numeroase prabusiri ce au avut ca efect mai ales formarea unor lacuri sarate. In acest context exploatarea prin solubilizare subterana cu sonde se remarcă de fapt ca un mod hiperintensiv de extragere a halitului, mod ce a avut deja urmari grave la Ocnele Mari (Teica).

Intre efectele dramatice datorate interventiei umane asupra ariilor cu masive de sare se remarcă, in afara prabusirilor relativ recente de la Ocnele Mari, modificarile de morfologie din aria



diapirului de la Ocna Mures, prabusirea relativ noua de la Gura Slanicului (Tg. Ocna) si Ocna Dej (sud) si prabusirile ce au marcat in timp exploatarea de la Slanic Prahova.

Intre ariile cu posibile evolutii nefavorabile in viitor se inscriu, in afara conului de surpare de la Teica, aliniamentul Telega-Doftana-Campina, Ocna Mures, Ocna Dej, Ocna Sibiului, Turda, Praida si Gura Slanicului (Tg. Ocna). Pentru aceste zone se impun studii geodinamice de mare detaliu, care sa includa si istoricul complet al exploatarilor, incepand cu cele mai vechi timpuri, precum si monitorizari care sa poata evidenta din timp eventuale evolutii ce ar putea avea consecinte dramatice asupra ariilor inconjuratoare.

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MINERALOGY OF THE SHEAR-ZONE RELATED GOLD ORES. I. GOLD COMPOSITION VS. HOST MINERALS

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As recently summarized by S. S. Udubasa (2004) the composition of gold in the shear-zone related ores at Valea lui Stan and Costesti, South Carpathians, is a function of both the global geochemistry of ores (Cu-dominated at Valea lui Stan, As-dominated at Costesti) and the nature of the host minerals.

Although the two gold ore occurrences belong to the same family or genetic category, they show several differences in many regards, i.e.:

- (1) monophasic (Costesti) against polyphasic (Valea lui Stan) shear-zone deformation events;
- (2) different major and minor ore elements, i.e. Au-As-Cu and Cr-Ni-Pb-Zn (Valea lui Stan) and Au-As-Bi and Cr-Ni-Se-Cd (Costesti), respectively;
- (3) quite different history: Valea lui Stan discovered at the beginning of the XXth Century and exploited until the 40^{ies} (Petrulian, 1936), Costesti discovered more recently (1982-1985) and partly explored (Apostoloiu et al., 1985);
- (4) higher gold grades (up to maximum 100 g/t) at Valea lui Stan and lower gold grades (up to 10-15 g/t) at Costesti;
- (5) both ore occurrences belong to the “small size shear zones” (SSSZ, acc. to Udubasa and Udubasa, 2002); however the known size of the shear zones is bigger at Valea lui Stan (more than 3-4 km) and smaller at Costesti (about 1.5 km).

Combination of items (1) and (5) may theoretically explain the relative richness of the Valea lui Stan gold deposit.

EPMA were carried out on gold from both occurrences. The natural alloy contain major Au and Ag, sometimes approaching the composition of the former electrum, and subordinate amounts of Cu, Ni, Mn, locally (mostly at Costesti) Bi, Fe and As. Some selected analyses are given in the Table 1 and plotted in the ternary diagrams (Figs. 1 and 2).

One can see that the natural alloys are richer in Au at Valea lui Stan (range 44.04 - 72.30 at. %, average 66.43 at. %) as compared to the Costesti occurrence (range 40.67 - 61.21 at. %, average 45.57 at. %). Taking into consideration the host minerals of the gold grains it becomes obvious that the gold grains included in chalcopyrite are poorer in Ag and those included in arsenopyrite are richer in Ag (Valea lui Stan). If present, the pyrrhotite takes the place of chalcopyrite, i.e. the included gold grains are richer in Au and the associated arsenopyrite contains gold grains richer in Ag, sometimes with significant Bi contents (Costesti). Bi always prefers the “electrum” compositions.

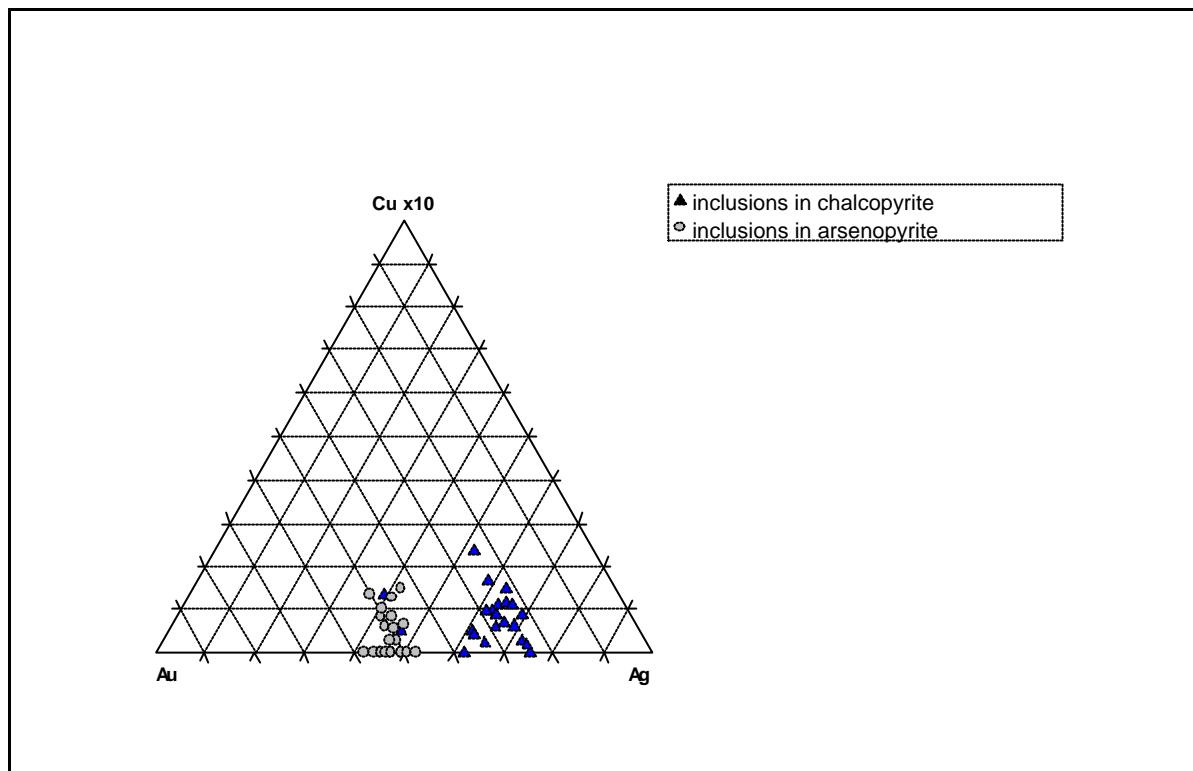


Fig. 1. Ternary plot Au-Ag-Cu·10 of the gold composition from Valea lui Stan.

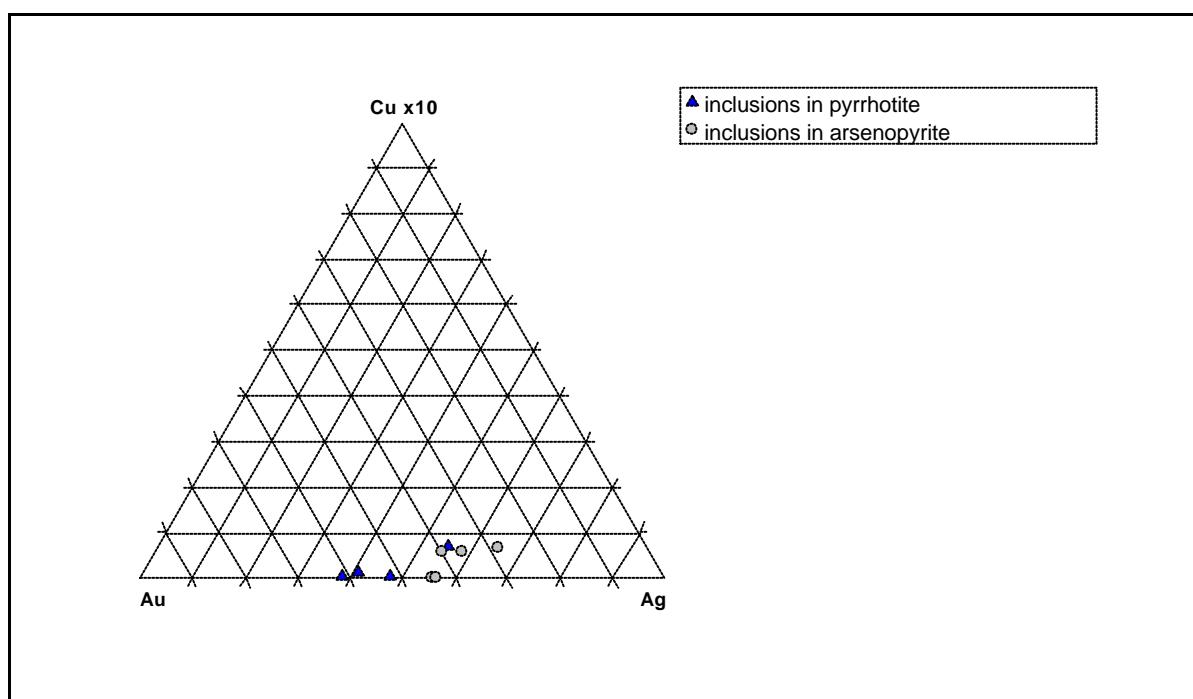


Fig. 2. Ternary plot Au-Ag-Cu·10 of the gold composition from Costesti.


Table 1. Gold composition (minimum and maximum contents) (at. %)*

Occurrence		Au	Ag	Cu	Host minerals
Valea lui Stan	1)	44.04	53.95	1.50	chalcopyrite
	2)	72.30	25.93	1.78	chalcopyrite
	3)	47.62	41.41	-	arsenopyrite
	4)	57.22	41.29	1.50	arsenopyrite
Costesti	5)	40.67	58.35	0.71	pyrrhotite
	6)	61.21	38.59	-	pyrrhotite
	7)	30.38	68.27	0.71	arsenopyrite
	8)	44.51	55.38	-	arsenopyrite

* Ni, Mn and Bi contents were omitted, being accidentally encountered and showing values of less than 0.3 at. % in both occurrences.

