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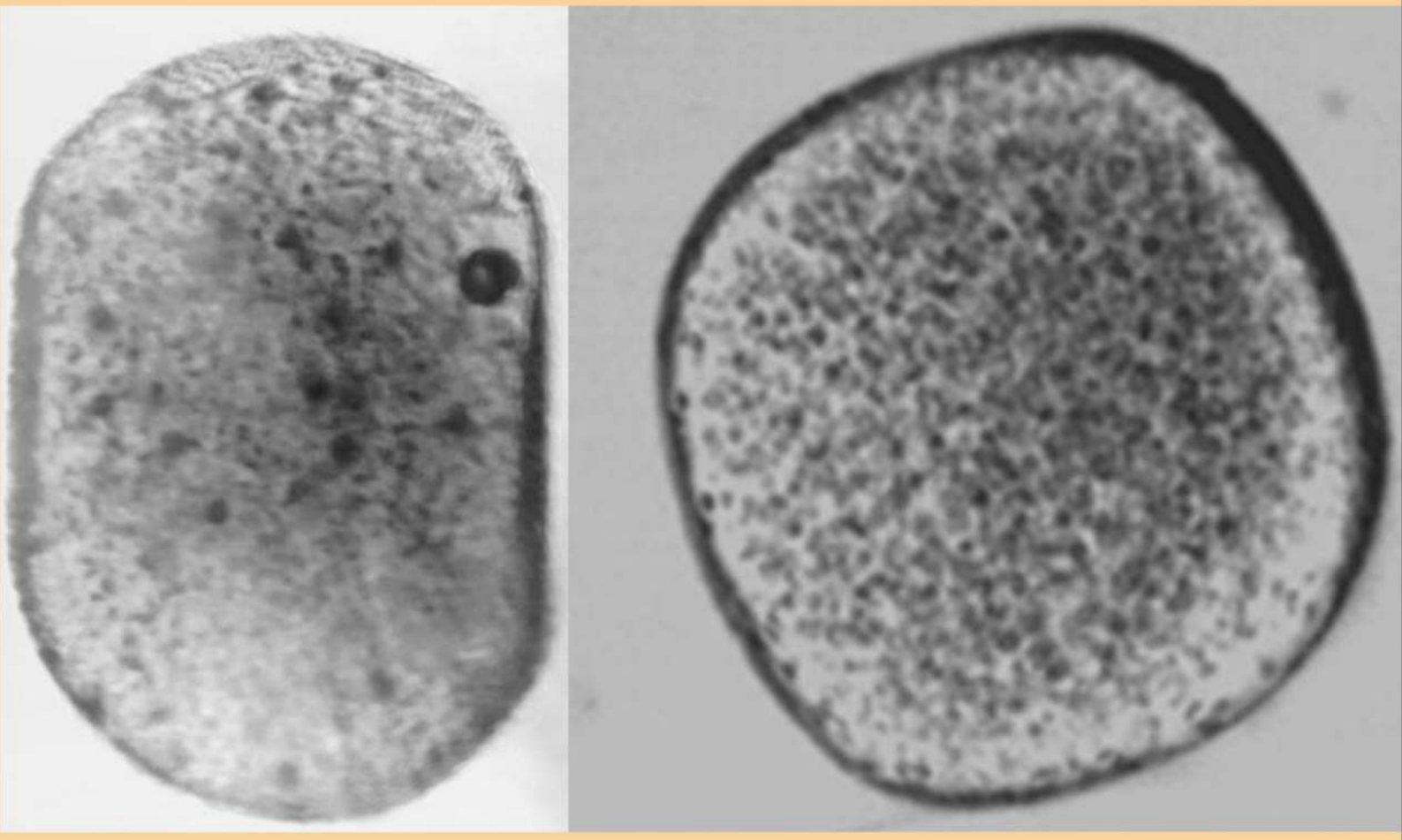

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Cover illustration: Foam-like melt inclusions in Dej tuffs. The base of each picture is 100 μm. Pictures taken by Ioan Pintea.

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New petrified woods from Căprioara Valley, Feleacu Hill, Cluj-Napoca, Romania

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Abstract: This paper presents the palaeoxylotomical study of some lignofossils collected from Căprioara valley, Feleacu Hill area, close to Cluj-Napoca, Romania, from a sedimentary formation of Early Sarmatian age, as indicated by the mollusks and foraminifers associations. From that area, a small Early Sarmatian Flora – known as "Feleacu Flora" – with species of *Pinus, Sequoia, Quercus, Daphnogene*, *Platanus, Engelhardia, Phragmites*, and *Cyperites*, was previously described. The new identified lignotaxa, as morphospecies of *Chamaecyparixylon, Sequoioxylon, Cinnamomoxylon, Rhysocaryoxylon,* and *Quercoxylon* confirm and complete the knowledge on Feleacu Flora, indicating also a mesophytic forested riverside and a warm temperate climate, thus making possible better paleoenvironmental and palaeoclimatic reconstructions.

***Keywords*:** Early Sarmatian, Feleacu Flora, petrified wood, palaeoclimatic significance

1 Introduction

A new collection of petrified woods coming from the area of Feleacu Hill (Fig. 1), close to Cluj-Napoca, Romania, was submitted to paleoxylotomical study in order to identify the original trees. The fossils were collected, partially by Iustinian Petrescu and Mircea Nicorici, and partially by Mirela Dumitrescu-Sabou, and are kept in the collection of the Geological Museum of the Faculty of Biology and Geology, Babeş-Bolyai University (UBB) of Cluj-Napoca.

As it is well known, at least in the south-western part of Feleacu Hill, the Early Sarmatian sediments present a littoral-neritic facies, being constituted of marly to sandy clays with intercalations of sands or sandstones with concretions, sometimes with pebble levels, overlaying the terminal Badenian sediments, probably after a gap of sedimentation (Givulescu, 1997).

The age of those sedimentary deposits is given by a brackish mollusk fauna, the list, revised by Givulescu (1997), indicating the Early Sarmatian: *Irus (Papirus) gregarius gregarius, Cerastoderma vindobonense vindobonense, Gibula (Rolandiana) picta, Gibula (Gibula) pusilla, Pirenella picta picta, Cerithium (Thericium) rubiginosum rubiginosum*.

In addition, fish remains and brackish foraminifers (Popescu, 1995) have also been described from that area. Suciu (2005a, b) made an areal study on the sedimentary rocks from around Cluj-Napoca, including Iris and Feleacu hills and the top of Lombi hill, which consist of already known Badenian and Sarmatian deposits (see Răileanu, 1955, 1959; Moisescu & Popescu, 1967; Mészáros & Clichici, 1988, Filipescu, S., 1999).

On the geological map edited in 1968 by the Geological Institute of Romania, at the scale 1:200,000, Cluj sheet (Răileanu & al., 1968), the Sarmatian deposits were assigned to the "Buglovian", as Earliest Sarmatian, and to Volhynian-Basarabian, but without reliable arguments. Based on biostratigraphical evidence (i.e. micropalaeontological content), the age of Miocene sediments from Transylvania was documented by Popescu (1995, 2000), Filipescu et al. (2005, 2008), Krézsek and Filipescu (2005) and Tóth et al. (2010).

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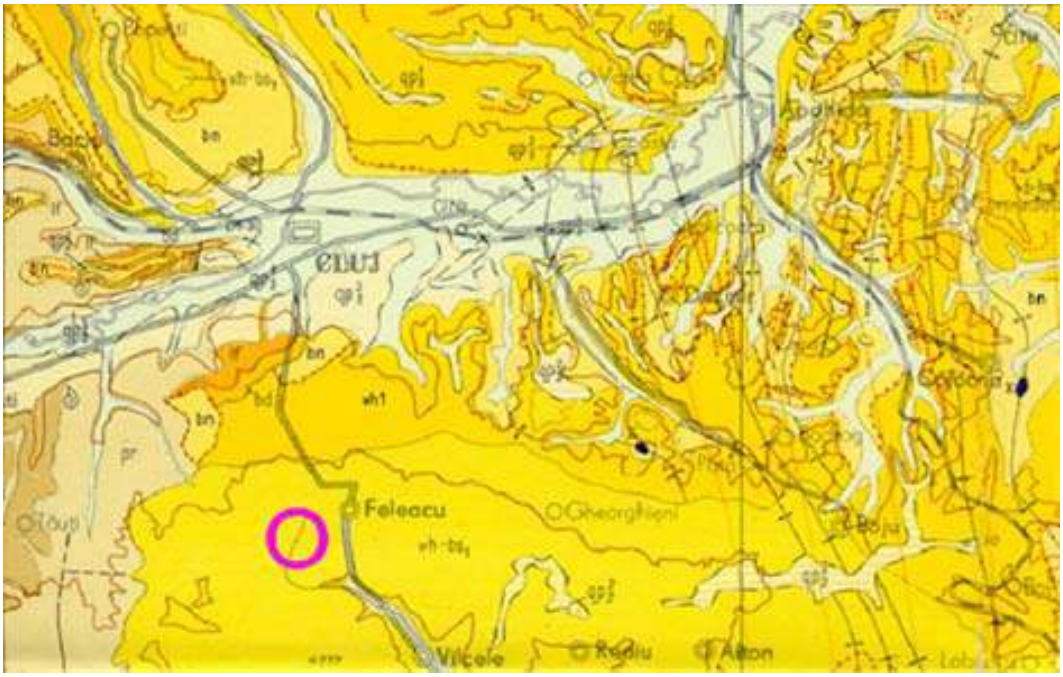

Fig. 1. Location of Căprioara Valley fossiliferous area (marked with red circle), on Cluj-Feleacu map. Excerpt of IGR Geological map 1:200000, Cluj sheet.

Nanno -zone	Chronostrati-graphical Units		Lithostratigraphical Units from NW of Transylvania Depression (Filipescu et al 1999)		Regional Biostratigraphic Units (Foraminifera biozones) Popescu, 1995, 2000)	
NN9 NN8 12	Pannonian s.s.		Lopadea formation / Guşterița formation		*Ammonia* acme (+ostracods) rare foraminifers	
NN7 13	Sarmatian	Basarabian	Feleac formation	Dobârca formation	*Porosononion aragviensis*	
					Dogielina sarmatica *	
		Volhinian	Iris formation		*Schakoinella imperatoria* *	
					Elphidium reginum *	*Articulina sarmatica*
					Varidentella reussi	
					Anomalinoides dividens	
NN6	Badenian		Mireș formation / Pietroasa formation		*Valpertina*	

Fig. 2. Late Mid-Miocene – Early Late-Miocene litho-biostratigraphic column in Transylvanian Basin, Cluj-Napoca area (modified after Suciu, 2005). * - plant remains, including fossil wood.

Around Cluj-Napoca, on Feleacu hill, some foraminifer assemblages were identified (Fig. 2): the Biozone with *Elphidium reginum,* representing the Late Volhynian and the Biozone with *Dogielina sarmatica,* representing the Early Basarabian (=Feleac Formation). These Elphidiids associations are typical for littoral facies and the association of the Rotaliids with the red algae thalli indicates brackish shallow waters, agitated, probably in an intertidal area. The sandy sediments with carbonate mud interlayers at several levels also suggest a littoral environment (Suciu, 2005a, b).

From the study site in Căprioara valley, located in the southwestern part of Feleacu Hill (Fig. 1), in some littoral-neritic sediments considered Early Sarmatian in age, Suciu (2005a, b) reported *Laevigatisporites haardti* (Polypodiaceae spores), and pollen of Taxodiaceae, of *Pinus, Abies, Tsuga, Picea* and *Ulmus, Carya, Pterocarya, Engelhardia, Betula, Quercus, Myrica, Alnus,* which indicate a mesophitic vegetation of warm temperate climate characterized by a mean annual temperatures (MAT) of 15°-19°C, and variable precipitations.

At Feleacu Hill, from the mentioned Sarmatian formation, Staub (1883, 1891) and Szadeczky (1917) described some vegetal taxa from levels with plants remains as leaf or fruit imprints. These have been completed and revised by Givulescu (1997). These older fossiliferous sites could be not identified during our study. Givulescu (1997) mentioned that he collected fossil material, in the course of time, from some small stone quarries located along the road Cluj Napoca – Turda, difficult to identify nowadays. Anyway, all the taxa previously described were revised by Givulescu (1997), resulting in a list of a small mixed flora – named by him "Feleacu flora".

This flora contains:

Cystoseirites partschi Sternberg
Pinus hepios Unger
Pinus felekiensis Staub (small branch)
Sequoia abietina (Brongn.) Knobloch
Abies cf. *alba* Miller
Engelhardia macroptera (Brongniart) Unger [not *Palaeocarya macroptera* (Unger) Mai]
Daphnogene polymorpha (Al. Braun) Ett.
Laurophyllum sp.
Ulmus pyramidalis Goeppert
Platanus cf. *leucophylla* (Brongn.) Knobloch
Acer tricuspidatum Brongn
Fraxinus sp.
Phragmites oeningensis (Al.Braun)
Cyperites senarius Heer
Phyllites sp.

It indicates a lowland mesophytic forested riverside and a warm temperate climate, as suggested by the predominating arctotertiary elements, with leaves having dentate margin.

The environment of sedimentation, at least locally, could be a sandy beach, taking into account the presence of *Cystoseirites*, a fossil Algae. A mountainous region probably was not very far, as indicated by the presence of *Abies* (Givulescu, 1997). The physiognomic analysis of this material concluded that the major part of the identified taxa were trees (83.33%) with predominant arctotertiary taxa (~75%), but the small number of described taxa was not enough to obtain a reliable paleoclimatic image for the time of that living vegetation (Givulescu, 1997).

The material studied here, represented by fragments of petrified trunks, was collected from Feleacu hill area, Căprioara valley, from the sedimentary deposits presented above. It seems that, based only on a preliminary unpublished study, Petrescu identified from this material some Conifers (*Sequoioxylon, Pinuxylon*) and some Dicots (*Quercoxylon, Laurinoxylon, Ebenoxylon, Rhisocaryoxylon*) mentioned later as a simple list (Petrescu, 2003, p. 115). Those taxonomic identifications have never been validly published and it is very possible that we had the chance to study, at least a part of that material, kept now in the Collection of the Geological Museum of the Biology and Geology Faculty of Babeş-Bolyai University of Cluj Napoca, registered under the specification "silicified wood, Feleacu, Cluj/Căprioarei valley, with the inventory numbers: 369, 370, 371, 375, 376, 377, 384, 385, 386, 387, 388, 389, collected by Petrescu and Nicorici". Two of the samples studied here were collected by Mirela Dumitrescu-Sabou and are deposited in the same collection (inv. nrs. 794, 795).

An adequate methodology of preparation and study of the petrified wood - the paleoxylotomy – was used, cutting standard oriented sections of petrographic type followed by microscopical observations on the oriented thin slides and a comparative study with similar structures published in the specialty literature in order to identify the original trees and their extant equivalents.

2 Systematics

Gymnosperms - Conifers

Family **Cupressaceae** Bartlett
Subfamilly **Cupressoideae** Rich ex Sweet
Genus ***Chamaecyparixylon*** Chudajberdyev
Chamaecyparixylon cf. *polonicum* (Kräusel) Chudajberdyev, 1958
Plate I, figs. 1-9.

Macroscopic description

The studied material was collected from Căprioara valley, southwestern part of Feleacu hill, from some littoral-neritic sediments considered Early Sarmatian in age and is now deposited within the Geological Museum of Department of Geology and Paleontology of UBB Cluj Napoca, under the numbers 376 and 386. The first specimen (nr. 376) is represented by a piece of silicified wood of centimetric size (11.2/5.8/1.8 cm). The second (nr. 386) is represented by 11 variably sized fragments, probably sampled from a big trunk, the biggest one having 9.5/5/2.5 cm, light grey-brownish color and fibrous texture.

Microscopic description

Growth rings - distinct, varied in width, with well-developed early-wood in gradual transition to the obvious thick-walled late-wood (5-9 tangential rows). No resin duct is present.

Tracheids with polygonal cross section, are sometimes slightly deformed, their lumina is polygonal with rounded corners and with 25-45 μm in diameter in the early-wood because they are slightly thick-walled (5-8 μm the double wall); in the late-wood, the cross section of the tracheids is smaller, the cellular walls thicker (8-12 μm the double wall); there are 2-7 radial rows of tracheids between two rays and the ray density is 1750-2050 tracheids on sq. mm. Tangential pitting is rarely present, as small pits (6-9 μm) with spaced arrangement and somewhat irregular. Radial pitting is spaced arranged in one single vertical row of round rather small pits with 8-10 μm in diameter and round apertures. Crassulae were not observed. The tracheids have rounded ends and are slightly curved.

Axial wood parenchyma is present in cross section, diffuse or slightly irregularly arranged in short tangential uniseriate groups, usually in the transitional wood. In the longitudinal sections it appears as thin walled rectangular cells separated by knotted horizontal thin walls with 3-6 nodules, and frequently with small balls or plugs of resin.

*Rays***,** in cross section view, are uniseriate and rectilinear or slightly curved, especially at the late-wood. In tangential view, short rays appear, having 1-9, usually 2-4 round to oval cells, without lateral spaces and with density of 8-12 rays on a tangential millimeter. In radial section, the ray cells are homogeneous, having 10-12 μm in height, the marginals slightly taller, with horizontal walls relatively thin and smooth, tangential walls thin and inclined; no indentures have been seen. Cupressoid cross fields are difficult to observe, with 1-2(3) cupressoid small pits, horizontally arranged or in vertical pairs, rather badly preserved, especially in the marginal fields.

Affinities and discussion

The xylotomical observations on the studied specimens allowed the outlining of a combination of features common for conifers devoid of normal resin ducts, like the Cupressaceae, whose morphology, structure and taxonomy were well studied in the recent time (see Gadek et al., 2000; Earle, 2013). Among them, the extant genera *Cupressus* and *Chamaecyparis* (see Greguss, 1955, 1967) are similar with our studied material, regarding the aspect of the tracheids, of the short rays, of the cross fields and of the axial parenchyma.

Compared with previously described similar forms attributed to the fossil correspondent morphogenus *Cupressinoxylon,* (see Greguss, 1967; Zalewska, 1953), or to *Chamaecyparixylon* (see Iamandei and Iamandei, 2000a) we confirm some general resemblances. Consulting the generic diagnoses revised by Kräusel (1949), Vogellehner (1967, 1968) and using the key of identification proposed by Vaudois and Privé (1971), it seems that we could assign our studied material to *Chamaecyparixylon* morphogenus, which belongs to Thujidae group, where the distinction between *Chamaecyparixylon* and *Thujoxylon* is difficult to be done, being reduced to the presence or absence of the parenchyma and of the indentures. As our specimens have rather abundant axial parenchyma and indentures were not seen, it is correct to attribute them to the *Chamaecyparixylon* morphogenus.

Beside these features, we noticed the rounded ends of the tracheids, which are slightly curved terminally and the low rays, resembling with the features of *C. polonicum* of Chudajberdyev (see

Vaudois and Privé, 1971; Iamandei and Iamandei, 2000a). Since it is difficult to talk about total identity, on this basis we confer our studied material to this species and we name it *Chamaecyparixylon* cf. *polonicum* (Kräusel) Chudajberdyev 1958.

Subfamily **Sequoioideae** (Luerss.) Quinn
Genus ***Sequoioxylon*** Torrey
Sequoioxylon gypsaceum (Goeppert) Greguss, 1967
Plate II, figs. 1-9.

Macroscopic description

The studied material was found in Căprioara valley, southwestern part of Feleacu hill, in Early Sarmatian littoral-neritic sediments and is now deposited in the Geological Museum of Department of Geology and Paleontology of UBB Cluj Napoca, under the numbers 384, 389 and 794. The sample 384 is represented by four pieces of silicified wood of centimetric size (the biggest has 22.2/9.5/6.5 cm); 389 is represented by five pieces of silicified wood of centimetric size (one of 5.5/4.5/1.2 cm and four smaller); 794 is represented by a trunk fragment, slightly larger (30/13/5 cm). All pieces probably come from larger trunks and have light grey-brownish color and fibrous texture.

Microscopic description

The growth rings are distinct, marked by 5-9(13) thicker-walled cells tangentially flattened in the late-wood. The early-wood has cells with big lumina and thinner walls. The rings are up to 60 cells thick, and the transition from early to late-wood is gradual. Normal resin ducts are absent.

The tracheids are polygonal in cross section, usually rectangular in the late-wood, with polygonal-rounded lumina of 30-60/25-35 μm radial/tangential diameters, more flattened tangentially in the late-wood (7-15/18-25 μm). The cellular wall is relatively thick in the early-wood (4-6 μm for the double wall), thicker in the final-wood (7-12 μm). There are 1-6 radial regular rows between two rays, some of them thinner. Their density is 806-930 cells on sq.mm. The tangential pitting, badly preserved, is rarely visible as small bordered pits of 7-9 μm in diameter having small round apertures, spaced or slightly irregularly arranged in a single vertical row. The radial pitting is badly preserved, is of abietineous type, and the pits have 8.5-11 μm in diameter and round apertures of 3.5-5 μm in diameter, show spaced arrangement in a single vertical row on the tracheids of transitional and late-wood but on the bigger tracheids of early-wood are arranged in 1-2(3) vertical rows, rather close to one another but not contiguous and with obvious *crassulae*.

The axial parenchyma is visible in cross section, as is filled with resin and diffusely distributed amongst the tracheids. It is also evident in the longitudinal sections (especially the tangential ones), as vertical strands of rectangular, thin-walled cells, of 1-2 μm the simple wall. The transverse end walls (horizontal walls) are thin (1-3 μm), usually nodular, with 1-3 nodules. The resin content appears as small or large glomerules and/or plugs.

The rays seen in cross section are linear, uniseriates in the early-wood, constituted of rectangular cells sometimes laterally crushed. Vertically, they are uniseriate, rarely with short biseriations, and are of 1-25, most frequently up to 18 cells tall. Tangentially seen, the ray cells are circular or vertically elliptic and usually empty. The density is of 12-17 rays on tangential mm. The rays are homogeneous, the parenchyma ray cells are 15-20 μm high, and the marginals sometimes taller (20-24 μm); the horizontal walls are relatively thick (2-4.5 μm double wall), smooth and probably pitted, the marginals with outer wall slightly wavy. Their tangential walls are usually thin (0.5-1.5 μm), smooth, oblique or slightly arcuate, difficult to observe because of poor preservation. Indentures were not seen. The cross fields are usually blurred, pitting is indistinct, as hardly can be guessed 1-2(3) blurred taxodioid pits 5-7(9) μm in diameter, with rounded to oval borders and small lens-like apertures, horizontally or slightly diagonally arranged.

Affinities and discussions

The xylotomical details observed on the standard oriented slides prepared from the studied specimens, suggest a conifer of "taxodiaceous" type, very similar to *Taxodioxylon* and *Sequoioxylon* genera. The distinction between the two genera seems to be difficult since the validity of *Sequoioxylon* genus, created by Torrey (1923), is contested by some palaeoxylolgists who consider that the diagnosis of *Taxodioxylon* genus (Kräusel, 1949) is comprehensive enough to include the

xylotomical features of all "taxodiaceous" woods (Prive, 1977). More recently, Dolezych et al. (2011) supposed some confusion, concerning the modern affinities, inside *Taxodioxylon* as fossil morphogenus. Prive (1977) identified this morphogenus in some Tertiary deposits, as a widespread taxon during the Tertiary in North Hemisphere (in Eurasia also, like in North America) with very diverse forms assigned to *Taxodioxylon gypsaceum* (Goeppert) Kräusel 1949, considering it as a correspondent of the extant *Sequoia sempervirens* (D.Don) Endlicher.

Evaluating the xylotomy of our studied specimens, regarding the distinct growth rings with thin walled tracheids in the early-wood, resinous parenchyma, rays of 2-18 cells high or more, tracheids with abietineous pitting in 1-2(3) vertical rows, with crassulae, and the horizontal walls in axial parenchyma that is thin and smooth, or weakly nodular, the cross fields with 1-3(5) taxodioid pits in 1-2 horizontal rows it can be observed the similitude with the extant *Sequoia sempervirens*.

Our specimens present distinct *growth rings*, up to 60 cells thick in gradual transition and without normal resin ducts; *tracheids* with polygonal cross section in the early-wood, badly preserved or absent tangential pitting, abietineous radial pitting arranged in 1-2(3) vertical rows, rather close to one another but not contiguous and with obvious *crassulae*; axial parenchyma with thin and smooth or weakly nodular horizontal end walls, and with resin content as glomerules or plugs; uniseriate rays, rarely with short biseriations, of 1-25 cells in height, more frequently up to 18 cells, their density of 12-17 rays on tangential mm, homocellular, with procumbent body cells, marginals taller, indentures not seen, taxodioid cross fields, with 1-2(3) blurred taxodioid pits, 3-5 in the marginal fields.

Comparing all these features with those of the forms of *Sequoioxylon gypsaceum* (Goeppert) Greguss, with forms described by Greguss (1967) and also by Petrescu & Dragastan (1971); Petrescu and Popa (1971); Petrescu and Nuţu (1971); Petrescu (1978); Iamandei (2002); Iamandei & Iamandei (2000); Iamandei et al. (2005, 2008a, 2012), we found numerous similarities; therefore, these studied specimens will be assigned to the fossil morphospecies *Sequoioxylon gypsaceum* (Goeppert) Greguss 1967, as a fossil correspondent of the extant genus *Sequoia*.

Angiosperms – Dicotyledons

Family *Lauraceae* Jussieu
Genus *Cinnamomoxylon* Gottwald, 1997
Cinnamomoxylon intermedium (Huard) Gottwald, 1997
Plate III, Figs. 1-9.

Macroscopic description

The studied material, represented by 3 specimens, was found in Căprioara valley, southwestern part of Feleacu hill, in some Early Sarmatian littoral-neritic sediments and is now deposited in the Geological Museum of Department of Geology and Paleontology of UBB Cluj-Napoca, under the numbers 369, 371, and 375. They have centimetric size: the sample 369 is represented by a piece of 14/9.5/4.5 cm and six smaller fragments; 371 is a piece of 7/3.8/2.5 cm and a smaller one; 375 is a piece of 10.5/7/3 cm and a smaller one. The samples show light brown to grey color, locally with yellowish tints, a fibrous texture with vessels, broad rays and visible growth rings suggesting a dicot wood structure.

Microscopic description

Growth rings, distinct, relatively wide, with distinct boundaries, defining a diffuse-porous to slightly semi-ring-porous wood structure.

Vessels - solitary or in radial multiples of 2-3(5). In cross section, the solitary pores have circular shape to slightly oval and thick walls (6-9 µm the double wall). The radial/tangential diameters for the big solitary pores in the early-wood are of 100-200/60-160 µm and, in the late-wood of 40-80/40-70 µm. Their density is 9-13 pores per mm^2. Sometimes, radially, ad- and abaxial to the vessels, vascular tracheids appear. The vessels have exclusively simple perforations on horizontal or tilted plates, and alternate, numerous, small, bordered pitting. The pits of 3-4 µm in diameter do not touch each other and have point like apertures. Some vessels have small and big thick-walled tyloses, sometimes with sclerotic aspect. Inside the vessel lumina on the cell walls, thickenings can appear, and also dark granular deposits or spherical agglomerates, more or less grouped, but not compact. The vascular elements are short and have 150-250 µm in length.

