GENESIS OF MUREŞ OPHIOLITIC SUTURE AND OF ITS N-TYPE MORB ROCKS AND ISLAND ARC VOLCANO-PLUTONIC ASSOCIATION*

HARALAMBIE SAVU

Geological Institute of Romania 1, Caransebeş St, 78344, Bucharest 32, E-mail geol. @ igr. ro

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The Mureş ophiolitic suture (MOS) resulted from the evolution of the Mureş Ocean that opened along an older continental pre-ocean rifting zone, accompanied by a WPB-type bimodal vocanism. The opening of the Mureş Ocean started in Liassic (ca. 180 Ma) and ended during the Upper Jurassic – Lower Cretaceous period. There was engendered an ocean crust, formed of tholeiitic rocks bearing a N-Type MORB signature. The closing of this ocean was determined by an asymmetric bilateral subduction, which determined the obduction of an ophiolitic plate over the margins of the Transylvanian and Apuseni Mountains convergent plates. The subduction was of Mariana-type on the northern margin of the Mureş Ocean, where the Drocea-Trascău subduction trench resulted from, and of Andean-type on the southern margin of this ocean, where the Metaliferi subduction trench occurred. At the same time, there started an island arc volcano-plutonic magmatism, which had a bimodal character in the Drocea-Trascău subduction trench and a normal calc-alkaline character in the southern Metaliferi subduction trench. Leucocratic (granitoid) intrusions (116-128 Ma) took place, too. The volcanism was accompanied by Mesozoic sedimentary deposits in the subduction trenches, among which limestones and red jaspers and argillites. The ophiolitic rocks have been affected by intra-oceanic and post-oceanic metamorphism processes. The MOS tectonics was strongly influenced by the subduction/obduction structure. It determined the occurrence of folding and thrust structures mainly in the marginal subduction trenches, which were of divergent-type at the beginning and changed into northwest-vergence thrust structures. The disjunctive fractures occurred soon after the Laramian collision, and belonged to two systems: a longitudinal system and a diagonal system. The longitudinal fracture system cut the MOS along its southern marginal part, without any horizontal shifting of the adjacent blocks. The TTFS, representing the most important system of longitudinal fractures, occurred concomitantly with the emplacement of the Transylvania mantle plume. It started from Paleogene (ca. 48 Ma) and continued up to Quaternary. It was accompanied by a hotspot volcanism. The diagonal *en échelon* fractures cut MOS into several blocks. The tendency of these blocks was to rise toward west, with the total erosion of the formations from the MOS superstructure, and to subside toward east, where a large part of the MOS bent zone formations was subsided into the Transylvanian Basin. In the end, by the closing of the Mureş Ocean due to the asymmetric bilateral subduction /obduction process, there resulted the MOS which is represented now by the asymmetric divergent Mureş Orogen.

Key words: Ophiolitic suture; Ophiolites; Arc magmatics; Petrology; Metamorphism; Tectonics.

INTRODUCTION

In 1952, when I began the researches on the “basic rocks” from the South Apuseni Mountains, the area of the MOS was represented on the geological maps as a violet-coloured area with a guide indicating there the presence of diabases, melaphyres and some gabbros, but not separated. Therefore, I started a systematic and lengthy study of these formations. Since 1983[1], when I started discussing this ophiolitic area in terms of a real ophiolitic suture up to 2001, in spite of all evidences,

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*Paper presented by Acad. R. Dimitrescu.

this concept was disavowed by some geologists[2], who considered MOS as an arc structure. During this period new data have been brought by myself and other authors on the evolution of this ophiolitic zone. Other data were referring to the igneous rocks like ophiolites and island arc magmatic rocks. Therefore, I considered of interest the elaboration of an up to date synthesis paper regarding all these new observations and the older problems. In doing so, an as rigorous as possible interpretation and definition were afforded them, so that some inadvertences referring to the geological evolution of the ophiolitic suture can be corrected, too, as it will be shown further down.

OCCURRENCE AND EXTENSION OF THE MUREŞ OPHIOLITIC SUTURE

To begin with, it is better to remember that the area referring to the MOS is known also as the South Apuseni Mountains[2] and the Mureş Zone[1]. But, if the genetic name of MOS is strictly referring to the structures and the formations resulted from the Mureş Ocean evolution, the other two terms are only geographic denominations, that refer to areas including, besides the ophiolitic rocks, many other formations. As for instance, the South Apuseni Mountains area includes both the MOS rocks and the Paleozoic crystalline schists and granitoid intrusions in almost equal proportions, as well as the banatitic, Paleogene and Neogene magmatic rocks.

MOS extends from the Transylvanian Basin at the east up to the north of Banat, south of Mureş River, that is running through its southern part. an opportunity from which its name derived, too (Fig. 1). It is oriented approximately east-west, its extremities being bent toward south and north respectively, directions in which MOS extends further away under younger formations (Fig. 1). The outcropping area of MOS extends over a length of about 200 km. and a width of maximum 40 km.

PRE-OCEAN RIFTING AND THE RELATED INTRA-PLATE VOLCANISM

MOS had a similar evolution to that of other structures of the sort. During its long evolution from the Triassic up to the Lower Cretaceous the area in which MOS has formed, passed through several stages. It started by an pre-ocean continental rifting, continued by the opening of the Mureş Ocean and ended by the bilateral subduction of the adjacent tectonic plates, that resulted in the occurrence of two intra-oceanic island arcs[1].

The pre-ocean continental rifting occurred by the end of the distension period, which lasted from the Late Paleozoic up to the beginning of Jurassic. During this period a distension regime was reigning