Empirical formulas:

- | | |
|---|--|
| 1) Au _{0.44} Ag _{0.54} Cu _{0.01} Ni _{0.003} Mn _{0.001} | 5) Au _{0.407} Ag _{0.583} Cu _{0.007} Ni _{0.002} Mn _{0.001} |
| 2) Au _{0.72} Ag _{0.26} Cu _{0.02} | 6) Au _{0.612} Ag _{0.386} Mn _{0.002} |
| 3) Au _{0.48} Ag _{0.41} Bi _{0.11} | 7) Au _{0.304} Ag _{0.683} Cu _{0.007} Ni _{0.003} Mn _{0.003} |
| 4) Au _{0.57} Ag _{0.41} Cu _{0.02} | 8) Au _{0.445} Ag _{0.554} Ni _{0.001} |

The gold fineness shows a larger variation at Valea lui Stan (246.5 – 836 %) as compared to Costesti (448.3 – 743.4 %) due to the greater availability of Ag in the ores at Valea lui Stan and of Au in the ores at Costesti (Table 2).

Table 2. Average contents of the ores.

	Au (g/t)	Ag (g/t)	Cu (%)	As (%)
Valea lui Stan*	1.4	17	0.48	0.23
Costesti	10.8	11	0.12	5.8
Valiug**	2.25	16.3	0.02	4.9

* Figures given refer to the recently explored parts of the deposit.

** Taken for comparison. Valiug is also a shear-zone related gold mineralization.

The bulk chemical analyses of technological samples show a great similarity between the two occurrences. The enrichment factors for gold (obtained by reporting to the average content of the host rocks) are given in the Table 3. The values are quite similar for the two occurrences but are one or two orders of magnitude smaller than the enrichment factors calculated for the Achean lode gold deposits.


Table 3. Gold enrichment factors.

	Au (g/t) in ores	Au (g/t) in rocks	Enrichment factors
Valea lui Stan	2	0.0167	120 ($\sim 10^2$)
Costesti	3*	0.0192	156 ($\sim 10^2$)
	11**	0.0192	573 ($\sim 10^2-10^3$)

* Average content for the entire zone.

** Average content of a technological sample from the richest zone.

Also some other properties of the mineralizing systems are similar to those of the Archean lode gold deposits, compiled by Groves and Foster (1993) and McCuaig and Kerrich (1994), suggesting the similarity to this type of deposits (Table 4).

Table 4. Comparison data - Archean lode gold deposits vs. Costesti & Valea lui Stan.

	Archean lode gold deposits (Groves & Foster, 1993; McCuaig & Kerrich, 1994)	Costesti	Valea lui Stan
T°C	160-700 (commonly 250-400)	180-385 450-550	350-400
P _{fluid} (kbar)	0,7-5,0 (commonly 1-3)	1,8-3,8	-
Salinity (wt.% NaCl eq.)	0-35 (commonly <6)	6-21	< 20
d ³⁴ S (‰)	-0,7 ... +9	-0,8 ... +5,8	-2,8 ... +15,22
Volatiles composition	H-C-O-S-N H ₂ O-CO ₂ ±CH ₄ ±H ₂ S±N ₂	H+C+O+N (-S) H ₂ O-CO ₂ -CH ₄ -N ₂	similar to Costesti
Gold enrichment factors	10 ³ -10 ⁴	10 ²	10 ²



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ASPECTE PRIVIND ALUNECAREA DE TEREN DE LA SLANIC NORD, JUD. ARGES

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Abstract

The paper presents the main aspects concerning the Slanic Nord landslides which affected the houses in the area and the nearby road and crop fields.

Rezumat

Sunt prezentate aspectele geomorfologice ale alunecarii de la Slanic Nord, în care au fost angrenate peste 950.000 m³ de material terigen. Aceasta a condus la distrugerea caselor aflate pe traseul alunecarii, a drumului de acces, a terenurilor agricole.

Introducere

In microcariera Slanic Nord, după ce procesul de extragere al carbunelui în aceasta a fost sistat, s-a hotărât ca excavatia rezultata să fie utilizată pentru depozitarea sterilului provenit din exploatarea carbunelui din zona. Acest lucru a dus la depozitarea unui volum urias de steril: cca 900.000 m³.

Datorita conditiilor de depozitare, dar și a factorilor naturali care au actionat și de care nu s-a tinut cont în procesul depozitarii, acest material a generat o puternică alunecare, canalizată de existența unei vai, denumita V. Tigancii. Începută în anul 1995, s-a manifestat moderat până în 2004, datorita anilor secetosi. Anul 2004, fiind un an cu precipitatii normale, a declasat o extindere fără precedent a alunecarii de teren, pe o lungime de 1,6 km, de-a lungul văii Tigancii. Acest lucru a condus la distrugerea completă a sase gospodării, inclusiv case de locuit, anexe și terenuri limitrofe, dar și compromiterea definitivă a mai multor terenuri agricole, a drumului de acces în zona și crearea unui lac prin blocarea cursului unui affluent al văii.

Date geomorfologice

Perimetru luat în considerație este situat în zona colinară, facând parte din Muscelele Argesului. Muscelele, ca forme de relief, sunt orientate nord-sud, cu pierderi de altitudine pe aceeași direcție, având cote cuprinse între 660 și 900m. ele se încadrează în dealuri montane, prezintând totusi culmi domoale, cu aspect rotunjit. Diferențele de altitudine din zona sunt date de piscul Slanicului (880m) și confluenta V.Slanicului cu V.Bratila (458m).

Hidrografic, perimetru se încadrează în bazinul râului Arges. De remarcat suprafața foarte mare a bazinului V. Tigancii, de peste 100.000 m².

Date geologice

Depozitele pe care este situat perimetru afectat de alunecare sunt formațiuni pontiene.

Pontianul, care se dispune transgresiv peste Badenianul inferior, a fost separat în trei subdiviziuni:

- 1.nivelul inferior, purtator de carbuni;
- 2.nivelul mediu;
- 3.nivelul superior.



Nivelul inferior. Acest nivel contine patru strate de carbune, note 0, I, II si III si debuteaza cu gresii microconglomeratice, urmate de nisipuri si alternante de argile cu marne, care suporta stratul 0 de carbune. În continuitate de sedimentare este un complex marnos-argilos, fosilifer, în grosime de 25-30m si care prezinta intercalatii de nisipuri sau gresii micafera. Peste acest complex se dispune stratul 1 de lignit, cu grosimi de 1,6 – 3,2m, principalul strat exploatat în zona.

Urmeaza depozite constituite dintr-o alternanta de marne si argile cu gipsuri si gresii cu caracter curbicortica, peste care se dispun stratele II si III de carbune, cu grosimi submetrice si cu dezvoltare lenticulara.

Nivelul mediu se dispune concordant peste nivelul inferior cu carbune, fiind alcătuit din nisipuri cu intercalatii de marne si pietrisuri.

Nivelul superior, în continuare de sedimentare, este constituit din marno-argile cenusii, compacte sau cu aspect foios, prezentând subordonat nisipuri cenusii fine.

Referitor la conditiile hidrogeologice existente în perimetru, se pot aminti, în așa-numitul “complex pontian”, două orizonturi acvifere care pot avea legătura cu producerea și întreținerea alunecarii de teren.

Orizontul acvifer din culcusul stratului I are dezvoltare continua și este format din 1-3 strate de nisip. De remarcat faptul ca primul strat de nisip se gaseste chiar sub stratul de carbune, strat ce a fost exploatat în cariera. Prin acest proces, în mod inevitabil, stratul acvifer situat sub stratul de carbune a fost deschis, apa impregnând sterilul până la saturatie chiar la interfata roca/steril depus.

Orizontul inferior din acoperisul stratului I prezinta o extindere generală fiind constituit din 1-2 strate de nisip, fiind poziționat la cca 15-30m deasupra stratului.

Tectonica este dominată de prezenta faliilor apartinând a două generații. Prima generație grupează faliile care afectează longitudinal zacamântul, cu direcție generală vest-est; cu cele mai importante fracturi, având sărituri de ordinul metrilor la zeci de metri. A doua generație de fali afectează transversal structura, imprimând un caracter disjunctiv tectonicii; ele sunt fali normale și au deplasări modeste pe verticală, de ordinul centimetrilor sau metrilor.

Descrierea alunecării.

Alunecarea, având obârsia în microcariera Slanic, se suprapune peste V. Tigancii, însuțind o lungime de 1.600m și o suprafață de 116.771m² (Fig. 1). În anul 1997 (mai) au început lucrările pentru stabilizarea alunecării, constând în construcția unui dig de anrocamente dublat de un dren, pe direcția punctelor 30-31 (Fig.1) și de canalizare a apelor. Datorită lipsei fondurilor, lucrările au fost abandonate în august același an. Alunecarea și-a urmat cursul, astfel încât pe perioada 1997-2004(aprilie) a avansat cca 600m, ducând la blocarea pârâului Tigancii și formarea unui lac la confluenta celor două văi. Începând cu 5 aprilie 2004, alunecarea s-a reactivat, înaintând în aceasta lună cu viteze de 10-12m/săptămâna. Până în toamna aceluiasi an, a ajuns la lungimea totală de 1.600m, fiind la cca 80m de casele din satului.

Fruntea alunecării este situată, așa cum s-a mai amintit, pe taluzul vechii cariere (Foto 1), suprafața de alunecare incipientă corespunzând cu marna din culcusul stratului de carbune (Foto 2).

În masa alunecatoare din cadrul carierei s-au format lacuri care colectează o parte din apă provenită din precipitații.(Foto 3, 4). Forajele efectuate în masa alunecatoare au arătat pentru acesta zona grosimi de 4,60-5,40m ale acesteia, cu nivele ale apei de -3,50m.

Alunecarea s-a canalizat pe o vale limitrofă incintei carierei, V. Tigancii, (Foto 5) care, înainte de utilizarea acesteia pentru depozitarea sterilului, aduna apele pluviale din bazinul carierei.