Axial parenchyma is abundant, of paratracheal - vasicentric type, as complete 1-3-seriate sheaths, but sometimes incomplete, when the rays touch vessels; also appear diffuse or as fine

bands, 1-2-seriate, in the late-wood. Vertically, the parenchyma cells do not have very thin walls (2 μm); longitudinally viewed they appear storied in strands of 4-12 rectangular cells, especially when adjacent to vessels or in the final wood. The vasicentric polygonal parenchyma cells are minutely pitted: pits of 2-3 μm, opposite, in horizontal rows or slightly irregular. Sometimes solitary crystals appear inside cells.

Rays - appear mostly uniseriate in cross section, but also bi and triseriate, composed of rectangular cells radially elongate; their trajectory is linear, sometimes touching vessels, slightly curling near the big vessels. They are not too tall, and have 3-10 cells in height (30-120 μm); the biseriate rays have 6-12 cells (100-152 μm in height). Their density is 12-14 rays on tangential horizontal millimeter. The rays are heterogeneous, with body cells all procumbent, followed by 1-2 rows or squared cells, of 16-20 μm in height, and marginal upright hypertrophied oil cells, tall, as flames, all pitted in the cross fields with vessels. The pits are big, oval, similar to those on the vasicentric parenchyma, round, small (3-3.5 μm), 1-3 in horizontal rows or slightly irregular, sometimes 2 superposed, or in vertical row of 3-4 on upright marginal cells.

The libriform fibres – slightly polygonal in cross section (12-25 μm), with rounded polygonal lumina and relatively thin walled (3-5 μm the double wall), are arranged in 2-9 radial regular rows, with small intercellular spaces. Longitudinally, the fibre cells are short, septate and relatively storied arranged in the early-wood. On the vertical walls there are rare, minute bordered pits, quite irregularly arranged on vertical rows.

Affinities and discussions

All the three specimens have similar xylotomical features, regarding: the aspect of the vessels, solitary and as short radial multiples in transverse view, thick-walled and with simple perforations; the aspect of the bi-triseriate rays, rarely broader, and with typical big oil-cells in radial view; and the vasicentric-parenchyma.

These features are characteristic to Lauraceae Family. The presence of oil-cells in rays and their absence in parenchyma suggests Perseoideae Group (Metcalfe and Chalk, 1950; Greguss, 1959). As it is known, the extant lauraceous species are spread mainly in the warm regions: tropical America, Brazil included, Southeast Asia, Australia and the Pacific islands. *Laurus* is the only European genus, appearing in the Mediterranean region.

Otherwise, Lauraceae is a rich family, having about 54 genera with 2000-2500 species. Among them, the most known are *Persea americana* by its fruit (Avocado pear), *Cinnamomum* spp. by cinnamon and camphor-oil, *Lindera* by the aromatic oil named benzoin, or *Sassafras*, with a fragrant wood used in cabinet-making. Xylotomically, two groups were admitted: Lauriodeae and Perseoideae, the first with tangential bands of parenchyma and only simple perforated plates, the second devoid of banded parenchyma and with scalariform or simple perforated plates (see Watson and Dallwitz, 1992).

Fossil woods of Lauraceae were described from the XIXth century, under different names, and were the object of many discussions and revisions. Names as *Laurinium* Unger was used for lauraceous devoid of secretory cells wood, (see Huard, 1967, p. 88), and also *Ulminium* Unger (Wheeler, Scott and Barghoorn, 1977; Privé-Gill and Pelletier, 1981). They remained in use but have been, gradually, abandoned for *Laurinoxylon*, name that shows clearly the appurtenance to Lauraceae Family. Süss and Mädel (1958) are those who have established the genus name for the lauraceous fossil wood as *Laurinoxylon* Felix 1883, the other lauraceous generic names falling into synonymy.

The xylotomical description of our studied specimens corresponds largely to the morphogenus *Laurinoxylon* Felix, to which we assigned the studied material. When we compared their features with those of the already described fossil species, we found many similarities, but also dissimilarities; therefore, it is difficult to identify it with one of the known species. These are the main characteristics:

- Vertical or radial secretory ducts appear in: *Cinnamomoxylon sp. 1-4*, *Laurinoxylon palfalvyi* Greguss, *Laurinoxylon* sp., all of them described by Greguss (1969);
- Terminal parenchyma and crystals in oil-cells appear in: *L.* cf. *hasenbergense* Süss, 1956;
- Aliform to confluent parenchyma appear in: *L. variabile* Priv**é**-Gill and Pelletier, 1981;
- Scalariform or simple and scalariform perforations appear in: *Laurinoxylon* sp. 2,

described by Petrescu and Blidaru (1972); *L.* (cf.) *litseoides* Süss, *L.* cf. *microtracheale* Süss (both quoted in Gottwald, 1992); *L.* cf. *compressum* Huard, 1967; *L. primigenium* Felix, 1894;

- Only simple in *Cinnamomoxylon intermedium* (Huard) Gottwald 1997; *Laurinoxylon bergeri* Süss and Mädel 1958; *L. neagui* Iamandei and Iamandei, 1997.
- Secondary phloem appears in: *L.* cf. *litseoides* Süss, *L.* cf. *seemanianum* Mädel (in Süss and Mädel, 1958); *Caryodaphnopsoxylon richteri* Gottwald 1992;
- Terminal fibre bands appear in: *Laurinoxylon cf compressum* Huard, 1967; *L. oligocenicum* Prakash, Brežinova and Awasthi, 1974;
- Oil-cells in parenchyma appear in: *Cinnamomoxylon* sp. 1-4, described by Greguss, 1969; *Laurinoxylon aniboides* (Greguss) Süss, 1956; *L. perseamimatus* Petrescu, 1978; *L. oligocenicum* var. *microporosum* (Mädel) Petrescu, 1978; *L. czechense* Prakash et al., 1971; *Caryodaphnopsoxylon richteri* Gottwald, 1992; *L. neagui* Iamandei and Iamandei, 1997;
- Terminal or subterminal botle-like oil-cells in rays appear in: *Cinnamomoxylon* sp.1-4 described by Greguss, 1969; *Laurinoxylon vadaszi* Greguss, 1969; *L.* cf. *californicum* (Platen) Süss, 1956; *L. müllerstolii* (Greguss) Süss, 1956; *L. süssi* Greguss, 1969; *L. neagui* Iamandei and Iamandei, 1997; *Cinnamomoxylon intermedium* (Huard) Gottwald, 1997.

Gottwald (1997) considered that genera such as *Perseoxylon* Felix, 1887, *Ocoteoxylon* Schuster, 1908 (see ING: ocoteoxylon) and *Machilusoxylon* (Bande) Ingle, 1978 (see insidewood), do not have enough xylotomic particularities to justify their separation from *Laurinoxylon* genus. Also, based on recent xylotomic studies on Lauraceae members, he admitted as valid genera coexisting with *Laurinoxylon* Felix, 1894, the new genera *Sassafrasoxylon* Brežinova and Süss, *Caryodaphnopsoxylon* Gottwald, 1992 and *Cinnamomoxylon* Gottwald, 1997, related to the extant correspondent taxa, suggested by name.

However, the list is open and new genera were described lately: *Mezilaurinoxylon* Wheeler and Manchester, 2002; *Olmosoxylon* Estrada-Ruiz, Martinez-Cabrera and Cevallos-Ferriz, 2010; *Paraperseoxylon* (Scott and Wheeler) Wheeler and Manchester, 2002; *Richteroxylon* Wheeler and Dillhoff, 2009; *Cryptocaryoxylon* Leisman, 1986 (see insidewood).

The fossil morphogenus *Cinnamomoxylon*, initially has been created by Greguss (1969), on a badly preserved material, without a good description and diagnosis for a type-species. Gottwald (1997) ignored this reality and described *Cinnamomoxylon* as a new morphogenus, taking *C. areolosum* Gottwald, 1997, as type-species. This is the genus diagnosis, given by Gottwald (1997): "Porous wood structure to half-ring-porous, with solitary vessels and in short radial multiples, with mean diameter of >80 µm, simple or simple and short scalariform perforated plates, alternate pitting of 8-12 µm in diameter, sometimes with thin-walled tyloses, rays usually slightly heterogeneous, of 2-4 cells wide and 250-450 µm high, cross fields of I/II type (Richter), secretory idioblastic oil-cells, usually variable marginal, obviously wider, vasicentric, aliform to confluent parenchyma, thin to moderately tick-walled fibers, usually unsepted, or very rare, sometimes secretory idioblast between fibers".

The extant correspondents of this genus are considered *Cinnamomumn* Schaeff., *Lindera* Thunb., *Litsea* Lam. and *Persea* Mill. Evaluating the already described species of *Laurinium/Laurinoxylon* Gottwald attributed to *Cinnamomoxylon* 12 species that "entsprechend vollständig und vergleichbar beschreiben und abgebildet" (Gottwald, 1997). In such conditions the morphogenus *Laurinoxylon* Felix comprehends all the fossil lauraceous wood that cannot be precisely attributed to a valid fossil genus, as it is specified upward. Gottwald (1997) considered that this genus is rather insufficiently defined by the original diagnosis of Felix (1883) when he described the type-species *L. primigenium* (wood with big solitary vessels, in pairs or in short; thin-walled or moderately thick-walled fibers in rather regular radial rows; parenchyma paratracheal aliform sometimes tending to confluent; rays usually multiseriate, the uniseriates with typical shaped cells; secretory cells sometimes between the fibers or within rays appear) even when the author repeated it in his papers (Felix, 1887a, b), when he described *L. aromaticum* and *Perseoxylon antiquum*, respectively.

Our specimens show the following features:

- diffuse-porous to slightly semi-ring porous structure, with meansized pores - from 100-200 μm radial diameters in the early-wood, lesser in the late-wood, thick-walled, solitary or in short radial multiples, density 9-13 vessels on sq. mm.; short vessel elements, exclusively simple perforations and alternate, numerous, small, bordered pitting (3-4 μm), pits not touching each other and with point like apertures; small and big thick-walled tyloses inside some vessels sometimes, with sclerotic aspect, and dark granular deposits or as spherical agglomerates;
- vasicentric-parenchyma, abundant as complete 1-3-seriate sheaths and in fine bands, 1-2-seriate, in the late-wood, without oil-cells; longitudinally viewed parenchyma appear storied, in strands of 4-12 rectangular cells, especially when adjacent to vessels or in the final wood; the vasicentric polygonal parenchyma cells are minutely pitted: pits of 2-3 μm, opposite, in horizontal rows or slightly irregular; sometimes solitary crystals appear inside cells;
- rays mostly bi- and triseriate, of average height, heterogeneous, with body ray cells all procumbent, plus 1-2 rows of squared cells, and one marginal row of upright, tall, bottle-like oil-cells, pitted in the cross fields with vessels in 1-3 in horizontal rows or slightly irregular, sometimes 2 superposed, or in vertical rows of 3-4 on upright marginal cells;
- septate fibres thick-walled and bordered pitted.

The presence in our studied specimens of the marginal upright, big, bottle-like oil ray cells, visible especially in radial sections, sustained by other xylotomical features, regarding the shape and the distribution of the vessels, fibers and parenchyma in cross sections, the perforations and intervascular pitting, the pitted parenchyma and septate pitted fibres are typical features for the lauraceae of Perseoideae group, represented as fossils by the morphogenus *Cinnamomoxylon* as it was defined by Gottwald (1997).

Consulting also InsideWood 2004-onwards and Wheeler (2011), it seems that we face a species of *Laurinoxylon* of Perseoideae type, therefore, the attribution to *Cinnamomoxylon* Gottwald, 1997 is correct, taking into account the above comparative analysis.

The comparative study of our material with many described species led our attention to the European Tertiary species with simple perforation plates, like *Laurinoxylon bergeri*, *L. endiandroides, L. litseoides*, *L.microtracheale* (see Süss and Mädel, 1958) described from the Tertiary deposits of Hasenberges (close to Wiesa, in Saxony, Germany); *L. intermedium* Huard (1967), from some Miocene lignite of Arjuzanx (Landes, France); *Laurinoxylon* cf. *microtracheale* Süss, found in 1992 by Gottwald in the Late Eocene of Helmstedt (Niedersachsen), *Laurinoxylon compressum* Huard 1967, *Laurinoxylon czechense* Prakash, Březinova & Bužek 1971 from northern Bohemia, Czech Republic, *Laurinoxylon ehrendorferi* Berger 1953 from Limnos (Greece), and also similar to the forms described by Petrescu and Lazăr (1970), Petrescu and Nuţu (1971), from Apuseni Mts., and Petrescu and Blidaru (1972) from NW Transylvania, we could assign the here studied material to a species with most similar features described initially by Huard (1967) and revised by Gottwald (1997), which we named *Cinnamomoxylon intermedium* (Huard) Gottwald 1997, having an extant correspondent in *Cinnamomum,* a tree widespread now in South-eastern Asia (Vietnam, India, Ceylon, Indonesia), a warm and wet region dominated by monsoons.

Familia **Juglandaceae** DC. ex Perleb
Genus ***Rhysocaryoxylon*** Dupéron, 1988
Rhysocaryoxylon transylvanicum Iamandei and Iamandei, 2003
Plate IV, figs. 1-9.

Macroscopic description

The studied material was found in Căprioara valley, southwestern part of Feleacu hill, in Early Sarmatian littoral-neritic sediments and is now deposited in the Geological Museum of Department of Geology and Paleontology of UBB Cluj Napoca, under the numbers 377 and 795. Each specimen is represented by pieces of centimetric sizes, most probably sampled from bigger trunks: 377 has 7 pieces (the biggest of 5.5/2.5/3.7 cm); 795 is a single piece of 15/10/2.5 cm. Both samples show light brown to grey color, locally with yellowish tints, a fibrous texture with vessels, broad rays and visible growth rings, suggesting a dicot wood structure.

Microscopic description

Growth rings have indistinct boundaries, but the structure is semi-ring-porous; the vessels tend to gradually decrease their lumina to the late-wood. The tangential parenchyma bands are long, regular and with slightly decreasing interval.

The vessels, in cross section, appear as solitary pores or grouped in small radial multiples of 2-3(5); the solitary pores have usually radial oval shape, commonly deformed and thick-walled (4-8 μm the simple wall). The solitary pores have 100-180/80-160 μm the radial/tangential in the early-wood, but in transitional to the late-wood their size is smaller (60-80 μm). Vessel density could be 20-30 vessels on sq. mm. The vessels have simple perforations on tilted plates and bordered alternate small pitting, pits of 8-10 μm in diameter and point-like apertures. The vascular elements are relatively short (150-200 μm). Often, big tyloses occupy the inside of the vessels.

Axial wood parenchyma appear in cross section as long tangential bands of (1)2-3 cells thick but of paratracheal type also, even if are fewer, as some flattened cells around the vessels; the parenchyma cells have 10-15(18) μm in diameter. In longitudinal view, the apotracheal parenchyma, sometimes chambered and slightly hypertrophied, bears crystals floating in a white-bright content. Nevertheless, almost all parenchyma, axial or radial, have normal big cells filled with dark content and have numerous polygonal or rounded, variably sized, smaller crystals in 1-2 rows, or irregularly arranged.

The medullary rays, in cross section, appear with linear or slightly curled trajectory, sometimes touching vessels and are formed from rectangular cells radially elongated and usually filled of dark matter. Tangentially seen, the major part of the rays are 1-2-3(-5)-seriate and are moderately tall, and appear fusiform with endings of 1-4 cells or more. The ray density is of 12-14-26 rays on horizontal tangential millimeter. Radially seen, the rays appear slightly heterogeneous, the body cells are procumbent, rather big, marginally taller or squared, and usually have dark content and numerous small or variably sized rounded crystals. Pitting in cross field is difficult to see because of the crystals, and appear as some round and small pits, on two horizontal rows.

The libriform fibers have rounded-polygonal cross section with diameter of 7-14 μm, are thick- to very thick walled (4-12 μm the double wall), with small rounded lumina and are arranged in radial regular rows. Longitudinally seen, the fibers are unpitted and unsepted; their length is difficult to appreciate due to bad preservation.

Affinities and discussion

Evaluating the xylotomical features of the studied material with regard to the general arrangement of the structural elements in cross section, the vessels with simple perforations and numerous bordered alternate pitting, the crystalliferous banded and ray parenchyma, it is quite obvious that we talk about an Juglandaceous wood, very similar to the extant species of ***Juglans***, as figured and described by Greguss (1959), Schweingrüber (1990), or Watson and Dallwitz (1992, onwards).

The extant Juglandaceae include 7 genera with relative disjunct living areas: South-East Asiatic distribution for *Engelhardia, Platycarya*, and *Carya*, Eurasiatic for *Pterocarya, Juglans,* and in the American continents, like *Oreomunea, Alfaroa, Juglans* and *Carya* (Manning, in Duperon, 1988; Orain et al., 2013). No occurrence was reported from Africa and Australia.

From xylotomic point of view, *Engelhardtia, Alfaroa,* and *Oreomunnea* usually have simple and scalariform perforations. *Platycarya* have spiral thickenings on the tracheal or tracheidal walls.

Juglans, Pterocarya and *Carya* have quite similar xylotomical features. However, the presence or the absence of crystalliferous parenchyma, the aspect or/and vessel wall thickness can separate the genera. *Carya* always has thick vascular walls (>3 μm the simple wall), crystalliferous parenchyma usually present and with solitary crystals in enormous cells, of barrel type and in short rows of 1-2-3, and the apotracheal parenchyma appear in continuous bands. *Platycarya* has ring-porous wood, spiral thickenings or narrow vessels and vasicentric tracheids. *Pterocarya* does not have crystalliferous parenchyma, vascular walls are always thin (<3 μm) and the parenchyma appear as short and thin bands, usually uniseriate. *Juglans* has 4 types (grey walnut, English walnut, black north temperate walnut, and black tropical walnut) with variations of the crystalliferous parenchyma (Miller, in Dupéron, 1988).

The fossil morphogenera of Juglandaceae, reviewed several times till now (see Blokhina, 2007; Sakala and Gryc, 2011), are the following:

- *Engelhardtioxylon* (Manchester) Dupéron, 1988, as correspondent of the extant *Engelhardia, Alfaroa* and *Oreomunnea* having quite similar features as it is difficult to separate them (see Blokhina, 2007);
- *Manchesteroxylon* Wheeler and Landon, 1992, proposed by Manchester (1983) for fossil wood of the Engelhardioideae, which seems to be a primitive juglandaceous taxon with simple and scalariform perforations, ring-porosity and vascular tracheids similar to the Platycaryoideae (see also Blokhina, 2007, p.1042);
- *Clarnoxylon* Manchester et Wheeler, 1993, does not have idioblasts with crystals in parenchyma like *Platycarya*, but has spiral thickenings on vessels, so it is a possible fossil correspondent of the latter (Wheeler and Manchester, 2002);
- *Pterocaryoxylon* Müller-Stoll and Mädel, 1960, is the fossil equivalent for species of *Pterocarya* and some species of *Juglans* (Asiatic species and an American species of grey walnut), having thin vascular walls, short and uniseriate parenchyma bands devoid of crystals;
- *Eucaryoxylon* (Müller-Stoll and Mädel) Dupéron, 1988, has porous structure, solitary vessels or short multiples, thick to very thick vascular walls, simple perforations, alternate intervascular pitting, rather big, paratracheal and apotracheal banded parenchyma, 1-2(4)-seriate, long, regular; sometimes crystalliferous parenchyma, large idioblasts with solitary crystal of "barrel" type, isolated or in 2-3 rows; rays 1-3(5)-seriate, and septate pith, corresponding to the extant species of *Carya*;
- *Rhysocaryoxylon* Dupéron, 1988 corresponds to the extant genus *Juglans* characterized by porous to semi-ring-porous structure, thick-walled vessels, solitary vessels or in small multiples, simple perforations, alternate intervascular pitting, banded apotracheal parenchyma, 1-2(4) seriate bands, long, regular, paratracheal few; crystalliferous parenchyma as short chains; rays 1-3(5)-seriate, heterogeneous and septate pith.

As a conclusion, the validity of the species described under the invalidated genera, previously used for fossil specimens like *Mirbelites*, *Juglandinium*, *Juglandoxylon*, *Juglansoxylon*, *Jugloxylon*, *Caryoxylon*, *Juglans*, *Carya* or *Pterocarya* (quoted by Dupéron, 1988), were, or had to be revised.

Blokhina (2007) made a new revision of the knowledge on Juglandaceous xylotomy and palaeoxylotomy and presents some questions on taxonomy, evolution and phylogeny of this group and gives a key for identification of wood anatomy in modern and fossil Juglandaceae. Using this key, but also the other information in the paper, it is clear that a wood structure with diffuse or semi-ring-porous structure, crystals only in axial parenchyma and thin to thick walled vessels define a *Rhysocaryoxylon* structure, completed by "solitary vessels or in radial multiples, vessel-ray and vessel-parenchyma pits with significantly reduced borders and large apertures, apotracheal parenchyma is in bands 1–2(4) cells wide, and rays are 1-3(-5)–seriate, homocellular to slightly heterocellular, and with or without crystals".

As to *Rhysocaryoxylon*, Sakala and Gryc (2011), describing a new species for this morphogenus, observe that "it was nomenclaturally superfluous when published and therefore illegitimate from the nomenclatural point of view, so either a proposal must be prepared for TAXON (the Journal of the IAPT) to conserve the name *Rhysocaryoxylon* against *Caryojuglandoxylon*, or new combinations must be published in the future". And this, because of the presence in this forms of crystalliferous idioblasts similar to *Carya* type, but more numerous, and because of the presence of narrower rays and smaller vessels and rays tending to be uniseriate in majority, features that did not correspond to the accepted diagnosis based on the type species *R. schenkii* (Felix) Dupéron.

Using the Dupéron's key of identification (1988) we can say that our specimens, by their xylotomical features are different of *Juglans regia* ("English Walnut" or "Persian Walnut"), from Dioscaryon section, which has the vascular walls not very thick and short-banded, non-crystalliferous parenchyma, but closer to the wood types from Rhysocaryon section **(**known as "Tropical or North-temperate black walnuts", because present thick-walled vessels, continuous crystalliferous parenchyma bands of 1-3 cells

wide and vertical short chains (no more than 5) chambers with solitary crystals, features very similar, for example, to the xylotomy of the extant *Juglans nigra* (see Greguss, 1959) showed, and consistent with the fossil types described by Müller-Stoll and Mädel (1960), Manning (1978), Dupéron (1988) and Blokhina (2007).

Taking into account the forms described till now having vertical short chains of crystalliferous hypertrophied idioblasts within parenchyma strands, different of the black tropical walnuts (see Iamandei and Iamandei, 2002), it seems that the studied specimens have features closer to a species previously described by us, as a black north temperate walnut, *Rhysocaryoxylon transylvanicum* Iamandei and Iamandei, 2003, to which we assign our studied specimens.

Rhysocaryoxylon caucasicum (Gayvoronsky) Dupéron, 1988,
Plate V, figs. 1-9.

Macroscopic description

The studied material was found in Căprioara valley, southwestern part of Feleacu hill, in Early Sarmatian littoral-neritic sediments and is now deposited in the Geological Museum of Department of Geology and Paleontology of UBB Cluj Napoca, under the numbers 385, 387 and 388. Each specimen is represented by some pieces of centimetric sizes, most probably sampled from bigger trunks: 385 has 6 pieces, the biggest has 23/10/5 cm; 387 has 2 pieces, the biggest has 20.5/9/8.5 cm; 388 has 6 pieces, the biggest has 13/6.5/3 cm. All the samples show light brown to grey color, locally with yellowish tints, a fibrous texture with vessels, broad rays and visible growth rings, suggesting a dicot wood structure.

Microscopic description

Growth rings have indistinct boundaries, but the structure is ring-porous to slightly semi-ring-porous as the first row of very big vessels suggests, followed by vessels which tend to gradually decrease their lumina to the late-wood. The tangential parenchyma bands are long, regular and with slightly decreasing intervals. The entire structure is compressed and difficult to compare with a normal one.

The vessels, in cross section, appear thick walled, as big solitary pores, of 80-230 μm tangential diameter, rarely as small radial multiples, on single first row of earliest wood arranged, and smaller (30-60 μm) to the late-wood, with a density of some tens on square millimeter. Simple perforations on tilted plates and alternate pitting can be observed, as small bordered pits (8-10 μm in diameter) with point-like apertures. Short vascular elements of 80-150 μm, and often big tyloses can be observed inside.

Axial wood parenchyma of apotracheal type appear in cross section as long tangential bands, 1-2 cells thick, and few of paratracheal type appear, as some flattened cells around the vessels, visible mainly in longitudinal view; the parenchyma cells have 12-15 μm in diameter. In longitudinal view, the apotracheal parenchyma – the bands – appear in strands of vertical rectangular cells. Sometimes, they appear as long rows of hypertrophied crystalliferous cells, easy to be mistaken, in tangential view, for uniseriate rays, which are filled with dark content and large solitary crystals, slightly rounded. The vasicentric parenchyma one is visibly simply pitted.

The medullary rays – in cross section appear thin, with linear or slightly curled trajectory sometimes touching vessels and are formed from rectangular cells radially elongated usually full of dark matter. In tangential view, the major part of the rays are uniseriate, probably rarely biseriate, and of average height. Their density is of 12-14 rays on horizontal tangential millimeter. Radially seen, the rays appear homogeneous, with all cells procumbent, relatively big, marginals slightly higher, or squared, and have usually dark content and big rounded crystals floating in. Pitting in cross field is so difficult to see because of the crystals, appear some on one or two horizontal rows, round, small.

The libriform fibers – even if the structure is compressed, they appear arranged in radial regular rows, have polygonal rounded cross section of 17-20 μm diameter, cellular thick- to very thick-walled (4-12 μm double wall), small round to point-like lumina. Longitudinally seen, the fibers are unpitted and unsepted but, due to rather bad preservation, their length is difficult to estimate.

Affinities and discussion

Evaluating the xylotomy of the studied specimens as having very similar wood structure, especially in radial view, with the above described material, it is simple to attribute

them to juglandceous wood type as it is described and figured by Greguss (1959) and Schweingrüber (1990), and consistent with the general description of wood anatomy of Juglandaceae given by Watson and Dallwitz (1992, onwards).

It is no need to repeat the discussion about the above described species, since the studied specimens seem to have long vertical chains of crystalliferous chambers, exceeding 5, and could be attributed to *Rhysocaryoxylon* morphogenus of type "Black Tropical Walnut", but it is a problem which is specific for this material, rather badly preserved, namely the presence of the uniseriate rays in a mainly ring porous (to semi-ring porous) wood structure. As it is well known, the morphogenus *Rhysocaryoxylon* Dupéron, 1988, is a correspondent of the extant genus *Juglans*, which has "porous to semi-ring-porous structure, thick-walled vessels, solitary vessels or in small multiples, simple perforations, alternate intervascular pitting, banded apotracheal parenchyma, 1-2(4)-seriate bands, long, regular, paratracheal few; crystalliferous parenchyma as short chains; rays 1-3(5)-seriate, heterogeneous and septate pith" genus diagnosis having as type species *Rhysocaryoxylon schenkii* (Felix) Dupéron, 1988, consistent with the fossil types described by Müller-Stoll and Mädel (1960), Manning (1978), Dupéron (1988), Iamandei and Iamandei (2002, 2003, and this paper), Blokhina (2007) and Sakala and Gryc (2011).