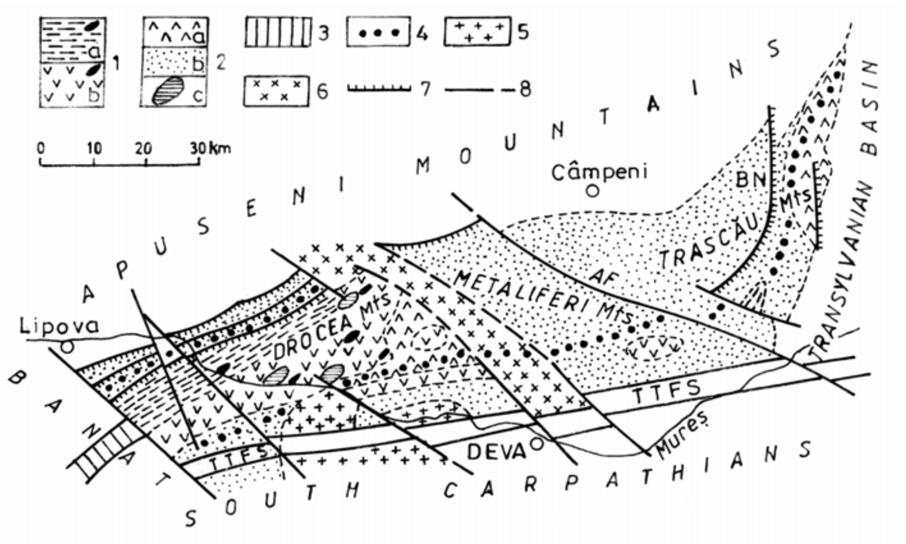

Fig. 1. Sketch-map showing the extension of MOS and its relationships with the south Transylvanian transcrustal fault system (TTFS): 1a, sheeted dyke complex (O_2); 1b, pillowed basalt complex (O_1): 1 a and b, (black spots), gabbro and ultrabasic small bodies; 2a, island arc volcanics (J_3-K_1); 2b, Mesozoic sedimentary deposits and island arc volcanics; 2c, Late Kimmerian island arc granitoid intrusions; 3, extension of MOS collision plane and slice of basic rocks toward Vardar Zone (Serbia); 4, alignment of MOS island arcs; 5, Laramian (banatitic) arc volcanics; 6, Neogene volcanics; 7, thrust; 8, faults: TTFS, Transylvanian trans-crustal fault system; AF, Ampoi fault; BN, Bedeleu Nappe.

over the entire European continent, which led to the strong erosion of the Pre-Variscan and Variscan structures, this area tending to become a large peneplain. But, by the end of Triassic, there started a pre-ocean continental rifting It extended parallel with the alignments of the scars of the old structures, distributed according to the branches of an old triple junction[3] extending along the actual East Carpathians, South Carpathians and along the North Dobrogea failed branch. Within the inner area comprised between the East Carpathians end the South Carpathians branches there was an other zone of pre-ocean continental rifting – the Mureş rifting zone – that was parallel with the first two branches of the triple junction. Along the parallel faults of this rifting zone a pre-ocean intra-plate volcanism manifested itself during the end of Triassic and the beginning of Liassic.

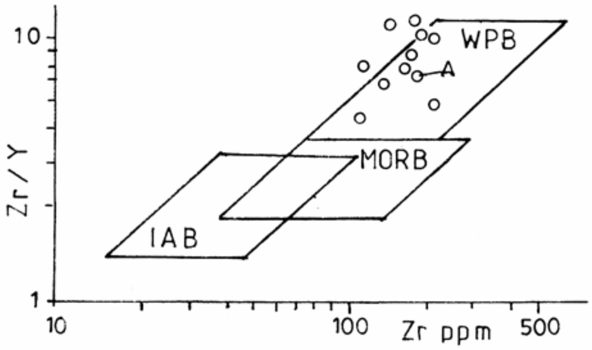

Fig. 2. Plot of the pre-ocean intra-plate Coştei basalts on the Ti *vs*. Zr diagram. Fields according to Pearce and Norry[4]. MORB, mid-ocean ridge basalts. Data from Savu *et al.*[5]

Vestiges of this volcanism occur now at Coştei, in the north of Banat, south of MOS[5].These consist of tholeiitic intra-plate basalts (Fig. 2). The lavas of this volcanism are associated with Triassic sedimentary deposits. Some dolerite dykes crossing the crystalline schists of Poiana Ruscă Mountains, described by Mureşan[6], represent products of the same Triassic intra-plate magmatism.

OPENING AND EVOLUTION OF THE MUREŞ OCEAN

The opening of the Mureş Ocean started in Liassic (ca. 180 Ma[7]) along the Mureş rifting zone (Fig. 3) by spreading processes. This ocean presented a bent shape with two extension branches, that were parallel to the Carpathian Ocean, which opened at the same time along the main branches of the Carpathian triple junction, as shown above. If the Mureş Ocean had an eastern failed branch in its bent zone, like the Dobrogea branch of the Carpathian Ocean, is now hard to say, because this part of the MOS was subsided into the Transylvanian Basin[8, 9]. However, a common characteristic of the two oceans is the fact that their southern branches opened along the north margin of the corresponding continental rifting branch.

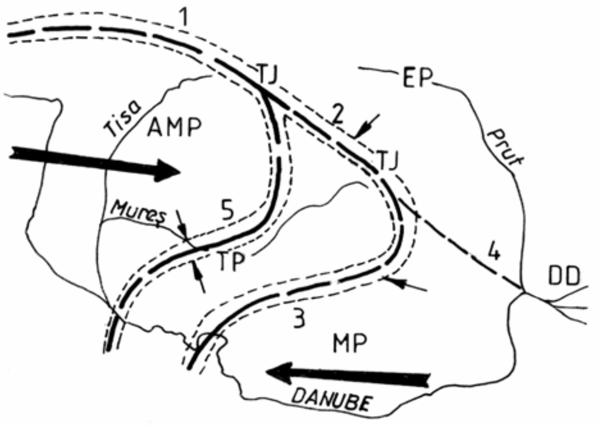

Fig. 3. Sketch-map (not to scale) showing the evolution of the Alpine pre-ocean continental rifting and ocean zones up to the two actual ophiolitic sutures from the Carpathian area: 1, West Carpathians; 2, East Carpathians; 3, South Carpatians; 4, Dobrogean failed branch; thick lines: 1, 2 and 3, the Carpathian Ocean (suture); 5, the Mureş Ocean (suture); broken lines mark the supposed boundary of the flysch deposits (Upper Jurassic-Lower Cretaceous) with or without island arc volcanics and olistoliths of ocean floor rocks; the thin arrows indicate the subduction directions of the convergent plates; thick arrows mark the clockwise rotation tendency of the AMP+TP and MP under the control of the EP and the African Plate (not represented); EP, East-European Plate; AMP, Apuseni Mountains Plate; TP, Transylvanian Plate; MP, Moesian Plate.