Masa alunecatoare din zona superioara a V. Tigancii (Foto 5) are grosimi relativ modeste, de 3-3,20m, (F 9, F 10), datorita pantei mari a vaili dar si faptului ca exista niste izvoare de coasta chiar la obârsia acesteia. Acest fapt explica fluidizarea continua a materialului terigen alunecat din cariera si curgerile de noroi observate in perioada de inceput a alunecarii. Chiar si in perioada cand au fost facute forajele (luna septembrie), secetoasa, umiditatile probelor recoltate au fost $w = 43,4-55,2\%$, cu un indice de consistenta $I_c = 0,28-0,46$.

In partea mediana a vaili, grosimea materialului alunecat creste, atingand in zona forajelor F5 si F6 la 5,70m, respectiv la 5,90m. Influanta apei din lacul format din blocarea parcului Tigancii (fig. 1) se resimte in masa alunecatoare, dand umiditati de peste 40% si indici de consistenta sub 0,5.

Fruntea alunecarii arata grosimi deosebit de mari pentru materialul alunecat, de 7,30m in F 1 si de 11,50m in F 2. Se constata in acelasi timp o scadere a umiditatii, la valori de 33-36% si indici de consistenta cuprinsi intre 0,36 si 0,66.

Concluzii

Halda de steril ce este cantonata in fosta cariera Slanic nord si care a fost antrenata in alunecarea de pe V. Tigancii, limitrofa Padurii Arsuri, a avut ca factor declansator si ca "motor" al deplasarii excesul de umiditate.

Trebuie remarcat faptul ca depunerea materialului terigen a fost facuta fara a exista un proiect de sistematizare si fara a se tine cont de existenta apei.

Cosideram ca originea apei care satureaza materialul depus in fosta cariera este 1) acviferul existent in culcusul stratului I, deschis ca urmare a exploatarii carbunelui si 2) precipitatii. Aportul pluvial este destul de important, daca se tine cont de suprafata carierei si a zonelor limitrofe acesteia, care se constituie intr-un bazin hidrografic foarte mare. In zona fostei cariere, prezenta lacurilor precum si nivelele hidrostatice intalnite in foraje, mai ridicate in jumatea sudica a acesteia, atesta acest lucru.

Pe traiectul alunecarii au mai fost identificate zone cu exces de umiditate: 1) izvoare de coasta in partea superioara a V. Tigancii; 2) infiltratii din lacul de baraj format la confluenta celor doua vai; 3) aport al apelor de siroire provenit de pe versantul Padurii Arsuri.

Propuneri

Pricipala masura care se impune este eliminarea excesului de umiditate atat in cariera, cat si pe traiectul alunecarii.

1. Amenajarea materialui terigen existent in vechea cariera sub forma unor trepte de cariera, a caror inaltime sa nu depaseasca 10. Luand in calcul situatia cea mai defavorabila (analizele geotehnice furnizate de F12), pentru un factor de siguranta $F_s > 1,2$, rezulta un unghi de taluz care nu trebuie sa fie mai mare de 8° .

2. Stabilizarea materialui taluzelor prin plantarea de specii vegetale cu radacini pivotante.

3. Aceasta amenajare trebuie dublata de amplasarea unui sistem de drenuri care sa preia atat apele din precipitatii, cat si cele provenite din acviferul stratului I. Un sistem de drenuri va fi paralel cu treptele de cariera, iar celalalt perpendicular pe directia acestora, pentru dirijarea si eliminarea apelor. Eliminarea apelor se va face pe traiectul V. Tigancii.

4. Executia unor drenuri de-a lungul vaili pe care sa canalizat alunecarea (cate unul pe fiecare latura), suficient de adanci ca sa poata prelua si izvoarele de coasta acoperite de alunecare.



5. Asecarea lacului de baraj format de traiectul alunecarii si regularizarea vaii Tigancii, astfel încât sa pota elmina debitele mari provenite din cumulul apelor pluviale si cursul permanent al pârâului Tigancii.

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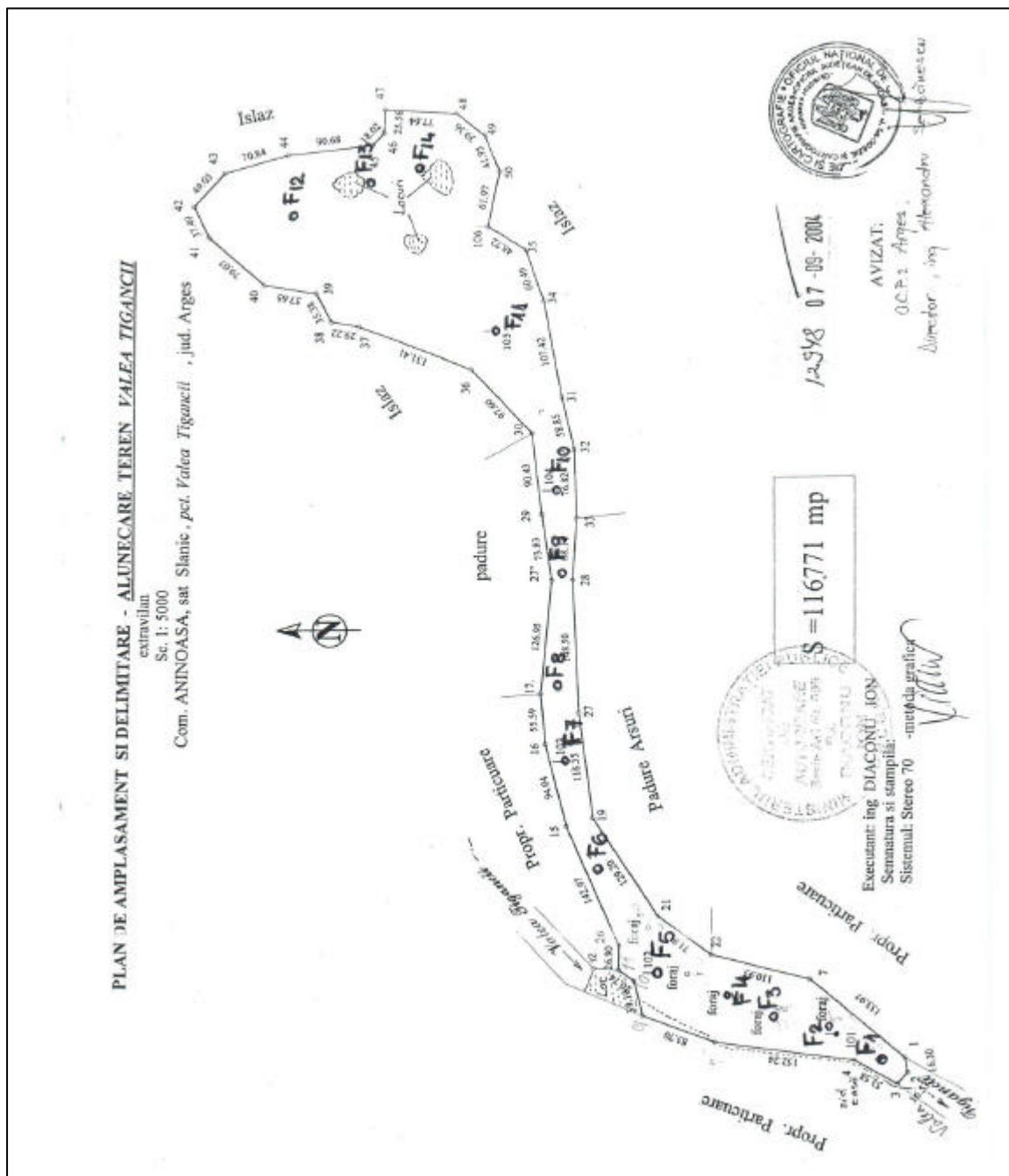


Fig.1. Plan amplasament si delimitarea alunecarii de teren Slanic Nord.



Foto 1. Fruntea alunecarii



Foto 2. Fruntea alunecarii, cu suprafața de alunecare pe marna de sub stratul de carbune



Foto 3. Imagine de ansamblu cu masa alunecatoare din microcariera Slanic.



Foto 4. Aspecte ale alunecarii în microcariera Slanic, cu formarea unor zone de baltire a apei



Foto 5. Canalizarea alunecarii pe V. Tigancii



Foto 6. Masa alunecatoare în zona mediana a V. Tigancii



Foto 7. Marginea nordica a alunecarii, care a afectat livezile limitrofe vail.



Foto 8. Ebulmentul alunecarii



TRANSFORMARI MINERALOGICE PUSE IN EVIDENTA IN SONDA K8, SUPLACU DE BARCAU, IN URMA PROCESULUI DE COMBUSTIE SUBTERANA A HIDROCARBURILOR

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Abstract

This paper presents the main mineralogical aspects concerning the rock transformation after the underground process, upon some minerals present in K8 well.

Rezumat

În prezența lucrare au fost trecute în revista unele transformări mineralogice ce putut fi puse în evidență în urma observațiilor efectuate cu ajutorul microscopului optic, electronic și analizelor Rx.

Introducere

Zacământul Suplacu de Barcau a început să fie exploatat prin procesul de combustie subterană începând cu anul 1967, pe un panou de 4ha, cu o sonde de injectie și 7 de reacție[1]. Stratul productiv, situat la o adâncime de 35 la 220m, are o grosime cuprinsă între 6 și 27m, fiind de vîrstă pliocena (pannoniana), constituind o serie detritico-pelitica[2]. Depozitele pliocene au fost împărțite în trei complexe: bazal, intermedian și superior, cel bazal continând trei strate.

Rocile colectoare prezintă variații de facies atât pe verticală, cât și lateral; ansamblul acestor roci cuprinde microconglomerate, gresii calcaroase, nisipuri slab consolidate de diferite sorturi, marne nisipoase, calcar oolitic[3].

Sonda K8 a fost sapată în luna mai 1982, fiind amplasată în partea de vest a zacământului; particularitatea ei constă în faptul că a fost sapată în spatele frontului de combustie, fiind într-un areal în care nu s-a injectat apă, astfel încât informațiile provenite din prelucrarea datelor se referă la propagarea frontului de combustie uscată, spre deosebire de informațiile obținute pentru celelalte sonde K. Ea a fost carotată continuu pe intervalul 64-84m, efectuându-se sapte marsuri.

