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On the cover: Pillow lavas in a Jurassic ophiolite complex in the quarry near Zam, Transylvania, Romania. Photo: Corina Ionescu and Volker Hoeck.

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Mesozoic ophiolites and granitoids in the Apuseni Mountains

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1. Geological introduction

The geology of Romania comprises several tectonic mega-units with different geological history and physio-geographic features, displayed in Fig. 1. Among them the most conspicuous features are the Eastern Carpathians, the Southern Carpathians and the Apuseni Mts. The Eastern Carpathians, the Southern Carpathians form a Z-like thrust-and-fold belt, which developed during the Alpine orogeny. The movement is directed towards the east in the Eastern Carpathians (Săndulescu, 1984), and the south and east in the Southern Carpathians, respectively (Berza *et al.*, 1994). The Transylvanian Basin separates the Carpathians and the Apuseni Mts.

In the following, we will briefly present the main features of the geology of the Romanian Carpathians and the Apuseni Mts. with emphasis on the Mesozoic to Cenozoic magmatites.

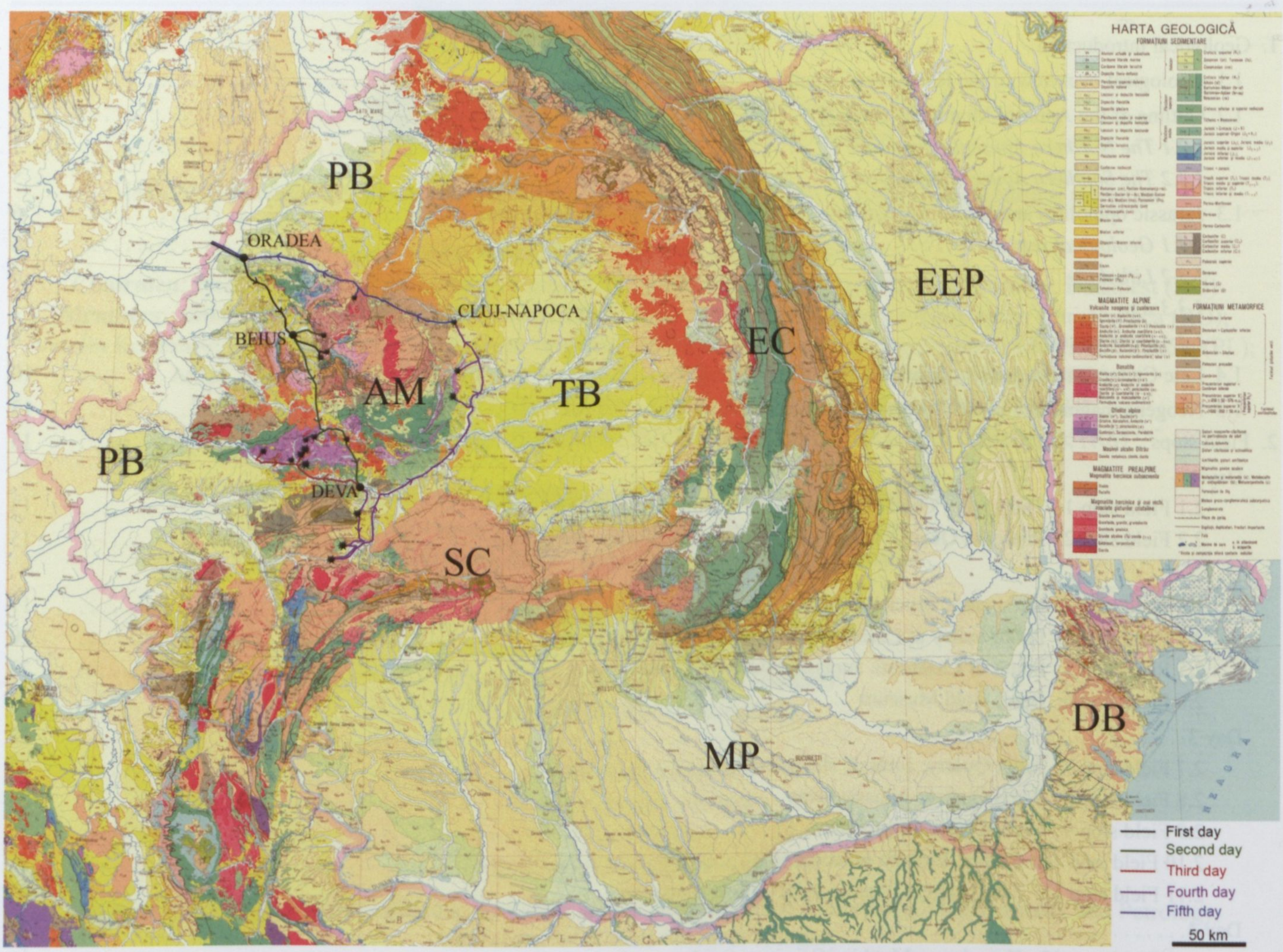

Fig. 1. Geological map of Romania (by Săndulescu *et al.*, 1978), modified with the RO-2 field trip route and major tectonic units.
Abbreviations: PB – Pannonian Basin, AM – Apuseni Mts., TB – Transylvanian Basin, EC – Eastern Carpathians, EEP – Eastern European Platform, SC – Southern Carpathians, MP – Moesian Platform, DB – Dobrogea.

1.1 Geological overview of the Romanian Carpathians

The tectonics of the Romanian Carpathians and their geodynamic relation to the Apuseni Mts. are a matter of debate. The most cited concept was developed by Săndulescu (1984, 1994). For the time being it is still the only unifying concept including the Apuseni Mts. and the Carpathians. It is schematically displayed in Fig. 2. The central feature of his concept is the so-called Main Tethyan Suture, which marks the trace of the Tethys and connects the Vardar Ocean (Vardar Zone) in the south, with the Piemont Ocean represented by the Pieniny Klippen belt, in the north-west. According to Săndulescu, two ophiolite nappes, both called "Transylvanides", originate in the Main Tethyan Suture: one is thrust to the north-west over the Northern Apuseni Mts. and the other towards the east over the nappes of the Eastern Carpathians (see below). Structurally, these units represent the uppermost nappe in the Apuseni Mts. and in the Eastern Carpathians, respectively. To avoid confusion among the two "Transylvanides", *i.e.* the Apuseni one and the Eastern Carpathians one, respectively, we will use the term Mureş Zone for the highest tectonic units in the Apuseni Mts. (see also Ionescu *et al.*, 2009b and Hoeck *et al.*, 2009).

The units beneath the "Transylvanides" are termed by Săndulescu (1984) "Dacides" and consist of a crystalline basement and a sedimentary cover. The "Dacides" form Mesozoic nappes and occur in the Apuseni Mts. as "Inner Dacides", and in the Eastern and Southern Carpathians as "Median Dacides", "Outer Dacides" and "Marginal Dacides", respectively (Fig. 2). The "Inner Dacides" form the nappe system of the Northern Apuseni Mts. (see Section 1.2). The "Median Dacides" in the Carpathians comprise the Supra-Getic and Getic nappes in the Southern Carpathians and the Bucovinic, Sub-Bucovinic and Infra-Bucovinic nappes in the Eastern Carpathians. The "Outer Dacides" consist of remnants of an older ocean, including its sedimentary cover. In the Southern Carpathians, they are represented by the Severin Nappe, the only one containing real ophiolitic remnants. In the Eastern Carpathians the "Outer Dacides" comprise the Black Flysch, the Baraolt and the Ceahlău nappes, with some volcanics. The "Marginal Dacides" are only found in the Southern Carpathians and are represented by the Danubicum. The outermost units are the Cenozoic "Moldavides" in the Eastern Carpathians, comprising the Convolute Flysch Nappe, the Audia and the Tarcău nappes, as well as the Subcarpathic Flysch Zone. The Molasse Foredeep forms the margin towards the Eastern European Platform.