The bent shape of the Carpathian Chain – including now the ophiolitic sutures (Fig. 3) – was imprinted since the end of the Precambrian, as shown by the old Precambrian and Paleozoic crystalline schist fold structures and the elongated granitoid plutons. These structures are oriented according to the direction of the Carpathian Chain branches[3]. Its genesis was controlled by the East-European Plate and the African Plate motion, that determined the occurrence of a couple formed of the microplates and oceanic zones from the region, which had a clockwise rotation tendency (see also Savu[3]). Recently, Burchfiel (Fig. 9),[10] Bortolotti *et al.* (Fig. 12*)*[11] and other authors considered this bent shape (Fig. 3) as being determined by the Miocene eastward shifting of the block situated north of the TTFS (Fig. 1), that was considered as a strike-slip fault. On this occasion the Mureş ophiolitic rocks which would have been formed somewhere at ca. 300 km west, have been shifted into the actual position. Such an opinion can not be supported, because TTSF is neither Miocene in age nor a strike-slip fault. It was formed starting from Paleogene (ca. 48 Ma)[12] and occurred due to the emplacement of the Transylvania mantle plume[13] as a fault system along which the adjacent blocks were moving in a vertical sense, thus generating there a thin continental rift, in which hotspot volcanics erupted.

The Mureş Ocean developed due to the spreading process. And, if the annual rate of this process would be considered of only 2 cm, it would result that the ocean width could have been at the end of about 400 km. The evolution of this ocean most probable led to the formation of a classical ocean crust. But, in the ophiolitic plate bilaterally obducted from this ocean there are to be found only the two upper complexes of such a classical ocean crust, as follows: the pillowed basalt complex and the sheeted dyke complex [1, 14] (Fig. 4).

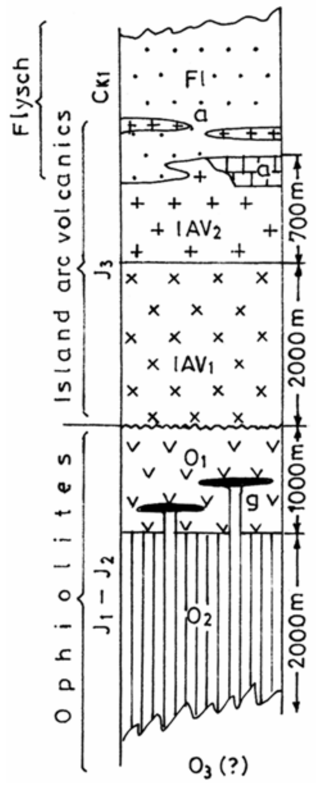

Fig. 4. Stratigraphic column showing the complexes of ophiolitic and island arc rocks from MOS. Fl. flysch deposits and island arc volcanics; IAV$_2$, island arc leucocratic volcanic complex, a, Upper Jurassic-Lower Cretaceous reef limestones; IAV$_1$, island arc andesite-basaltic complex; unconformity; O$_1$, ocean floor pillowed basalt complex; g, gabbro and ultrabasic small bodies intruded into the basalt complex; O$_2$, sheeted dyke complex.

It is notable that Saccani *et al.*[15] also presented a stratigraphic column of the MOS formations, in which the basic and ultrabasic bodies were not considered as small bodies, but were represented as forming the lower basic and ultrabasic layers of a classical complete ocean crust.

a. *The pillowed basalt complex (O_1).*The basaltic rocks of this complex extend about all over the outcropping area of the obducted ophiolitic plate (Fig. 1) and occur, also, as olistoliths in the island arc volcanics from the east of the ophiolitic suture, there forming a *mélange* with pyroclastic matrix[16]. This complex consists of submarine lava flows of basic rocks like basalts, anamesites, hyalobasalts (tachylites) and seldom variolites and amigdaloidal basalts, which usually occur in pillow lava facies. Between the lava flows there rarely occur intercalations of basaltic tuffs and agglomerates associated with tachylites. Some jasper layers and recrystallized limestones are associated with. The basaltic rocks exhibit intersertal, intergranular and radial textures.

Within the basaltic complex six small layered gabbro bodies are located[17 18], consisting of diopside gabbro, olivine gabbro, troctolite and Ti-magnetite gabbro, rarely hyperites, as well as three small ultrabasic bodies formed of peridotite and picrite[19, 20.] The structure of these layered bodies is that of elongated lacolith, (Julița, Cuiaş-Toc), intrusive sheet (Căzăaneşti-Ciungani) or of swollen dyke (Almăşel, Almaş Seliste). Their length does not exceed 6 km; the ultrabasic bodies are much smaller These intrusive bodies have been supplied with magma by a feeding dyke intruding the basalt complex (see Fig. 4).

b. *The sheeted dyke complex (O_2).* This complex I formulated as a sheeted dyke complex during the field researches in the Dumbrăvița–Baia area in the summer of 1981 (unpubl. rep.) and published it in a special paper in 1984[21]. But the presence of dyke structures in this area I pointed out as far as in 1953 (unpubl. rep.). In the meantime references to this complex as well as to the pillowed basalt complex have been made in a general paper issued in 1981[14]. This complex is situated beneath the pillowed basalt complex, as shown in Figure 4. It outcrops in the west part of MOS (Fig. 1), where it thrust over the Upper Jurassic–Lower Cretaceous formations occurring in the northwestern trench of the ophiolitic suture, but a slice of sheeted dykes is tectonically crossing the island arc volcanics at Ampoița, in the Trascău Mountains. This complex extends over a length of about 35 km. and a width of ca. 10 km, between Bata in Banat and Saturani Valley in the Crişul Alb basin. It consists of a sequence of parallel dykes oriented close to one another like the leaves in a book These dykes are generally trending NE –SW and are strongly dipping with 70^0 S–86^0 N. The thickness of dykes usually ranges between 0.5 to 2 m rarely more. They consist mostly of intergranular basalts, dolerites and albite dolerites, but within the northwest part of this area, between Dumbrăviţa and Baia, dykes of diopside gabbro and Ti-magnetite gabbro occur, too[22]. The dykes of basic rocks like dolerites or intergranular basalts show on one or on either margin a narrow chill zone, consisting of hyalobasalt.

Sometimes, between the thick basic dykes there occur very thin dykelets, which consist, like the chill margins of the big dykes, of hyalobasalt. The elongated plagioclase and augite phenocrysts in these dykelets are disposed parallel to their walls, according to the magma motion.

As shown by Savu[21] and Savu *et al.*[23], the rocks of both basaltic and sheeted dyke complexes are tholeiitic rocks, which are bearing a N-Type MORB signature (Figs. 5 and 6). The tholeiitic character of the MOS ophiolites has recently been confirmed by Saccani *et al.*[15] and Bortolotti *et al.*[11] by analysing the clinopyroxenes from the ophiolitic rocks like gabbros, ultrabasics and ocean floor basalts, data which plot on the diagram in Figure 7 in the tholeiitic domain.