There are some species described with narrow rays, such as *Rhysocaryoxylon treibelii* (Caspary) Dupéron, 1988, described also by Müller-Stoll and Mädel (1960) and Van der Burgh (1973); *R. pilinyense* Greguss (1969), recently reviewed by Sakala and Gryk (2011, p. 48), who describe a new morphospecies *R. madsenii* Sakala and Gryc 2011; *R. fryxellii* (Prakash and Barghoorn) Dupéron, 1988; *R. tertiarum* (Prakash and Barghoorn) Dupéron, 1988 were described also with narrow homocellular rays. Otherwise, Wheeler and Dillhoff (2009) recently described some woods from the Middle Miocene Vantage Flora (Washington, USA) and reviewed the type species of Prakash and Barghoorn (1961a, b).

Very similar to our material is *R. caucasicum* (Gayvoronsky) Dupéron, 1988, which has uni- and biseriate rays (see Gayvoronsky, 1962). Since the xylotomical features of our specimens are almost identical with the species of Gayvoronski, especially the fine rays, after this short analysis we assign our specimens to this morphospecies, named *Rhysocaryoxylon caucasicum* (Gayvoronsky) Dupéron, 1988.

Familly **Fagaceae** Dum.
Genus ***Quercoxylon*** (Kräusel) Gros 1988
Quercoxylon sp. cf. *Quercus frainetto* Ten.
Plate VI, figs. 1-9.

Macroscopic description

We had in study some samples of petrified wood found in Căprioara valley, southwestern part of Feleacu hill, in Early Sarmatian littoral-neritic sediments, which is now deposited in the Geological Museum of Department of Geology and Paleontology of UBB Cluj Napoca, under the number 370. It consists of 8 fragments of centimetric size; the biggest one has 8/6.5/3 cm; all come from a single trunk. The samples show light grey color, occasionally with brownish tints, has a fibrous texture with vessels, broad rays and visible growth rings suggesting a dicot wood structure.

Microscopic description

Growth rings distinct, marked by the first 1-2 rows of big solitary vessels of the early-wood, which also marks the ring-porous structure of the studied fossil wood, as the late-wood is marked by abrupt transition to the small vessels that show a slightly gradual size decrease and are arranged in radial to slightly dendritic pattern.

The vessels, in cross section, appear usually solitary, lesser as small multiples of 2(3) in the early-wood, on 1-2 rows, as well as in the transitional and late-wood where are radially arranged or slightly dendritic. The solitary pores, in fact the cross section of the vessels, large or smaller, having radial oval shape, in the late-wood, slightly radially elongate. The vessels are moderately thick-walled (5-7 μm the double wall) and the lumina of the solitary vessels have the radial/tangential diameters of 90-200/40-100 μm in the early-wood and of 40-70/20-30 μm in the late-wood; their density is difficult to evaluate because of the poor preservation and its value could not be significant, since there are interradial fascicles devoid of vessels. All the vessels have simple perforations on tilted plates and intervascular pitting bordered, small (3.5-5.5 μm in diameter), round to oval opposite, subopposite to alternate, numerous, spaced, with

horizontal small elliptic apertures of 1-2 µm. Pitting to ray cells is similar. The vascular elements appear to be short, around 200-300 µm, or more. Tyloses present but, due to the bad preservation, it is difficult to appreciate other details.

Axial parenchyma – is of diffuse apotracheal type, mixed with libriform or tracheids together in broad bands. Vasicentric parenchyma also present, visible in longitudinal view, and with numerous large simple pits, round or oval. In longitudinal view, the parenchyma cells bear crystals and tanninous dark content.

The rays have a linear trajectory and are two-sized: numerous uniseriate rays, and fewer multiseriate-compact, sometimes dissected by fibers, of 8-18 cells broad or more spaced arranged, and more than 1 mm tall. The ray density is of 10-22 rays per tangential horizontal millimeter. The rays are homogeneous-homocellular; in radial view, the ray cells are all procumbent. The cross fields with vessels are poorly preserved and have rectangular pits (palisade). The cross field pitting is difficult to observe since the ray cells are usually full of polygonal rounded crystals and tanninous dark remains.

The libriform fibers have a polygonal cross section of 8-13 µm in diameter, relatively large lumened, thick walled (4-5 µm the double wall), are unsepted and unpitted.

Fibro-tracheids and vascular tracheids are rather unclear in cross section also because of the bad preservation. Longitudinally, they present small pitting, bordered, round or slightly elliptic, with small apertures, and in 1(-2) vertical rows.

Affinities and discussion

The cross sections of the studied specimen is typical for Fagaceae – ring porous wood structure and two sized rays – broad and tall multiseriates and numerous fine rays, especially uniseriates, and also the aspect and arrangement of the vessels in the early-wood and in the late-wood, are typical for Quercineae.

After Hadziev and Mädel (1962) four types of oak structures could be separated:

1. "Weisseichen" (white oak type), comprising most of the species of section *Lepidobalanus* (in fact *Mesobalanus*, a group present in Europe, Asia, North Africa, see http://en.wikipedia.org/wiki/List of Quercus species), characterized by ring porous structure with small, polygonal, thin walled late-wood vessels.
2. "Roteichen" (red oak type), comprising the species of section *Eritrobalanus* (*Lobatae* section), http://en.wikipedia.org/wiki/List of Quercus species) and some species of *Lepidobalanoideae*, characterized by ring porous structure and relatively large round thick walled late-wood vessels.
3. "Semperviren oak" type, comprising species of *Quercus* and *Lithocarpus,* characterized by diffuse porous or half-ring-porous structure, relatively small and spaced vessels, often radially disposed.
4. "Root wood" type, corresponding to all types of oaks, characterized by diffuse-porous structure and crowded large pores.

Privé-Gill, in 1975, in perfect accord with Hadziev and Mädel, observed that the oak wood is characterized by vessels in radial rows more or less dendritically distributed and showing simple perforations. She also emphasized that two-sized rays (multiseriates very broad and finer, 1-2-seriates) are usually present in the extant *Quercus* L. and in *Lithocarpus* Bl. (= *Pasania* Oerst.) genera with some tens of species growing in temperate to warm regions of the northern hemisphere, islander as well. She specified that, generally, the oak wood structure could be characterized by vessels in radial rows, more or less anastomosed, or dendritically distributed, and with simple perforations. Their structure is diffuse porous, characteristic for the evergreen species since the ring-porous one is typical to the deciduous species of *Quercus* L. and to the most septentrional species, islander also, of *Lithocarpus* Bl. (= *Pasania* Oerst.).

The diffuse distribution of the vessels characterizes the semperviren species for both genera. The ring-porous distribution characterizes the deciduous oak species and the more septentrional species of *Lithocarpus*. As about the root wood, even the deciduous species tend often to lose the ring-porousness, becoming similar to the evergreen species, the broad rays become divided in "false rays", i.e. aggregate rays (Privé-Gill, 1975).

New attempts to establish the taxonomy of the fossil oaks, based on their xylotomy, was made by Gros (1988) and, independently, Suzuki and Ohba (1991), who made a comparative synthesis of the essential xylotomical characters of the extant fagaceous taxa, based on the previous anatomic studies and on the phylogenetic relations between the genera of Fagaceae.

A very useful systematization of the xylotomical details of Fagaceae, was made by Petrescu (1976), and even if the taxonomy of the group evolved, it is still a good key of identification of the fossil fagaceous wood (Table 1).

In fact, a lot of studies on oak wood remains were done till now (we cite only the one used for the documentation of this study): Müller-Stoll and Mädel, 1957; Brett, 1960; Gottwald, 1967; Greguss, 1969; Nagy & Petrescu, 1969; Hadziev & Mädel, 1962; Huard, 1966; Privé(-Gill) 1975, 1984, 1990; Privé & Brousse, 1976; Nagy and Petrescu 1969; Petrescu, 1969 1976, 1978; Petrescu et al., 1970 - 1974; Starostin & Trelea,

Table 1. Key for generic identification of fagaceous fossil woods (Petrescu, 1976)

Type A	Structures with solitary vessels or grouped (2-3 or more), with simple and scalariform perforations.	Fine rays (1-3-seriate) and multiseriate compact.	*Fagoxylon* *Nothofagoxylon*
Type B	Structures with solitary vessels, simple perforated, sometimes some scalariform, with few bars.	Exclusively uniseriate rays.	*Castanoxylon*
		Uniseriate and pluriseriate compact, sometimes compact-composed or partially aggregate rays.	*Quercoxylon*
		Uniseriate and pluriseriate aggregate rays.	*Lithocarpoxylon*

1969, 1984; Lupu, 1984; Samuel, in Lamouille et al., 1985; Selmeier, 1971, 1997; Suzuki & Ohba, 1991; Sakala and Gryk, 2011; Iamandei, 2002; Iamandei et al. 2008b, 2011, 2012. In fact, a lot of studies on oak wood remains were done till now, we cite the consulted ones: Müller-Stoll and Mädel, 1957; Brett, 1960; Gottwald, 1967; Greguss, 1969; Nagy & Petrescu, 1969; Hadziev & Mädel, 1962; Huard, 1966; Privé(-Gill) 1975, 1984, 1990; Privé & Brousse, 1976; Nagy and Petrescu 1969; Petrescu, 1969 1976, 1978; Petrescu et al., 1970-1974; Starostin & Trelea, 1969, 1984; Lupu, 1984; Samuel, in Lamouille et al., 1985; Selmeier, 1971, 1997; Suzuki & Ohba, 1991; Sakala and Gryk, 2011; Iamandei, 2002; Iamandei et al. 2008b, 2011, 2012.

Using at least these keys, it seems that our here studied structure suggests a type of white oak close to the extant Mesobalanus group that is assigned to the corresponding fossil morphogenus Quercoxylon, created by Hofmann (1929), but correctly defined by Kräusel (1939), and emended by Müller-Stoll and Mädel (1957), by Gros (1983, 1988), and by Suzuki and Ohba (1991) (see also Iamandei et al., 2008b, p.203-204).

After Gros (1988) accepted emendation, the Quercoxylon diagnosis is: porous or ring-porous structure, usually solitary vessels simply perforated, alternate vascular pitting, pitted parenchyma, banded or diffuse, libriform, tracheids and small vessels in groundmass and two-sized rays.

As to the wood structure of the extant species of “white oak”, Privé-Gill (1975) observed a big intraspecific variation, confirmed for the entire genus Quercus L., which shows a great interspecific anatomic homogeneity. This makes it very difficult to ascribe fossil species, thus terms that have only a descriptive value, as for morphospecies being used.

Selmeier (1997) remarked that vessel diameter, ray size, ray frequency and ray distribution are variable: even in the same described fossil species, the biometric values may differ. For this reason, in xylotomical description of extant wood, no updated measurements are given, and any accurate measurements made by paleoxylologists cannot be always used to identify an unknown species. Moreover, the rules from "IAWA List of Microscopic Features for Hardwood" (Wheeler at al., 1989) impose different ways to convert older measurements.

However, the ray structure is homocellular and the cross fields with vessels show quadrangular rounded bordered pits, disposed vertically, as palisade, very similar to those in the extant species Quercus frainetto TEN. (known as the Hungarian Oak or Italian Oak, or Gârniţa in Romanian). This form is classified into the genus Quercus, section Mesobalanus,

and is a species of oak native in Southeastern Europe and Turkey. The atlas of Schweingrüber (1990), the sites Schoch et al. (2004) and InsideWood (2004) also showed us much similitude of this type of structure with our studied specimens. Under these circumstances, based on the observed xylotomical details, we appreciate that the studied specimen could be assigned to Quercoxylon sp. cf. Quercus frainetto Ten.

Table 2. Feleacu Flora

Leaves, fruits (in Givulescu, 1997)	Spores, Pollen (Suciu, 2005a, b)	Petrescu 2003, unpublished	Iamandei et al. 2011 and 2013 (present paper)
-	*Laevigatisporites haardti* (Polypodiaceae)	-	-
Sequoia abietina (Brongn.) Kn.	Taxodiaceous pollen	*Sequoioxylon*	*Sequoioxylon gypsaceum* - 2 specimens
-	-	-	*Tetraclinoxylon romanicum* – 1 specimen
-	-	-	*Chamaecyparixylon* cf. *polonicum* – 2 specimens
Pinus hepios Unger *Pinus felekiensis* Staub	*Pinus*	*Pinuxylon*	-
Abies cf. *alba* Miller	*Abies,*	-	-
-	*Picea, Tsuga*	-	-
Laurophyllum sp., *Daphnogene polymorpha* (Al.Br.) Ett.	-	*Laurinoxylon*	*Cinnamomoxylon intermedium* - 3 specimens
Platanus cf. *leucophylla* (Brongn.) Kn.	-	-	-
-	-	*Ebenoxylon*	-
Engelhardia macroptera (Brongn.) Unger	*Carya, Pterocarya Engelhardia*	*Rhysocaryo-xylon*	*Rhysocaryoxylon pravalense* - 2 specimens *Rhysocaryoxylon caucasicum* - 3 specimens
Quercus mediterranea Unger	*Quercus*	*Quercoxylon*	*Quercoxylon sp.* cf. *Quercus frainetto* – 1 specimen
Ulmus pyramidalis Goeppert	*Ulmus*	-	-
Acer tricuspidatum Brongn.	-	-	-
Fraxinus sp.	-	-	-
-	*Betula*	-	-
-	*Myrica*	-	-
-	*Alnus*	-	-

3. Conclusions

In this paper, newly identified lignotaxa from Feleacu hill are presented, partially confirming the list of Staub (1883, 1891), revised formally by Givulescu (1997) and Petrescu (2003) as characterizing the Early Sarmatian, Feleacu flora. Petrescu (2003), after a preliminary survey of the new material, quoted some lignotaxa as *Sequoioxylon, Pinuxylon*, *Quercoxylon, Laurinoxylon, Ebenoxylon* and *Rhisocaryoxylon*, but he never published them correctly.

More recently, a new lignotaxon (*Tetraclinoxylon*) was described from the same site (Iamandei et al, 2011) and the present paper partially confirms the generic identification of Petrescu. The material studied here was collected from a sedimentary formation comprising gravels, sands, sandstones, clays, which indicate a continental fluvio-deltaic environment to a littoral one, and this small flora seems to indicate, for the Early Sarmatian, a warm temperate paleoclimate in this region. A comparative record of the previously reported Feleacu paleoflora and the additions provided by our research of is given in Table 2.

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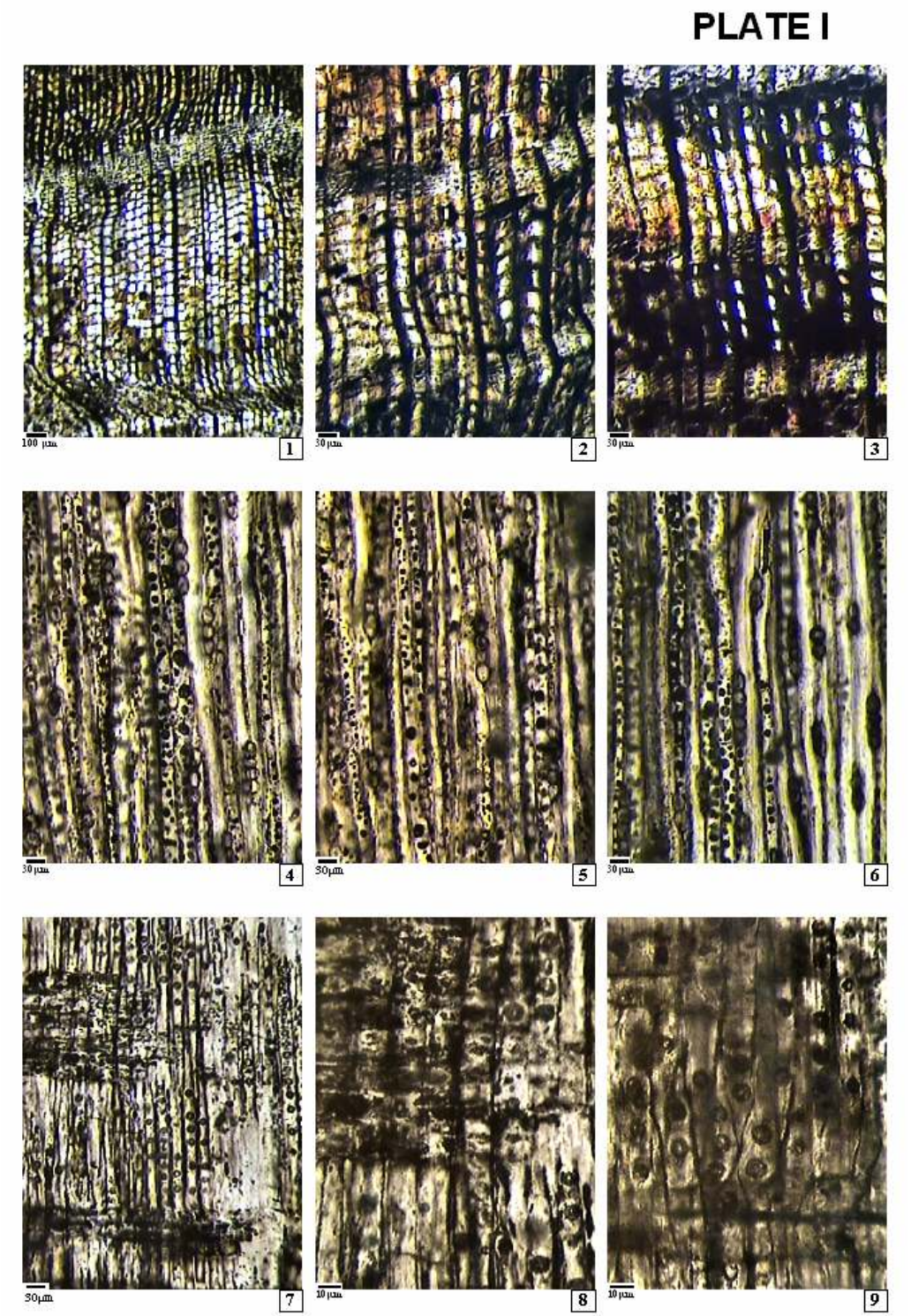
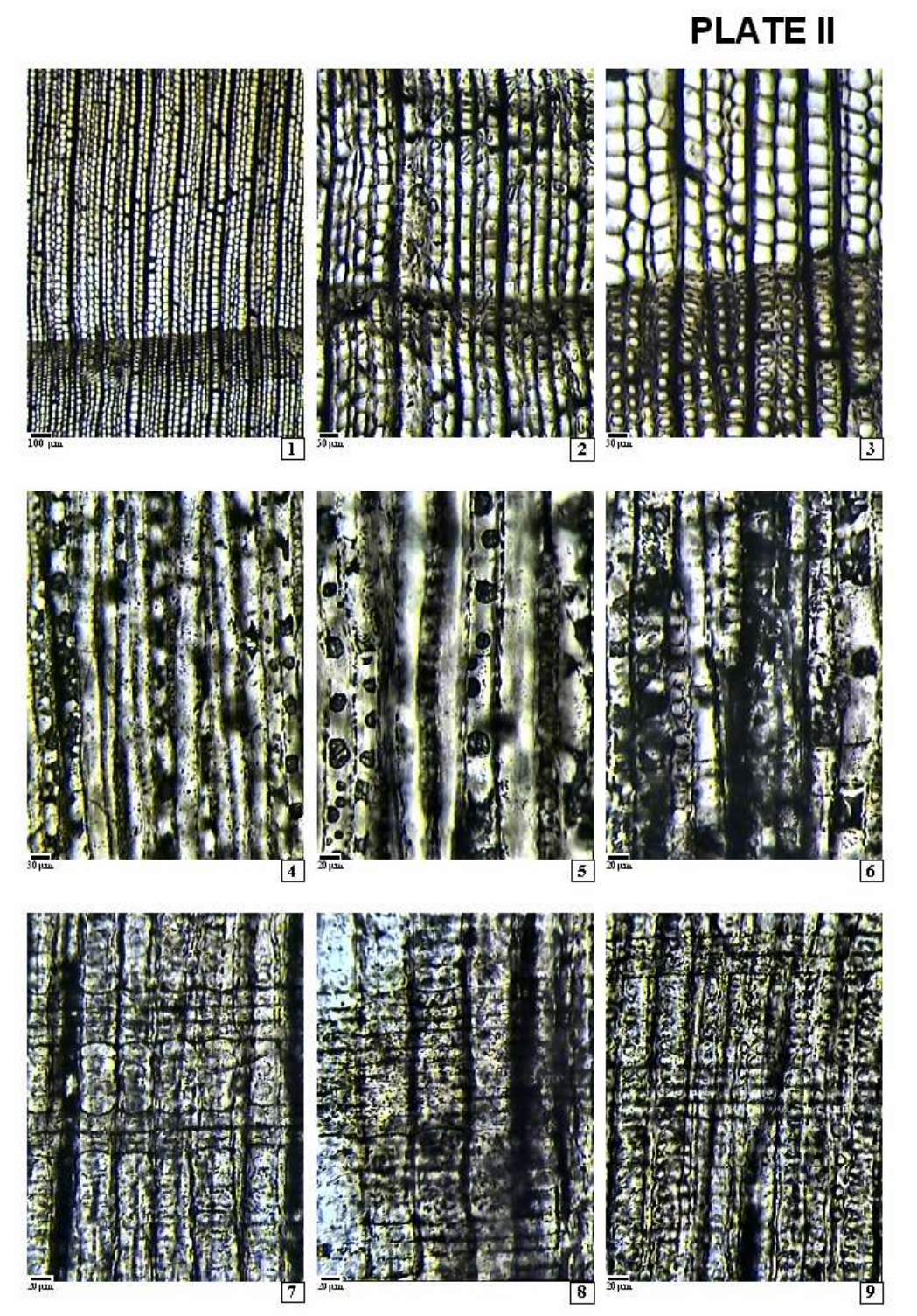
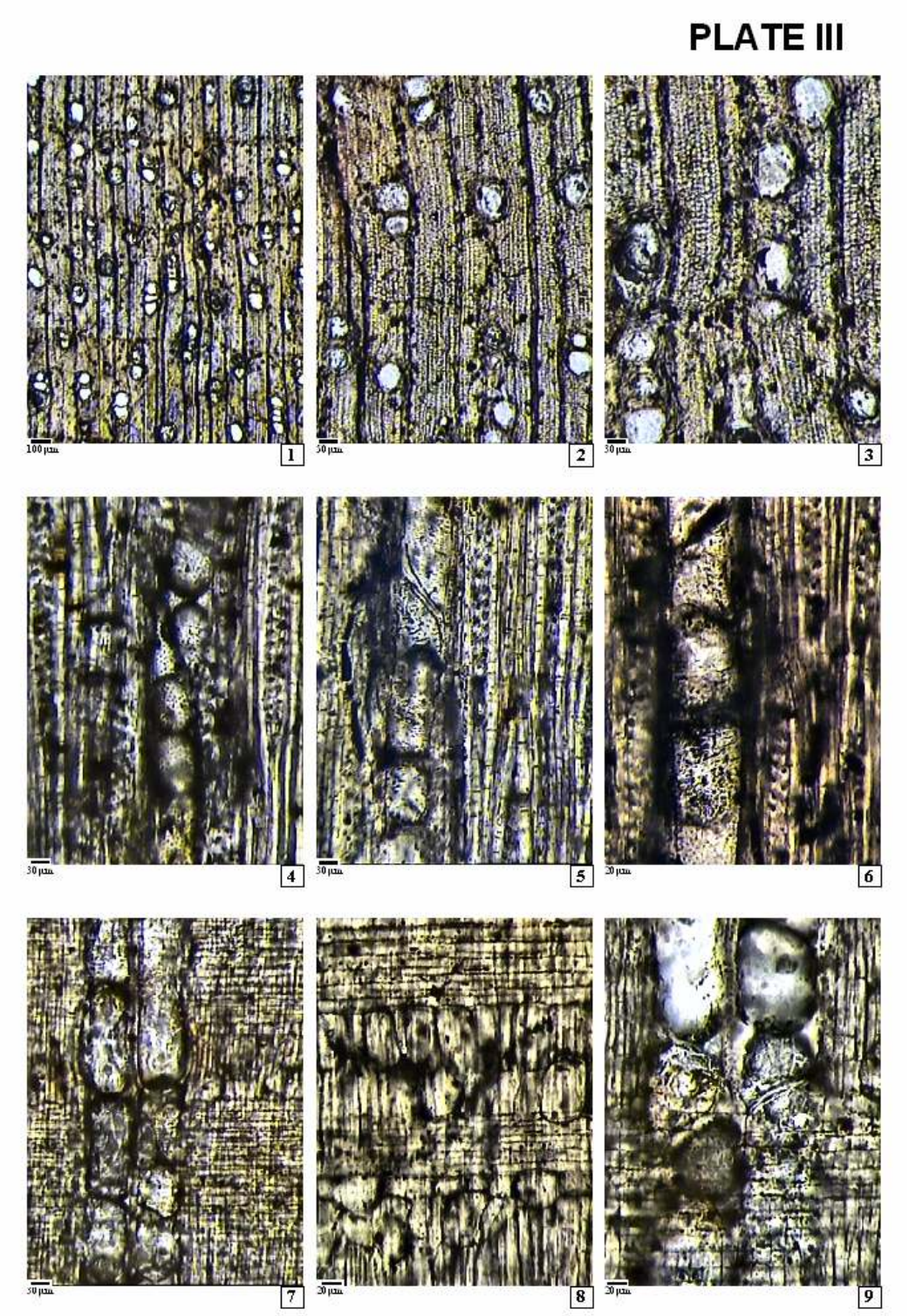


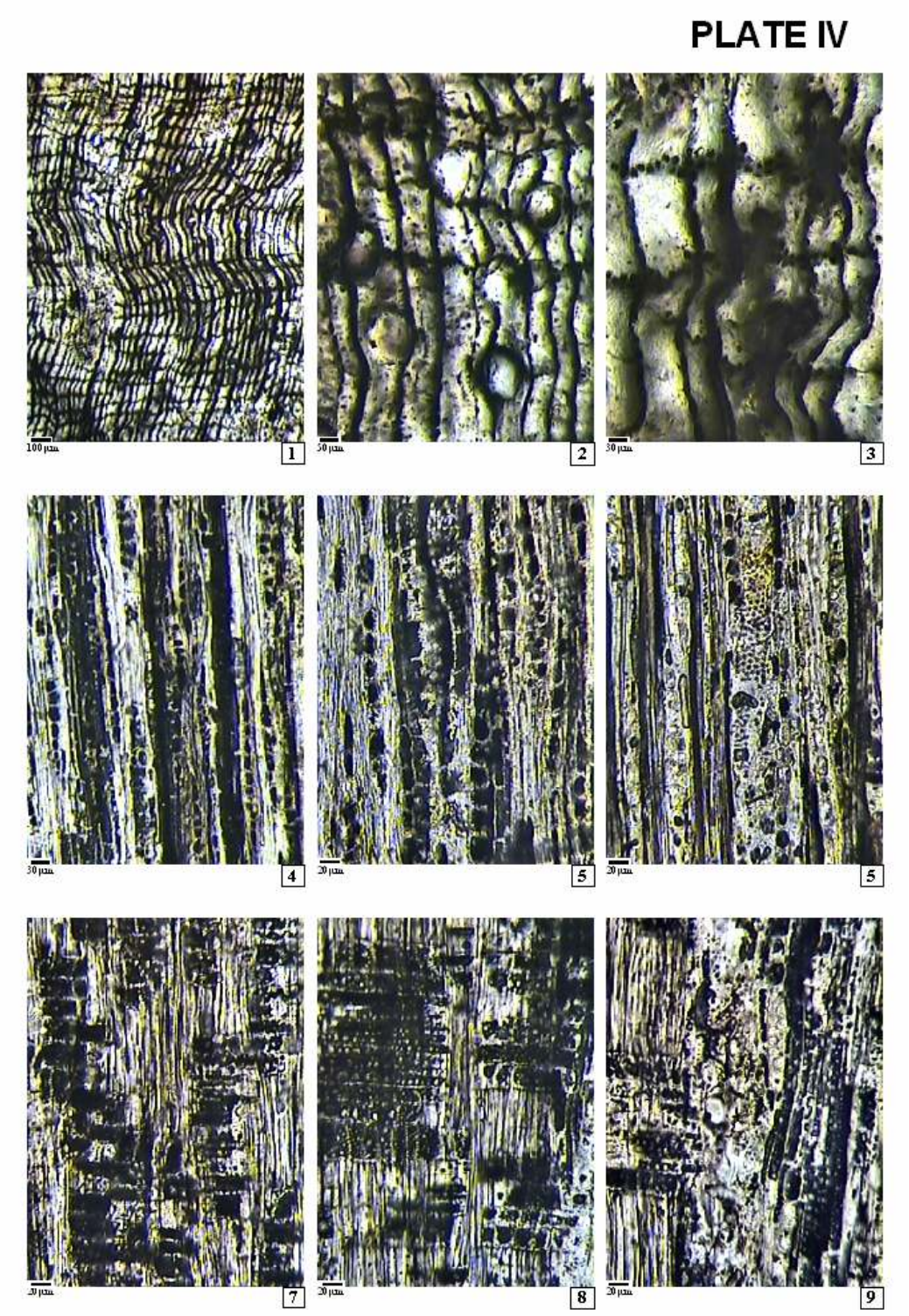
Figs. 1-9. *Chamaecyparixylon* cf. *polonicum* (Kräusel) Chudajberdyev 1958, material; Sm.1; Căprioara valley – Feleacu, Cluj; stored under no. 386 in Geological Museum - UBB Cluj Napoca.
Figs. 1-3. Cross section – distribution of relatively thickwalled polygonal tracheids, growth rings with early -wood, late-wood, distinct boundary and scattered parenchyma cells.
Figs. 4-6. Tangential section – abundant parenchyma with resinous glomerules, short uniseriate rays.
Figs. 7-9. Radial section – uniseriate spaced abietinean pitting, homocellular rays, blurred cupressoid cross fields.