Proba pe care autorul a supus-o observațiilor provine din marsul nr. 5 (77,14 – 77,95m) și acoperă intervalul 77,2-77,5m.

Pentru evidențierea aspectelor mineralogice, au fost efectuate observații la microscopul optic (Amplival), la microscopul electronic (Remma 202) și au fost facute analize Rx și spectrale.

Descrierea mineralogică a probei

Macroscopic, roca este o gresie de culoare cenusiu-negricioasă, cu bobul mediu la fin, slab micaferă, cu ciment calcaros. Textura este psamitica, aspectul compact, fiind lipsită de stratificare. Granulele componente se încadrează în categoria subangular-subrotunjite.

Microscopic, roca este constituită din granule de cuart, feldspati, calcit, mice, clorit, fragmente litice, toate prinse în ciment calcitic.(fig. 1). Dupa propoția constituentilor mineralogici



si natura liantului, având în vedere clasificarea lui Pettijohn, Potter, Siever(1973), roca se încadreaza în categoria gresiilor arcoziene[4].

Cuartul apare cu frecventa cea mai ridicata, atingând proportii de 50-55%. Este prezent atât sub forma monocristalina, cât si policristalina, granulele având morfologia de la angular la rotunjit. Se constata un grad de rotunjire mai înalt la granulele de dimensiuni mari.Granulele sunt în general bine delimitate de ciment, foarte putine prezentând o slabă coroziune marginală calcitică.

Feldspatii apar ca granule detritice, însumând peste 25% din totalul mineralelor. Sunt prezente atât forme de feldspat plagioclaz, cât si ortoclaz. Granulele sunt fie rotunjite, fie apar cristale “clivate”. De remarcat, indiferent de tipul genetic sau morfologic, gradul înalt de corodare marginală a cristalelor de către calcit. Feldspatii plagioclazi sunt mult mai alterati decât cei ortoclazi.

Fragmentele litice sunt reprezentate prin granule de cuatite si mai rar filite si micasisturi. Prezinta un grad de rotunjire de la angular la subrotunjit si sunt afectate – cu exceptia cuartitelor – de procese de alterare mai intense decât ceilalți constituenți.

Micile, reprezentate prin muscovit si biotit, se prezinta sub forma agregatelor minerale de dimensiuni reduse; nu se poate sesiza o orientare preferentială a lamelelor. Muscovitul se prezinta în general nealterat (doar un slab proces de hidratare) în timp ce la biotit acest proces se manifesta prin cotururi slab rotunjite si un incipient proces de cloritizare.

Calcitul apare atât alogen cât si autigen; formele alogene sunt reprezentate prin aggregate granulare normale, având aspect mozaicat. Cele autigene apare ca produs de alterare în feldspati, fragmente litice, rar marginal pe quart. Calcitul mai apare prezent, relativ rar, sub forma resturilor scheletice neidentificabile.

Cimentul este calcitic, înglobînd si minerale argiloase, fiind singenetic; după tipul structural se poate defini ca ciment bazal. Cimentul este caracterizat printr-un intens proces de recristalizare, mai intens în apropierea cristalelor de dimensiuni mari. Aceasta este însotit de un proces de “purificare”, în sensul migrării materiei organice (titei) din mediul poros-permeabil al cimentului si acumularii în anumite zone, usor sesizabile microscopic, prin aspectul brun-roscat. De remarcat si aparitia de zone de extrema concentrare, unde titelui s-a transformat în “semi-cocks”. Cimentul este fragmentat de aparitia unui adevarat sistem de fisuri, caracterizat prin “bordarea” lui cu material organic intens transformat.

Microscopia electronica a pus în evidenta gradul ridicat de alterare a mineralelor feldspatice, ce sunt afectate nu numai de procesul de calcitizare observabil la microscopul optic, ci si de procese de argilizare. În foto nr. 2a se poate observa un sistem de microfisuri ce grupează de-a lungul sau zone de intensa concentrare a materiei organice, sub forma de “semi-cocks”[5]. Un astfel de granul este detaliat în foto nr. 2b, caracterizat de faptul ca “semi-cocks” adera la peretii fisurilor, formând un adevarat edificiu dezvoltat tridimensional. Fata de granulele de “semi-cocks” observate în sonda K2, care au forma aproape sferică, aceste prezinta aspecte filiforme, care sunt rezultate în urma unei temperaturi mai ridicate decât în primul caz si formate în lipsa aportului de apa.

Difractia Rx a fost efectuata atât pe proba initială, cât si pe o probă supusă unei extrageri a materiei organice cu aparatul Soxlet a cărei difractogramă este prezentată în fig. 3. Operatiunea de extragere a materiei organice a fost necesara deoarece stabilirea compozitiei mineralogice semicuantitative este dificila în prezenta acesteia.

Wollastonitul, $\text{Ca}_3[\text{Si}_3\text{O}_9]$, un ciclosilicat, se poate forma la temperaturi de 350-400°C, la presiuni normale, asa cum arata Dobrovă citându-l pe Kalimin [6]. A fost pus în evidență termenul triclinic, de temperatură scăzută, care este citat în primul rând în metamorfismul de contact, dar și în

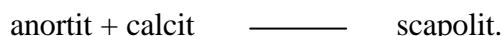


metamorfismul regional de presiune scazuta sau medie. Prezenta wollastonitului este justificata de conditiile de formare ale acestuia din calcit, în prezenta siliciului, fapt favorizat de existenta apei si a bioxidului de carbon.

Wairakitul, $\text{Ca}_2[\text{Al}_{16}\text{Si}_{32}\text{O}_{96}] \cdot 16\text{H}_2\text{O}$, un zeolit cu lanturi tetraedrice simple, este specific metamorfismului regional de intensitate scazuta. El apare în asa numitul facies zeolitic, caracterizat de presiuni scazute si temperaturi cuprinse între 100-350°C. El este citat ca aparând în metamorfismul sedimentelor carbonatice, în cazul participarii concomitente a apei si a bioxidului de carbon [6], luînd nastere ca urmare a instabilitatii feldspatilor plagioclazi (anortitul):



Scapolitul, $(\text{Na},\text{Ca},\text{K})_4[\text{Al}_3(\text{Al},\text{Si})_3\text{Si}_6\text{O}_{24}](\text{Cl},\text{F},\text{OH},\text{CO}_3,\text{SO}_4)$, poate lua nastere, conform lui Dobretov si colab.(1977), prin transformarea felspatilor plagioclazi bazici în faciesul metamorfismului regional de temperatura scazuta, prin prin asa numitele reactii de deanortizare, în conditiile unui exces de boxid de carbon.



Tabel 1. Compozitia mineralogica semicantitativa a probei K8

COMPONENTI	PROPORTIE
Cuart	15%
Feldspati	10%
Minerale illito-micacee	20%
Calcit	50%
Clorit	<5%
Silicati de calciu, zeoliti	<5%

Analizele spectrale semicantitative au aratat tediua observata în celelalte sonde K, privind abundenta mari pentru elementele zircon, plumb, mangan si titan, explicabile prin procesele de concentrare a lor în materia organica.

Tabel 2. Rezultatele analizelor spectrale semicantitative pentru proba K8

Element	Zr	Cu	Pb	Sn	Mn	Ga	Cr	Ni	V	Ti
Limita detectie	30	3	3	3	30	3	10	3	3	30
Continut	130	12	100	3	370	7	22	5	17	>3000

Concluzii

Procesul de combustie subterana a hidrocarburilor conduce, în mod cert, la transformari privind aspectul fizic si în compozitia mineralogica a rocilor gazda. Principalele modificari observate sunt urmatoarele :

1. procese de alterare, mai pregnante la feldspati, mice, fragmente litice;
2. cimentul basal de tip calcitic, este: a) recristalizat; b) brazdat de un adevarat sistem fisural; c) afectat de procese de concetrare a materiei organice;
3. microscopia electronica a pus în evidenta granule de “semi-coce” cu aspect filiform, dat de temperaturi de formare mai mari, si de lipsa apei, diferite de celelalte granule evidențiate în sondele K, aproape sferice;



4. analizele Rx au pus în evidență minerale specifice metamorfismului regional de temperatură scăzuta : wollastonit, scapolit, wairakit.
5. analizele spectrale au detectat abundente ridicate pentru o serie de elemente, cum ar fi plumb, mangan, titan, zircon, explicabile prin procesele de concentrare ale lor în materia organică.