A very characteristic aspect of the sheeted dyke complex is the presence among its rocks of the ocean floor trondhjemitic rocks[28], which are to be found within the Dumbrăviţa, Baia and Lupeşti area. Such trondhjemitic dykes also occur in association with the gabbro bodies of Juliţa, Cuiaş-Toc and Almăşel. It clearly shows that their original acid magma differentiated from the tholeiitic parental magma the gabbroic bodies came from. These dykes consist of quartz-diorites, albite granophyres, plagiogranites, plagiaplites and felsites (Fig. 8). In their composition enter quartz, albite and a green hornblende, which occur in long crystals. It is notewothy that, whereas these ocean floor trondhjemites usually contain as melanocratic mineral a green hornblende, the island arc trondhjemitic plutonic rocks or quartz-keratophyres contain biotite. More rarely an amphibole is associated with, as it will be shown farther down. In plagiogranites and aplites these minerals occur in isometric crystals, and in albite felsites they form garlands or rosettes of albite and very thin

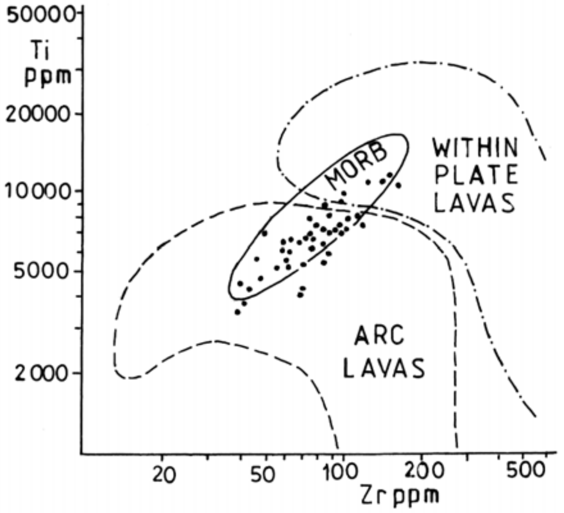

Fig. 5. Plot of the ophiolitic rocks on the Ti *vs.* Zr diagram. Fields according to Pearce[24]. Data from Savu *et al.*[23] MORB, mid-ocean ridge basalts.

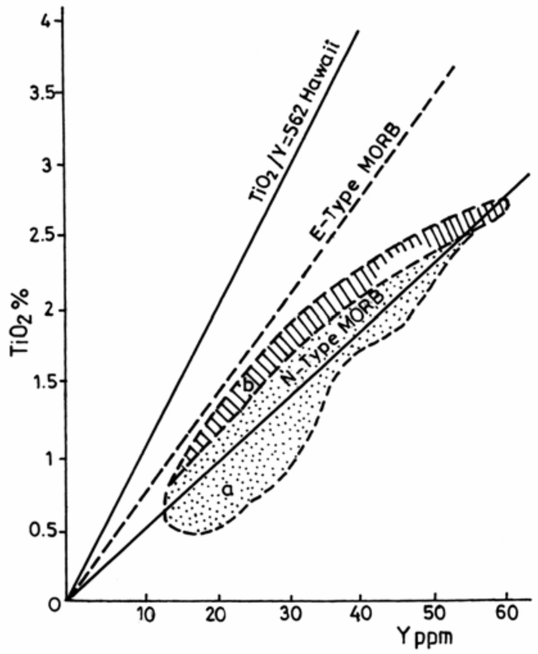

Fig. 6. Plot of the ophiolitic rocks on the TiO_2 *vs.* Y diagram. Fields according to Perfit *et al.*[25]. Data from Savu *et al.*[23]: a, ophiolites from the western ridge segment; b, ophiolites from the eastern ridge segment.

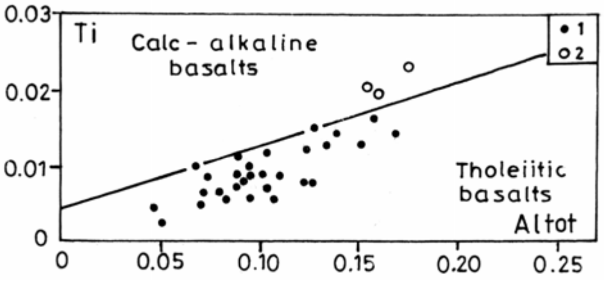

Fig. 7. Plot of the clinopyroxenes from the ophiolites and island arc volcanics of MOS on the Ti *vs.* Al tot diagram. Fields according to Leterrier *et al.*[26]: 1, clinopyroxenes from the Drocea Mountains N-Type MORB ophiolites (data from Saccani *et al.*[15] and Bortolotti *et al.*[11]); 2, clinopyroxenes from the Metaliferi Mountains island arc calk-alkaline basic volcanics (data from Berbeleac and David[27]).

and elongated augite and hornblende crystals. In association with these trondhjemitic dykes there are to be found some albite dolerite dykes, which are similar to some dykes crossing the old Scandinavian shield, where they were called spilitic rocks[29]. It is noteworthy to show that such ocean floor trondhjemites also occur in association with the ophiolites from Oregon, Newfoundland, Corsica et cetera (see Coleman[30] and Thayer[31]).

According to the geochemical characteristics of the ocean floor (MORB) rocks from MOS,, this ophiolitic suture could be separated into two segments: a western segment including the rocks from the Drocea Mountains and an eastern segment including the rocks from the Metaliferi and Trascău Mountains. It seems that the area between Căzăneşti and Zam represents a transitional zone between these segments. The rocks from the western segment are richer in CaO, Ba and Rb and show high values of Rb/Sr ratio. Those from the eastern segment are richer in Na_2O and Sr and show higher values for the La/Yb, (La/Sm)N, (Ce/Sm)N and (La/Ce)N ratios. The chondrite-normalized REE patterns show a Tb negative anomaly in the western segment rocks and an Eu negative anomaly in those from the eastern segment[22,33].