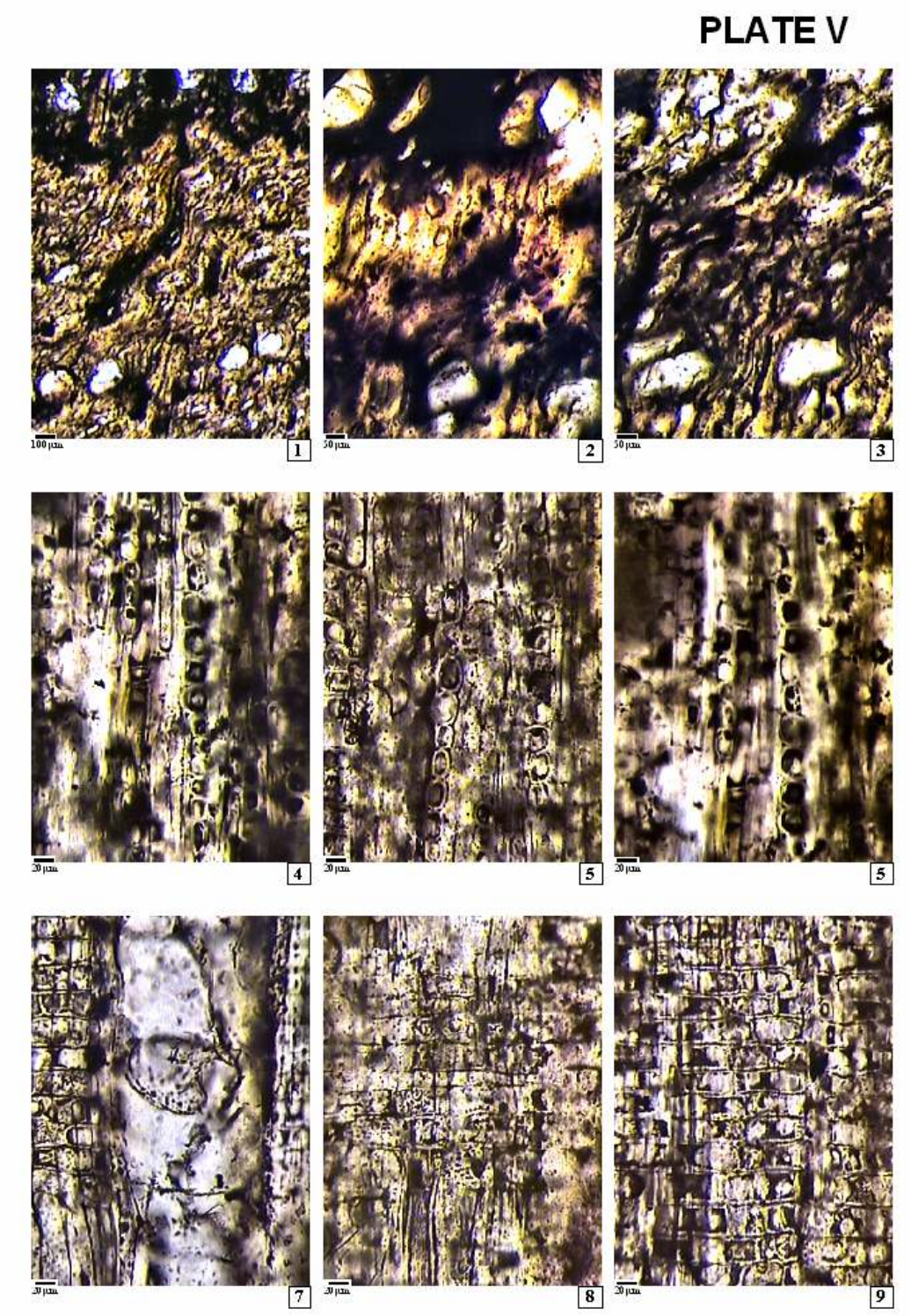
Figs. 1-9. *Sequoioxylon gypsaceum* (Goeppert) Greguss 1967, material; Sm.1; Căprioara valley – Feleacu, Cluj; stored under no. 797 in Geological Museum - UBB Cluj Napoca.
Figs. 1-3. Cross section – distributions of polygonal tracheids, thin walled in early-wood, thick walled in late-wood, distinct ring boundaries, scattered resinous parenchyma cells.
Figs. 4-6. Tangential section – axial parenchyma with thin horizontal walls and resin balls or plugs, tangential tracheidal pitting absent, high rays.
Figs. 7-9. Radial section – 1-3 tracheidal radial abietinean pitting, homocellular rays, taxodioid cross fields with 2-3 pits in horizontal row, marginal fields taller, with 4-6 pits in 2 horizontal rows.



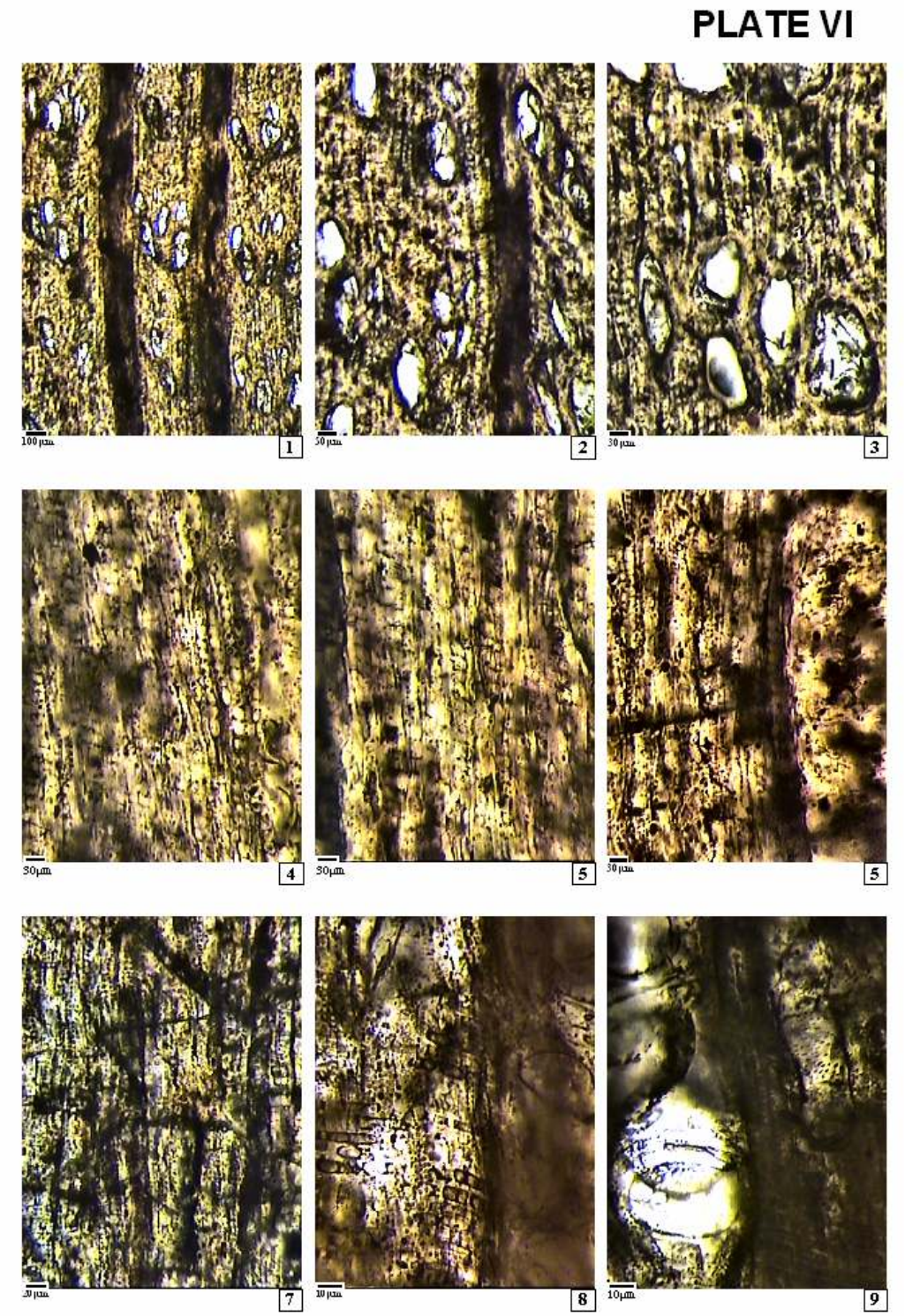
Figs. 1-9. *Cinnamomoxylon intermedium* (Huard) Gottwald, 1997, material; Sm.1; Căprioara valley – Feleacu, Cluj; stored under no. 369 in Geological Museum - UBB Cluj Napoca.
Figs. 1-3. Cross section – thick-walled vessels solitary or in small radial groups, thick walled fibers and linear 1-3 seriate rays.
Figs. 4-6. Tangential section – thick walled tylosed vessels with alternate pitting, bi- triseriate rays, septate fibers and axial parenchyma.
Figs. 7-9. Radial section – vessels with simple perforations, heterocelular rays with marginals hypertrophied.



Figs. 1-9. *Rhysocaryoxylon transylvanicum* Iamandei and Iamandei, 2003, material; Sm.1; Căprioara valley – Feleacu, Cluj - stored under no. 377, in Geological Museum - UBB Cluj Napoca.
Figs. 1-3. Cross section – solitary pores or in small radial multiples, long tangential bands of (1)2-3 parenchyma cells, giving scalariform aspect.
Figs. 4-6. Tangential section – alternate pitting on vessels, 1-3 - seriate rays.
Figs. 7-9. Radial section – vessels with simple perforations, crystalliferous radial and axial parenchyma, homocellular rays.



Figs. 1-9. *Rhysocaryoxylon caucasicum* (Gayvoronsky) Dupéron, 1988, material; Sm.1; Căprioara valley–Feleacu, Cluj; stored under no. 388, in Geological Museum - UBB Cluj Napoca.
Figs. 1-3. Cross section – ring porose structure, big vessels in the early-wood, long tangential fine bands of parenchyma.
Figs. 4-6. Tangential section – crystalliferous parenchyma, uniseriate rays.
Figs. 7-9. Radial section – vessels with simple perforations, thick walled ray cells all procumbent and full of crystals.



Figs. 1-9. *Quercoxylon* sp. cf. *Quercus frainetto* Ten., material; Sm.1; Căprioara valley – Feleacu, Cluj; stored under no. 370, in Geological Museum - UBB Cluj Napoca.
Figs. 1-3. Cross section – ring porose structure, 2-3 rows of big vessels in the early-wood, short tangential bands of parenchyma.
Figs. 4-6. Tangential section – uniseriate and multiseriate rays.
Figs. 7-9. Radial section – vesels with simple perforations, alternate pitting, vasicentric pitted parenchyma, crystalliferous radial parenchyma.

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Melt inclusions texture and thermal history in minerals from the Dej Tuff, Transylvania basin, Romania

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Abstract: This paper is based upon petrography and microthermometry and shows the presence of "foamy/bubbly" silicate melt inclusions in quartz, feldspar and zircon from the "Dej Tuff" that crystallized in magma chamber before and/or during the plinian- and plinian- phreatomagmatic explosive volcanic activity occurred during early Badenian (14.8-15.1 Ma) in the NW-Transylvania Basin. The "foam-like" silicate glass inclusions" were trapped as homogeneous and/or heterogeneous silicate melt, now consisting of a package of tiny bubbles occupying up to 75-80% in the silicate glass volume. The homogenization temperature determined in the heating stage at one atmosphere, by vapor bubble(s) disappearance (i.e. the minimum trapping temperature), in the silicate melt phase ranged between 710°-900°C (±20°C), with maximum of frequency between 750° to 830°C (n=344). These values are in very good agreement with temperature calculated by common geothermometers such as Fe-Ti oxide, Zr saturation temperature, $\Delta^{18}O$ (Qz-Mt) etc, applied in some classical rhyolitic magmas. The trapping pressure of the "foam-like" silicate glass inclusions" from "Dej Tuff's phenoclasts and phenocrysts was estimated between 200 MPa and 20 MPa, based upon spontaneous homogeneous vs. heterogeneous bubble nucleation during heating in the microthermometric stage. It is suggested in this study that "foam-like" silicate glass inclusions is an adequate term for the silicate melt inclusions described in "Dej Tuff" phenoclasts, beside "hourglass inclusions" and common silicate melt ones. The silicate melt was probably trapped during crystals growing in a shallow foam layer before eruption in the top of a molten magma chamber that ultimately was erupted explosively as probably a result of a hotter mafic influx at its bottom region.

Keywords: "foam-like" glass inclusions", microthermometry, Badenian, "Dej Tuff"

1 Introduction

The silicate melt inclusions containing more than one shrinkage bubble at the room temperature conditions were described early in the history of the subject since Sorby (1858), Vogelsang (1869) and Zirkel (1893). It should be noticed that the serial drawings no. 90 and 91 from the XIXth plate of Sorby's mentioned paper could be taken as "the first microthermometric cycle" presented in a published paper worldwide about silicate melt inclusions. There was estimated a temperature range of 800° -1000°C, based upon the color of the experimentally-heated augite crystal from Vesuvius. More recently Roedder (1979) presented a piece of feldspar from the "Icelandic tuff" in which a "foam-like" silicate melt inclusion assemblage was formed. The most important and convincing data on the topic were published on microthermometry and petrography of silicate glass inclusions trapped in volcanic quartz worldwide by Clocchiatti (1975). This author presented original serial drawings and pictures of sequential microthermometry on "ponce" silicate glass inclusions, and recorded their features during heating, including, in many cases, the final homogenization.

Kamenetsky and Danyushevsky (2005) studied quartz crystals from the Taupo Volcanic

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Zone (New Zealand) and revealed that bubbles nucleated in the melt inclusions at the α-ß transition temperature (573°C) and dissolved completely at 820-850°C (homogenization). In this study we also observed at the quartz transition temperature an enhancement of bubble nucleation, but many bubbles were earlier nucleated coincidently with Tg values around 500-530°C (glass transition temperature in rhyolites, e.g., Bagdassarov et al., 1996), and the microtexture of the silicate foam became visible even in the apparently "bubble-free" glass inclusions at room temperature. The "Bishop Tuff" formation in the western USA is another classic example of rhyolitic explosive products with a large amount of published data during the last decades by Anderson (1991, 2003), Skirius et al. (1990), Peppard et al. (2001), Wallace et al., (2003) on glass inclusions trapped in quartz phenocrysts and phenoclasts. The term phenoclast used in this study, mainly from the quartz and feldspar broken crystals, was taken from Best and Christiansen (1997} which stated that: "Rapidly erupted Plinian pyroclasts that form ash-fall deposits mix with cool atmosphere and thus generally quench before volatile dissolved in melt inclusions can nucleate bubbles and blow their host crystal apart". Further information about phenoclasts generated by fragmentation of the magmatic crystals was taken from Bindeman (2005), who reported that during explosive volcanic eruptions phenocrysts fragmentation was caused mainly by decrepitation of silicate melt inclusions and the resulted crystal fragments were defined as phenoclasts. A volcanic rock sample containing fragments of feldspar and rarely quartz from Amiata volcano, Italy described firstly as "reoignimbrite" by Rittman (1936), was used here for comparison. Data on melt inclusions in rhyolite or dacite worldwide were also given by Johnson et al. (1994), Manley (1996), Lowenstern (1995, 2003), Frezzotti (2001), Naumov et al. (1993), Cesner (1998) and many others.

The study of "foam-like" glass inclusions from "Dej Tuff" phenoclasts shown that they could be used to estimate trapping temperature by homogenizing in the heating stage under the microscope, if we related them to an enclosed silicate foam rather than a common silicate melt, which normaly shown only one contraction bubble at room temperature condition

2 Geological setting

The geological formation known as "Dej Tuff" was defined by Pošepny (e.g., Szakács, 2000) in the NW part of the Transylvanian Basin (Fig. 1) and was deposited during the climactic plinian or plinian-phreatomagmatic volcanic activity in the Badenian time (14.8-15.1 Ma - Szakács et al., 2012). There are three different acidic tuffs sequences spread in the Transylvanian Basin (NW-Romania). Their lithofacial features suggest that majority of the "Dej Tuff" tephra has been reworked and redeposited after their original deposition in seawater environment (Szakács, 2000). The caldera (s) that generated such large volume volcaniclastic material is localized probably in the NW-Transylvanian Basin between the Vihorlat Mountains (Ukraine) and Gutâi Mountains from Romania (Szakács, 2000; Fülop, 2002), Data on petrochemistry mineralogy, litho-stratigraphy and timing, were presented by Mârza and Meszaros, 1991; Mârza et al., 1991a,b; Szakács, 2000, Szakács et al., 2012 and references therein.

The mineralogical composition of the "Dej Tuff" includes quartz, plagioclase as main rock forming minerals and minor components represented by K- feldspar and plagioclase, amphibole, pyroxene, and with Fe-Ti oxides, zircon, apatite and allanite as accessory minerals, rarely completed by monazite and xenotime. Various lithoclasts of magmatic or nonmagmatic origin (metamorphic, sedimentary) are frequent observed and described (Szakács, 2000, 2003).

For this study, quartz and feldspar phenoclasts and rarely phenocrysts were extracted simply by crushing of pumice fragments, ignimbrites and unwelded tephra collected from Şoimeni, Pâglişa, Jichişul de Sus,

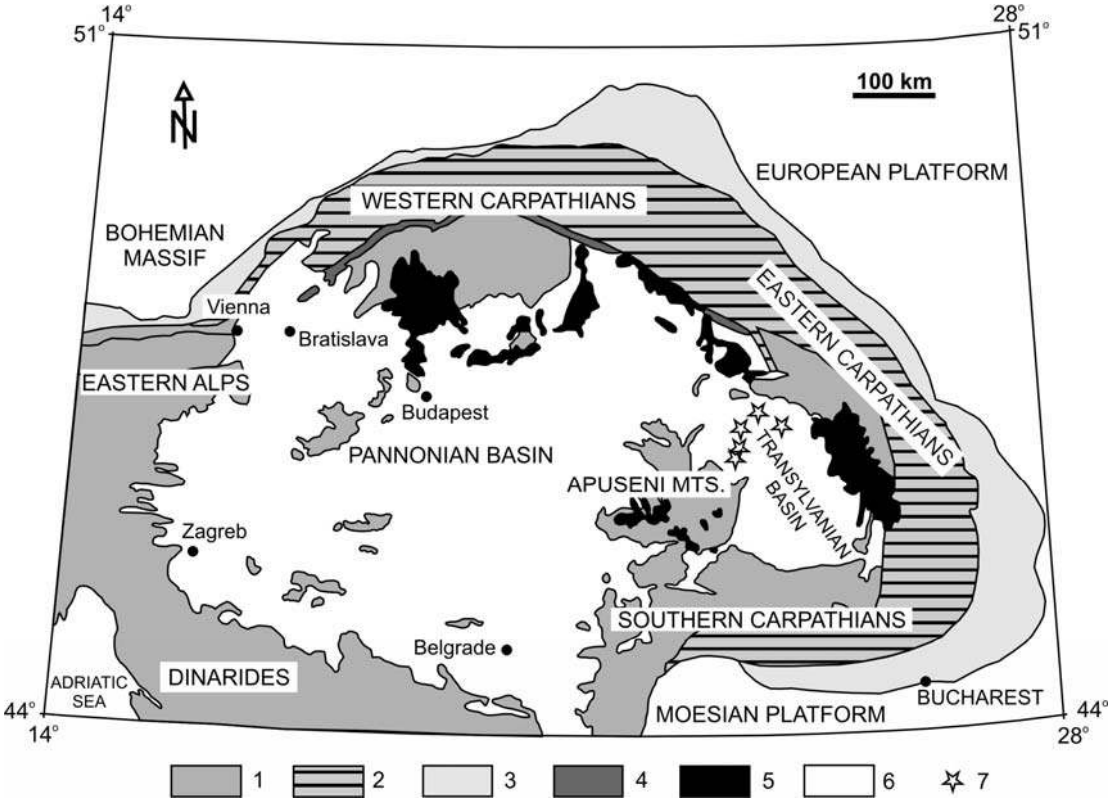

Fig. 1. Geological sketch map of Carpathian-Pannonian region (after Seghedi et al., 2004), showing the location of the samples used in this study. 1: Inner Alpine-Carpathian belt and Dinarides; 2: Alpine-Carpathian Flysch belt; 3: Molasse; 4: Pieniny Klippen belt; 5: Outcropping calc-alkaline volcanic rocks; 6: Neogene-Quater-nary sedimentary deposits; 7: Sample location.

Măgura Ciceului and Cepari localities. Zircon microcrysts were also collected from several ignimbrite and pumice samples. All these samples were selected to identify and describe melt inclusion microtexture and thermal history features by using petrography and microthermometry. Further silicate melt inclusion study is needed to elucidate the thermal history of each sequence in the "Dej Tuff" tephra.

3 Melt inclusion petrography

Doubly-polished thin sections were made on rock samples. and/or phenoclasts. The crystals were simply hand-picked under the binocular microscope. They consist of quartz and/or feldspar grains (0.5 to 3 - 4 mm) and zircon microcrysts (200 – 500 microns in length).

Five types of melt and fluid inclusions were observed and described at the room temperature (phase assemblage), as followings: mono-, and biphasic glass inclusions, multiphase glass inclusions, vapor-rich inclusions, opaque globular inclusions and microphenocrysts (microlites).

3.1 Monophasic and biphasic glass inclusions

A common silicate glass inclusion type described in many papers (e.g., Anderson, 2003) is a glass inclusion apparently "bubble-free" at room temperature. In this study it is proved that this kind of glass inclusions contain homogeneous silicate foam trapped before eruption. I based this observation mainly upon the fact that during microthermometry the apparently "bubble-free" inclusions shown silicate "foam-like" microtexture. During the heating procedure in the stage, above 500^{o}C the silicate glass inclusion has been transformed into a "foam-like" or "cloudy bubbly" microtexture. This behavior is robust being reproducible for several repeated cycles in the same inclusion.

The primary assemblages of "foam-like" glass inclusions trapped mainly in quartz and feldspar phenocrysts in the "Dej Tuff" display

cavities of different size (10 to more than 200 microns) and various shapes, from irregular to negative crystals forms. Generally, they are primary, distributed as zonal melt inclusions in feldspar, and as random clusters in quartz (Fig. 2 and Fig. 3). It should be noticed that the "foam-like"silicate melt inclusions described in this study are either monophasic (glass) or biphasic (glass + one or more bubbles), sometimes occurring in the same crystal zone, and both were frequently homogenized during microthermometry. After complete micro-thermometry analysis, it is concluded that the components of the inclusions studied acts as a compact homogeneous package made by bubbles and silicate melt. A contracting bubble is released during cooling, after trapping, and the package silicate + bubbles remained undisrupted in certain conditions (the ideal case). The same contraction bubble(s) ultimate disappeared during heating in the stage and allowed to measure the final homogenization temperature (e.g., *bc* in E, Fig. 2).

Severes et al., (2007) demonstrated experimentally that H_2O-loss during heating from the glass inclusion from quartz in Bishop Tuff is important, e.g., from 4wt % to 1 wt %, so we can expect that glass-bubbles assemblages resulted from thermal decrepitation and/or water-loss, become a disrupted silicate foam. Although in our reliability experiments by repeated microthermometric runs (around 30) the homogenization temperature was almost the same for the same inclusion (Fig. 7), suggesting that water-loss was not so important in this case. We can emphasized that the ideal "foam-like" glass inclusion" are apparently "bubble-free" at room temperature, and any glass inclusions with variable amount of bubbles suggest that the content was disrupted by thermal decrepitation and losses variable amount of water. It is evident from this study that 'foam-like" glass inclusions are thermosensitive and they decrepitated frequently during heating in the stage (Fig. 5 and Figs. 9 to 16). Moreover the darkness phenomena starting around Tg (glass transition temperature) and enhanced at $\alpha \rightarrow \beta$ quartz transition is probably just an optical feature of the enclosed "foamy-like" silicate melt, under the microscope (Fig. 14). Microtexture features of melt inclusions (glass ± bubbles) in quartz in the "Bishop Tuff" were interpreted by Wallace et al. (2003) to be related to water speciation in the silicate glass inclusions during cooling of the plinian-fall and pyroclastic flow deposits. In this study reheating episodes seem to be predominantly responsible for the generation of the observed microtexture features of the melt inclusions.