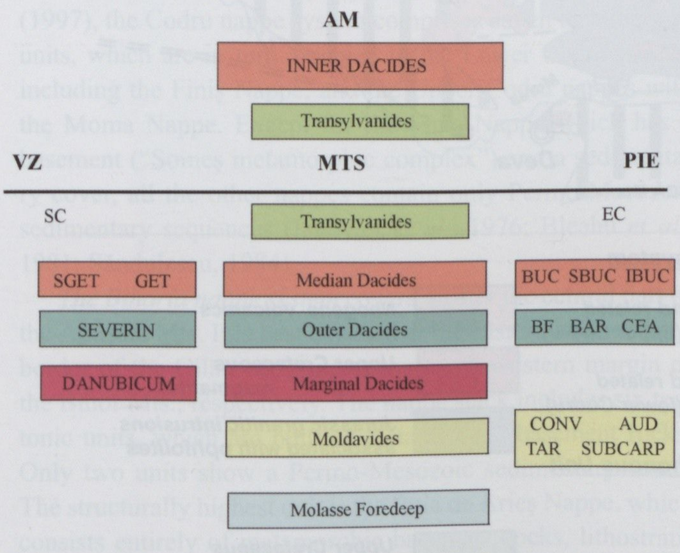

Fig. 2. Tectonic scheme of the Romanian Carpathians based on Săndulescu (1984).

Abbreviations: AM – Apuseni Mts., VZ – Vardar Zone, MTS (Main Tethyan Suture), PIE – Pieniny Klippen Belt, SC – Southern Carpathians, EC – Eastern Carpathians, SGET – Suprageticum, GET – Geticum, BUC – Bucovinicum, SBUS – Subbucovinicum, IBUC – Infrabucovinicum, BF – Black Flysch, BAR – Baraolt Nappe, CEA – Ceahlău Nappe, CONV – Convolute Flysch, AUD – Audia Nappe, TAR – Tarcău Nappe, SUBCARP – Subcarpathian Flysch.

1.2 Outline of the Apuseni Mountains geology

The interpretations over the last decades by Bleahu (1976), Burchfiel (1976), Ianovici *et al.* (1976), Săndulescu (1984), Balintoni (1994, 1997), despite different views in details, agree in the overall concept. The specific tectonic configuration is due to the Alpine orogeny and is classically divided in two large structural areas, namely the tectonic units of the Northern Apuseni Mountains (NAM) and the Southern Apuseni Mountains (SAM). The former includes the Bihor autochthonous unit, the Codru nappe system and the Biharia nappe system, the latter the Mureş Zone (Fig. 3). Each of these units has a different composition, lithostratigraphy, geological history and origin. Both are characterized by a complex nappe structure resulting from the Mid- and Late Cretaceous orogenic phases (*e.g.* SAM) and an intra-Turonian phase (*e.g.* NAM). The NAM roughly include geographically, the Plopiş, Meseş, Pădurea Craiului, Bihor, Vlădeasa, Codru Moma and Highiş Mts., whereas the SAM consists of the Metaliferi, Trascău and Drocea Mts. (Bleahu *et al.,* 1981).

The overall architecture of the Apuseni Mts. includes Jurassic, Late Cretaceous and Miocene to Quaternary magmatism. The Middle Jurassic ophiolites and the Late Jurassic Island Arc Volcanics (IAV) are restricted to the SAM, whereas the Late Cretaceous calc-alkaline intrusives and extrusives (banatites) as well as the calc-alkaline Miocene and Quaternary volcanics are found both in NAM and SAM.

1.2.1 The Northern Apuseni Mountains

The tectonic units that build up the northern segment of the Apuseni Mts. are termed as *Apusenides* (Fig. 3) and are part of the Tisia unit (Kovács, 1982; Fülöp *et al.*, 1987; Csontos, 1995; Csontos & Vörös, 2004; Császár, 2006). According to Haas *et al.* (2001) these units cover a large area ranging from Croatia via northern Serbia and Hungary to western Romania. It is