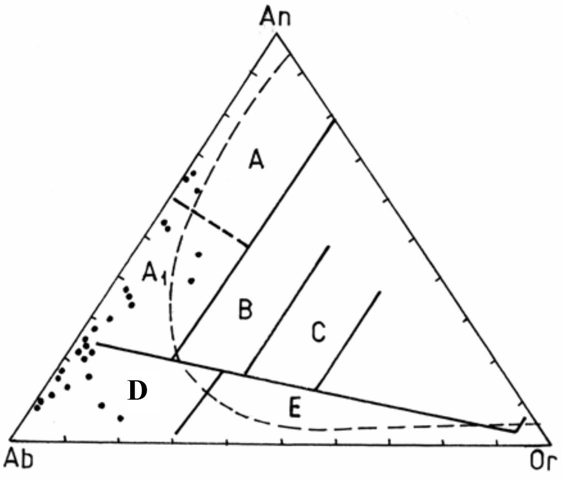

Fig. 8. Plot of trondhjemitic rocks and related quartz rocks from the MOS area on the Ab-An-Or diagram. Fields according to O'Connor[32]. Data from Savu[28]. A, melanocratic quartz rocks; A_1; intermediate quartz rocks; B, granodiorites, C, adamellites, D, trondhjemites, E, granites. The broken line separates, to the left, the low pressure feldspars field.

The geochemical differences between the rocks of these segments of MOS could have been determined by different factors like the mantle heterogeneity, the melting conditions of the mantle source and the differentiation of the parental tholeiitic magmas.

CLOSING OF THE MUREŞ OCEAN AND THE OCCURRENCE OF THE RELATED ISLAND ARC VOLCANISM

The closing of the Mureş Ocean started by the end of the Oxfordian and was due to a bilateral subduction process[1] of the ocean crust attached to the convergent continental tectonic microplates like the Transylvanian Plate and the Apuseni Mountains Plate (Fig. 9). Started at the end of Oxfordian, the bilateral subduction process continued up to the Lower Cretaceous, there resulting the actual MOS, that was affected by some tectonic processes, later on. The subduction process was an asymmetric one. As shown in Figure 9, the subduction on the two flanks of the Mureş Ocean was different. It was of Mariana-type on the northern flank and of Andean-type on the southern flank (see Uyeda[34] for nomenclature). The Mariana-type subduction manifested itself on an almost vertical plane. On the contrary, the Andean-type subduction from the southern flank of this ocean manifested itself on a slopping plane of about 45^0. Due to these two different subduction processes the rocks from the axial zone of the Mureş Ocean have been bilaterally obducted like an ophiolitic plate which, in fact, represents the so-called Căpâlnaş-Techereu Nappe described by Lupu[35]. The actual thickness of the obducted ophiolite plate was established by geophysical methods at about 3ooo m (Andrei, oral comm.).

By the collision of the two convergent plates beneath the obducted ophiolitic plate, the last got suspended over the collision plane, situated between the convergent plates, along which a slice of basic rocks was caught. This situation led, in the past, to the supposition that the Mureş ophiolitic rocks would be located in the initial place of eruption. Now, it is hard to say if the obducted ophiolitic plate is still connected to the slice of basic rocks caught between the two convergent plates. The collision plane and its basic rock slice have been detected by geophysical investigations (Andrei, oral comm.) under the Neogene deposits farther toward southwest up to the Serbian border. This observation shows that the connection between the Mureş Ocean and the Vardar Zone was made by northeast. Which is in evident contrast with the supposition of Bortolotti *et al.*[11], that the connection between the MOS ophiolites and those from the Dinarides was made by northwest, where the original place of the Mureş ophiolites was supposed by these authors.

The extension toward northeast of both the Mureş Ocean and the MOS, respectively, was confirmed by the presence of Upper Jurassic volcanics at the base of the Transylvanian Basin deposits[36] and by the olistoliths of the same rocks occurring in the sedimentary deposits from the Maramures and Transcarpathia areas[37].

The bilateral subduction that took place within the Mureş Ocean generated an island arc volcanism. It is better to show that in 1955 (unpubl. rep.) I separated the MOS vocanics from the Troaş-Pârneşti area into two rock series: an older ophiolitic series, which was described above, and a volcanic series associated with the Upper Jurassic-Lower Cretaceous deposits, that derived from central-type volcanoes (see also Savu[38]). Later on the last series was considered as an island arc bimodal volcanic series[39]. The bilateral subduction directly influenced the character of the following island arc volcanic activity, which generated a volcano-plutonic association. Thus, according to the two types of subduction, there occurred two types of island arc calc-alkaline volcanism: a Mariana-type volcanism and an Andean-type volcanism.

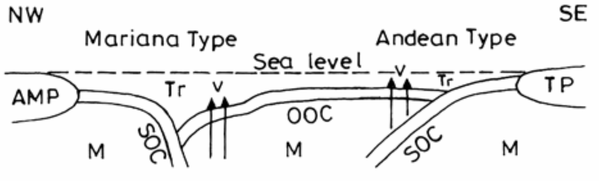

Fig. 9. Model (not to scale) showing the closing of the Mureş Ocean by a bilateral subduction and the obduction of the median ocean crust as an obducted plate. According to Savu[1], revised. AMP, Apuseni Mountains Plate;TP, Transylvanian Plate; SOC, subducted ocean crust; OOC, obducted ocean crust; M, mantle.

a. *The Mariana-type island arc volcanism* In the northern subduction trench, where a Mariana-type subduction manifested itself, a bimodal island arc volcanism was active[39]. Its products were associated with Stramberg limestones and red argillites. This volcanism generated basic rocks and acid rocks (quartz-keratophyres), sometimes accompanied by paleo-trachytes (orthophyres), volcanics between which there are not any transitional rocks, like andesites (Fig. 10)[39]. The quartz-keratophyres resulted from magma fractions of trondhjemitic character, as it results from the diagram in Figure 11[40]. In the genesis of the both trondhjemitic magmas of different origin (MORB and island arc) from the MOS, a convergence of phenomena intervened there. Thus, the ocean floor trondhjemitic magma derived from the differentiation of the N-Type MORB tholeiitic magma; and the island arc trondhjemitic magma derived from an arc calc-alkaline parental magma. It is notable that the usually characteristic melanocratic mineral in the arc trondhjemitic rocks is biotite, very rare an amphibole.