Pseudosecondary trails decorated with "foamy-like" glass inclusions were also observed in quartz and feldspar phenoclasts. In zircon they are elongated and seem to be mainly secondary melt inclusions trapped during overgrowth. Hourglass-shaped inclusions (Anderson 1991), are also frequent in quartz and feldspar phenoclasts in the pumiceous "Dej Tuff" pyroclastic deposits, and probably their initial content was also a "foam-like" silicate melt which was disrupted by depressurization on the narrow neck side of the hourglass inclusion.

3.2 Multiphase glass inclusions

This kind of melt inclusions contains one or more vapor bubbles and daughter minerals, or accidental trapped solid phases plus silicate glass. They are more frequently distributed in the inner core of the feldspar phenocrysts and are very rare in quartz. Sometimes the glass is a bubble-rich, "foam-like" silicate melt and the solid phases obviously did not melt during the heating cycles in the stage. The daughter minerals are represented by quartz or feldspar, sometimes biotite or amphibole. Their melting temperature is higher than our micro-thermometric capabilities (the used microther-mometric device is working only up to 1064°C), and they remain partially melted in the silicate melt, together with the vapor bubble(s). Probably these silicate melt inclusion are no longer "foamy-like" glass inclusions showing just one contracted bubble at room temperature.

3.3 Vapor-rich inclusions

A vapor rich phase formed mainly by H_2O, and perhaps some trace amount of CO_2, was

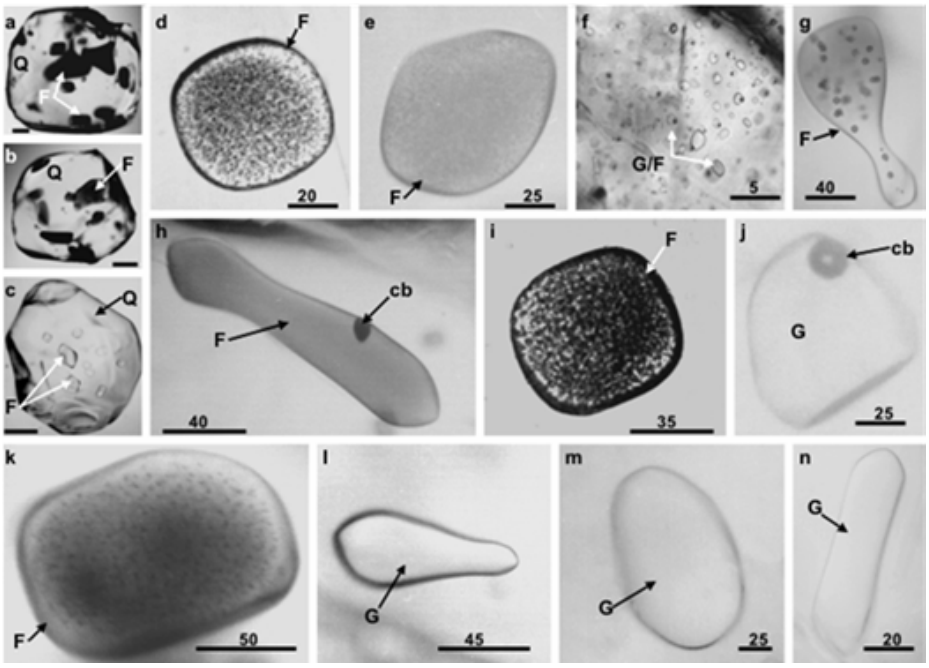

Fig. 2. "Foam-like"glass inclusions types in quartz phenocrysts from "Dej Tuff" formation, Transylvania basin, Romania. Quartz phenoclasts containing glass foam inclusions(F) partially decrepitated in **a.** and **b.**, **c.**-clear "bubble-free" glass inclusions; **d., e., i., k**.-natural reheated glass foam inclusions revealing silicate foam microtexture; **g.**-biphasic glass inclusions with glass+contraction bubble; **f.**-cluster of primary glass foam inclusions (F) in sanidine, some of them contain contraction bubble (G); **l., m., n.,**-silicate glass inclusions bubble-free which shown obviously homogeneous foam microtexture above 500^{o}C, and they homogenized on further heating in the stage (minimum trapping temperature).It should be noticed that during repeated microthermometric cycles some of them lost all bubbles and become "real" "bubble-free" silicate glass inclusions. Scale bar: a, b, c-250 μm; d, e, f, g, h, I, j, k, l, m, n-in μm.

trapped as monophasic gas inclusions. The CO_2 content was probably low and it is hard to be detected by microthermometry and/or crushing tests mainly because the abundance of tiny water vapor bubbles in the silicate glass. Sometime solid grains were observed inside the contraction bubble suggesting the presence of new crystallized solid phases. They are difficult to be distinguish from the false "vapor-like" microcavities formed by decrepitation of the former "foam-like" silicate melt inclusions which leaved behind empty voids that should be avoided for the study (e.g., Fig. 3g).

3.4 Opaque globular inclusions

This melt inclusion type is generally represented by Fe (-S-O) immiscible melt (Larocque et al. 2000; Pintea 2002), and suggest the presence of a mafic component input in the magma chambers. They could be recognized by their rounded shape and opacity under the microscope. A silicate blackish component (obsidian) could be also envisaged and more sophisticated analytical facilities are needed to be described. In the "Dej Tuff" phenoclasts the globular inclusions were trapped frequently in feldspar and zircon, rarely in quartz crystals and originated probably in the mafic melt batches, introduced time to time in the lower part of the magma chamber as immiscible blebs. They were also observed inside the silicate glass inclusions as immiscible globules (e.g., Fig. 3d). Fe-Ti oxide show similar opaque appearance under the microscope but they have polygonal external shape being totally crystallized, probably by fractional crystallization rather by immiscibility. Sometime only the shape and optic features are not enough to separate between immiscible globule and crystallized opaque solid microinclusion.

3.5 Microphenocrysts

Apatite, zircon and feldspar micro-phenocrysts are the most frequent microcrysts

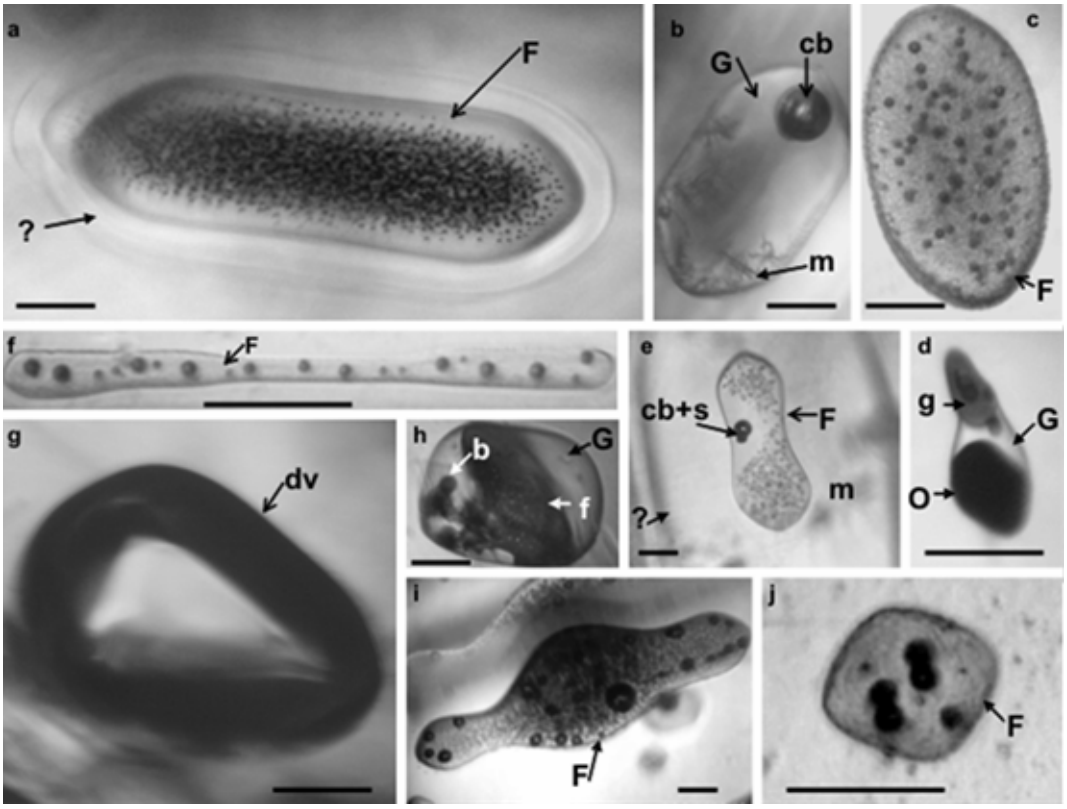

Fig. 3. More microtexture features of the "foam-like" glass inclusions from "Dej Tuff" trapped in quartz phenoclasts. **a.** Foam glass inclusion (F) surrounded by unknown halo(?), perhaps another silicate melt (?), **b.** Glass + contraction bubble (+ CO_2-?) and microlites (m), **c.** Partly devitrified foam glass inclu-sion (F), **d.** Multiphase glass inclusion containing glass (G), opaque globule (O) and probably another droplet of immiscible silicate melt (mafic-g), **e.** Foam glass inclusion with one contraction bubble and an attached solid phase (cb+s) surrounded by an unknown ellipsoidal halo (silicate glass -?), **f.** Elongated secondary glass foam inclusion (F), partially decrepitated showing several ripened-? bubbles(b), **g.** Empty void(dv) left behind after complete decrepitation of a glass foam inclusion, **h.** Glass foam inclusion partially decrepitated containing glass(G), fissure (f) and bubbles (b), **i., j.** Typical partially decrepitated glass foam inclusion which never homogenized during the heating procedure in the stage.

trapped as solid inclusions beside the "foamy like" silicate glass inclusions", mainly in zircon and feldspar, rarely in quartz. Some of them look-like tubular solid inclusions and contains silicate glass + bubbles which in several cases homogenized during heating procedure in the stage around 830-840°C (n=5), sometime up to 900°C in a plagioclase phenoclast from Şoimeni, for example. They were also described elsewhere by Davidson (2004) which pointed out some of their characteristics as such: microprobe analyses indicate an $Ab_{75}An_{20}Or_5$ component; euhedral form; uniform distribution in all samples; often decorating the growth zones; they are not daughter crystals; they have macrophenocrysts correspondent and can be used to distinguish between microphenocrysts and melt inclusion. The microphenocrysts enclosed in magmatic mineral is a common feature and were described as very tiny microcrysts (<20μm), randomly distributed in macrophenocrysts, decorating growth zones and most often were trapped accidentally in silicate glass inclusion. The acicular feldspar microphenocryst were also described by Ştefan et al., (1982) in the upper Cretaceous quartz-monzodioritic porphyry in the Birtin dyke-like intrusion, and probably they are related to the pressure variation during melt decompression in ascending magma. Similar acicular product was observed by Keppler and Audetat (2005), by dissolving, in pure water, a piece of andesite in the diamond anvil cell at 854°C and 11.7 kb. Squeezing interstitial melt from a deep-seated batolith (granodioritic or andesite-dacite) and rapid ascension during decompression from high

pressure may be the origin of these microphenocrysts, probably flushed-up by the new hot melt batches in the crystallizing molten magma chamber (e.g., Bachmann and Bergantz, 2004).

4 Microthermometry of glass foam inclusions

The microthermometric features of the "foam-like" silicate glass inclusion during the temperature variations as observed under the microscope are depicted in two examples in Fig. 4 and Fig. 5, respectively. In the first example (Fig. 4) a clear bubble-free glass inclusion is present at room temperature conditions, while in the second one (Fig. 5), the devitrified natural silicate foam was remelted. The spontaneous nucleation of a cloud of bubbles in the silicate melt, i.e. "foam-like" silicate melt, started usually at temperatures ranging between 500°C and 550°C with a maximum number of bubbles (up to 80% in the cavity volume) observed around 590°C and 600°C. It should be noted that the bubble cloud becomes clearly visible early around 500°C and 530°C, and in quartz crystals it is more enhanced during the alpha/beta transition around 573°C. In feldspar phenoclasts the behavior of melt inclusions during heating in the stage is almost the same with no change in the volume ratio between bubbles and melts. In this case bubble nucleation cannot be related to the transition above mentioned for quartz. It may be concluded from these observations that the "foam-like" silicate melt formation is influenced mostly by the temperature and pressure inside the melt inclusions. In addition, the cloud of bubbles seems to be, in many cases, the remaining silicate foam microtexture with darker appearance because the total reflexion under the optical microscope. In the open cavity under SEM microscopy, Clocchiatti (1975) evidenced a perfect "alveolaire" microtexture showing a channel network of very fine glass walls and named them "ponces". A comprehensive physical description of the silicate foam texture was given by Cashman and Mangan (1994).

On further heating the "foam-like" microtexture becomes unstable and complete homogenization in the silicate melt state was recorded between 710°C and 900°C, with a maximum frequency between 750°C and 830°C, for n = 344 measurements (Fig. 6). It is important to mention that the "foam-like" microtexture is generally stable between 500°C and 600°C, observed in the microthermometric stage, and the final homogenization is due by the disappearance of the contraction bubble which generally has the largest diameter above Tg (around 530°C, coincidently with glass transition temperature for rhyolitic magma, i.e. Bagdassarov et al. 1996) and it remains frequently the only one bubble above 600°C. In the ideal case we suggest that the "foam-like" cell-microtexture would release by contraction a single vapor bubble, similar with the normal homogeneous silicate melt which shown only one bubble during contraction, as temperature decrease below the homogenization value. Correspondingly, we described here homogeneous silicate foam with one contracting bubble, which become homogeneous by heating up in the stage or heterogeneous (disrupted, decrepitated) foam when there are more than one bubble and they cannot be homogenized during microthermometry.

Each microthermometric measurement was replicated twice (sometime more) and generally the results obtained appear robust (Fig.7). Although it is observed that final homogenization temperature tends to decrease after multiple microthermometric cycles (on the contrary, if decrepitation occurred, the homogenization temperature increased in "normal" silicate melt or fluid inclusion). It is presumed that these phenomena could be related to the "foam-like" characteristics rather to the post-trapping event or thermal decrepitation in the heating stage. Anyhow the silicate "foam-like" seems to be fragile and sensitive to many external factors including P-T variations, or even during sample preparation procedure. This could be easily evidenced during microthermometry by heating quartz sample containing "foam-like"

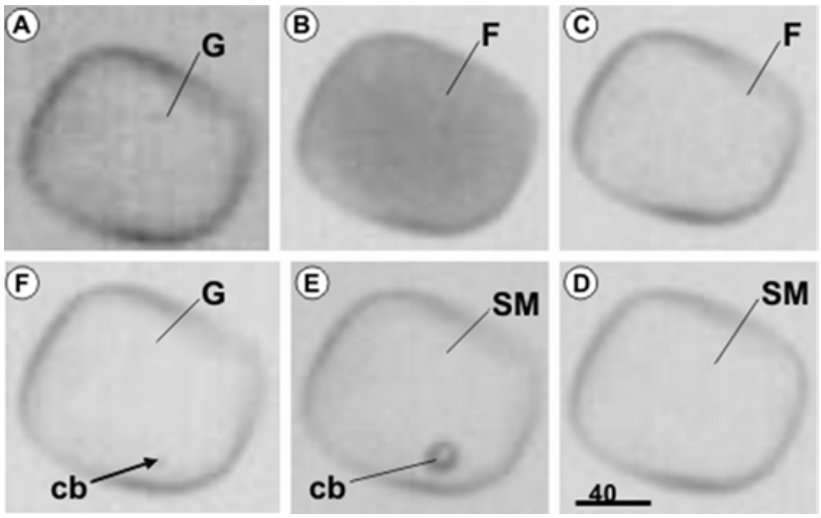

Fig. 4. Microphotographs sequence of a microthermometric cycle in a "foam-like" glass inclusion apparently monophasic ("bubble-free") at room temperature from magmatic quartz phenocrysts from Dej tuff. **A.** Clear silicate monophasic glass inclusion, G- glass; **B.** Silicate foam revealed during the heating procedure in the stage at 580°C by spontaneous bubbles nucleation starting around 500°C, F - silicate foam; on further heating the bubbles vanished progressively showed in **C** at 770°C and final homogenization temperature was reached at 810°C in **D**, SM- silicate melt; **E.** On cooling back by cutting- off the power supply, a single contraction bubble (cb) nucleated in the silicate melt (SM) around 590°C, became more shrunken (cg) in the glass (G) at room temperature conditions as it shown in picture **F**. In the replicated cycle the same temperature values were recorded for the mentioned phase transitions. Scale bar in μm.

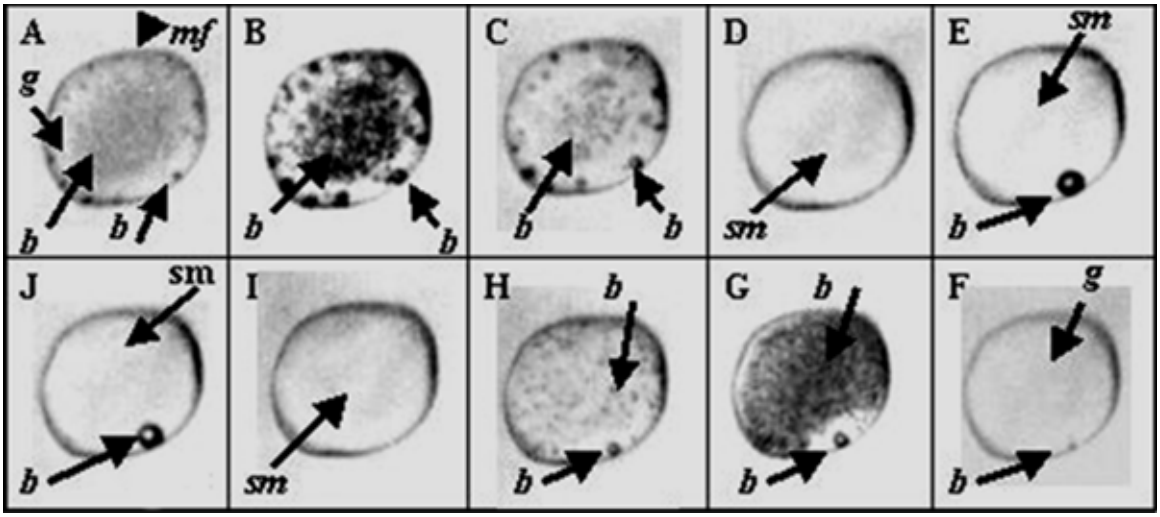

Fig. 5. Microthermometry sequences in a natural devitrified "foam-like" glass inclusion (FGI) from magmatic quartz phenocryst in ignimbrite from "Dej Tuff" formation.

A. At T= 25°C, the natural magmatic foam (***mf***) is composed by a large number of gas bubbles (***b***) as cloudy area and white glass (***g***); **B.** and **C. -** intermediate values at T=611°C and T=681°C, respectively; On further heating the foam is unstable and progressively the bubbles disappeared and the cavity contains at the final homogenization temperature a clear silicate melt showed in picture **D** at Th=764°C (***sm-*** homogeneous silicate melt). During cooling back, at T=662°C a single bubble was renucleated in cavity, and the foam became invisible as shown in picture **E** (***b*** - gas bubbles, ***sm*** - silicate liquid). Back to the room temperature, the cavity contains now only one small shrunk bubble (***b***) and glass (***g***), showed in picture **F.**

In a replicated microthermometry cycle, after five days, the silicate foam renucleated spontaneous around 500°C with a maximum of bubbles number around 570°C; the pictures **G** and **H** were taken at 557oC and 696°C, respectively. The final homogenization in the silicate melt was measured now at Th=735°C, shown in picture **I**, ***sm***- silicate melt; a shrinkage bubble was renucleated during decreasing of temperature in the stage, below 700°C showed in picture **J** (***b*** - bubble, ***sm***- silicate melt). ***FGI*** length = 60 μm.

silicate glass inclusions" depicted in the serial microphotographs from Fig. 8 to Fig. 16.

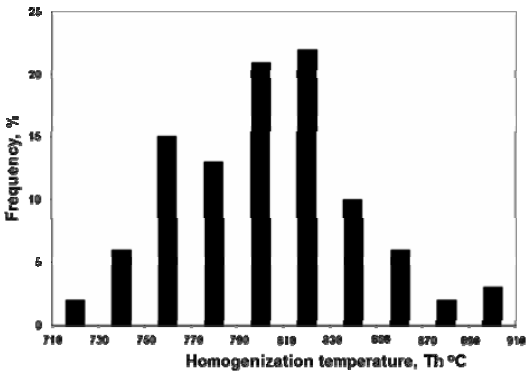

Fig. 6. Histogram of the homogenization temperature values from primary "foam-like" silicate glass inclusions in quartz, feldspar, and secondary in zircon, from the rhyolitic "Dej Tuff" formation from Transylvania basin. All measurements were done in a "home-made" microthermometry stage calibrated with potassium dichromate (398°C), halite (800°C), silver (961°C) and pure gold (1064°C), the precision of measurements ranged between +\- 20°C.

5 Discussions

Theoretically the silicate foam was defined by Cashman and Mangan (1994) as containing a dispersed gas phase of > 74% in volume, and more strictly the conceptual model of the foam structure-cells claims that rigid spheres attained maximum packing from 74.05% to 85% filled space. The bubble number density and nucleation rate depend on the supersaturating pressure and water content of the melt (Mangan and Sisson, 2000). For example, bubble number density is about 9 x 10^6 at 175 MPa in a rhyolitic melt (Mangan et al., 2004). The same authors mentioned that the range in bubble number density observed in Plinian air-fall pumice ranged between 10^8 - 10^{11}/cm^3. It is obvious that such data cannot be obtained by direct counting of bubbles number in the "foam-like" silicate glass inclusions by visual estimation under the microscope. Anyhow we estimated that up to 75 - 80% volume of the gas phase (H_2O vapor bubbles) in the studied cavities was distributed in the necessary number of bubbles to be structured as foam (see Figs. 2, 3, 4). Consequently, it is more practical to name them "foam-like" silicate glass inclusions instead of normal silicate melt inclusions, because the silicate foam released a contracting bubble during decreasing temperature, suggesting a compact package of cell- microtexture between bubbles and silicate melt. Any change in this behavior would imply the presence of a heterogeneous or disrupted silicate foam. To be more convincing the reader is also advised to see Fig. 1c, 1g, as well as Fig. 2e, 2j in the papers of Skirius et al. (1990) and Fig. 1g, e, I, k, l, m, from Wallace et al. (2003) which shown similar feature on bubble – melt relation as resulted in our study during cycling microthermometry, but the mentioned authors have different explanation for the showed microtexture features. Similar pictures with "foam-like" silicate glass inclusions were presented from various volcanic samples e.g., by Clocchiatti (1975), Roedder (1979), Anderson (1991, 2003), Anderson et al. (2000), Frezzotti (2001), and Lowenstern (2003).

The first attempt to introduce bubbles dynamics of a viscous silicate melt (i.e. up to silicate foam) as a possible mechanism of volcanic eruption was mentioned long time ago (e.g., Rittmann, 1936). Since then a lot of data on silicate foam were accumulated mainly from glass technology (e.g., v deer Schaaf and Beerkens, 2002), theoretical approaches (Proussevitch et al. 1993) and experimental study on bubbly silicate melts, presented by Hurwitz and Navon (1994), Sparks et al. (1994), Cashman and Mangan (1994), Mangan and Sisson (2000), Mangan et al. (2004), Martel and Bureau, (2001) and references therein.

Based on microthermometry data and petrographic observations presented in this study, such silicate foam would be formed in the upper part of the magma chamber, very close before eruptions, as it was suggested elsewhere by Stix (2007), as the main effect of the episodically hotter mafic magma influx in the

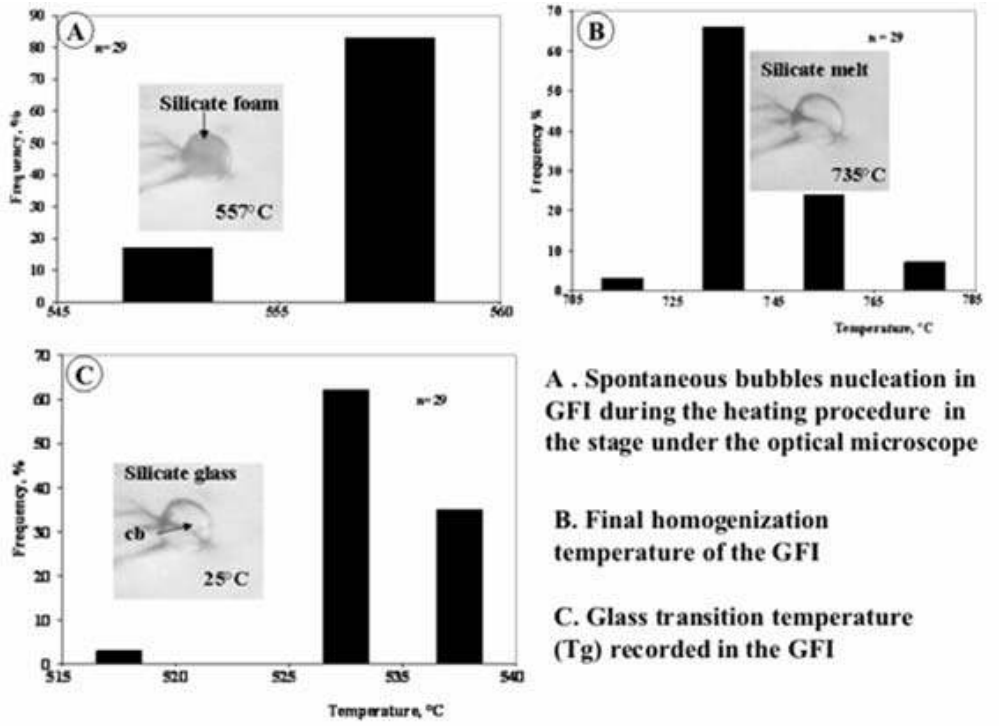

Fig. 7. Reliability data for a "foam-like" glass inclusion in quartz phenoclasts from the "Dej Tuff" formation. The same inclusion was repeatedly heated up until final homogenization temperature was achieved and then it was cooled- back to room temperature (n= 29). During heating, spontaneous bubble nucleation temperature in **A** and final homogenization temperature in **B** were recorded. Glass transition temperature (Tg) was recorded during cooling- back and was based upon the sudden collapsed bubble (s) in the silicate melt/glass around 530°C, depicted in **C**.