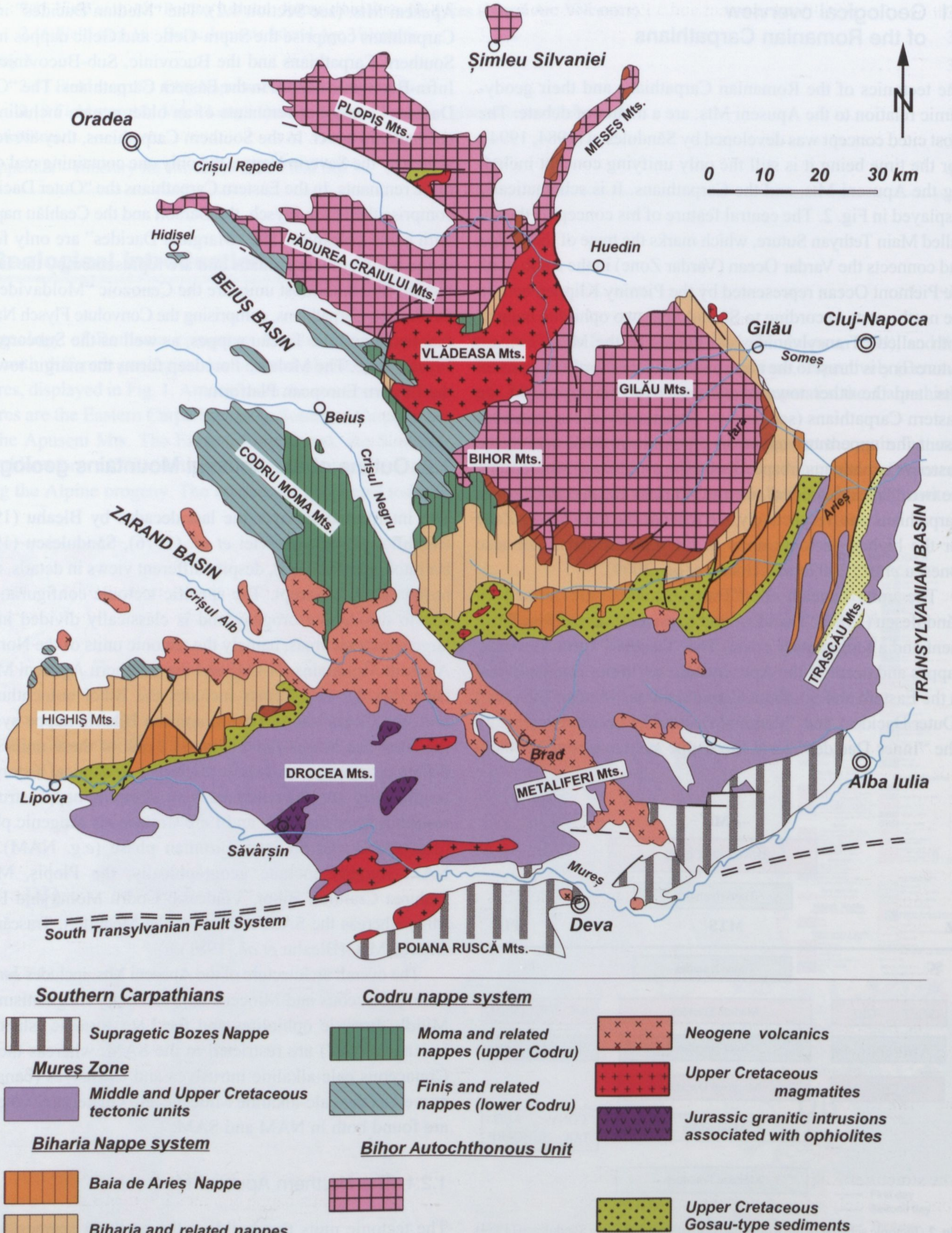

Fig. 3. Simplified Alpine structure of the Apuseni Mountains [from Ionescu *et al.* (2009a) compiled by C. Balica from papers by Ianovici *et al.* (1976), Bleahu *et al.* (1981), Săndulescu (1984), Kräutner (1996), and Balintoni & Puște (2002)].

built up of a crystalline basement and a Permo-Mesozoic sedimentary sequence ranging up to the Late Cretaceous. The Tisia unit (micro-continent) is believed to be originally part of the Variscan Europe, from which it derived and rotated in a complex way (Pătraşcu *et al.*, 1994) until it reached the present position (Haas *et al.*, 2001).

The architecture of the Apuseni Mts. contains, as the structurally deepest unit, the "Bihor Autochthonous Unit" or the "Bihor Unit". It is of a regional extent and has a relative autochthonous position in respect to the higher nappe systems, covering a good part of the northern segment of the Apuseni Mts. (Fig. 3). The structurally higher units can be grouped, according to their origin and lithological content, in two nappe systems, thrust on top of the "Bihor Autochthonous Unit" (see below): the deeper Codru nappe system and the tectonically higher Biharia nappe system. Each system is believed to originate in different areas of the Tisia micro-continent.

"The Bihor Autochthonous Unit" consists of a metamorphic basement and a sedimentary cover. The basement, known as the "Someş metamorphic complex", is a lithostratigraphic unit comprising medium-grade micaschists and quartzites, gneisses, amphibolites and ultramafics. One of its dominant features is the frequent occurrence of pegmatites, large-scale migmatization and the presence of large granitic intrusions such as the Codru and Muntele Mare plutons. The cover is represented by Permo-Mesozoic siliciclastic and calcareous sediments. It has to be noted here that rocks of the "Someş metamorphic complex" can also be found in the basement of some Alpine nappes of the Codru and Biharia systems (Balintoni, 1997; Balintoni & Puşte, 2002).

The Codru nappe system is located at the south–western side of the "Bihor Autochthonous Unit" and covers a smaller area than the Biharia nappe system. According to Balintoni (1997), the Codru nappe system comprises seven tectonic subunits, which are in turn assigned to the Lower Codru nappes including the Finiş Nappe, and the Upper Codru nappes with the Moma Nappe. Except for the Finiş Nappe which has a basement ("Someş metamorphic complex") and a sedimentary cover, all the other nappes contain only Permo-Mesozoic sedimentary sequences (Ianovici *et al.*, 1976; Bleahu *et al.*, 1981; Săndulescu, 1984).

The Biharia nappe system covers mostly the central part of the Apuseni Mts. It is best exposed at the eastern and southern border of the Gilău Mts. and at the south-western margin of the Bihor Mts., respectively. The nappe stack includes six tectonic units, which are built of metamorphic basement rocks. Only two units show a Permo-Mesozoic sedimentary cover. The structurally highest unit is the Baia de Arieş Nappe, which consists entirely of metamorphic basement rocks, lithostratigraphically summarized as "Baia de Arieş metamorphic complex". It comprises medium- to high-grade Variscan micaschists with cm-sized garnet and staurolite porphyroblasts, paragneisses, metamorphosed limestones and dolomites. Furthermore, eclogites as inclusion in some marbles (Mărunţiu *et al.*, 2004) and high grade amphibolites (Radu, 2003) have been described. Several granitoids occur as well.

Most of the other basements rocks are lithostratigraphically assigned to the "Biharia metamorphic complex", which is lithologically very different from the other basement rocks. It is defined by the presence of extensive metabasites (greenschists with albite porphyroblasts, probably metatuffs) accompanied by more acidic rocks and metamorphosed ultramafics. Marbles, terrigenous sediments as well as black quartzites occur subordinately (Giuşcă, 1960; Ianovici *et al.*, 1976).

1.2.2 The Southern Apuseni Mountains

Mid-Jurassic ophiolites, Upper Jurassic island arc volcanics, Jurassic and Cretaceous sediments build up the Southern Apuseni Mts. In this chapter we will concentrate on the main structural units, a detailed discussion of ophiolites and Island Arc Volcanics being given below.

Regarding the formation of the Jurassic magmatics, Săndulescu (1984) envisaged a major ocean between the continental crust of the Northern Apuseni Mts. (Tisia micro-continent) on one side and the continental crust of the Carpathians (Getic nappes in the Southern Carpathians and the Bucovinian nappes in the Eastern Carpathians, respectively) on the other side (Fig. 2). The remnants of this ocean, which disappeared in the "Main Tethys Suture Zone" are the ophiolites in the Southern Apuseni Mts., including the island arc sequence and their continuation beneath the Transylvanian Depression, and also the small ultramafics and basaltic to andesitic blocks in the Eastern Carpathians. These remnants have tectonically the highest position in the Apuseni Mts. as well as in the Eastern Carpathians, and were included by Săndulescu (1984) into the term "Transylvanide" or "Transylvanian nappes". The related nomenclature problems were already outlined in Section 1.1 (see also Hoeck *et al.,* 2009 and Ionescu *et al.,* 2009b).

The tectonic subunits of the Mureş Zone have been emplaced during two orogenic phases, the intra-Albian ("*Austrian"*) phase and the intra-Maastrichtian ("*Laramian")* phase, respectively. Accordingly, two sets of tectonic units were generated. The "Baia de Arieş metamorphic complex" acted as basement for the intra-Albian tectonic units. Among the many Mid-Cretaceous tectonic units, the Rimetea–Bedeleu Nappe in the north-eastern part of the SAM is important. It is built up by a thick set of Upper Jurassic Island Arc Volcanics and the overlying Upper Jurassic limestones.

The second Late Cretaceous orogenic event, *i.e.* the "Laramian" phase, generated a new system of nappes, whereby each Upper Cretaceous nappe contains fragments of the previously generated units. The most important "Laramian" nappe containing the Jurassic ophiolites and the IAVs is the Căpâlnaş–Techereu Nappe in the western part of the SAM (Balintoni & Iancu, 1987) (Fig. 3). The post-Albian sedimentary cover is also deformed and thrust to the NW during this phase.