Noteworthy, too, is the fact that, sometimes, between the bimodal basaltic rocks of this volcanism high-Mg basalt flows are to be found.[41]. These basalts exhibit numerous large phenocrysts of augite and have a tholeiitic signature (Fig. 10), which is in evident contrast with the calc-alkaline character of the surrounding arc basic rocks. The Figure 10 shows also that the parental calc-alkaline basaltic magma underwent a fractionation process. Thus, by the formation and accumulation of augite crystals from the parental basaltic magma, there resulted a tholeiitic mesh, consisting of melt and augite crystals, the high-Mg basaltic rocks resulted from. And, by extraction of Fe and Mg from the parental basaltic magma for the augite crystallization it got depleted in such elements, there occurring a trondhjemitic residual magma rich in Si and alkalies, in which the quartz-keratophyres and the paleo-trachytes originated.

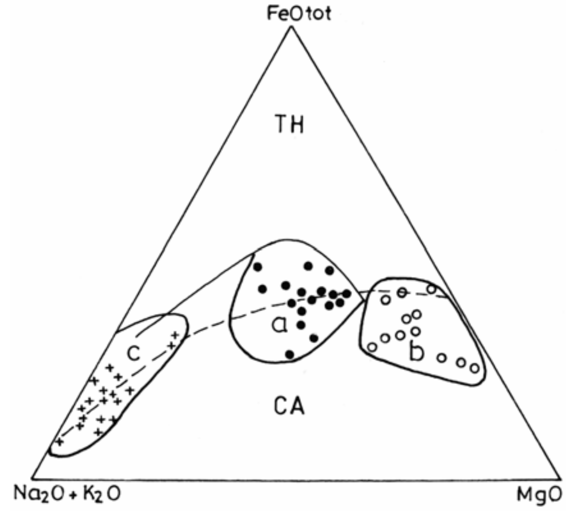

Fig. 10. The differentiation of the Mariana–type parental magma on the FeOtot-Na$_2$O+K$_2$O-MgO diagram Fields according to Irvine and Baragar[42] Th, tholeiitic; CA, calc-alkaline; a, normal arc parental basic magma; b, high-Mg tholeiitic mash (magma); c, residual acid magma. Data from Savu *et al.*[39], and Savu[41].

b. *The Andean-type volcanism.* Within the southern subduction trench of the MOS, which was associated with the Andean-type subduction, the island arc volcanics had a normal calc-alkaline character, being represented by the triad basalt-andesite-rhyolite.

The island arc volcanism was accompanied by a plutonic activity, too. It generated some laccoliths, formed of diorites and granodiorites of 121–128 Ma (Temeşeşti), granodiorites and granites (Cerbia, Pietroasa[43]) and only granites (Săvârşin). The Săvârşin granites include large soda-potash phenocrysts,

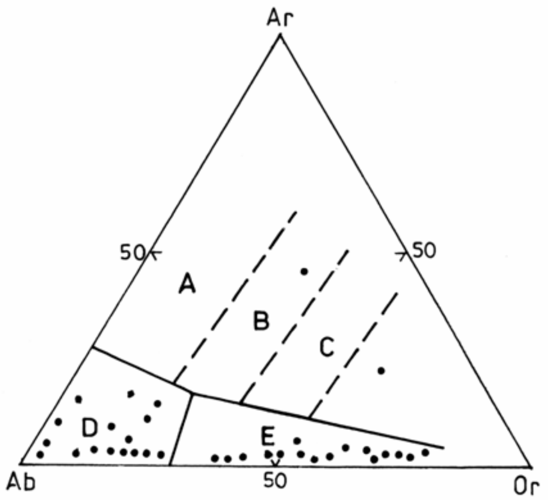

Fig. 11. Plot of the acid rocks of the Mariana-type volcanism on the Ab-An-Or diagram. Fields adapted after O'Connor[32]: A, basic quartz rocks; B, dacites, C, rhyodacites, D, quartz-keratophyres; E, rhyolites. The broken line separates, to the left, the low pressure feldspar field (< 5 kbar), according to Coleman[30].

which show a zoning structure. There occurred also a system of swarm dykes (116–120 Ma) like those situated between Vărădia and Troaş.

METAMORPHISM

The most characteristic and important metamorphic processes, which affected the MOS rocks, manifested themselves during the eruption of the ocean floor and island arc rocks. Less important metamorphic phenomena occurred during and after the closing of the ocean. The types of metamorphism and their effects on the MOS rocks have been summarized by Savu (Table 1)[45]. According to their position as against the Mureş Ocean structure these metamorphic phenomena have been separated in this table into several types: intra-ocean crust, super-ocean crust and syn- and post-ocean closing types.

The contact or the magmatic metamorphism was determined by both the consanguineous gabbro bodies and the granitoid plutons belonging to the island arc volcano-plutonic magmatism, as well as by the post-collision banatitic intrusions. At the contact with the gabbro bodies of Juliţa, Cuiaş-Toc, Almăşel, Almaş-Selişte and Ciunngani-Căzăaneşt the ophiolitic rocks have been transformed at 1000–700^0 C into basic hornfelses (beerbachites) of the pyroxene hornfels-type, in which the clinopyroxene was the main characteristic mineral[45,46]. It is remarkable that at the contact with the Almaş-Selişte body formed of diopside and hornblende gabbro, there resulted hornfelses containing both pyroxene and hornblende as index minerals[18].

The metamorphism of the ophiolitic rocks determined by the late-magmatic and hydrothermal solutions released by the magma during the emplacement of the sheeted dyke complex and by the gabbro bodies as well, led to the occurrence of the following three metamorphism facies, pointed out as far back as in the middle of the last century[47]: epidote-amphibolitic (650–580^{0}C); epidote-chloritic(greenschist) at 400–375^{0}C and calcite-zeolite (375–200^0 C) facies, as shown by Savu[47,48]. Such metamorphic phenomena are to be found especially within the sheeted dyke complex and in the gabbro bodies, but also in the pillowed basaltic complex, where hot solution released from the magma chambers penetrated through.

The metamorphic processes that manifested themselves on the ocean floor have been of two categories. First of all, there must be mentioned the spilitization process. This process was determined by the reaction between the superheated basic lavas, erupting on the ocean floor, and the ocean salty water This reaction rose the temperature of the ocean floor water up to 400^0 C, at least. Thus, it determined the substitution of the original melanocratic minerals (pyroxene, hornblende, biotite) by chlorite and eventually epidote, and of the basic plagioclase by albite, there resulting spilites *sensu stricto.* The process was consistent with the experiment performed by Eskola *et al.*[49]. However, 15 per cent of the basaltic rocks from this area were not affected by the spilitization process (Fig. 12). At the same time the layers of Liassic limestones occurring in the basaltic complex, on the alignment situated between Căpâlnaş, Zam and Vălişoara recrystallized.