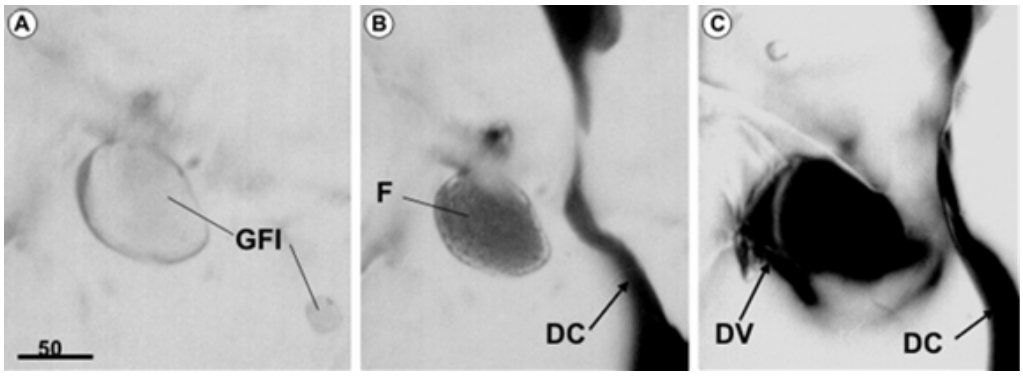

Fig. 8. Violent decrepitation of a "foam-like"silicate glass inclusion" (**F**) during heating procedure in the stage. A fissure (DC) was developed starting at 460°C and he silicate foam is completely revealed at 590°C (**A**, **B**). Increasing pressure and the proximity of the fissure caused explosion before homogenization temperature (**C**).**A**. 25°C, **B**.590°C, **C**.700°C.

rhyolitic magma chamber (Wark et al. 2007). In the Dej Tuff majority of the minerals were crystallized from this kind of rhyolitic silicate foam because they contain more than 95% of the melt inclusions in the form of "foam-like" silicate melt inclusions as primary assemblages. Many of them decrepitated partially or completely during rising temperature induced probably by the new hotter vapor rich mafic magma influxes in the crystallizing rhyolite magma, as it was mentioned above. In fact this would be the principal mechanism of crystal fragmentation as it is documented in the literature (e.g., Bindeman, 2005) and demonstrated in this study by microthermometry (e.g., Fig. 14). The presence of microphenocrysts

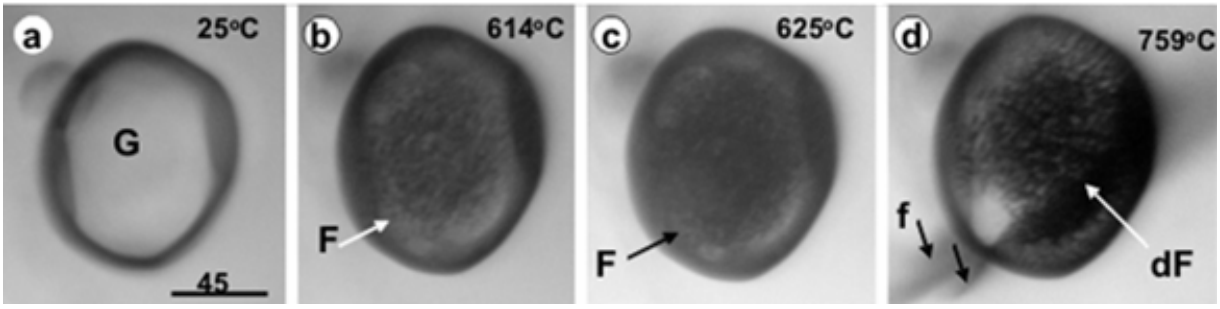

Fig. 9. "Foam-like" silicate glass inclusion "bubble-free" at room temperature (**a**) developed on heating a blackish microtexture, probably because total reflexion of the silicate foam microtexture (**b** and **c**).On further heating this inclusion decrepitated around 750°C (**d**) and a microfissure system developed (f). Cannot be homogenized on further heating and cooled-back below 500°C in this microtexture- feature as indicated in **d**.

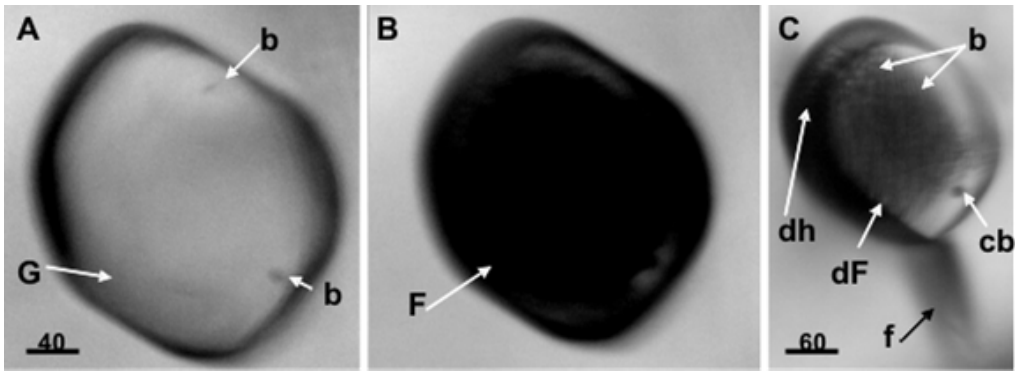

Fig. 10. Silicate glass inclusion with several bubbles (b) visibile at room temperature conditions (A) revealing a dark foam microtexture at 625°C (B) decrepitating around 750°C and formed a characteristic halo(dh) and micro-fissure system (f) back to the room temperature (C). A separate contraction bubble (cb) and more bubbles (silicate globule? + vapor bubbles) - b in the heterogeneous silicate foam microtexture are shown in C. This remained unchanged back to room temperature.

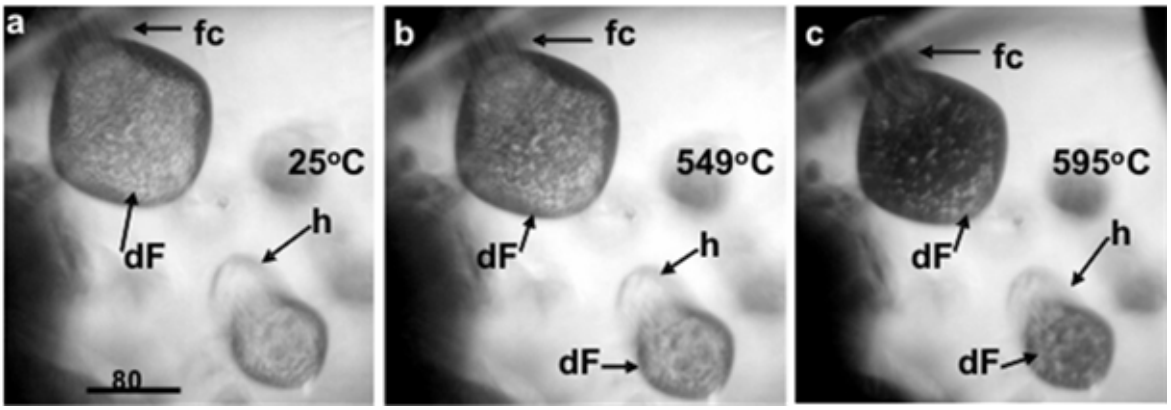

Fig. 11. Naturally decrepitated "foam-like" silicate glass inclusion showing a globular microtexture in (a) at room temperature. fc-microfissures, h-natural decrepitation halo. During re-heating in the stage the microtexture become darkness (dF) and globular microtexture was more evident after glass transition temperature (Tg) and temperature (α→ß) of quartz transition (b and c). On further heating cannot be homogenized.

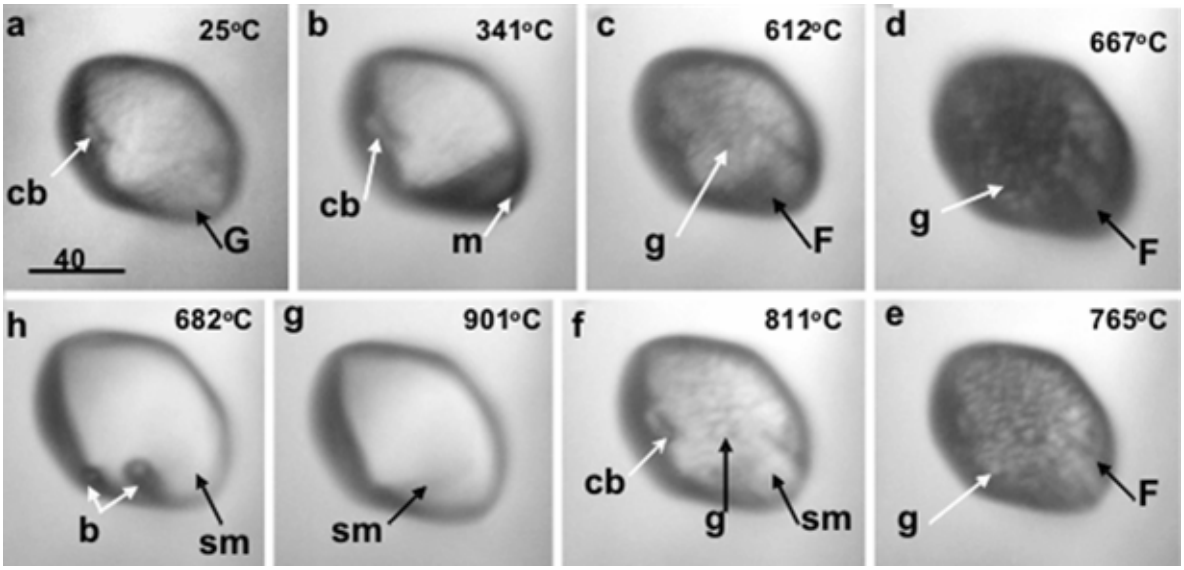

Fig.12a. Complete cycle of microthermometry in a "foam-like" silicate glass inclusion which is almost "bubble-free" at room temperature (**a**), but shown a microglobular microtexture on the surface and contained a contraction bubble-cb. The optic-dark silicate foam microtexture-F was revealed starting in (b) showing globular microtexture in (**c**), (**d**), (**e**), and (**f**) .The contraction bubble-cb**,** homogenized at presumed trapping temperature of 901°C in (g). Back to room temperature, around 685°C two bubbles-b renucleated showed in (**h**). Other notations: m- meniscus between glass and wall cavity, g- silicate globule -?, sm- silicate melt.

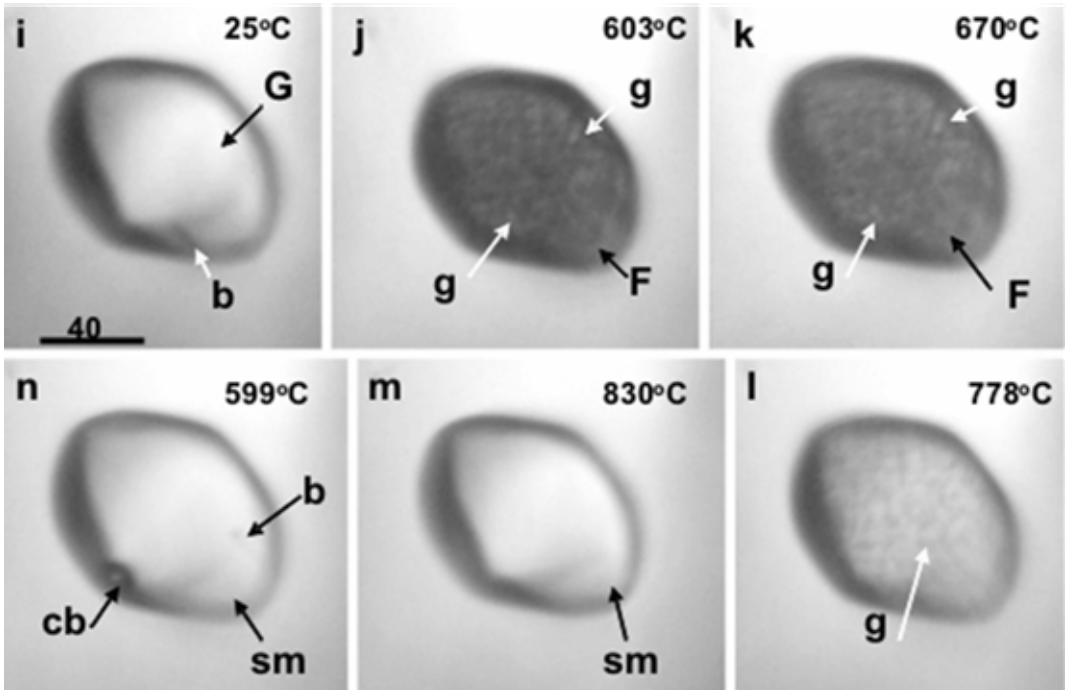

Fig. 12b. After one day the replicated microthermometric cycle showed the same behavior. At room temperature in (**i**) glass-G and a single contraction bubble-b (?) is present. The foam is revealed between 600°C and 700°C (**j** and **k**), and final homogenization temperature was recorded at 830°C (*71 °C lower than in the first cycle. This is a characteristic feature during repeated microthermometry in "foam-like" silicate glass inclusions, that is the temperature become lesser probably because of the foam characteristic*). Notations: F- silicate foam,. S- silicate globule-?, silicate – melt, cb- contraction bubble, b- another bubble.

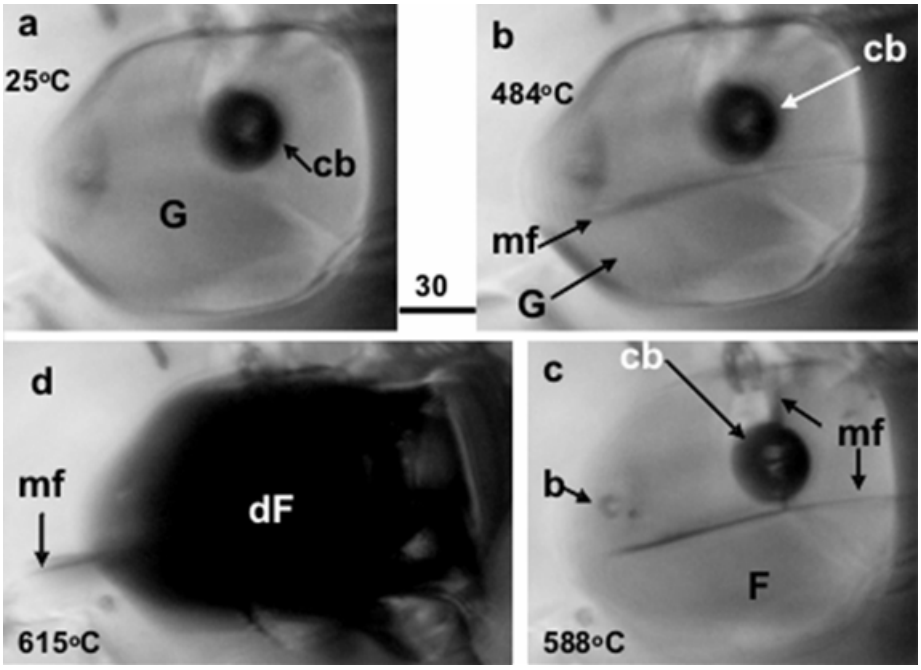

Fig.13. Experimental decrepitation of a "foam-like" glass foam inclusion during heating procedure in the stage. Notations: G-silicate glass, mf- microfissure, F- silicate foam, b- bubbles, dF- decrepitated silicate foam.

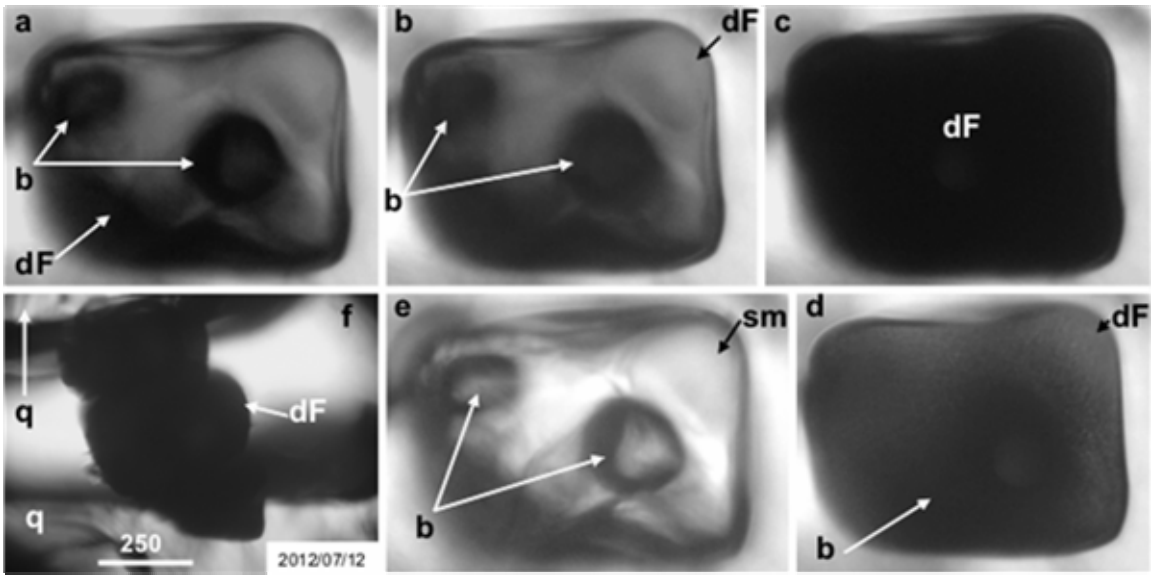

Fig. 14. Naturally decrepitated "foam-like" silicate glass foam inclusion containing two bubbles in a brownish glass microtexture at 25°C (a). During heating in the stage the silicate foam microtexture is reactivated between 625°C (b) and 686°C (c). This start clearing around 738°C (d) and become colorless at 837°C (e), and the two bubbles reduces slightly their size. At 841°C (e) the inclusion decrepitated so violently than the quartz (q) was broken in two pieces. The size of "foam-like" silicate glass inclusion did not change, it was open on the edge and the silicate foam (df) liberated and increased volume about *five times*, but all the stuff remained attached to the host quartz broken pieces in the stages, and looks like a "micro-/mini eruption".

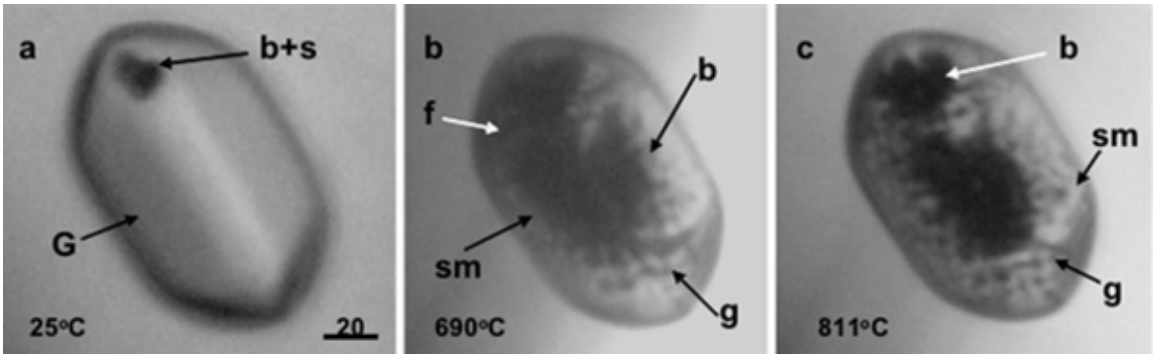

Fig. 15. “Foam-like” silicate glass inclusion contain glass and bubble plus attached solid microlites at room temperature (a).During heating in the stage a microfissure (f) was formed at 321°C and between 558° and 690°C (b) a mixture of bubbles (b), silicate melt (sm) and silicate globules (g) were released. Around 811°C (c) some vapor bubbles (b) coagulated and are coexistent in the heterogeneous silicate foam with silicate globules (g) and silicate melt (sm). It cannot be homogenized and on further heating decrepitated around 882°C along the c-axis and could not be relocated again.

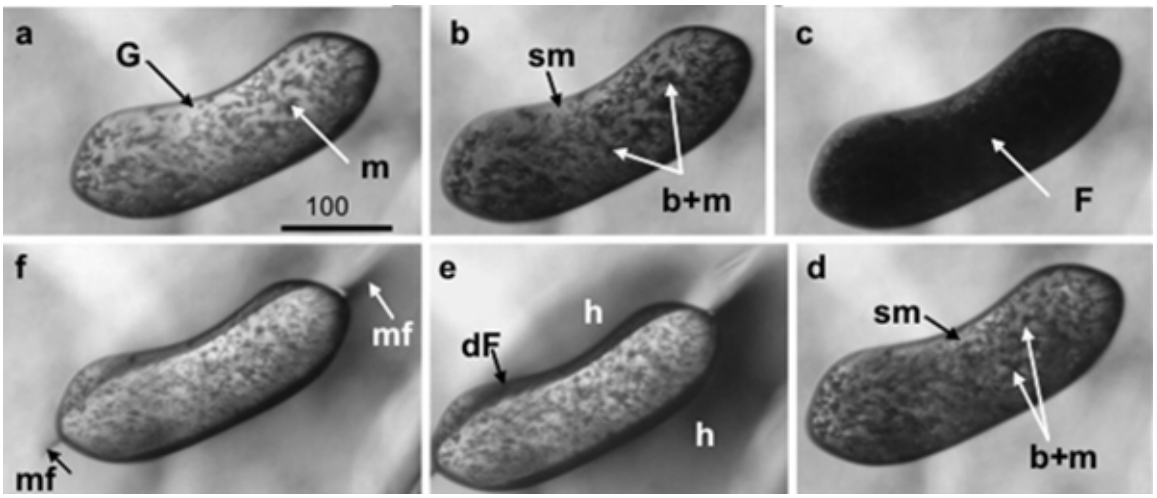

Fig. 16. Large glass “foam-like” silicate glass inclusion contains glass (G) and microlites (m) at room temperature (**a**). It start blackish around 597°C (**b**) and become complete dark around 611°C (**c**). On further heating, around 762°C (**d**) the silicate foam (F) is almost clear again and now contained silicate melt-sm, bubbles-b and microlites-m. The inclusion decrepitated violently at 776°C and a decrepitated halo (DF) was formed along a microfissure (mf), showed in (**e**). Back to room temperature the inclusion contains glass + microlites, but a lot of gas bubbles were lost. Microphotographs (**e**) and (**f**) were fixed at different size to see the fissure (mf) and the decrepitated halo (h).

as microinclusions in the same phenoclasts suggest rapid pressure changes during melt accessioned toward the shallow layers. The microthermometric experiments undertaken in this study show that the volatile-rich silicate melt “foam-like” was stable only up to 500° to 600°C in the microthermometric stage and by similarity at the same temperature in the magma chamber. Above these values under increased internal pressure in the cavities, during heating in the stage (or naturally heating in the pressurized magma chamber by the hotter mafic magma) the volatile-rich ”foam-like” silicate melt would be homogenized in a single melt phase, and decrepitated on further heating. As a consequence, we presume that a volatile supersaturated and homogeneous silicate melt was trapped in minerals growing in magma chamber prior eruption between 730° and 850°C, which is in good accord with other temperature estimation based on Fe-Ti oxide by Szakács (2000) in the Dej Tuff (more data for

Table 1. Comparative data between homogenization temperatures measured in "foam-like" silicate glass inclusions from the Dej Tuff and calculated temperatures by various geothermometers including melt inclusion microthermometry in several well known acidic tuff formations.

Acidic tuff formation	**Fe-Ti oxide, T, ^{o}C**	**Zr sat. T, ^{o}C**	**Δ^{18}(Qz-Mt)T, ^{o}C**	**Titan i-Q, ^{o}C**	**Melt Inclusions Th, ^{o}C**	**Author(s)**
Dej Tuff, Romania	604-892	-	-	-	"foam-like"silicate glass inclusions" 710-900 (+/-20) (n=344)	Szakacs, 2000; Pintea, 2005, 2007 and this study
Toba Tuff, Indonezia	701-847 2.6-3.4 kb	-	-	-	~ 960 in Toba quartz, sanidine and plagioclase (Beddoe-Stephens et al., 1983)	Cesner,1998
Bishop Tuff, USA	714-818	762 -801	714 -817	712-742 (cores) 700-810 (rims)	800-900,~ 2kb, (quenching experiments)	Bindeman and Valley,2002; Hildreth and Wilson, 2007; Wark et al., 2007
Taupo Volcanic Zone	690-990 (Shane, 1998)				820-850	Szakacs,2000; Kamenetsky and Danyushevsky, 2005

comparison in Table 1). The trapping pressure was roughly estimated by comparing the microthermometric behavior of the included "foam-like" silicate melt during heating procedure in the stage with the experimentally results published by Mangan and Sisson (2000) on homogeneous and heterogeneous bubble nucleation processes. In this respect the pressure was estimated between 200 MPa and 20 MP, respectively. The depth of magma chamber where the melt inclusions were formed could be estimated around 7 Km in the case of homogeneous trapping.

In this study the presence of silicate foam was documented mainly on the microtexture changes observed during heating in the microthermometric stage. Another method such as SEM microscopy would be necessary to reveal the complete imaging on these microtextures inside the phenoclasts, as Clocchiatti (1975) have already done in many others quartz samples. At larger scale, relicts of foam-like microtexture could be observed in the groundmass in thin sections of the rhyolitic ignimbrite and pumice lapilii, where the silicate glass is composed by different types of collapsed vesicles (see also Seghedi and Szakács, 1991). Such microtexture was described earlier by Istrate (1975) in the rhyolitic ignimbrites in the Laramian formation in the western part of the Vlădeasa massif from the North Apuseni Mountains (western Romania).

The breaking of phenocrysts by decrepitation of the "foam-like"silicate glass inclusions was another important feature, recorded during overheating in the microthermometric stage (Fig. 8 to Fig. 16), suggesting that this process enhanced the crystal fragmentation during the explosive volcanic eruptions. The accumulation mechanism of a silicate foam layer and their stability was pointed out at Masaya volcano (Nicaragua) by Stix (2007). Similarly, we suggest in the case of Dej Tuff, that the silicate foam is accumulated in the top of the magma chamber (trapped in the crystallizing minerals which were fragmented under the heating

episodes from below) at the base of the volcanic conduit(s), as an immiscible layer, and exploded when some critical parameters are reached. It was demonstrated experimentally by Cottrell et al. (1999) at Santorini that the episodic intrusion of magma, provided the necessary heat and perhaps contributed to the ascent of the volatile-rich magma to shallow crustal depth and probably the mafic injection continued until trigger the cataclysmic eruption by decompression. An analogue behavior is suggested in this study by the trapped "foam-like" silicate melt in quartz, feldspar and zircon (majority becoming phenoclasts after fragmentation because "foam-like" silicate foam decrepitation) by petrography and microthermometry at a "micrometric magma chamber scale".