Regarding the Mid-Cretaceous and Late Cretaceous orogenic phases and the resulting nappes, some unsolved problems remain, as for instance: the precise age of the pre-Albian and post-Albian sedimentary formations, the correlation between the Mid-Cretaceous tectonic units incorporated into the Upper Cretaceous nappes, the distance of the tectonic displacement or the relation between the Apuseni Mountains and the Southern Carpathians.

1.3 Jurassic ophiolites and Island Arc Volcanics in the Southern Apuseni Mountains

1.3.1 Geological setting

The Mureş Zone represents a complicated nappe system (Bleahu *et al.*, 1981; Balintoni, 2003) and consists of ophiolites and superimposed on these, calc-alkaline island arc volcanics (Fig. 4). Apart from the detailed mapping by Savu (1962), a number of excellent petrographic description and geochemical data were published by Savu (1982), Savu *et al.* (1982), Nicolae (1995), Savu & Udrescu (1996) a.o. During the last decades, both complexes, ophiolites and Island Arc Volcanics, were again investigated by Saccani *et al.* (2001), Bortolotti *et al.* (2002) and Nicolae & Saccani (2003), who presented a large set of modern mineralogical and geochemical data.

The geological situation in the Căpâlnaş–Techereu Nappe, along the western boundary of the IAVs, which are deposited on top of the ophiolites, is somewhat different from their northeastern boundary in the Rimetea–Bedeleu Nappe. At the western margin, for example in the vicinity of the Zam Quarry (Field stop 11), the Island Arc Volcanics are clearly deposited on top of the ophiolites (see also Nicolae & Saccani, 2003). In the Trascău Mts. (Rimetea–Bedeleu Nappe) slices of continental basement are still preserved beneath the IAVs (Giuşcă *et al.*, 1967a,b). Whether the contact between these crustal fragments and andesitic rocks is tectonic or depositional remains an open problem.

1.3.2 Lithology

A mantle section is entirely missing in the ophiolites (see also Saccani *et al.*, 2001). The sequence comprises only crustal portions beginning with rare ultramafic cumulates at the base, followed by layered and isotropic gabbros including Fe-Ti gabbros (Fig. 5). A sheeted dyke complex represents the transition from the plutonic to the volcanic sequence and consists of basaltic to basaltic-andesitic dykes. The volcanic sequence includes basaltic lava flows, pillow basalts and basaltic pillow breccias.

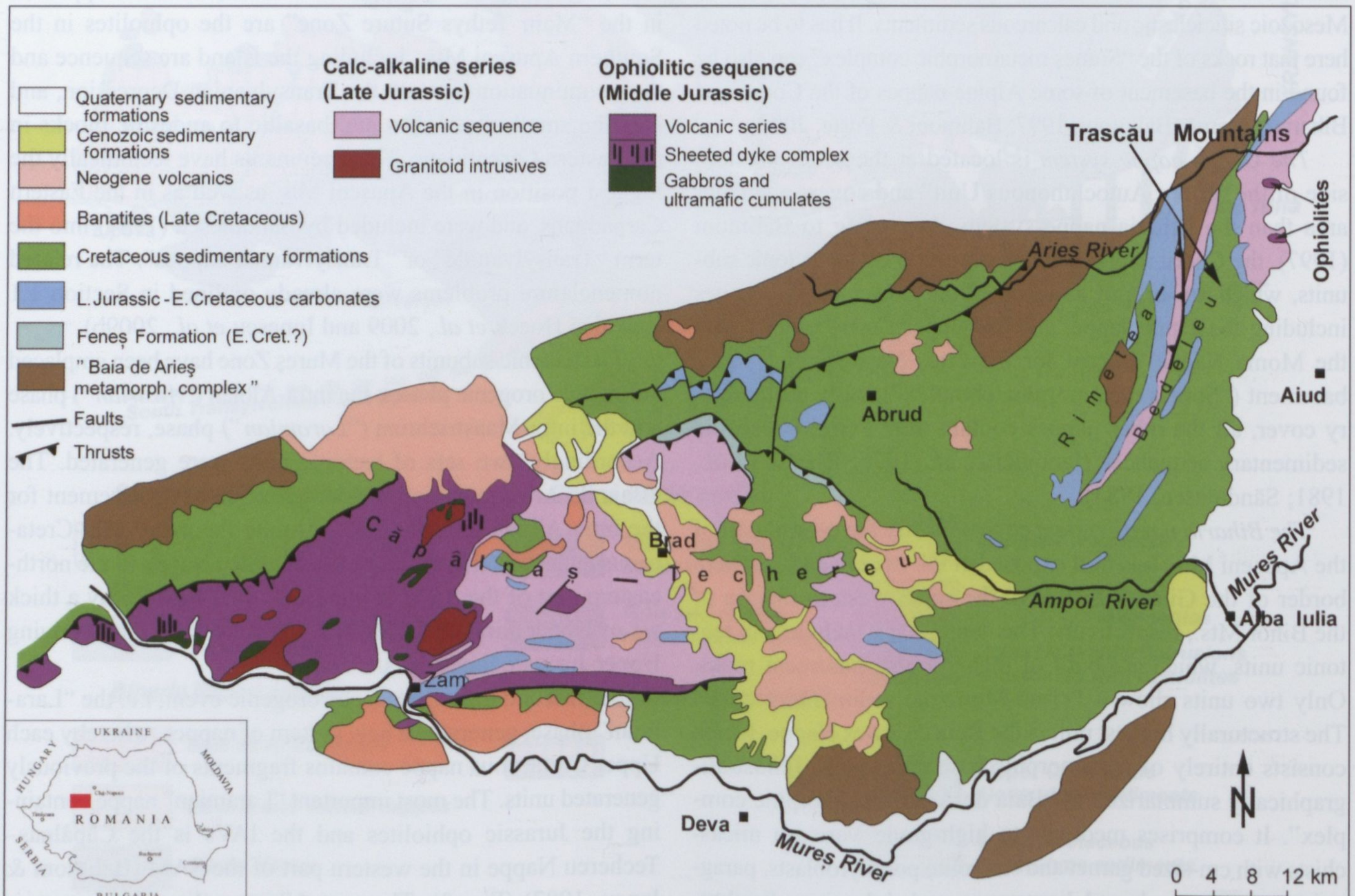

Fig. 4. Simplified geological map of the Southern Apuseni Mountains (redrawn and modified after Saccani *et al.*, 2001). The insert in the bottom left shows the position of the map within Romania.

The short lithological and petrographic description is based on the conception of Saccani *et al.* (2001), Nicolae & Saccani (2003) and Bortolotti *et al.* (2004). The deepest preserved parts of the ophiolites are cumulates and isotropic gabbros. Occasionally, the cumulates display an ultramafic composition. The isotropic gabbros are predominantly clinopyroxene gabbros, rarely with pseudomorphs after olivine. Gabbros are very often enriched in Fe and Ti, which eventually leads to the formation of Fe-Ti gabbros (Cioflica, 1962; Savu, 1996). Gabbros grade upwards in the ophiolite pseudostratigraphy into a sheeted dyke complex consisting of basalts to basaltic-andesites (Fig. 5). The dyke complex can be well studied in some quarries *e.g.* Julița (Field stop 7). The distribution of the outcrops indicates a relatively large area of NE–SW striking sheeted dykes in the western part of the ophiolites (Fig. 4). The pillow lavas above the dykes are partly inter-fingering with the latter as can be seen for example in the northern part of Zam Quarry (Field stop 11). The pillow lavas and also the massive lava flows are aphyric to scarcely porphyritic, with plagioclase and clinopyroxene appearing mainly as micro-phenocrysts. The groundmass of the basaltic rocks consists of acicular plagioclase, clinopyroxene and Fe oxide. The former glassy groundmass is completely altered into chlorite and some clay minerals.