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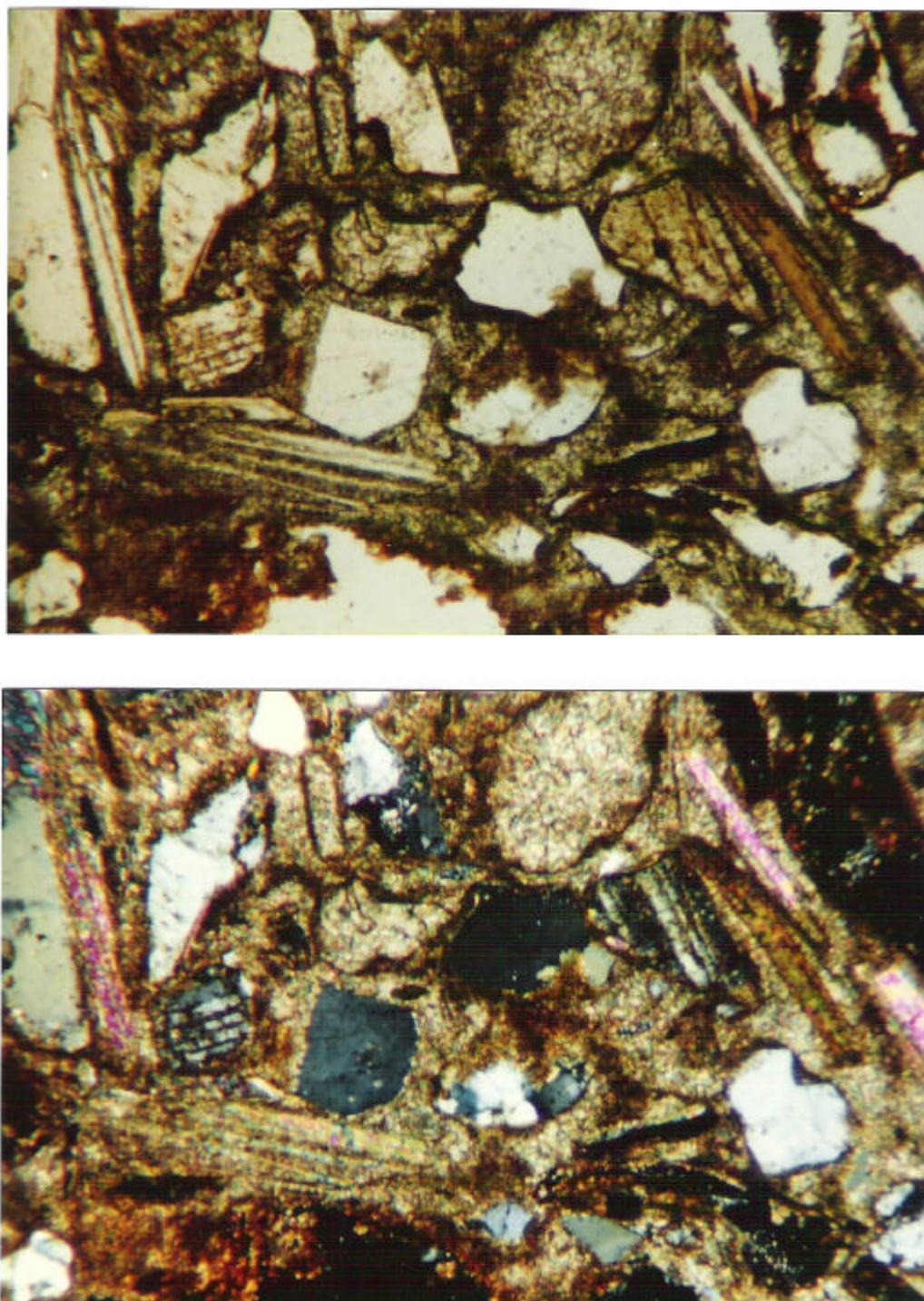


Fig. 1. Observatii microscopice pentru proba sondei K8;
a) N II, x25. b) N +, x 25.



Fig. 2. Fotografii microscopie electronica pentru proba sondei K8;
a) sitem de fisuri cu zone de concentrare a materiei organice.
b) aspecte filiforme ale granulelor de “semi-cocs”.

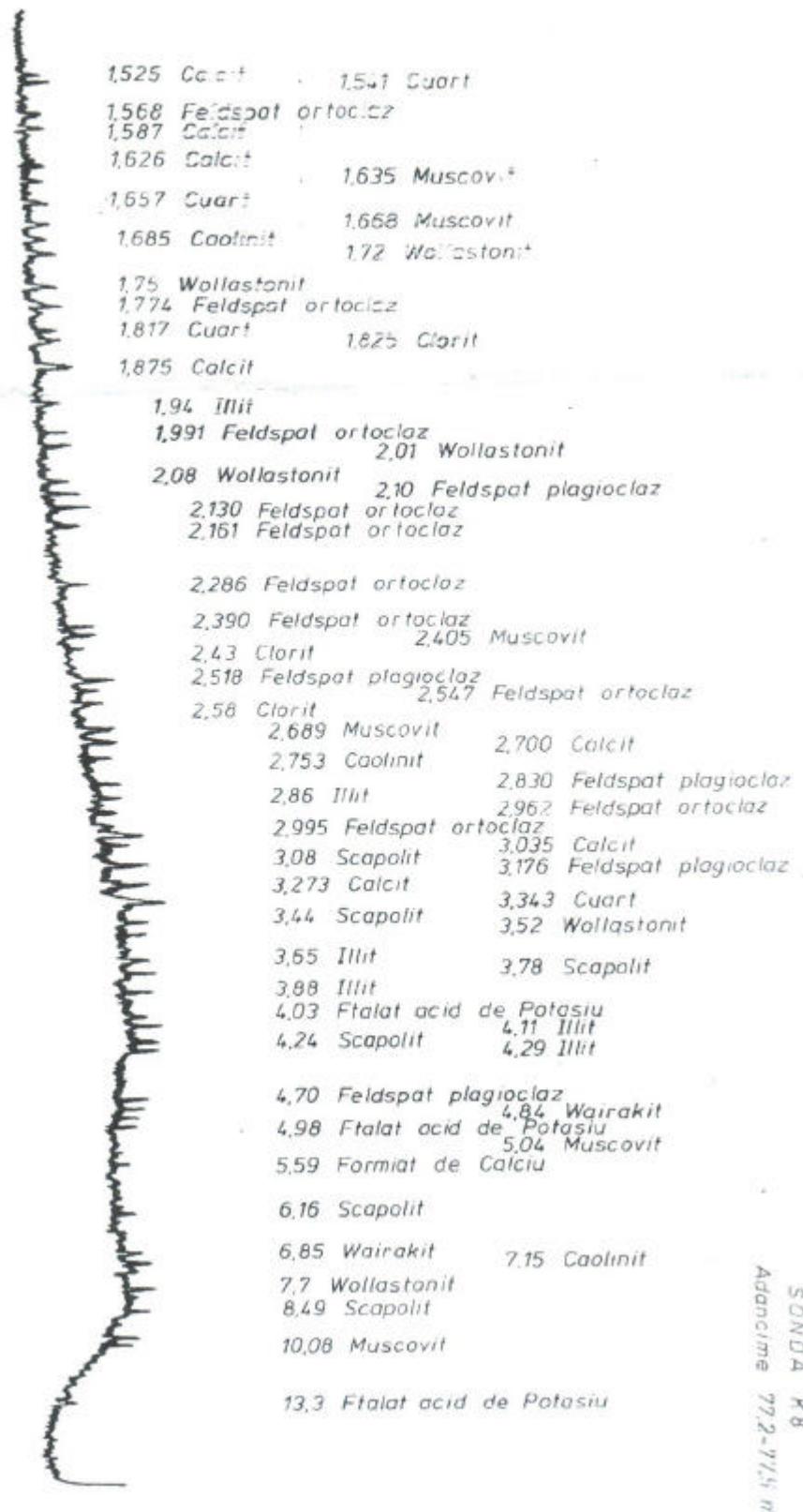


Fig. 3. Diffractogramma Rx pentru proba sondei K8, adâncime 77,2-77,5m



FIELD TRIP GUIDE



THE ROSIA MONTANA ORE DEPOSIT

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Abstract

The Rosia Montana gold deposit is located in the historical gold mining district known as the Golden Quadrilateral within the Apuseni Mountains of Transylvania in western Romania. Historical gold mining has occurred at Rosia Montana since pre-Roman occupation of the area during the Dacian period over 2000 years ago. Gold at Rosia Montana occurs within an intermediate-sulphidation style epithermal mineralized system, hosted in Tertiary age dacitic intrusions and associated phreato-magmatic breccias. The diatreme is situated within an area dominated by Cretaceous shallow marine and deltaic sediments. Recent exploration of the deposit by Gabriel Resources has outlined a total resource (measured, indicated and inferred) of 400.41 million tonnes at an average grade of 1.3 grams per tonne gold and 6.0 grams per tonne silver for a total contained resource of 16.1 million ounces of gold and 73.3 million ounces of silver.

Regional Geology

Romania includes three major Alpine and older orogens, namely the Carpathian chain that comprises the Southern Carpathians and the Eastern Carpathians, the Apuseni Mountains and the Northern Dobrogea. Tertiary sediments were deposited in the intervening Pannonian and Transylvanian Basins, as well as on the Scythian and Moesian Platforms. Two principal areas of Tertiary volcanic rocks, of predominantly calc alkaline affinity, intrude and overlie these sequences. First one spreads in the Eastern Carpathians from the north in the Baia Mare area (Oas-Gutâi mountains) to the south (Calimani-Gurghiu-Harghita mountains), containing also a subvolcanic median sector (Tibles-Toroiaga-Rodna-Bârgau mountains). The second area with Tertiary volcanic rocks is the Apuseni Mountains in central-western Romania.

The famous mining districts of the Metaliferi Mountains of Transylvania, which represent the southern part of the Apuseni Mountains, comprise a 500km² region, immediately to the north of the city of Deva, commonly referred to as the Golden Quadrilateral (Fig. 1). The Golden Quadrilateral has remained Europe's most important centre of gold production for more than 2000 years since Geto-Dacian (pre-Roman) times, with the Roman conquest of Dacia in 105AD-106AD predicated on gaining control over this important goldfield. The district reached peak production during the period of the Austro-Hungarian Empire at the end of the 17th Century to 1918 as well as before World War II.

The Golden Quadrilateral lies within the Apuseni Mountains, which consist of Mesozoic, shallow marine and non-marine sedimentary rocks overlying Palaeozoic and Precambrian sedimentary and metamorphic basement. North-directed thrust faulting during the late Cretaceous resulted in a series of nappes that are unconformably overlain, and intruded, by Tertiary volcanics associated with high-level gold-silver mineralisation and porphyry copper deposits of the Golden Quadrilateral.