In the island arc basic volcanics, especially, amygdaloidal basalts occurred, the vesicles of which have been filled with secondary minerals resulted from the spilitization process, like chlorite, calcite, zeolite and secondary iron oxides[48].

Under almost the same conditions, but at a lower temperature (< 200^0 C), the diagenesis phenomena, took place in which a migration of the chemical elements was active.

During the Mureş Ocean closing and after that, the metamorphism processes continued. Thus, due to the pressure determined by the nappes superposition, there resulted phenomena of an incipient load metamorphism (400–200^{0}C) with prehnite and pumpelyite, as for instance in the

Trascău Mountains[50, 45]. On the thrust plane due to the friction between the Bedeleu Nappe and its autochthone there manifested itself, in the same area, a dynamo-thermal metamorphism with sericite and chlorite, and the Stramberg limestones at the base of the nappe recrystallized[45].

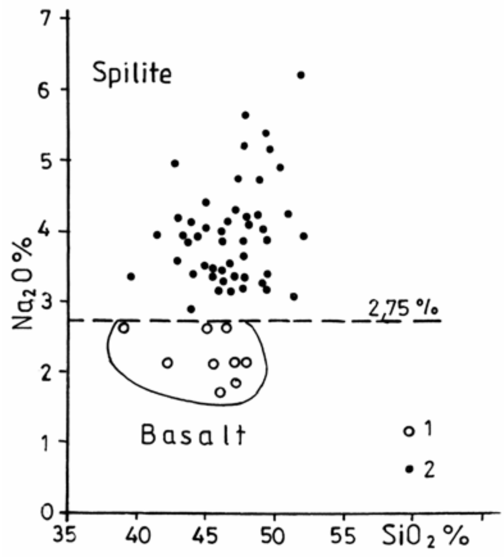

Fig. 12. Na_2O *vs.* SiO_2 discriminant diagram for the N-Type MORB basalts and related spilites from the MOS area. Fields according to Savu:[48] 1, tholeiitic basalts; 2, related spilites.

Along the fracture planes of the TTFS from the south of MOS, due to the friction between the adjacent blocks, moving in a vertical sense, phenomena of a dynamic metamorphism manifested themselves and pseudo-conglomerates and fault argillites formed there.

TECTONICS

The MOS tectonics was strongly controlled from the beginning by the bilateral subduction, that led to the obduction of the ophiolitic plate on the margin of the two convergent tectonic plates (Fig. 9)[40], a process that ended at the beginning of Lower Cretaceous. As shown above, the actual remnant of this obducted ophiolitic plate was called the Căpâlnaş-Techereu Nappe[35]. The Subhercynian and Laramian movements that followed, affected the MOS structure especially on its margins, causing the formation of the nappe-scales from the northwest subduction trench of Drocea and Trascău Mountains, as well as the Tisa Nappe in the south of MOS. Thus was outlined the base structure of the asymmetric Mureş Orogen which formed on the initial MOS structure[51].

Soon after the Laramian collision, the Apuseni Mountains, the Mureş Orogen inclusively, have been affected by two fracture systems: an east-west longitudinal fault system and a diagonal fault system.

In the Mureş Orogen area the longitudinal fault system determined the occurrence of a series of faults mostly in the south part of it, e.g. the Căpâlnaş –Vorţa fault, on which the Vorţa copper mineralization was deposited. Due to the later movements of the adjacent blocks this mineralization was sheared. On the occasion of the Transylvania mantle plume emplacement during the Tertiary period, there took place the TTFS which cut the MOS formations along the south margin of this structural unit (Fig. 1)[13].

The diagonal system of *en échelon* fractures has segmented the Mureş Orogen into several blocks, the tendency of which was to rise toward west and to subside toward east-northeast. Therefore, by the rising of the westernmost blocks the Mureş Orogen formations, including the obducted ophiolitic plate from the superstructure, were strongly eroded. There rested only the infrastructure represented by the two convergent plates, formed of the old crystalline schist and granitoid bodies, and the above mentioned ophiolitic rock slice caught between these plates, which extended thus toward the Serbian territory, under the Neogene deposits (Fig. 1).

Toward east-northeast the block tendency was to subside, so that a big part of the bent zone of MOS was subsided into the Transylvanian Basin[36] Consequently, toward east the MOS Mesozoic formations, including the obducted ophiolitic plate, were not so deeply eroded. As for instance, in the Trascău Mountains the obducted ophiolitic plate is deep situated beneath the island arc volcanics and the Mesozoic sedimentary deposits.

CONCLUSIONS

MOS resulted from the evolution of the Mureş Ocean, the opening of which started in Liassic During the evolution of this ocean an ocean crust was formed, consisting of tholeiitic rocks bearing a N-Type MORB signature. It is remarkable that the Coştei pre-ocean intra-plate (OFB) basalts are tholeiitic rocks, too, like the ocean floor basalts which erupted into the following Mureş Ocean. This fact is in an evident contrast with the post-collision Tertiary intra-plate basalts which, if not alkaline, are calc-alkaline like their forerunner island arc basalts.

The closing of this Mureş Ocean, which started by the end of Upper Jurassic, was determined by an asymmetric bilateral subduction, due to which an ophiolitic plate was obducted over the margins of the Transylvanian and Apuseni Mountains convergent plates.

Concomitantly, an island arc volcanism manifested itself within the Mureş Ocean area, which engendered bimodal volcanics of Mariana-type in the Drocea-Trascău subduction trench and normal calc-alkaline volcanics of Andean-type in the southern trench. It determined also a volcano-plutonic rock association.

The MOS tectonics was strongly influenced by the subduction/obduction structure, as the initial folding structures occurred mostly within the marginal subduction trenches, thus resulting the base structure of the Mureş Orogen. These marginal structures were affected later on by the Subhercynian and Laramian movements, when the main thrust structures occurred.

Soon after the Laramian collision the Mureş Orogen was affected by longitudinal and diagonal fracture systems. The longitudinal fractures cut the southern part of the Mureş Orogen without any horizontal shifting of the adjacent blocks. The diagonal fractures determined the segmentation of the Mureş Orogen into several blocks, the tendency of which was to rise toward west, with the erosion of the initial MOS formations, and to subside toward east, where a large part of the MOS bent zone was subsided into the Transylvanian Basin.