Gabbros were subject to an ocean-floor metamorphism, inferred from the replacement of clinopyroxene by amphibole, plagioclase by calcite, prehnite and epidote. The sheeted dykes also show some replacement of pyroxene by amphibole, whereas in the volcanics the typical alteration minerals are albite, chlorite and calcite.

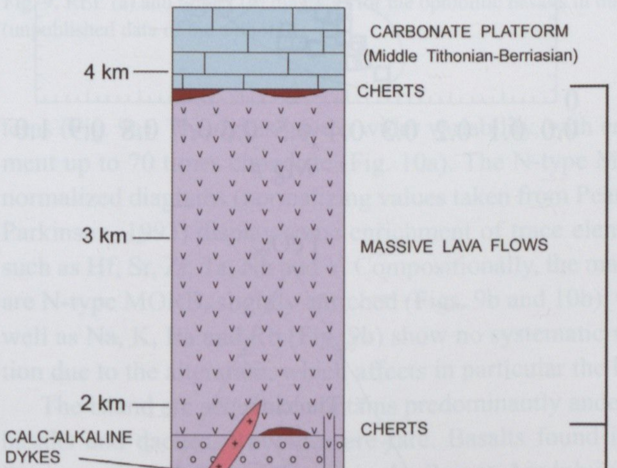

Fig. 5. Stratigraphic column of the Jurassic ophiolites and IAVs in the Southern Apuseni Mountains (redrawn and modified from Bortolotti *et al.*, 2002, 2004).

In few places, the ophiolites are overlain by thin radiolarites, in others directly by the Upper Jurassic island arc volcanics. The latter consist predominantly of calc-alkaline basalts and andesites, but dacites and rhyolites, as well as tholeiitic basalts occur as well. The volcanics represent remnants of an island arc developed above the oceanic crust. By contrast to the ophiolitic lavas, the IAVs are often highly porphyritic, with phenocrysts of plagioclase, clinopyroxene and amphibole. The plagioclase composition ranges from oligoclase to labradorite (Fig. 6). Secondary feldspars, *i.e.* orthoclase and albite (Fig. 6), are frequent. The rhyolitic and dacitic dykes found in the island arc sequence are partly porphyritic, with phenocrysts of feldspar, quartz and biotite in a quartz-feldspar groundmass.

1.3.3 Age

The ophiolites are at several locations intruded by Upper Jurassic granites, granodiorites and diorites. They postdate the ophiolites and are interpreted by Bortolotti *et al.* (2002) as the plutonic equivalents of the calc-alkaline island arc sequence. Zircons from the intrusions were dated by Pană *et al.* (2002) with U/Pb as Oxfordian–Kimmeridgian in age (152–156 Ma). Recently, Zimmerman *et al.* (2008) published Re-Os ages on molybdenite from granitoids near Săvârşin and Cerbia with an age of 159.1 to 159.8 Ma. Thus, the ophiolites are older, which is in accordance with the biostratigraphic results by Lupu *et al.* (1995), who found a Callovian to Oxfordian age for radiolarites on top of the ophiolites. The biostratigraphic findings in limestone deposited on top of the IAVs indicate a Tithonian age for the beginning of the sedimentation (Săsăran, 2006).

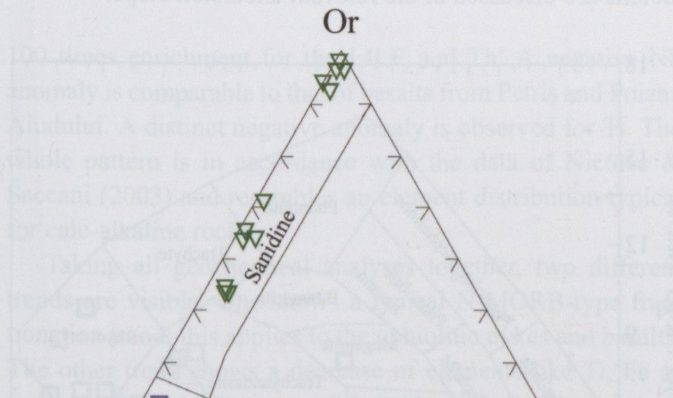

Fig. 6. The distribution of feldspar compositions in the IAVs from the Trascău Mts. Primary feldspars (circles: plagioclase; squares: anorthoclase) and secondary feldspars *i.e.* alkali feldspars, including albite (triangles). Electron Microprobe Analyses (EMPA): unpublished data of the authors.

The distribution of the ophiolites is mirrored in the conspicuous geomagnetic pattern in the Southern Apuseni Mts., which is due to the high content of Ti-bearing magnetic iron oxides (Ionescu *et al.,* 2009b). The whole lithologic composition and in part also the magnetic anomaly support the model connecting these oceanic remnants with the Eastern Vardar Zone (Andelcović & Lupu, 1967), which can be traced from Northern Greece, Makedonia and Serbia in the south, to the Mureş Zone in the north. Their continuation further to the east and north-east beneath sediments of the Transylvanian Basin was drilled in several boreholes. Towards the northern part of the Transylvanian Basin any sign of Mesozoic magmatic activity disappear (Ionescu *et al.,* 2009b). In particular, the widespread intrusion of Jurassic granitoids into the ophiolites is a characteristic feature for the Eastern Vardar Zone. The Dinaridic ophiolites, which were regarded by some authors as being connected with the Mureş Zone (Saccani *et al*., 2001; Bortolotti *et al*., 2002, 2004) are completely different in that they are built up predominantly of large ultramafic units with only occasionally a thin crustal sequence on top of the mantle section (Shallo, 1994; Hoeck *et al*., 2002; Pamić *et al*., 1998, 2002).

1.3.4 Geochemistry

For the following discussion of the geochemistry we rely mainly on our own unpublished data as well as on the papers by Ionescu & Hoeck (2004, 2006) and Ionescu *et al.* (2009 a,b). For the analytical conditions of the electron microprobe analyses and the bulk rock chemistry the reader is referred to Ionescu *et al*. (2009b). Our data are in agreement with those presented by Saccani *et al*. (2001) and Nicolae & Saccani (2003). More details are discussed at the relevant excursion stops.