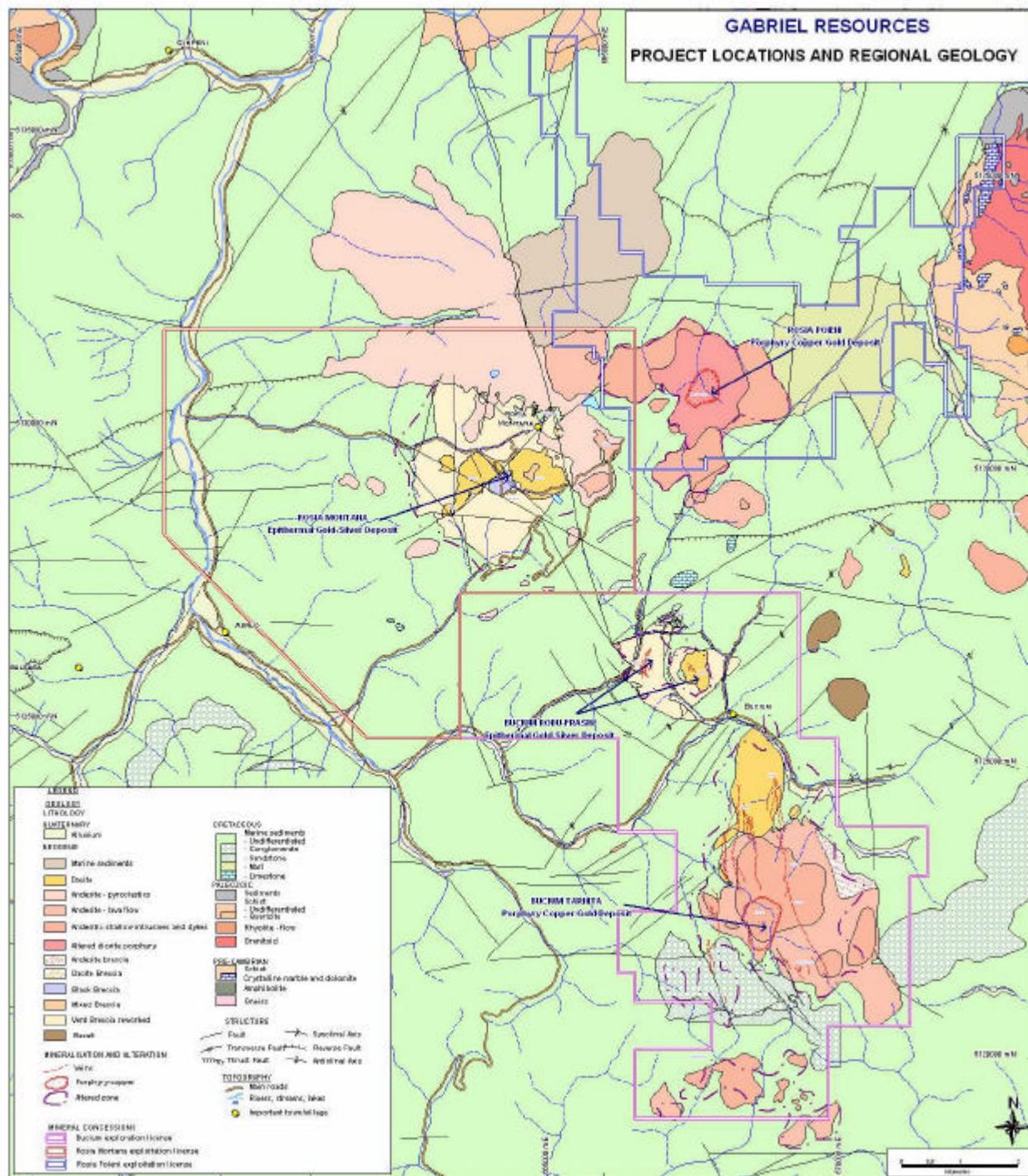


Fig. 1. Rosia Montana regional geology



According to the classical view regarding the Tertiary volcanism from the Apuseni Mountains three cycles has been distinguished (Ianovici et al., 1976). The earliest cycle is interpreted as lower Badenian age and comprises rhyolitic ignimbrite overlain by rhyodacitic and andesitic volcanics. Volcanogenic sediments occur throughout this cycle and widespread hydrothermal alteration overprints all rock types.

The rocks of the second cycle outcrop extensively and are characterised by andesite and dacite overlain by a very thick sequence of quartz andesite that is, in turn, overlain by pyroxene andesite. The sequence is interpreted to be late Badenian – Sarmatian and Pannonian age. The middle (dacite) and upper (quartz andesite) sequence of this cycle represents the principal host to gold-silver mineralisation currently being mined in Romania, as well as significant occurrences of copper, lead, zinc and mercury.

The third and final cycle of volcanism continued into the Quaternary era and is characterised by pyroxene andesite, basaltic andesite and potassic basalt.

According to available K-Ar datings (Pécsay et al., 1995), the main volcanic activity from the Apuseni Mountains range between 14.7 and 7.3 Ma, and ended in Quaternary (1,6 Ma).

Three major northwest-trending belts of volcanism (Brad – Sacărâmb, Zlatna – Stanija, and Rosia Montana-Bucium) and associated mineralisation are identified within the Golden Quadrilateral, with the Rosia Montana Complex representing part of the northernmost belt.

Project Geology

The local Rosia Montana deposit is interpreted as a maar-diatreme complex of Neogene age emplaced into a sequence of Cretaceous sediments, predominantly black shales, sandstones and conglomerates. Mărza et al. (1997) documented the Rosia Montana ore deposit as a low sulphidation epithermal deposit. More recently, Rosia Montana has been interpreted as intermediate sulphidation epithermal (Sillitoe and Hedenquist, 2003).

The three dimensional geometry of the area is well established due to the extensive network of underground development that has been undertaken since the Austro-Hungarian Empire period, and from the extensive surface and underground drilling completed in the last 25 years.

Lithologies within the diatreme complex are dominated by breccias, including magmatic-phreatic and sub-aqueous reworked breccia, intruded by porphyritic dacitic sub-volcanic intrusives. These intrusions are interpreted as Neogene age and are informally named the Cetate dacite (Cetate and Cârnica massifs). The dacite bodies are interpreted to have intruded vertically through the diatreme breccias and to have spread laterally at shallower levels forming surface domes (Fig. 2). An alternative interpretation is that only one major dacite intrusion has occurred and that this has been split into the now separate Cârnica and Cetate dacite bodies by a northeast trending strike-slip fault.

The majority of the Rosia Montana diatreme is made up of a lithology locally referred to as the ‘vent breccia’. This is a diatreme breccia produced by numerous magmatic-phreatic eruptions produced as hot rising dacitic magma interacted with ground water. The vent breccia hosts the dacitic intrusives and, in the case of Cârnica, forms a sub-vertical, ‘ribbon like’ NE-SW oriented structure inside the dacite. This is referred to as the ‘internal vent breccia’ for descriptive statistical and estimation purposes. It is of variable composition with clasts of dacite, Cretaceous sediments and basement schist and gneiss. The clast size, degree of rounding, and the proportion of matrix, vary widely. Texturally it exists as both massive breccia units and sub-aqueous reworked breccia indicating the breccia has erupted into a shallow lake or maar.

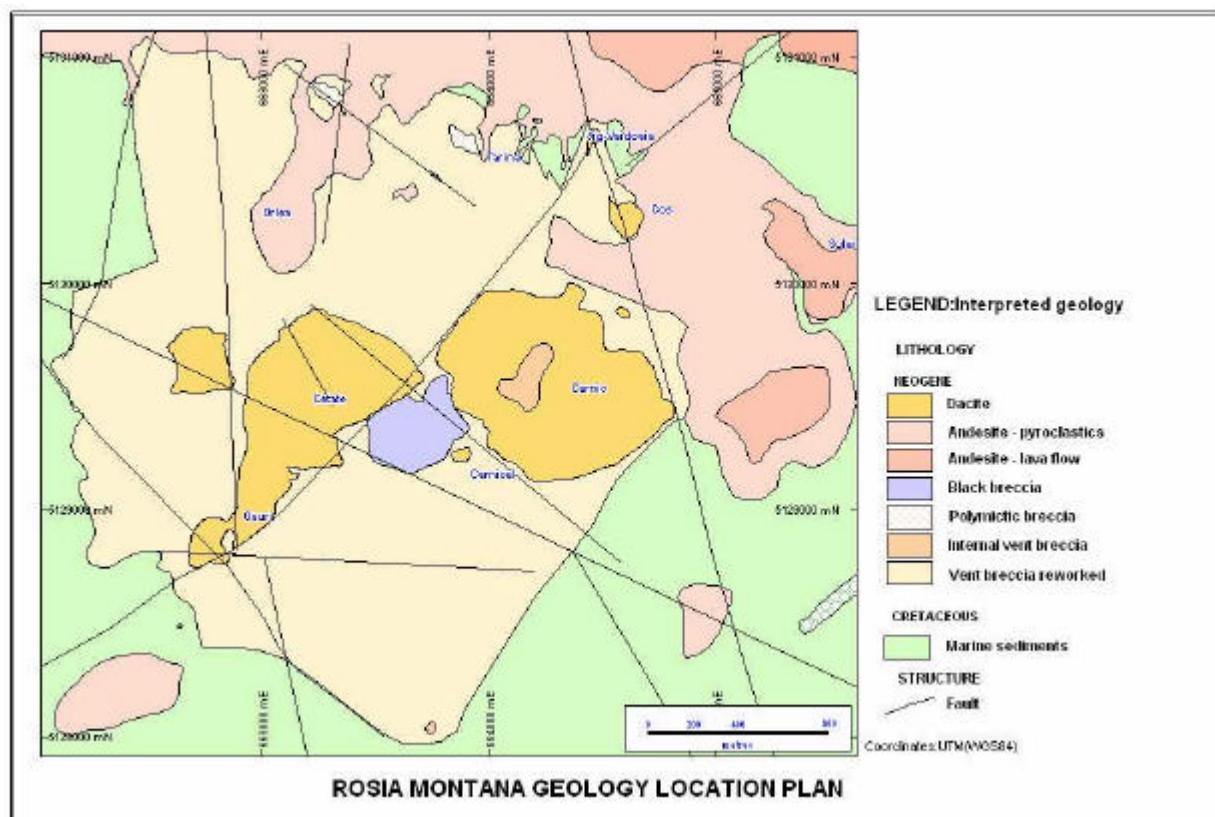


Fig. 2. Rosia Montana geology

This reworked vent breccia is fine to coarsely bedded and varies from clay rich to more common sandy and gravelly beds through to beds containing poorly sorted, cobble sized clasts. Graded bedding is common and cross bedding and ripple marks have also been observed.

A breccia, locally termed the ‘Black Breccia’, forms a sub-vertical pipe within the complex core, emplaced adjacent to the dacitic intrusions. The Black Breccia consists of clasts of Cretaceous black shale from the surrounding sedimentary sequence incorporated into the explosive deposits in the centre of the maar complex, along with clasts of ‘basement’ lithologies such as garnet-bearing schist and gneiss. The unit only hosts significant gold-silver mineralisation along its northeastern margin. The matrix is dominated by pulverized (rock flour) Cretaceous shale.