6 Conclusions

Based on petrography and microthermometry of "foam-like" glass inclusions" in minerals of the "Dej Tuff", the following conclusions can be drawn:

- the studied minerals from the Dej Tuff contain more than 95% "foam-like" silicate glass inclusions as mainly primary assemblages in quartz and feldspar phenoclasts and secondary "foam-like" inclusions in zircon;
- the "foam-like" silicate glass inclusions are various in shape, size and distribution and they are "bubble-free" (glassy monophasic) or contained various ratio between glass and bubbles at room temperature conditions;
- their formation temperature ranged between 710-900°C (±20°C) with a maximum of frequency between 750° to 830°C (n=344) in very good agreement with the Fe-Ti- oxide geothermometer used at Cepari by Szakács (2000) and calculated the formation temperature between 604°C and 892°C;
- the estimated trapping pressure ranged between 200 MPa and 20 MPa, suggesting a depth for the magma chamber around 7 Km based upon homogeneous bubble nucleation in "foam-like" silicate glass inclusions;
- the silicate foam was identified firstly in this study by visual observation during heating in the microthermometric stage of the apparently "bubble-free" silicate glass inclusions and named "foam-like" silicate glass inclusions;
- many of the "foam-like" glass inclusions decrepitated during the heating experiments, mainly the bigger ones (i.e. more than 150 microns) and coalesced bubbles were formed and the cavities never homogenized again. Frequently the empty decrepitated "foam-like" glass inclusions appearing opaque cavities (empty) under the optical microscope. The darkening phenomenon revealed during heating in the microthermometric stage seems to be the effect of the foam skeleton (microtexture) totally reflected the transmitted light (e.g., Fig.14);
- during reliability test the "foam-like" glass inclusions", apparently "bubble-free" at room temperature revealed a compact package microtexture into silicate foam around 500-600°C during heating, and the recorded temperature were generally reproducible (Fig.7). No leakage was observed during replicated measurements, but the partial decrepitated cavities never homogenized again. It must be stressed that often during microthermometry, the repeated cycles in the same cavity shown homogenization temperature less than the precedent one, which is contrary in the normal silicate melt inclusions, and this is perhaps a specific characteristic to the trapped silicate foam or because the water-loss during experimentally (or naturally) heating and partial decrepitation (decompression);
- there are many reasons for decrepitation of "foam-like" silicate glass inclusions" including natural reheating episodes and partial decrepitation during the explosive events or artificially during sample handling and preparation; The larger "foam-like" silicate glass inclusions (more than 150 microns) decrepitated easily and this seems to be the main cause of mineral fragmentation (see also Bindeman, 2005);
- the entrapped silicate foam behave in a similar mode at a micrometric scale such as the silicate foam accumulated in the top of the magma

chamber at the base of the volcanic conduit as an immiscible layer, and would explodes when some critical parameters are reached. The hotter volatile-rich mafic recharging generates more and more silicate foam in the upper level of the crystallizing rhyolitic magma chamber, perhaps triggering the cataclysmic volcanic eruption. In conclusion, the "foam-like" silicate glass inclusions behaved during the microthermometric experiments in this study as a micro-magma chamber analogue.

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The Căpâlnaş-Vorţa-Nevoiaş (Pb-Zn-Cu) metallogenetic district from the Banatitic Province, Romania

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Abstract: The Căpâlnaş-Vorţa-Nevoiaş metallogenetic district is superposed over the northern part of the Lăpugiu Basin Banatitic volcanics. Its mineralizations belong to the Pb-Zn-Cu–type. They are located along some E-W trending fractures. Along the fractures the host ophiolitic and island arc volcanics from the basement have been strongly affected by the hydrothermal solutions, released by some Banatitic plutonic intrusions from the depth of the district. It is argued that the tectonic movement on the faults was active during the emplacement of the mineralization, causing the occurrence of broken pyrite crystals cemented by chalcopyrite.

Keywords: Metallogenetic district; occurrences; mineral assemblages; chemistry; origin.

1 Introduction

The Banatitic metallogenetic province extends from Romania through Serbia to the southeastern part of Bulgaria, accompanying the Banatitic magmatic province, described by Cotta (1861), Giuşcă et al. (1965; 1966) and Ciobanu et al. (2002). On the territory of Romania, it includes several metallogenetic districts. The Căpâlnaş-Vorţa-Nevoiaş metallogenetic district extends along the Mureş River, so that it could be called the Mureş district, as well. It is located within the boundaries of the Banatitic Metallogenetic Province (Fig. 1). This district is characterized by a Pb-Zn-Cu mineralization, which will be described below. The data presented here support the fault-related genesis of the mineralization from Vorţa district and argue for its emplacement while the tectonic movements were still active.

2 Geological structure of the district and its related magmatic rocks

As the mineralizations from this metallogenetic district are related to the Laramian (Banatitic) magmatic rocks, which erupted by the end of the Upper Cretaceous, they affected the geological formations from the southern part of the Mureş ophiolitic suture (Savu, 2007). In the structure of the district, there are two structural levels: an infrastructure (basement) and a superstructure. The oldest host rocks from the infrastructure are the Jurassic ophiolitic rocks, covered by Late Kimmerian (Upper Jurassic) island arc volcanics. The ophiolites are represented by tholeiitic basalts, anamesites and dolerites, rocks with a MORB signature, and layers of red jaspers and limestones. These rocks make a pile of magmatic and sedimentary formations, intruded by layered small bodies of gabbro and ultramafic rocks.

The Upper Jurassic island arc volcanics consist of calc-alkaline basalts, andesites and acid rocks (dacites and rhyolites). These rocks are associated with red jaspers and argillites, Stramberg limestones and rare sandstones. The Stramberg (Upper Jurassic) limestones form a sort of reef barrier. The remnants of this reef barrier extend from Căpâlnaş, in the west, through Zam, to Vălişoara, in the east. According to Kadič (1906) and Papiu (1954) the reef limestones was formed during the Dogger. Neocomian sedimentary formations occur mostly in the eastern part of the district, where they are represented by the marly-argillaceous

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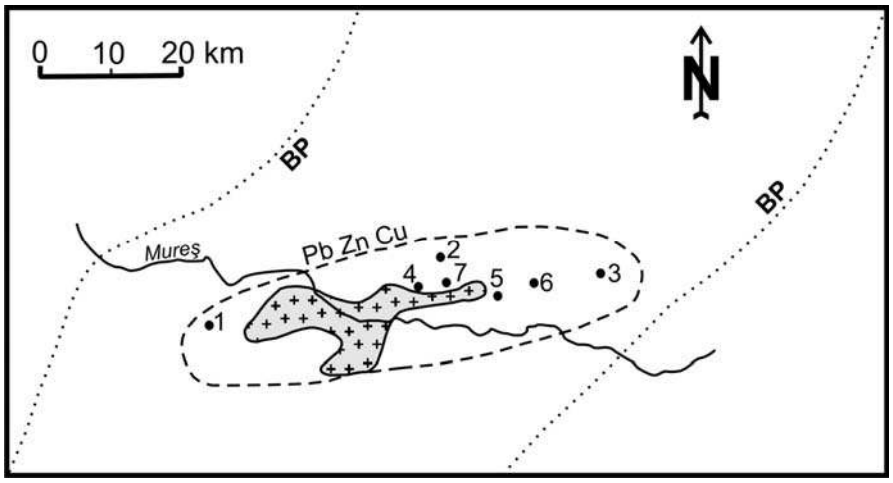

Fig. 1. Sketch map showing the position of the metallogenetic district within the median segment of the Banatitic Metallogenetic Province and the distribution of the Pb-Zn-Cu mineralizations. Compiled after the Metallogenetic Map of Romania scale 1:1,000,000 (1969) and the Tectonic Map of Romania scale 1:1,000,000 (1970), with the author's additions. BP, boundary of the Banatitic Metallogenetic Province; Pb-Zn-Cu broken line marks the boundary of the metallogenetic district; occurrences of Pb-Zn-Cu mineralization: 1, Căpâlnaş; 2, Vorţa; 3, Nevoiaş; 4, Vlădeşti; 5, North Târnava; 6, Sulighet; 7, Valea Mare; +, Banatitic volcanics.

deposits mentioned above; manganese accumulations and red jaspers are associated with the Neocomian sedimentary and were designated as a manganese metallogenetic district on the Metallogenetic Map of Romania, scale 1:1,000,000 (1969). These formations are covered by Cenomanian and Senonian deposits, which support the Laramian volcanics from the superstructure.

The Early Laramian tectonics of the region is represented by one ESE-dipping reverse fault, which extends from Groşi and Tisa to the Gurasada Brook. This fault was also indicated by the geophysical research, as a gravimetric axis (Andrei and Cristescu, 1966). Along this fault, the anchimetamorphic Groşi Formation is thrust over the Mureş ophiolitic suture. It is covered by the Upper Cretaceous Banatitic volcanics. This fault was cut by a north-south trending fault, which brought in tectonic contact the Laramian volcanics with the Neocomian sedimentary deposits along the Visca-Gurasada meridian. This north-south trending fault should be post-Laramian. The Laramian faults are represented by the Căpâlnaş-Zam-Vorţa fault, trending almost E-W with northerly dip, and several associated shorter faults (Savu et al., 1992), resulting in an *en échelon*-like fault system. The superstructure consists of the Laramian (Banatitic) volcanic rocks from the Lăpugiu Basin, to which the Pb-Zn-Cu mineralizations of the CVN district are related. The Laramian magmatic rocks are represented by volcanics, mostly pyroclastics (Savu, 2011). Pyroclastics of andesites and trachyandesites dominate, with rare basalts and basaltic andesites, as more basic rocks, and dacites, trachydacites and even trachytes, as leucocratic rocks (Fig. 2). The initial area of the Laramian volcanics must have been larger than the actual one, probably covering the entire surface of the metallogenetic district, but it was extensively eroded. The radiometric ages of these rocks vary between 60 Ma, Rb/Sr (Herz et al., 1974) and 70 Ma, K/Ar (Savu et al., 1992).

3 Characteristics of the Pb-Zn-Cu mineralization

The Pb-Zn-Cu mineralized zones (Fig. 1) occur within the metallogenetic district under three circumstances: (1) as isolated skarn-type

occurrence like that of Căpâlnaş, (2) as lenses situated along a fault, like that of Vorţa and (3) as small occurrences dispersed within the Aptian-Cenomanian sedimentary formations.

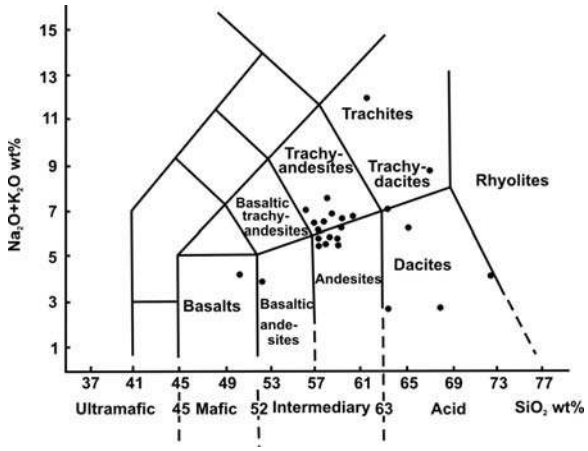

Fig. 2. Plot of the Banatitic volcanics from the Lăpugiu Basin on the $Na_2O + K_2O$ vs. SiO_2 diagram. Trends according to Le Bas et al (1986). Data from Savu et al. (1992).

3.1 Căpâlnaş-type mineralization

The mineralized zone occurs at the springs of Pârâul Plumbului (Lead Creek), a tributary of the Căpâlnaş Brook. This occurrence is situated at a small distance west of the western margin of the Banatitic volcanic area, south of the Căpâlnaş Village. The host rocks are represented by arkosian sandstones and marbly limestones intercalated in the Lower Jurassic tholeiitic basalt flows, and Upper Jurassic limestones, which are crosscut by dykes of porphyries with violet quartz phenocrysts, which generated small scale contact metamorphism of the limestones and a low-grade mineralization of galena and sphalerite.

In a small mine ditch, it can be seen that the marbly limestones are associated with brown-yellowish grossular-andradite and greenish grossular-epidote skarn veins (Savu, 1966). All the garnets show a zonal structure. The limestones are white or grey and, under the microscope, show a saccharoidal texture, rarely a porphyroblastic texture. The limestones underwent an intense fracturing and even brecciation. This permitted the circulation of the hydrothermal solutions, which partly substituted the limestone and created a network of metasomatic skarn veins, up to 10 cm thick. The mineralization is represented by a vein up to 5 cm thick. It consists mostly of galena and sphalerite, with rare crystals of chalcopyrite and pyrite. Although the mineralization does not represent any economic importance, the occurrence is interesting as regards the genesis of the infiltration skarns.

3.2 Vorţa-type mineralization

As it was mentioned above, at the beginning of the Laramian tectonic movements within the CVN district, an ENE-WSW trending crustal fault system was active (Andrei and Cristescu, 1966; Savu et al., 1992). Along these fractures, the Laramian volcanic rocks from Lăpugiu Basin erupted and the Pb-Zn-Cu sulfides were deposited, such as, for instance, those from the Heiuş Creek and Băii Creek, near the Vorţa village. The latter is the most characteristic mineralization of the district and is better known due to the fact that it was explored by the Certej and Deva Mining Companies and exploited by the Deva Mining Company for a long period before 1990. Deva Mining Company produced numerous assay data on the mineralization at Vorţa, some of these data being made available to the author of this study. Therefore, this mineralization will be presented here in more detail.

The ore at Vorţa exhibits a massive structure, sometimes with porphyritic texture determined by large xenomorphic crystals of chalcopyrite and sphalerite. The characteristic mineral assemblage from Vorţa ore is made up of pyrite, galena, sphalerite and chalcopyrite ± bornite ± covellite, minerals deposited approximately in this succession. Chalcopyrite usually occurs as grains with irregular margins, which may reach up to 1.5 cm in size. Chalcopyrite hosts pyrite and sometimes sphalerite and galena. Thin veins of sphalerite and galena cross the large crystals of chalcopyrite. Bornite and covellite are partly substituting the chalcopyrite crystals. Bornite occurs in crystals with irregular shape, which are located at the margin of the chalcopyrite grains.

Galena forms xenomorphic grains, which can include other minerals from the sulfide assemblage. Sometimes, sphalerite, shows very weak pleochroism and anisothropy. Broken crystals of sphalerite are seldom present in the Vorţa mineralization.

The chemical composition of the Vorţa ore, determined on numerous assays by the Deva Mining Society, shows that it is rich in Zn, which prevails over Pb and Cu (Fig. 3). Among these metals, Cu presents the lowest contents. A positive correlation between Zn and Pb is obvious (Savu and Nicolae, 1975b). Tămaş-Bădescu (2010) showed that beside these major elements, the Vorţa sulphide ore contains up to 0.38 g/t gold.

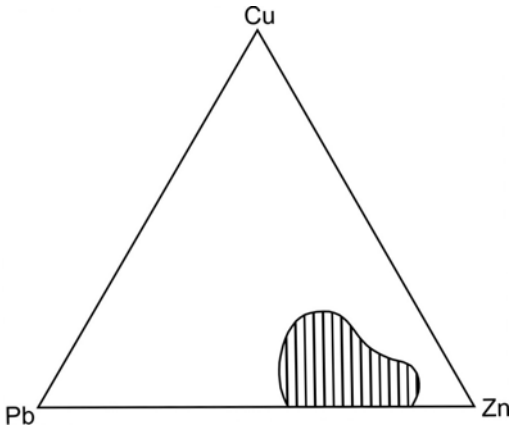

Fig. 3. Plot of 55 samples of mineralization from the Vorţa mine on the Cu-Pb-Zn diagram. Data from the archives of Deva Mining Society.

3.3 Nevoiaş-type mineralization

As Ghiţulescu and Socolescu (1941) showed on their map, in the eastern extension of the Banatitic volcanics from the northern part of the Lăpugiu Basin and north of Mureş, there are four ore fields situated along an E-W alignment, which could also be related to a fracture, parallel to the one on which the Vorţa mineralization is located. As shown in Fig. 1, this alignment extends between Vlădeşti (4), Nedeli Summit (5, north of Târnava), Sulighet (6) and Nevoiaş (3). On the map made by Ghiţulescu and Socolescu (1941), all these mineralized zones are located in the Cretaceous sedimentary deposits, the youngest succession of which belongs to the Senonian. The Nevoiaş mineralization (3) is the most important among these occurrences and is represented by short and thin veins, without any economic importance. West of Valea Mare Village (7), south of Vorţa (2), the author found blocks of sulfide ore dispersed in the dump heap of a small prospecting shaft dug in the Cretaceous formations near a small creek.

All these occurrences show Pb-Zn-Cu mineralization and have been attached by Ghiţulescu and Socolescu (1941) to the Neogene mineralizations from the Metaliferi Mountains. Nevertheless, since this mineralization alignment is E-W oriented, it is not conformable to the NW-SE trending alignments of Neogene volcanics and mineralizations from the Metaliferi Mountains. Therefore, they should be considered as related to the Laramian (Banatitic) volcanism, which would be consistent with their spatial association with the Banatitic rocks. The Vlădeşti mineralization, for instance, is located in the Banatitic volcanic field but Ghiţulescu and Socolescu (1941) considered they are associated with Neogene volcanics.

4 Discussions

The first information about the mineralized zones in the CVN district occurred on the geological map published by Ghiţulescu and Socolescu (1941), followed by the unpublished reports by Cioflica et al. (1957) and Teodoru et al. (1964). According to Ghiţulescu and Socolescu (1941), the hydrothermal mineralization at Vorţa is Neogene, i.e. it has the same age as the volcanic rocks in the Lăpugiu Basin, an idea shared by Cioflica et al. (1957), and Teodoru et al. (1964). Borcoş and Stanciu (1968) considered the hydrothermal mineralizations from the region as related to the Neogene volcanism from the Metaliferi Mountains. This idea cannot be supported because the Neogene volcanism manifested itself along NW-SE faults whereas the mineralizations from the CVN district occurred on an almost E-W fault system.

Among the numerous opinions on the genesis of Vorţa mineralization two concepts are to be remarked – the discordant hydrothermal model

and the Kuroko stratiform deposition model. Savu and Nicolae (1975) showed that the Vorţa hydrothermal mineralization was deposited along an E-W vertical fracture and was affected by shearing. Udubaşa et al. (1978) considered that Vorţa mineralization can be of Kuroko-type. However, in a subsequent paper (Udubaşa and Gaftoi, 1986), this character was considered as uncertain.

Cioflica et al. (1984) observed that, between Vorţa and Furcşoara, zones mineralized with sphalerite, chalcopyrite, and pyrite occurred, which were strongly brecciated and altered. Cioflica et al. (1985) showed that a belt of hydrothermal alterations extends between Vorţa and Furcşoara, which generated low-grade zeolite deposits. As it results from their sketch map, the hydrothermal solutions circulated along an *en échelon* fault system. Popescu et al (1997), referring to the papers by Udubaşa, considered that the Vorţa mineralizations have a Kuroko metallogenetic character and have been affected by a burial metamorphism, a conclusion that is in disagreement with the location of the mineralization along a vertical fault, discordant on the structures of the basement. Vlad and Tămaş (1998) associated its formation with an Early Cretaceous magmatism, which would have produced granitoid intrusions. Mârza (1999), referring to the published data on the Vorţa mineralization, supposed that they are rather related to the Banatitic (Late Cretaceous) magmatism. The idea of Kuroko-type mineralization at Vorţa was adopted by Ciobanu et al. (2001), who asserted that the mineralization was generated in submarine conditions, as a syngenetic and stratiform ore formation at a sea mountain with alkali basalt flows. Nevertheless, in the Vorţa-Dealul Mare area there are no seamount remnants or any alkali basalts, thus the mineralization at Vorţa is not of Kuroko-type, being discordant relative to the structures of the basement, although it has been generated by hydrothermal solutions probably released from Banatitic intrusions or from volcanic chambers at the end of the Laramian volcanism. The fault-related genesis of the mineralization from Vorţa district is also supported by Vlad and Tămaş (1998).

Two explanations were proposed with regard to the deformations which affected the mineralization: shearing (tectonic) deformation and burial metamorphism. The burial metamorphism concept was presented first by Şeclăman (oral communication), but no minerals characteristic for such a metamorphism-type could be found in the region. Moreover, considering both the mineralization-type and ore deformation, the burial metamorphism is not consistent with the geological relations observed in the field and this can be argued based on several features of the ore and of the country rocks that could be remarked in the adits of Vorţa mine and at the microscope.

Based on the evidence produced by Savu and Nicolae (1975a, b) and more recent observations (Savu, 2007) it is obvious that the Vorţa mineralization has a structural control, being located along the mentioned E-W trending fracture, which extends from the northern Banat, through Căpâlnaş and Zam to Vorţa and cuts the Upper Jurassic–Lower Cretaceous island arc volcanics from the basement of the Mureş ophiolitic suture. These volcanics are lying over the ophiolitic rocks of MORB-type. They are represented mostly by pyroclastics of basaltic andesites, hornblende–pyroxene–andesites and quartz-andesites, oligophyres and dacites (see the map by Savu and Nicolae, 1975a).

Within the Vorţa area four alteration and mineralization zones occur, related to the E-W fracture system. The hydrothermal solutions passing through this fracture zone affected both andesite pyroclastics and the dacitic rocks. The most important zone, which is 50 to 200 m wide, extends eastward from the Cărmăzineşti Valley, by the Heiuş Creek, to the Băii Creek. The mineralization, consisting of galena, sphalerite, pyrite and chalcopyrite was deposited along the median zone of the subvertical fracture. Along this zone, all the minerals from the volcanic rocks were transformed into secondary minerals. In the andesitic rocks, the plagioclase crystals have been strongly argillized and the mafic minerals have been substituted by chlorite,

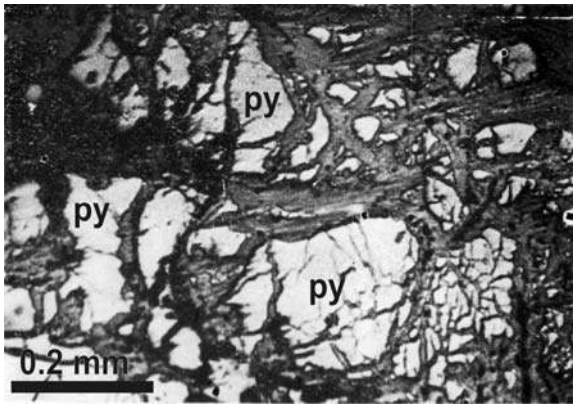

Fig. 4. Broken crystal of pyrite the fragments of which are cemented by chalcopyrite. Vorţa mine.

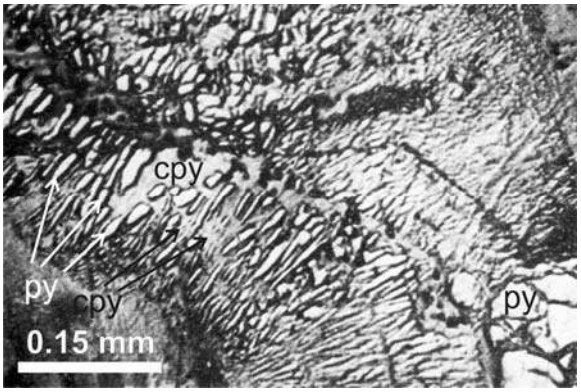

Fig. 5. Elongated fragments of a broken cryslal of pyrite cemented by chalcopyrite. Vorţa mine.

sometimes associated with epidote. The altered rocks are impregnated with sulphides associated with quartz, calcite and zeolite. The latter mineral is widespread in the region (Cioflica et al., 1985). The dacites from the Homorod Hill have been strongly affected by the hydrothermal solutions, being impregnated with pyrite crystals. Locally, the rocks have been completely transformed into kaolinite with quartz relics. Along the fracture that supposedly served as main passageway for the hydrothermal fluids, the andesites were intensely silicified.

The fracture hosting the mineralization kept being tectonically active after the formation of the sulphide ore, causing the shearing of the mineralization to small lenses or boudins (see Fig. 2 in Savu and Nicolae, 1975a).

In the hydrothermalized host rocks, not affected by the shearing, pyrite occurs in euhedral (cubic) crystals, sometimes with irregular margins. In the ore, the pyrite is included in chalcopyrite, in galena and sphalerite. Broken crystals of pyrite often occur as inclusions in chalcopyrite. These probably were produced by the tectonic movement along the fracture on which the mineralization was deposited. The breaking of the pyrite crystals took place under two conditions: (1) when the strain was not very strong, the fragments of the broken pyrite crystals kept a rugged shape (Fig. 4); (2) at more intense efforts, the resulting pyrite fragments got an elongated shape (Fig. 5). In both cases, the pyrite fragments have been cemented by the malleable chalcopyrite. The *Kataklase* phenomenon presented by Ramdohr (1960) represents, in fact, the incipient stage of this process of pyrite breaking.

5 Summary and conclusions

The Căpâlnaş-Vorţa-Nevoiaş metallogenetic district is superposed over the northern part of the Lăpugiu Basin Banatitic volcanics. All the mineralized zones of the district consist of the Pb-Zn-Cu assemblage. Although the mineralization occurrences from the district may give the appearance of scattered distribution, they are located on E-W trending Laramian fractures, a good example in this respect being the Vorţa mineralization. The movements along these faults were effective during and after the sulfide deposition, thus causing the breaking of the crystals of the brittle pyrite, cemented by the ductile chalcopyrite. Along the fractures involved in the genesis of the mineralization, the basement ophiolitic and island arc volcanics have been strongly altered by the hydrothermal solutions of mesothermal nature. It seems that in the depth of the district plutonic intrusions of Banatitic acid rocks were emplaced, the apophyses of which are the dacitic bodies occurring around the mineralized areas at Vorţa. More probably, these intrusions released the hydrothermal solutions that precipitated the Pb-Zn-Cu mineralization.

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Cârlibaba-Broşteni strike-slip fault - a possible old segment of Bogdan Vodă-Dragoş Vodă fault system: implications for the structural and metallogenetic correlations in the Crystalline-Mesozoic Zone of the Eastern Carpathians (Romania)

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Abstract: This study proposes the ramification of the Dragoş Vodă fault to the south-east, on the line Cârlibaba-Zugreni-Crucea-Holdiţa, reaching the eastern limit of the Crystalline-Mesozoic Zone of the Eastern Carpathians north of Broşteni. This segment is a strike-slip fault, which can be named the Cârlibaba-Broşteni fault. The proposed structural arrangement indicates the presence of a distinct tectonic compartment of the Crystalline-Mesozoic Zone, comprising Rodna Mountains and Bistriţei Mountains, delimited, to the north and north-east, by the Dragoş Vodă fault and the Cârlibaba-Broşteni fault, and, to the south and south-west, by the Dârmoxa-Grinţieş fault. The tectonic compartment between these two strike-slip faults has a southeasterly offset of ca. 75 km and comprises all tectonic windows where the Infrabucovinian Nappes (Anieş, Ştiol, Valea Vinului, Rusaia, Iacobeni, Borcut-Ulm and Arsiţa Barnarului) are cropping out. Because of its southeastward movement, the structural features of the tectonic compartment Rodna Mountains-Bistriţei Mountains differ from those of the adjacent compartments to the north and to the south. It is inferred that the Leşul Ursului and Baia Borşa ore fields were initially parts of the same metallogenetic zone, the former being displaced to the south-east along the Cârlibaba-Broşteni fault.

Key words: strike-slip fault, Eastern Carpathians, Rodna Mountains, Bistriţei Mountains

1 Introduction

Many of the arguments for the delimitation of the structural units and tectonic compartments of the Crystalline-Mesozoic Zone used in this work are based on the detailed study of the pre-Alpine basement of the Alpine tectonic units. The traditional structural delimitations and modeling in a coherent system for the entire Carpathian area were based on the Alpine evolution deduced from the knowledge on the sedimentary formations. In 1907, Uhlig evidenced the Marginal Syncline of the Eastern Carpathians – the Rarău Syncline, in the north, and the Ghilcoş-Hăghimaş-Ciuc syncline, in the south, both constituted of Mesozoic formations. The lack of continuity between the two synclines was explained by an axial raise of the geological structure between the Rarău Mountains and Hăghimaş Mountains. The model largely accepted since the late 1900s is the outcome of numerous studies resulted in synthetic works on facies, sedimentation zones and geological evolution accomplished by Patrulius et al., (1969) and Săndulescu et al. (1989). A remarkable presentation, for the Carpathians and the neighboring areas, is comprised in the book Geotectonics of Romania (Săndulescu, 1984).