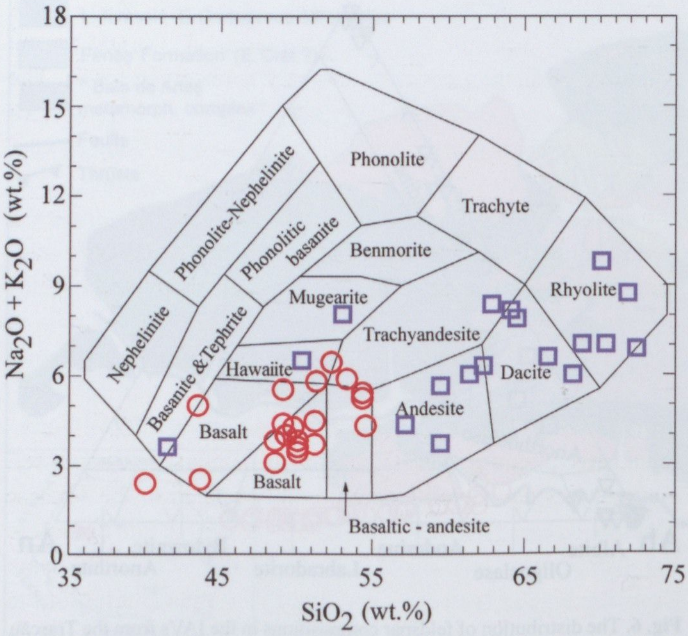

Fig. 7. Na_2O+K_2O *vs.* SiO_2 diagram showing the petrographic range of ophiolitic (circles) and calc-alkaline + tholeiitic island arc volcanics (squares) in the SAM (unpublished data of the authors). Diagram from Ionescu *et al.* (2009a).

The geochemical composition of the ophiolitic dykes and lavas overlap to a wide extent (Saccani *et al*., 2001, and our own unpublished data). Based on the petrographic composition, mineral chemistry and the whole-rock chemistry, Bortolotti *et al*. (2002) concluded that the ophiolites represent a normal ocean crust profile with some gabbros and basalts showing a high-Ti affinity.

The ophiolites combined with the island arc sequence span a range from basalts to rhyolites (Fig. 7). The sheeted dykes and the ophiolitic lavas display an evolutionary trend which is typical for MOR-type basalts, with increasing Ti and Fe content combined with a decreasing Mg# which might be used as fractionation index. In Fig. 8a the negative correlation between TiO_2 and Mg# for the ophiolitic volcanics is clearly visible. In turn, the Fe-enrichment trend of the ophiolitic basalts is displayed in Fig. 8b. In general, the MOR-type basalts and the sheeted dykes span approximately the same range of trace and RE elements (Figs. 9a,b and 10a,b). Basalts show a variable enrichment in the REE, up to 30 times chondrite, and flat pat-

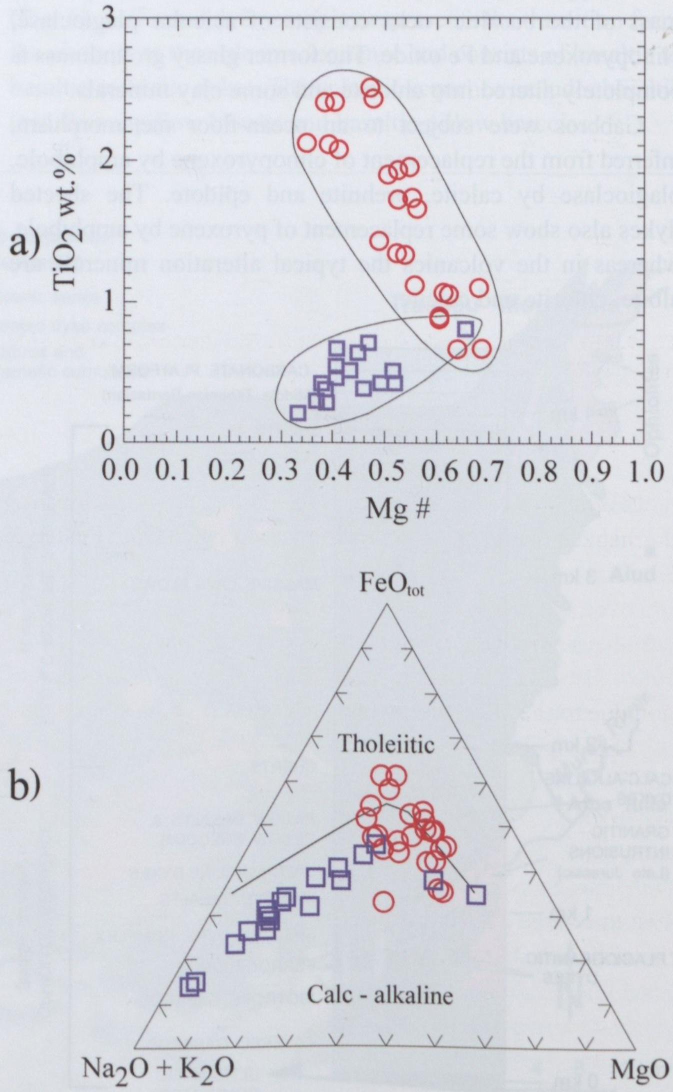

Fig. 8. a) TiO_2 *vs.* Mg# and b) $FeO_{tot} - Na_2O + K_2O - MgO$ diagrams for ophiolitic (circles) and calc-alkaline volcanics (squares) in the SAM (unpublished data of the authors).

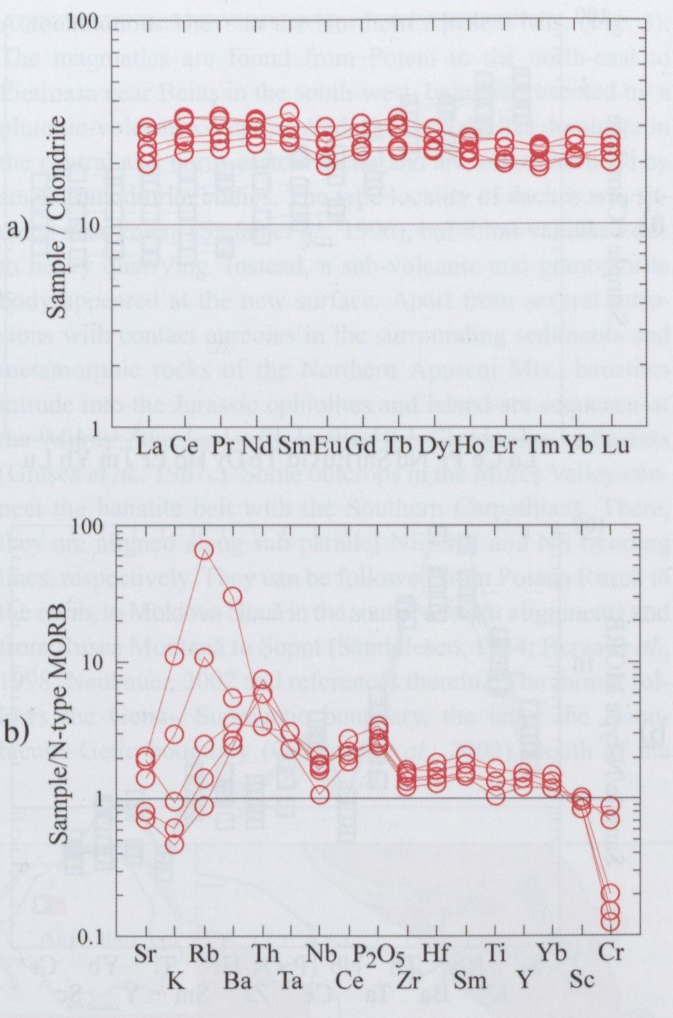

Fig. 9. REE (a) and Spider (b) diagrams for the ophiolitic basalts in the SAM (unpublished data of the authors).