A number of intrusive polymictic diatreme breccia bodies that crosscut the reworked vent breccia have been identified between Tarina and Jig. These are composed of matrix-supported breccias with sub-rounded clasts of dacite, Cretaceous sediment and crystalline schists. They vary from poorly sorted to moderately sorted and generally have a uniform maximum clast size (0.5 to 5.0cm) within a specific eruptive unit. The matrix is made up of pulverized sand sized dacite, Cretaceous sediments and schist (rock flour). These are the product of magmatic-phreatic eruptions at depth (within the schist), and have followed steep fault structures to the surface giving them a sub-vertical orientation and are either cylindrical or elongated along the structure in shape. They are up to 150m in width and are generally well mineralised.

In the Igre area dacite and polymictic breccia dykes have also been identified within the Cretaceous sediment. The diatreme breccia dykes are usually 0.5m to 2m in width and are lithological similar to the diatreme breccia described above. These are interpreted to be offshoots of the diatreme breccia material along smaller faults. The dacite dykes are usually 0.5m to 1m in width



and are composed of fine-grained, silicic dacite, with coarse sand-sized phenocrysts of quartz. Dykes containing both polymictic breccia and dacite have also been observed.

Andesitic extrusive rocks are mapped mantling the northern and eastern parts of the project area, forming a thin to moderately thick cover over the maar complex. The lowest units within the sequence are pyroclastic block and ash flows, further north and east andesitic lavas overlie the pyroclastics.

Structure

Structure has played an important role at Rosia Montana, firstly supplying dilation for the emplacement of the maar-diatreme complex, and secondly the structural permeability up which the mineralising fluids have flowed.

The Rosia Montana diatreme is interpreted to be emplaced at the intersection of two sub-vertical structures that trend north-northwest and northeast. The north-northwest trend is interpreted to be the earlier and the larger of the two structures. This is a basement structure and of regional scale that has produced the broad zone of fracturing and veining that is seen on the surface within the dacites, vent breccia, Cretaceous sediments and in the andesites at Bucium. It can be traced from Orlea North through Orlea, Cetate and Cîrnic and down through the Bucium exploration licence to the south; along this 13 km trend the dominant vein orientation strikes parallel to it and is either sub-vertical or dipping steeply to the west.

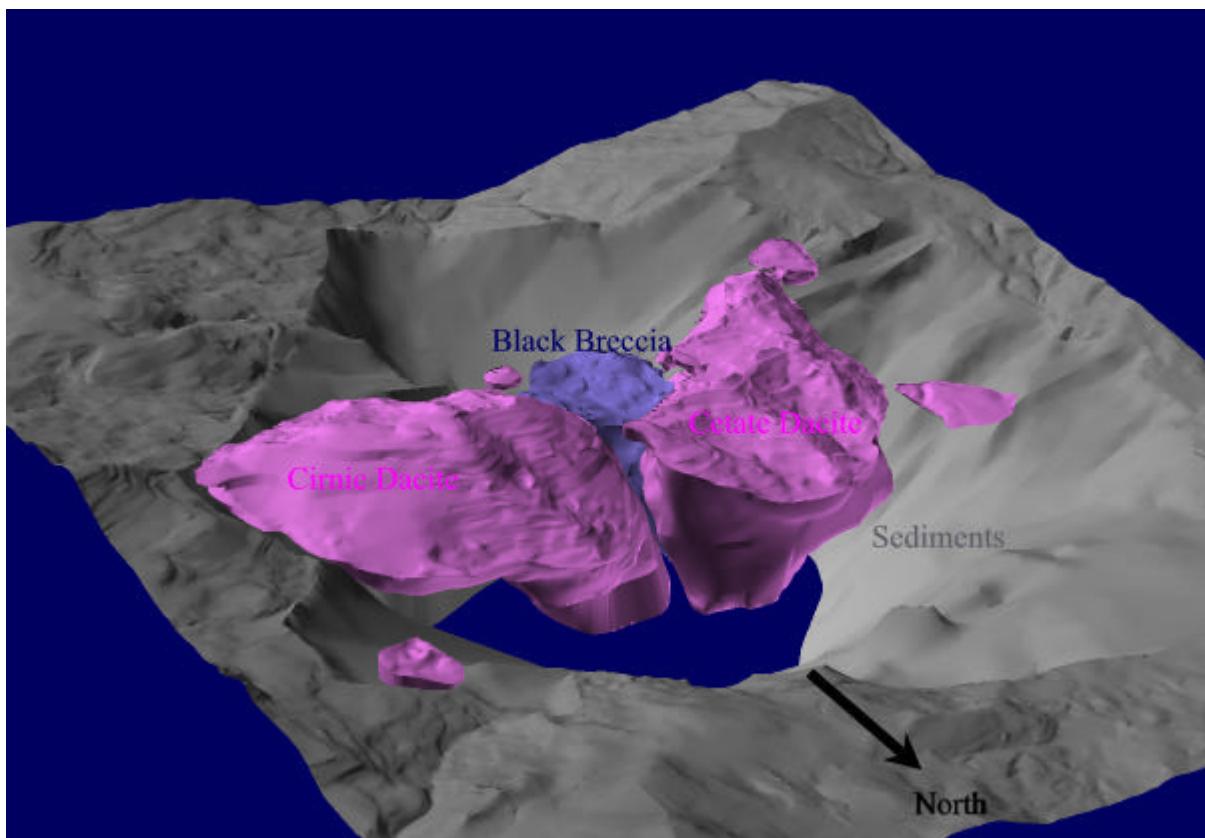


Fig. 3. Perspective view of the geological model.



Across the north-northwest structural trend cuts the northeast trending faults, interpreted to be dominated by left-lateral strike slip movement. The intersection of these regional structures has produced dilation and a focus for the breccias, intrusions and mineralisation at Rosia Montana. Steeply southwest dipping northwest trending structures have also been identified in the Igre area. These are interpreted to have a right lateral sense of movement and may be the conjugate pair to the northeast trending structures.

At Orlea, a large east-west trending clay rich vein (the Crucii vein) dipping at 45° to the south, cuts across the vent breccia. This structure is interpreted to be related to diatreme collapse immediately following eruption of material from the vent. Similar structures have also been interpreted in the Tarina and Igre areas and also tend to dip moderately to steeply towards the centre of the diatreme.

Alteration

An extensive zone of strong hydrothermal alteration hosts the Rosia Montana deposits. The distribution of alteration assemblages is quite complex, however, it can be simplified down to the following groupings (Manske 2004):

- i) Chlorite-carbonate-smectite alteration. Peripheral to the resource, there are not many locations where this alteration has been preserved except for a few small outcrops on Cârnici hill. Although a “green” alteration that superficially resembles a propylitic imprint, this zone lacks epidote or appreciable albite (it may preserve relicts of the original igneous plagioclase, andesine-labradorite). It most closely matches the “intermediate argillic” alteration type in the sense of Meyer and Hemley (1967).
- ii) Phyllitic-argillic alteration. The most widespread alteration at Rosia Montana, creating most of the “bleached” exposures in dacite porphyry. Rocks affected by this alteration show abundant fine-grained sericite (illite) and murky, birefringent clays (smectites) in thin section. XRD analyses and glycolation tests indicate the presence of mixed-layer, illite-smectite minerals as well (*cf.* Tamas, 2002). Supergene argillization certainly exists in the weathering profile over the pyritic orebodies, but the presence of phyllitic-argillic altered rock in the sulfide zone shows that much of the clay is hypogene.
- iii) QSP (quartz-sericite-pyrite) alteration. QSP appears as a subtype of the general phyllitic-argillic imprint, around some quartz veins and also as local zones of more pervasive alteration. It imparts a bluish-gray color to the rock and may be confused with moderate silicification, but may be easily scratched with a knife blade or steel dental pick ($H < 5$). Silica may or may not have been added to QSP rock, as the quartz may be generated by release of SiO_2 during hydrolytic alteration of feldspar (*e.g.*, Barton *et al.*, 1991). This alteration type may destroy much of the primary igneous texture of the rock, except for quartz phenocrysts.
- iv) Quartz-adularia replacement. Adularia \pm quartz appears in vein halos and as more pervasive alteration in dacite and vent breccia. Fine-grained adularia in some cases effects a wholesale replacement of the dacite matrix and creates a pale rock with ceramic-like texture. In extreme cases, original quartz phenocrysts are attacked and partly replaced with adularia along grain boundaries and micro-cracks. As rock alteration, adularia commonly is partly overprinted by sericite (illite) or clays. As a vein mineral in dacite or a vug-filling phase in breccia, paragenetic studies show that adularia was introduced into the system at two or three distinct stages.



- v) Silicification occurs in some quartz vein halos but is not volumetrically widespread in unbrecciated dacite, at least in the Cetate area. Strong silicification is much more characteristic of the margins of breccia zones within or along the edges of the dacite flow-domes. Brecciated dacite but especially the breccia matrix may be densely silicified, replaced by more than 90 vol. % fine-grained quartz as seen in thin section, and much harder than a knife blade in hand sample ($H \sim 7$).

XRD and TEM analyses (Mărza et al, 1997, Tamas, 2002) confirmed for Rosia Montana the following alterations: potassium silicate assemblages, phyllitic assemblages, intermediate argillic, and advanced argillic assemblages.

Mineralisation

Mineralisation within the Golden Quadrilateral district includes porphyry-related gold-silver, copper-gold and copper deposits associated with Badenian-Pliocene (Neogene) andesitic to dacitic volcanic rocks, and associated intrusive rocks. The gold-silver mineralisation outlined at Rosia Montana is interpreted to represent a mid to shallow-level, intermediate sulphidation epithermal system. The mineralisation is dominantly disseminated, with associated stockwork and breccia hosted gold-silver mineralisation.

Gold-silver mineralisation at Rosia Montana is hosted by the following lithologies:

1. Dacite-hosted mineralisation:

Characterised by wide zones of finely disseminated sulphide hosted within dacite porphyry. Silicic-adularia alteration combined with very fine-grained disseminated pyrite are distinctive features of the mineralised dacite and the best indicator of gold and silver grade. Narrow, usually widely spaced stockwork veining is always present but is minor in terms of contained gold and silver. The veins are generally steeply-dipping, discontinuous and less than 1m wide. Significant gold mineralisation of this style occurs at Cetate, Cârnic, Carpeni, Gauri, Lety-Cos and parts of the Vaidoaia zone.