The geological data presented here support an interpretation according to which, the continuity of the Alpine structures of the Crystalline-Mesozoic Zone is interrupted by a sinistral

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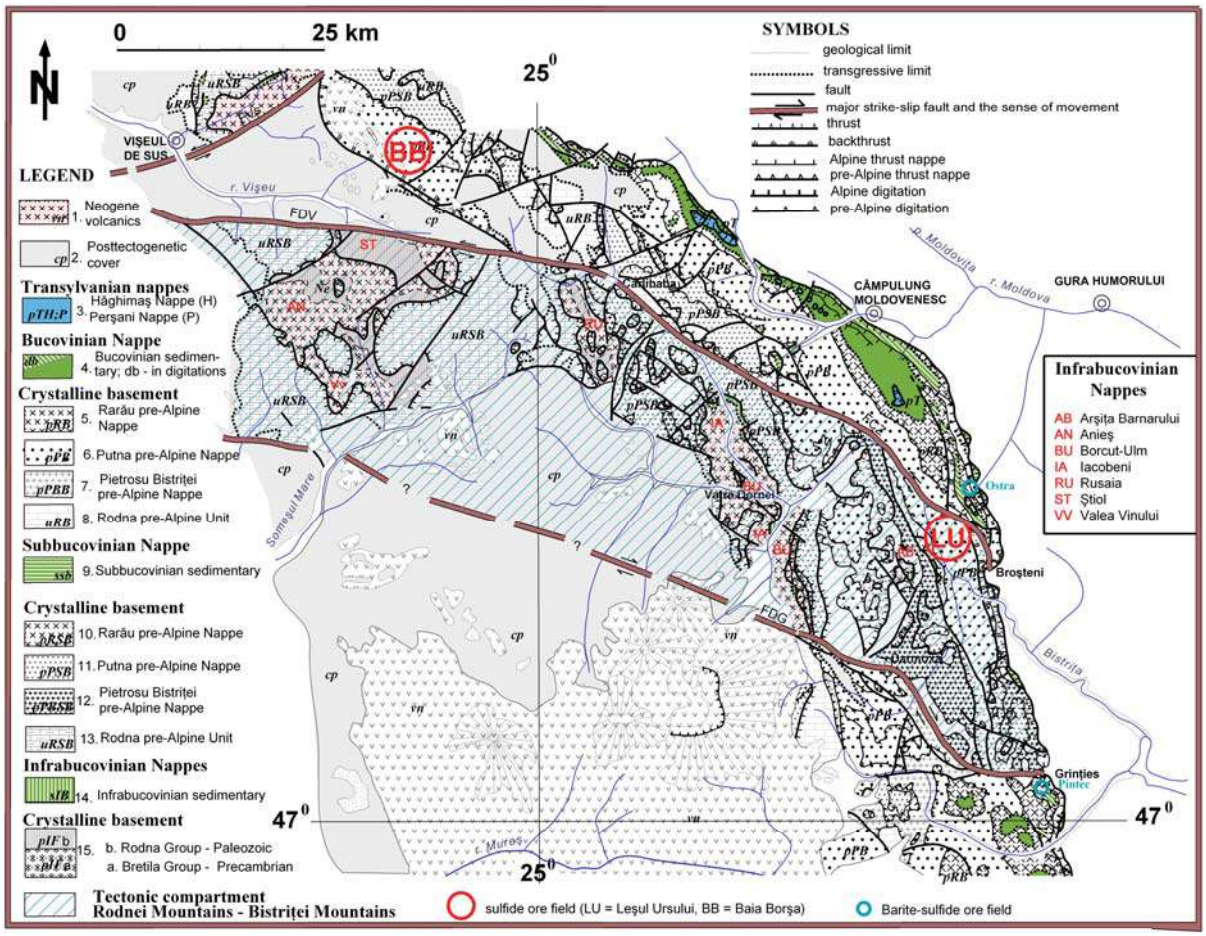

Fig. 1. Simplified geological map of the central and northern parts of the Crystalline-Mesozoic Zone of the Eastern Carpathians, including the tectonic compartment delimited by the strike-slip faults Cârlibaba-Broşteni and Dârmoxa-Grinţieş.

strike-slip fault, which probably is an old segment of the Bogdan Vodă-Dragoş Vodă fault system. South of this fault, the Rodna Mountains and most parts of the Bistriţei Mountains constitute a tectonic compartment that extends to the Dârmoxa-Grinţieş dextral strike-slip fault (Vodă, 1999). The compartment between the two strike-slip faults, was interpreted as having a great south-eastward displacement, which could explain the absence of the structural features known at the eastern border of the Crystalline-Mesozoic Zone to the north, in the Rarău Mountains, and to the south, in the Hăghimaş Mountains

2. Geological evidence

2.1 The geological situation at Cârlibaba

The analysis of the stratigraphic and structural context from the eastern sector of the Dragoş Vodă fault, made by L. Szasz during the elaboration of the Geological Map of Romania, scale 1:50.000, Dragoş Vodă Sheet (Kräutner et al., 1993; Vodă et al., 1997) argued for the extension of this fault to the left slope of the Ţibău valley, at ca. 1 km upstream from the confluence with Bistriţa river, based on the sedimentary formations of the posttectogenetic cover, Cenomanian-Oligocene in age. The sequence of the sedimentary deposits contains numerous gaps of sedimentation and erosion, which mark the movement of the tectonic compartments along the Dragoş Vodă fault.

At Gura Ţibăului, south of the eastern extension of the Dragoş Vodă fault, within the geological reserve from Stânca Ţibăului, polymictic conglomerates, sandstones and fossiliferous calcarenites occur across a small area, lying directly on the metamorphic formations. These sedimentary formations are missing north of the Dragoş Vodă fault and do not extend in the eastern side of Bistriţa valley,

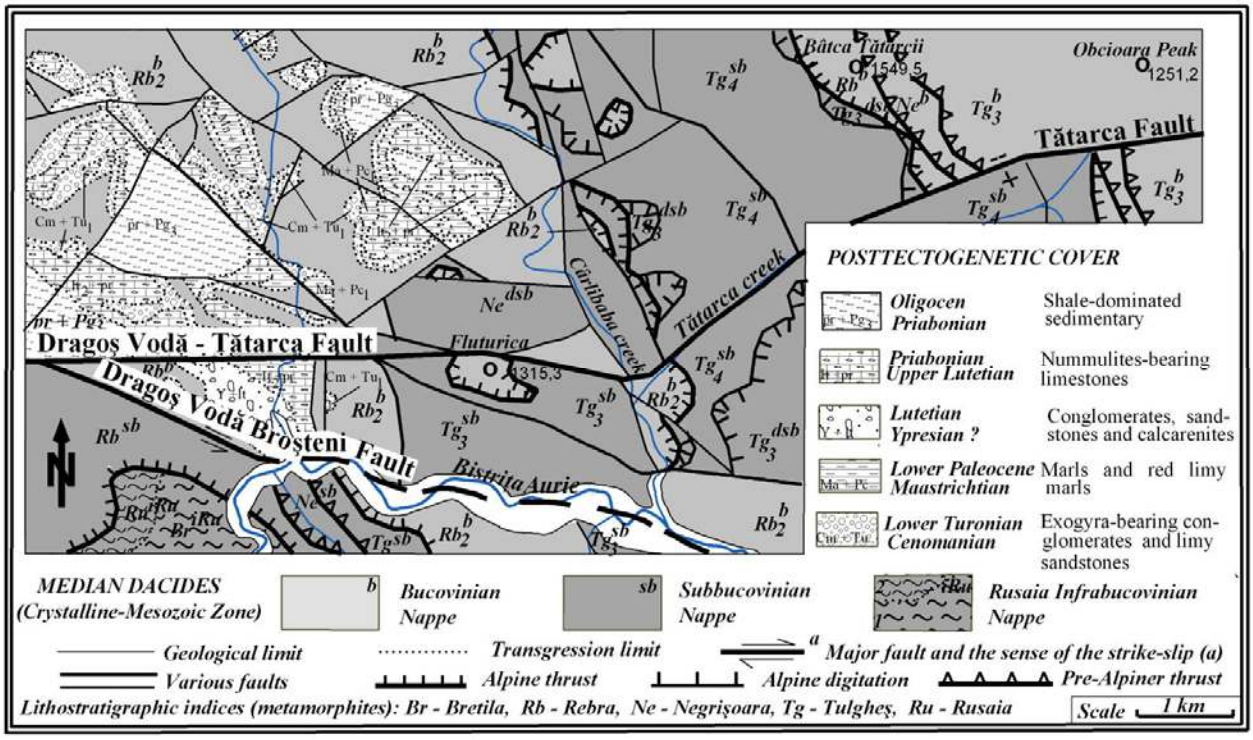

Fig. 2. Simplified geological map of the Cârlibaba-Ţibău zone

obviously from tectonic reasons, i.e. uplifting that induced the retreat of the outcropping zone of the posttectogenetic sedimentary cover far to the west, in an area upstream from the confluence between the valleys Bistriţa and Bretila. East of Ţibău valley, the direct continuation of the Dragoş Vodă fault passes north of Fluturica Peak (1345.3 m), follows the course of the Tătarca Cârlibabei valley, the crest between the hydrographic basins of Bistriţa and Moldova rivers, south of Obcioara Peak (1251.2 m) and reaches Moldova valley downstream from the confluence with Tătarca creek, tributary of Moldova river from the right hand side (Kräutner et al., 1993; Vodă et al., 1997).

The cartographic shape of the area between the Bistriţa river, at Cârlibaba, and the ridge toward the Moldova river, close to its confluence with Tătarca tributary, shown in Fig. 2, indicates an uplift of the tectonic compartment south of the eastern extension of the Dragoş Vodă fault (named Tătarca fault), which is coherent with the structural situation to the west, at the northern limit of Rodna Mountains.

South of Tătarca fault, the geological structure continues to the Bistriţa valley, but cannot be correlated with the structure in the western slope of Bistriţa valley, characterized by the presence of lithostratigraphic sequences and tectonic features of the Bucovinian and Subbucovinian nappes different from those in the eastern slope of Bistriţa River. The domain with different features from the western side of Bistriţa valley also includes the Rusaia Infrabucovinian Nappe occurring in a tectonic window. An analysis of the geological structure and of the lithostratigraphic content of the structural units from this area as far to the south as south of Broşteni, led to the tracing of the strike-slip fault Cârlibaba-Broşteni, named in Fig. 2 Dragoş Vodă-Broşteni fault, to highlight the possible relation with the strike-slip fault system Bogdan Vodă-Dragoş Vodă.

2.2 The geological situation at Broşteni

Fig. 3 contains a geological map with the interpretation of the Cârlibaba - Broşteni fault between Crucea and Holdiţa. In the area between the Holda (Puzdra) creek and Crucea, the fault was revealed by the surface mapping correlated with the underground works from the Leşul Ursului mining field (Kräutner et al, 1975; 1978; 1986). The strike-slip character of the fault was

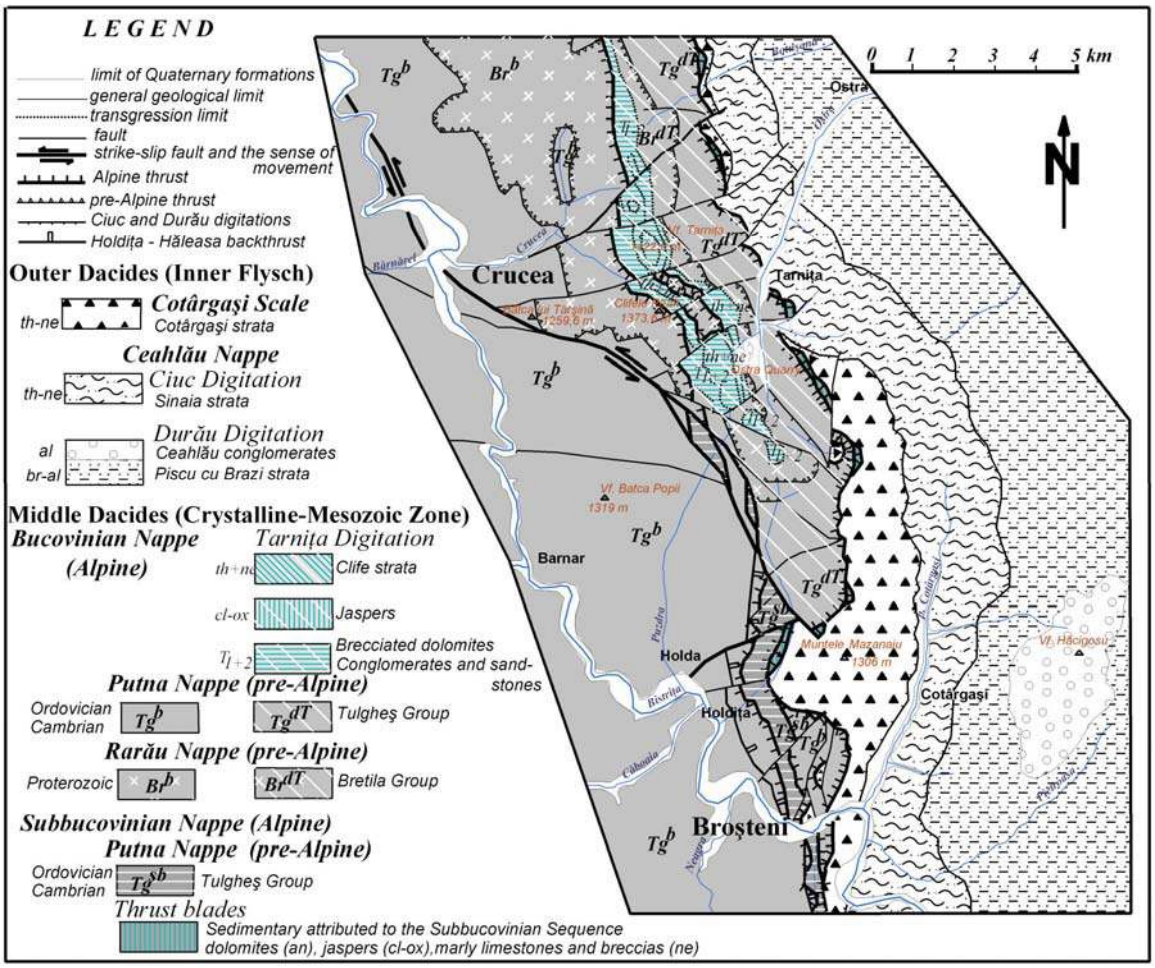

Fig. 3. Simplified geological map of the Broşteni-Ostra zone. The eastern part, the Cotârgaşi Scale and the digitations of the Ceahlău Nappe after Săndulescu (2008).

neglected. The continuation of the fault to the left hand slope of the Holdiţa creek, at the contact between the Crystalline-Mesozoic Zone with the Inner Flysch Zone, was easy to draw due to the occurrence, on the fault, of several outliers of black graphitic quartzites similar to those from Holdiţa-Broşteni and Borca, from the Subbucovinian Nappe (Vodă et al., 1999). Previous mapping results (Vodă, 1980) outlined a small half-window in the right hand slope of the Holdiţa creek, at the confluence with the Haştei creek, the border of the Crystalline-Mesozoic Zone being indented westward with more than 1 km. On the contact with the Inner Flysch Zone, there is a thrust blade made up of Anisian dolomites, in the base, and a Neocomian sequence composed of red jasperoid argillites with thin intercalations of coarse sandstones. Recently, Săndulescu (2008) identified the Ciuc digitation, on the Ceahlău Nappe, and the Cotârgaşi scale, under the thrust of the Crystalline-Mesozoic Zone over the Inner Flysch, with the greatest extent in the right hand side of the Cotârgaşi creek and on the Haştei creek, tributary of the Holdiţa creek from the left side, in the area where the strike-slip fault Cârlibaba-Broşteni reaches the border of the Crystalline-Mesozoic Zone. Fig. 3 contains this structural interpretation of the Inner Flysch Zone (after Săndulescu, 2008). The structure of the Alpine nappes of the Crystalline-Mesozoic Zone obviously exhibits peculiar features in this sector:

- the tectonic compartment north-east of the Cârlibaba-Broşteni strike-slip fault is characterized by the presence of the Bucovinian Nappe and its frontal digitation, Tarniţa digitation (Săndulescu, 1981);
- the tectonic compartment south-west from the strike-slip fault contains the Bucovinian Nappe and, beneath it, the Subbucovinian Nappe, on a zone uplifted by a reverse fault, named Holdiţa-Hăleasa fault (Vodă, 1982).

The strike-slip fault ends to the contact with the Inner Flysch Zone, which indicates the freeze of the tectonic movement

after the deposition of the Cotârgaşi strata, whose age is Tithonian-Neocomian (Săndulescu, 2008).

3 Discussions

3.1 The trace of the Cârlibaba-Broşteni strike-slip fault

The lack of sedimentary formations of the posttectogenetic cover made possible the observation of the great differences between the lithostratigraphic sequences and structure of the metamorphic formations in both slopes of the Bistriţa River. The Cârlibaba-Broşteni strike-slip fault follows Bistriţa valley from Gura Ţibăului downstream, to Ciocăneşti, where it passes in the eastern slope, downstream of the confluences with the tributaries Pârâul Rece and Colacu, then crossing the Brezuţa creek at ca. 1 km from the confluence with Bistriţa and Puciosu creek, at the margin of Mestecăniş village. It continues by the source of Pârâul Roşu creek (tributary of the valley Putna of Pojorâta), passes the Giumalău Mountain (1856.5 m) near the sheepfold from Poiana Ciungi (1513.1 m) and reaches Bistriţa valley again, in Bistriţa Gorges from Zugreni. From Zugreni to Crucea, the fault follows Bistriţa valley, cutting locally through the left hand slope (most obviously at « Dâmbul Colacului »), then remains in the left hand (eastern) slope of Bistriţa, passing by the source of Holda (Puzdra) creek and stops at the tectonic contact between the Crystalline-Mesozoic Zone and Inner Flysch Zone, on the Holdiţa creek, at the confluence with Haş creek.

3.2 The tectonic compartment Rodna Mountains-Bistriţei Mountains

In Fig. 1, it cam be seen how the Dragoş Vodă fault and its extension, the Cârlibaba-Broşteni fault, delimit, to the south, a tectonic compartment that extends southward to the Dârmoxa-Grinţieş fault. Between the two strike-slip faults, the entire Rodna Massif and most of the Bistriţei Mountains are comprised. To the east, this tectonic compartment is limited by the border of the Crystalline-Mesozoic Zone between Broşteni and Grinţieş. To the west-north-west, the Dârmoxa-Grinţieş fault is overlain by the formations of the posttectogenetic cover and by the Neogene volcanic rocks of the Călimani Massif from Păltiniş. Because of this, its extent and location at the southwestern margin of the Rodna Massif are just inferred.

The main feature of the compartment Rodna Mountains-Bistriţei Mountains consists of the occurrence, over extended areas, of the lowermost Alpine tectonic units of the Crystalline-Mesozoic Zone. In this compartment, the Subbucovinian Nappe has the greatest extent of all parts of the Crystalline-Mesozoic Zone. From beneath the Subbucovinian Nappe, the following Infrabucovinian structural units occur (from north-west to southeast): Anieş, Ştiol, Valea Vinului, Rusaia, Iacobeni, Borcut-Ulm and Arşiţa Barnarului.

The eastern border of the Crystalline-Mesozoic Zone of the Eastern Carpathians, at the contact with the Ceahlău Nappe (Inner Flysch Zone), represented by the Ciuc Digitation, is a continuous tectonic line from Maramureş, in the north, to Ciuc Depression, in the south. Along this major tectonic contact, there are different structural situations, induced by the specific features of the compartment Rodna Mountains-Bistriţei Mountains:

- North of the Cârlibaba-Broşteni fault, from the left hand slope of the Holdiţa creek, in the south, to Slătioara (south of Câmpulung Moldovenesc), in the north, along the contact with the Ceahlău Nappe (Inner Flysch Zone), the Crystalline-Mesozoic Zone is represented by the Tarniţa digitation of the Bucovinian Nappe.
- In the compartment Rodna Mountains-Bistriţei Mountains, from Holdiţa-Broşteni, in the north, to the Grinţieşul Mare creek, in the south, the border of the Crystalline-Mesozoic Zone is constituted of the Subbucovinian Nappe. Between Broşteni and Borca, outliers and scales that could be attributed to the Bucovinian Nappe occur along the border of the Crystalline Mesozoic Zone.
- South of the Dârmoxa-Grinţieş fault, the Bucovinian Nappe reappears, with a frontal digitation (Chicera Digitation), best visible between the Bistricioara and Bicaz valleys.

The lack of continuity of the Mesozoic sedimentary formations of Rarău and Hăghimaş type in the segment between Broşteni and Grinţieş, and, also, the lack of the frontal digitations of the Bucovinian Nappe in the same segment, can be considered as resulting from the southeastward movement of Rodna Mountains-Bistriţei Mountains compartment, between the strike-slip faults Cârlibaba-Broşteni and Dârmo-

xa-Grinţieş. Regarding the evolution of this compartment, it can be conceived that the onset of the strike-slip movements was triggered by the blocking of the movement in the northeastern compartment of the Dragoş Vodă-Cârlibaba-Broşteni fault (which could be named Maramureş-Rarău compartment) and the escape in the east-south-east direction of the land from the southwestern side of the fault. Several studies concerning the Tertiary displacement of the major structural units of the Alps and Carpathians (Tischler, 2005; Tischler et al., 2007; Groger et al., 2008) allow the consideration of the strike-slip and rotation along the Cârlibaba-Broşteni fault, coupled with the Bogdan Vodă-Dragoş Vodă fault system and with the possible connection to the west and south-west (to tectonic system of the Mid Hungarian Line).

3.3 Structural and metallogenetic correlations

Out of all the possible structural and metallogenetic correlations in the presented framework, maybe the most interesting is the possible correlation between the Leşul Ursului and Baia Borşa pyrite and polymetallic sulfide deposits (pyrite + copper and pyrite + lead, zinc, copper, hosted in the formation of acid metavolcanics (Tg3) of Tulgheş Group from the Bucovinian Nappe. The geologic and metallogenetic models, based on numerous geochemical and depositional studies, were elaborated by Kräutner (1965) and Kräutner et al. (1975. 1976, 1986), for the deposits in the Leşul Ursului mining field, and by Zincenco (1971, 2000), for the ore field from Baia Borşa.
Leşul Ursului is situated near Broşteni, in the Bucovinian Nappe, in the tectonic compartment Rodna Mountains-Bistriţei Mountains, south of the Cârlibaba-Broşteni fault. The lithostratigraphic sequence from Leşul Ursului, with numerous ore layers and lenses exposed in outcrops and known in detail from numerous adits and drill holes in the left hand slope of Bistriţa river downstream of Crucea village, could not be found north from a fault that cut the structure, from the left hand slope of Crucea creek, the middle courses of Leşului and Ursului creeks, reaching, north of Bâtca Popii, at the confluence of Grebin and Puzdra creeks (Kräutner, in Săndulescu et al., 1989). That fault is part of the Cârlibaba-Broşteni fault.

At Baia Borşa, there is the same geologic-metallogenetic context as that from Leşul Ursului, but the two ore fields could not be correlated because of the great distance between them (ca. 75 km), because of some complications generated by the Neogene intrusions from Toroiaga-Ţiganu complex, of the posttectogenetic sedimentary cover from some areas and of the structure, characterized by shallow dips, which poses difficulties in detecting the effects of the faults. In the structural conditions presented here, the two regions can be easier correlated, with an initial (pre-fault) position of the Leşul Ursului ore field in the southern extension of the Baia Borşa ore field. It is of note the similar geological context of the two areas: from Broşteni to the south, at Borca, and from Baia Borşa to the north-west, in the Bardi-Bardău zone, beneath the sequences of Tulgheş Group with sulfide deposits, from the Bucovinian Nappe, there are sequences of the Tulgheş Group with manganese ores, from the Subbucovinian Nappe. This correlation would lead to the estimation of an offset of ca. 75 km along the Cârlibaba-Broşteni fault.

The structural arrangement proposed here also favors the correlation between the epigenetic mineralization (barite and sulfides) of Ostra-Gemenea-Slătioara type, which occurs in Tarniţa digitation, north of the Cârlibaba-Broşteni fault, and the barite mineralization from the right hand slope of the Bistricioara valley, upstream of the confluence with Pintec creek, in the tectonic compartment south of the Dârmoxa-Grinţieş fault. The barite + sulfides mineralization of Ostra-Gemenea-Slătioara type consists of unconformable vein-type bodies. These cut the metamorphic basement and the Mesozoic sedimentary sequence from Tarmiţa Digitation (at Ostra), are in contact with the Triassic sedimentary and the Callovian-Oxfordian jasper but do not reach the flysch-type sequence of the Clifele strata, Neocomian in age. At Pintec, the barite mineralization identified by Moga and Podaşcă (1988) occur in a similar geological context, in a frontal digitation of the Bucovinian Nappe (Chicera Digitation), which correlates with Tarniţa Digitation.

It can be remarked that, similar to the areas with barite and sulfides from the frontal digitations of the Bucovinian Nappe (Tarniţa and Chicera), the uranium accumulations of Crucea type also occur only north of the Cârlibaba-Broşteni strike-slip fault (in Crucea zone) and south of the Dârmoxa-Grinţieş strike-slip fault (at Tulgheş). In the tectonic compartment between the two faults there is no indication of either Ostra-type or Crucea-type mineralization.

The movement along the Cârlibaba-Broşteni sector ceased before the deposition of the posttectogenetic cover of the Crystalline-Mesozoic Zone, while the segment represented by the Tătarca fault remained active.

4 Conclusions

• The strike-slip fault Cârlibaba-Broşteni was traced; this can be considered as a continuation of the fault system Bogdan Vodă-Dragoş Vodă. The fault does not cross the eastern border of the Crystalline-Mesozoic Zone and has no effect on the nappes of the Inner Flysch.
• The sinistral displacement of the Cârlibaba-Broşteni fault is subsequent to the emplacement of the Alpine nappes in the Crystalline-Mesozoic Zone and precedes their thrust over the nappes of the Inner Flysch.
• The tectonic compartment south of the Cârlibaba-Broşteni strike-slip fault is limited, to the south, by a dextral strike-slip fault, the Dârmoxa-Grinţieş fault. This tectonic compartment was pushed eastward, most likely by the blocking of the movement in the compartment north of the Cârlibaba-Broşteni fault.
• The structural interpretation derived from the identification of the Cârlibaba-Broşteni fault, imposes structural correlations and metallogenetic at regional scale, the most interesting outcome being the correlation between the Leşul Ursului and Baia Borşa polymetallic sulfide deposits, which are located at more than 75 km from each other. Also of interest are the correlations between the barite and sulfide mineralization of Ostra type and of the uranium mineralization (Crucea-type), occurring in Tarniţa Digitation and Chicera Digitation, both frontal digitations of the Bucovinian Nappe, which are missing from the compartment Rodna Mountains-Bistriţei Mountains, displaced ca. 75 km eastward, between the strike-slip faults Cârlibaba-Broşteni and Dârmoxa-Grinţieş.

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