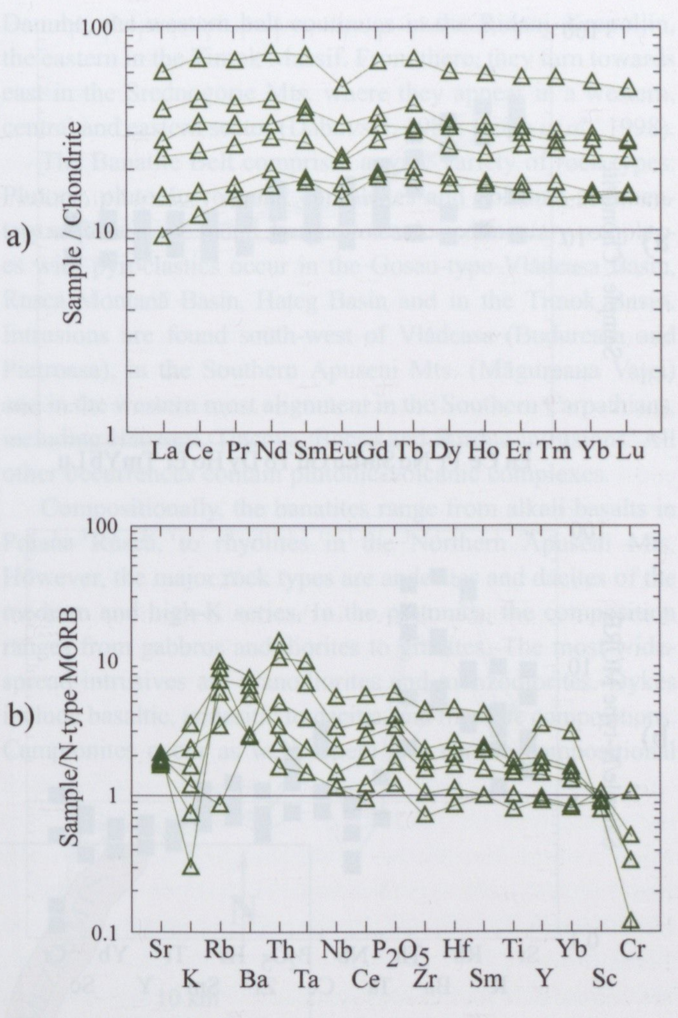

Fig. 10. a) REE (a) and Spider (b) diagrams for the sheeted dykes in the SAM (unpublished data of the authors).

terns (Fig. 9a). The dykes have a wider variability, with enrichment up to 70 times chondrite (Fig. 10a). The N-type MORB normalized diagrams (normalizing values taken from Pearce & Parkinson, 1993) display some enrichment of trace elements, such as Hf, Sr, Zr, Ta, Nb and Y. Compositionally, the magmas are N-type MORB, slightly enriched (Figs. 9b and 10b). Ca as well as Na, K, Ba and Rb (Fig. 9b) show no systematic variation due to the alteration, which affects in particular the lavas.

The island arc sequence contains predominantly andesites, basalts and dacites. Rhyolites are rare. Basalts found in the Petriş quarry (Field stop 8) and in the Poiana Aiudului (Field stop 15) are variably enriched in the LREE over the HREE, which are in the range of N-type MORB (Fig. 11a). Normalized to N-type MORB (Fig. 11b) they show a pattern close to tholeiitic arc basalts, with depleted HFSE (< 1) and slight enrichment in LIL elements.

The REE diagram for andesites (Fig. 12a) displays a pattern with 30-50 times chondrite enriched LIL elements, whereas the MREE and the HREE are close to N-type MORB. This pattern is reflected by a $(La/Yb)_N$ ratio larger than 1. The N-type MORB normalized spider diagram (Fig. 12b) shows a 10 to 100 times enrichment for the LILE and Th. A negative Nb anomaly is comparable to that of basalts from Petriş and Poiana Aiudului. A distinct negative anomaly is observed for Ti. The whole pattern is in accordance with the data of Nicolae & Saccani (2003) and resembles an element distribution typical for calc-alkaline rocks.

Taking all geochemical analyses together, two different trends are visible. One shows a typical N-MORB-type fractionation trend; this applies to the ophiolitic dykes and basalts. The other trend shows a decrease of elements like Ti, Fe as well as of the Mg#, compatible with that of more evolved rocks in which amphibole and magnetite play an important role as fractionating phases. This trend can be observed in the Island Arc tholeiitic basalts, and in the calc-alkaline andesites. Rhyodacites and rhyolites are possibly related to the Jurassic granitoids (Field stop 11). The high Ti content in the ophiolitic basalts and the relatively low content in the IAVs, at the same Zr concentration, are not compatible with an evolution of the island arc basalts from Fe-Ti rich ophiolitic magmas. Two magma sources might explain the different evolution trends among ophiolites and IAVs.

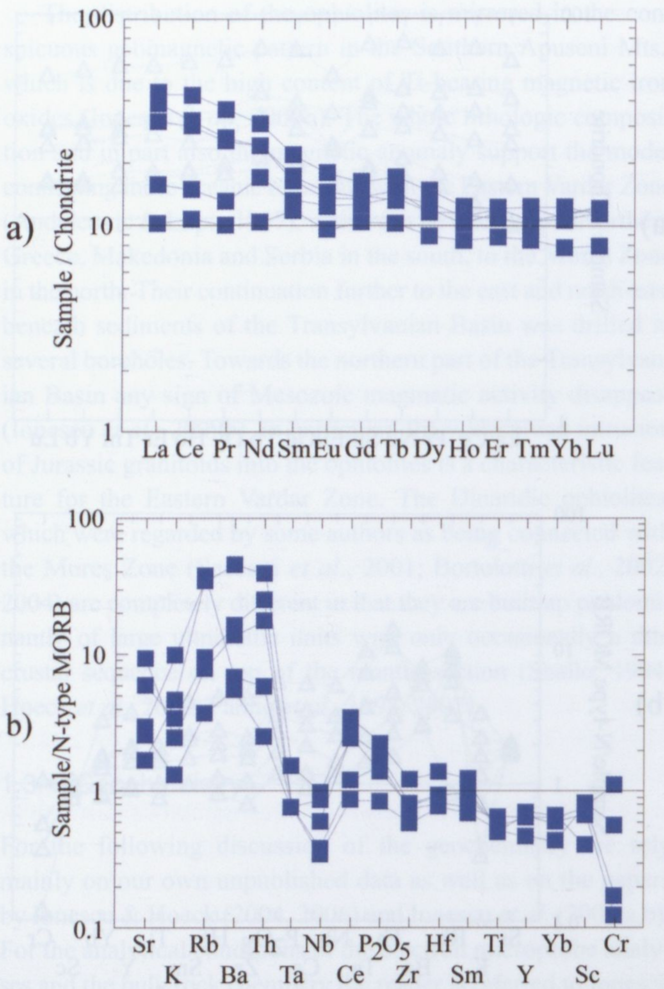

Fig. 11. a) REE (a) and Spider (b) diagrams for the Petriş and Poiana Aiudului tholeiitic basalts (IAV), based on unpublished data of the authors.