2. Sub-vertical breccia zones cross-cutting dacite intrusive bodies:

Breccias are commonly of mixed lithology and are considered to represent structurally controlled phreato-magmatic breccias. Mineralisation occurs within strongly, to intensely, silicified alteration zones and contain low to moderate amounts of disseminated fine-grained sulphide within both the matrix and breccia clasts. Relevant examples of this type are known in Cetate and Cârnic massifs.

3. Disseminated and vein hosted gold-silver mineralisation within vent breccia:

Significant gold-silver mineralisation is hosted by the vent breccia surrounding the dacitic intrusions. The mineralisation is characterised by silicification and finely disseminated pyrite with veining infrequent, and generally narrow (less than 1m). Examples of this style of mineralisation are at Cârnicel, Vaidoaia, Jig (also known as Lespedari), Igre, Orlea and Tarina.

4. Diatreme breccia pipe hosted mineralisation:

This mineralisation hosted by the sub-vertical diatreme breccia pipes at Igre/Jig. It is characterised by intense, pervasive silicification of both the breccia matrix and the diatreme breccia clasts, disseminated pyrite is also pervasive within the matrix and clasts and will sometimes completely replace the black shale clasts. Zones of rhodochrosite have also been identified, occurring within the matrix of the diatreme breccia.



5. Cretaceous sediment hosted mineralisation:

This mineralisation has been identified at Igre, Gauri and Cos. The mineralisation occurs directly below the vent breccia-Cretaceous sediment contact and is usually hosted by shale, sandstone and less frequently by conglomerate beds. The mineralisation is characterised by both silicification and pervasive fine-grained disseminated pyrite and in some areas (Igre and Gauri) by hydrothermal crackle brecciation that varies from mm-width widely spaced spidery crackle breccia through to more intense mosaic (jigsaw) brecciation. Clasts are always very angular and made up of locally derived sediment. The brecciation can be over 50m thick and tends to be most intense close to the vent breccia-Cretaceous contact. The breccia matrix is typically vuggy and crystalline, some coliform banding has been observed and up to five phases of mineralisation can be present. Mineralisation is dominated by carbonate (both calcite and rhodochrosite), quartz and pyrite with galena and sphalerite not uncommon and rarer chalcopyrite.

Gold has been identified by petrography in numerous samples as electrum. Occurrences were noted as minute (4 µm) inclusions in pyrite, as minute grains (up to 25 µm) intergrown with, and overgrowing Ag-sulphosalts and tellurides. It has also been observed as coarser grains (up to 100 µm) intergrown with carbonate and barite. The electrum is also associated with quartz, galena, and sphalerite. The electrum had a fineness ranging from .537 to .763 (Leach & Hawke 1997).

Fluid inclusion studies indicate that mineralisation was precipitated from a dilute NaCl (0.35 to 7.85 wt. %) hydrothermal solution at temperatures between 200 to 340°C. The lack of vapour-rich inclusions indicates boiling was not the major trigger for gold precipitation. Gold mineralisation may have formed due to the rising mineralising fluid mixing with oxidising CO₂-rich meteoric waters. (Leach & Hawke 1997, Tamas 2002)

Veining

Although veining is secondary to alteration in terms of contained gold and silver, they are important as fluid conduits and therefore control the location and distribution of the gold-silver mineralisation. The veins often contain significantly higher grade gold and were the focus of historical mining.

Multiple phases of mineralisation have been identified at Rosia Montana that can be subdivided into 3 main vein types:

1. ‘Chinga’ veins

The ‘Chinga’ is a black, very fine-grained argilic material, intensely silicified. These ‘veins’ are the earliest phase of veining at Rosia Montana and their textures indicate that the argilic material has been injected into fractures rather than precipitating from a hydrothermal solution. Chinga material can also occur as breccia matrix and the texture is typically massive, veins are often overprinted by later phase mineralisation. This material was probably derived from pulverised Cretaceous black shales and has been emplaced soon after the diatreme eruption. The chinga veins are most common in the upper levels of the Cârnic and Cetate dacites. The grade of chinga material is typically 0.5 to 3g/t Au.

2. Quartz-carbonate-sulphide veins

These veins are associated with the main mineralising phase at Rosia Montana. Veining occurs as multiphase, quartz-carbonate-sulphide (pyrite) ± adularia and rare base metals, texture varies widely from massive to banded, often vuggy and from very fine grained to coarsely crystalline.



Carbonate dominated veins are interpreted to have been deposited late in this mineralised phase and are often associated with base metal sulphides. Carbonates are predominately calcite and rhodochrosite, with rarer dolomite and siderite. Base metal sulphides are dominated by galena, low Fe sphalerite and chalcopyrite. The veins are typically narrow (up to 30cm) and more common in the deeper levels at Cârnici and Cetate. This also appears to be the main mineralising phase in the Igre Cretaceous sediments and also the commonest vein type around the margins of the Black Breccia. Textures are typically banded and vuggy, sometimes massive. The very last phase is dominated by calcite with marcasite and very little gold or silver mineralisation. Two tellurides occurrences in rhodochrosite-rhodonite high grade silver veins were recently made (Tamas et al., 2004, Cook et al., this volume).

3. Clay veins

These ‘veins’ are generally considered late stage, possibly retrograde. Locally referred to as veins as they contain some gold-silver mineralisation many of these could be classified as shear or fault structures. Clays vary from illite-smectite to illite and often occur with pyrite and marcasite. They usually display a massive or sheared texture.

Veins are usually discontinuous, widely spaced and only millimetres to centimetres in width, but have been observed up to 1m wide. They generally become less frequent, but thicker and more continuous at depth.

Resources and Reserves

Recent exploration of the deposit by Gabriel Resources has outlined a measured and indicated resource of 352.27 million tonnes at an average grade of 1.3 grams per tonne gold and 6.0 grams per tonne silver for 14.6 million ounces of gold and 82 million ounces of silver. The total resource (measured, indicated and inferred) is 400.41 million tonnes for a total contained resource of 16.1 million ounces of gold and 73.3 million ounces of silver (using a 0.6g/t Au cut-off) (Gossage 2003). Proven and probable reserves total 217.96 million tonnes at an average grade of 1.52 grams per tonne gold and 7.5 grams per tonne silver for total reserves of 10.6 million ounces of gold and 52.3 million ounces of silver (using a 0.6g/t Au cut-off).

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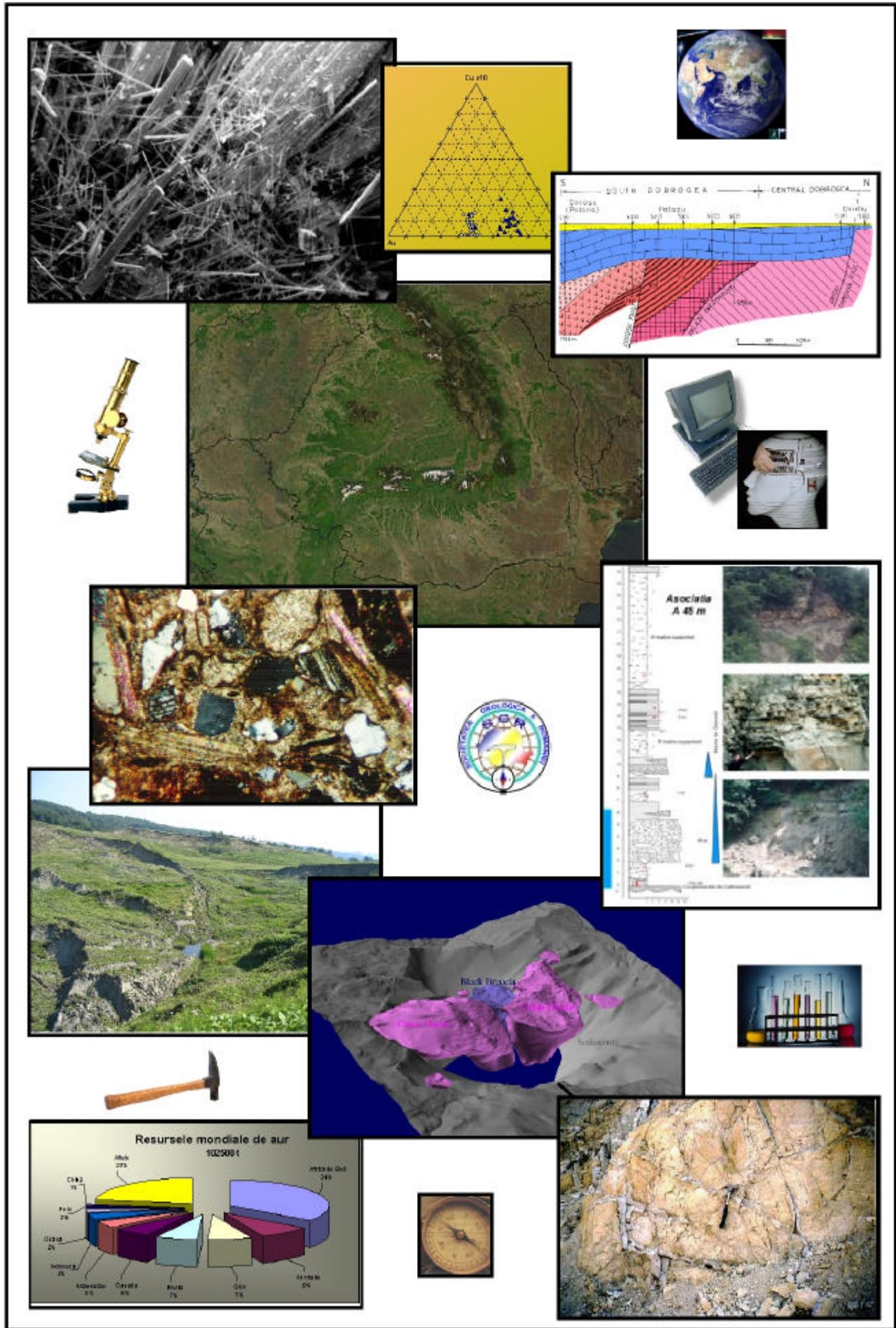
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