The final conclusion is that the closing of the Mureş Ocean by an asymmetric bilateral subduction/obduction process resulted in the asymmetric divergent Mureş Orogen, formed on the initial MOS structure.

REFERENCES

1 Savu H., *An. Inst. Geol. Geofiz.*,**1983**, *LXI*, 253

2\. Cioflica G., Lupu M., Nicolae I., Vlad Ş., *An. Inst. Geol. Geofiz.*, **1980**, *LVI*, 79.

3\. Savu H., *Geo-Eco-Marina*, **2005**, *9–10*, 116.

4\. Pearce J.A., Norry M.J., *Contrib Mineral. Petrol.*, **1979**, *69*, 33.

5\. Savu H., Udrescu C., Neacşu V., Stoian M., *Rev. Roum. Géol.*, **1992**, *36*, 35.

6\. Mureşan M., *An. Inst. Geol.*, **1992**, *XLII*, 7.

7\. Herz N., Jones M., Savu H., Walker R.L *.Bull. Volc.* **1974**, *XXXIII – 4*, 1110.

8\. Savu H., *Rev. Roum, Géol*.**,** **1990**, *34*, 13.

9\. Savu H., *An. Univ. Bucureşti (Geologie)*, **1995,** *XLIV*, 17.

10\. Burchfiel B.C., *Tectonophysics*, **1980**, *63*, 31.

11\. Bortolotti V., Marroni M., Nicolae I., Pandolfi L., Principi G., Saccani E., *Ofioliti*, **2004**, *29, 1*, 5.

12\. Savu H., Udrescu C., Neacşu V.*, Rom. J. Petrology*, **1996**, 77, 83.

13\. Savu H., *Proc. Rom. Acad., Series B*, **2004***, 1,* 33.

14\. Savu H., Udrescu, C., Neacşu, V., *Ofioliti,* **1981**, *6 (2–3),* 269.

15\. Saccani E., Nicolae I., Tassinari R., *Ofioliti*, **2001**, *26*, 9.

16\. Savu H., Udrescu C., Neacşu V., *Rom. J. Petrology*, **1992**, *75*, 53.

17\. Cioflica G., Savu H.*, Stud, cerc.geol.*, **1960**, *7, 1*, 693.

18\. Savu H., Udrescu C., Neacşu V., *Miner. Slovaca,* Kosice, **1984**, *16, 1*, 43.

19\. Savu H., *D.S. Com. Geol.*, **1962**, *XLV***,** 52.

20\. Savu H., Vasiliu, C., Udrescu, C., *D.S. Inst. Geol.,* **1970**, *LVI, 1*, 219.

21\. Savu H., *Rev. Roum. Géol.,* **1984***, 28*, 29.

22\. Savu H., Stoian M., Tiepac I., Grabari G., Popescu G., *Rom. J. Petrology*, **1995**, 76, 77.

23\. Savu H., Udrescu C., Neacşu V., Ichim M.*, Rom. J. Petrology.*, **1994**, *76*, 67.

24\. Pearce J. A., In: A. Panayotou (Ed.), *Ophiolites, Proc. Intern.Ophiol. Symp.,* **1979,** Nicosia, 261.

25\. Perfit M.R., Gust D.A., Bence A.E., Arculus P.J., Taylor B.R., *Chem. Geol*., **1980,** *30*, 227.

26\. Leterrier J., Maury R.C., Thoron P., Girard D., Marshal M., *Earth Planet. Sci. Lett.* **1982**, *59*, 139.

27\. Berbeleac I., David M., *Rev. Roum. Géol., Géophys., Géogr., (Géol.)*, **1977,** *21*, 63.

28\. Savu H., *Rev. Roum. Géol*., **2002**, *46*, 29.

29\. Merilinen K., *Acad. Dissertation,* Helsinki, **1961**, 75 p.

30\. Coleman R.G., *Ophiolites,* Berlin, **1977**, 229 p.

31\. Thayer, T.P., Geotectonics, Moscow, **1977,** 6, 32.

32\. O'Connor J.T.*, U.S. Geol. Surv. Prof. Paper,* Washington, *D.C.*, **1965**, *525,* 79.

33\. Savu H, Stoian M., *Rev. Roum.Géol., Géophys., Géogr., (Géol.*), **1988**,*32*, 37.

34\. Uyeda S., *Geol. Surv. Japan,* **1981** *261*, 1.

35\. Lupu M., *Rev. Roum.Géol., Géophys., Géogr*., (Géol.), **1976**, 20, 21

36\. Savu H., Bombiţă G., Udrescu C., Grabari G.*, An. Univ. Bucureşti, (Geologie),* **1995,** *XLIV*, 17.

37\. Bombiţă G., Savu H., *Ann. Soc. Geol., Poloniae***,** **1986**, *56*, 337.

38\. Savu H., *D.S. Com. Geol.*, **1962,** *XLIV*, 137.

39\. Savu H., Udrescu C., Neacşu V., *D.S. Inst. Geol. Geofiz*., **1986**, *70–71*, *1*, 153.

40 Savu H., *Proc. Rom. Acad., Series B*, **2003***, 5, 1–2*, 39.

41\. Savu H., *Proc. Rom. Acad., Series B***,** **1999**, *1, 1*, 45.

42 Irvine T.N., Baragar W.R.A.*, Can. J. Earth Sci*., **1971**, *8*, 523.

43\. Cioflica G., *An. Com. Geol.*, **1962**, *XXXII*, 257.

44\. Savu H., *Rom. J. Tectonics and Reg,Geology.*, **1995**, *76*, 73.

45\. Savu H., *An, Univ. Bucureşti, (Geologie),* **2000**, *XLIX*, 3.

46\. Savu H., Vasiliu C.,Udrescu C., *D.S. Inst. Geol. Geofiz***.,** **1981**, *LXVI, 1*, 127.

47\. Savu H., *Acta Geol. Acad. Sci. Hung*., **1967**, *11, 1–3*, 59.

48\. Savu H., *Proc. Rom. Acad., Series B*, **2002**, *4, 3*, 157.

49\. Eskola P., Vuoristo U., Rankama K., *Soc. Géol. Finlande,* **1937**, *10*, 61.

50\. Seghedi A., Oaie G., Mărunţiu M., Nicolae I., Rădan S., Ciulavu M., Vanghelie I., *An. Inst. Geol.Rom.,* **1996**, *69, (Supl. 1),* 181.

51\. Savu H*., An. Univ. Bucureşti, (Geologie),* **1999**, *XLVIII*, 3.