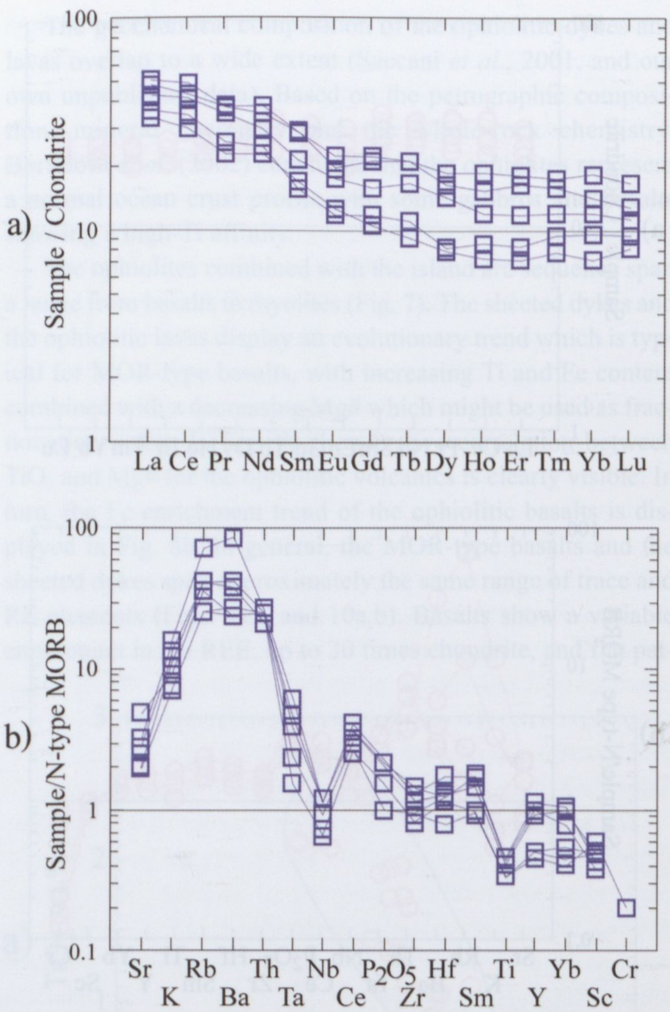

Fig. 12. REE (a) and Spider (b) diagrams for the andesitic IAVs in the SAM (unpublished data of the authors).

1.4 Late-Cretaceous magmatic complexes (banatites)

We will discuss here shortly the further magmatic activities. The next magmatic event after the formation of the IAVs, took place in the Late Cretaceous with the formation of the so-called "banatites" (Fig. 13). These comprise a sequence of calc-alkaline intrusions ranging from granite to diorite, accompanied often by volcanics. The rock name dacite for example is derived from the adaequate banatitic volcanics in the Northern Apuseni Mts. The banatitic magmatism gives rise to some important mineralisation, for example in skarns which formed along the contact of the intrusions with mainly Mid-Trassic limestones (see also Ilinca, 2010). A characteristic feature of the banatites and in particular granodiorites is the frequency of the xenoliths of metamorphic, sedimentary and magmatic nature of various sizes, ranging from 1-2 cm to several cubic meters.

At the end of Cretaceous, during the "Laramian" orogenic phase, widespread intrusives, subvolcanic bodies, as well as volcanics, were formed. This magmatic event can be followed from the Northern Apuseni Mts. (Fig. 13) through the Southern Carpathians into the Srednogorie in Bulgaria. Von Cotta (1864) termed the whole sequence of volcanic and plutonic rocks displaying various compositions as "banatites", according to their occurrence in the Banat, *i.e.* from northern Serbia to south-western Romania. Later on, the whole area of these Upper Cretaceous magmatites was called "Banatitic Province" (Giuşcă *et al.*, 1965, 1966), "Laramian Province" (Cioflica & Vlad, 1973) or "Banat–Srednogorie Rift" (Popov, 1981). The most widespread notation for this belt today is "Banatitic Magmatic and Metallogenetic Belt" (BMMB) according to Berza *et al.* (1998) or "Apuseni– Banat–Timok–Srednogorie Belt" (ABTS) according to Popov *et al.* (2003). The whole belt is L-shaped and extends over 900 km, with a maximum width of 70 km. In the last fifty years the interest of many geologists focused on these magmatic rocks, because of their skarn, porphyry and epithermal deposits, mined in some places since many years (Cioflica & Vlad, 1973; Cioflica, 1989; Popov, 1981, 1995; Ianovici *et al.*, 1976; Vlad, 1979, 1997; Popov *et al.*, 2003).

Over its whole extension, the banatites cross several tectonic units from the north to the south. The northernmost occurrences are the Vlădeasa banatites crosscutting the "Bihor

Autochthonous Unit" in the Northern Apuseni Mts. (Fig. 3). The magmatics are found from Poieni in the north-east to Pietroasa near Beiuş in the south-west, being represented by a plutonic-volcanic complex. Andesites and dacites dominate in the central and north-eastern sector and are in turn intruded by small granodiorite bodies. The type locality of dacites was situated near Poieni (Ştefan *et al.*, 1996), but it has vanished due to heavy quarrying. Instead, a sub-volcanic and granodiorite body appeared at the new surface. Apart from several intrusions with contact aureoles in the surrounding sediments and metamorphic rocks of the Northern Apuseni Mts., banatites intrude into the Jurassic ophiolites and island-arc sequence of the Mureş Zone, as well as into the Cretaceous sediments (Giuşcă *et al.,* 1967c). Some outcrops in the Mureş Valley connect the banatite belt with the Southern Carpathians. There, they are aligned along sub-parallel NE–SW and NS trending lines, respectively. They can be followed from Poiana Ruscă in the north, to Moldova Nouă in the south (western alignment) and from Rusca Montană to Şopot (Săndulescu, 1984; Berza *et al.*, 1998; Neubauer, 2002 and references therein). The former follows the Getic– Supragetic boundary, the latter the Danubicum–Getic boundary (Ciobanu *et al.,* 2002). South of the Danube, the western belt continues in the Ridanj–Krepoljin, the eastern in the Timok Massif. From there, they turn towards east in the Srednogorie Mts. where they appear in a western, central and eastern sector (Dabovski, 1980; Berza *et al.,* 1998).

The Banatite Belt comprises a wide variety of rock types. Plutons, plutonic-volcanic complexes and volcanic-sedimentary sequences are found. Larger volcanic-sedimentary complexes with pyroclastics occur in the Gosau-type Vlădeasa Basin, Rusca Montană Basin, Haţeg Basin and in the Timok Basin. Intrusions are found south-west of Vlădeasa (Budureasa and Pietroasa), in the Southern Apuseni Mts. (Măgureaua Vaţei) and in the western most alignment in the Southern Carpathians, including Hăuzeşti, Tincova, Bocşa and Surduc intrusions. All other occurrences contain plutonic-volcanic complexes.

Compositionally, the banatites range from alkali basalts in Poiana Ruscă, to rhyolites in the Northern Apuseni Mts. However, the major rock types are andesites and dacites of the medium and high-K series. In the plutonics, the composition ranges from gabbros and diorites to granites. The most widespread intrusives are granodiorites and monzodiorites. Dykes include basaltic, andesitic to dacitic, and rhyolitic compositions. Camptonites occur as well. There is a certain compositional

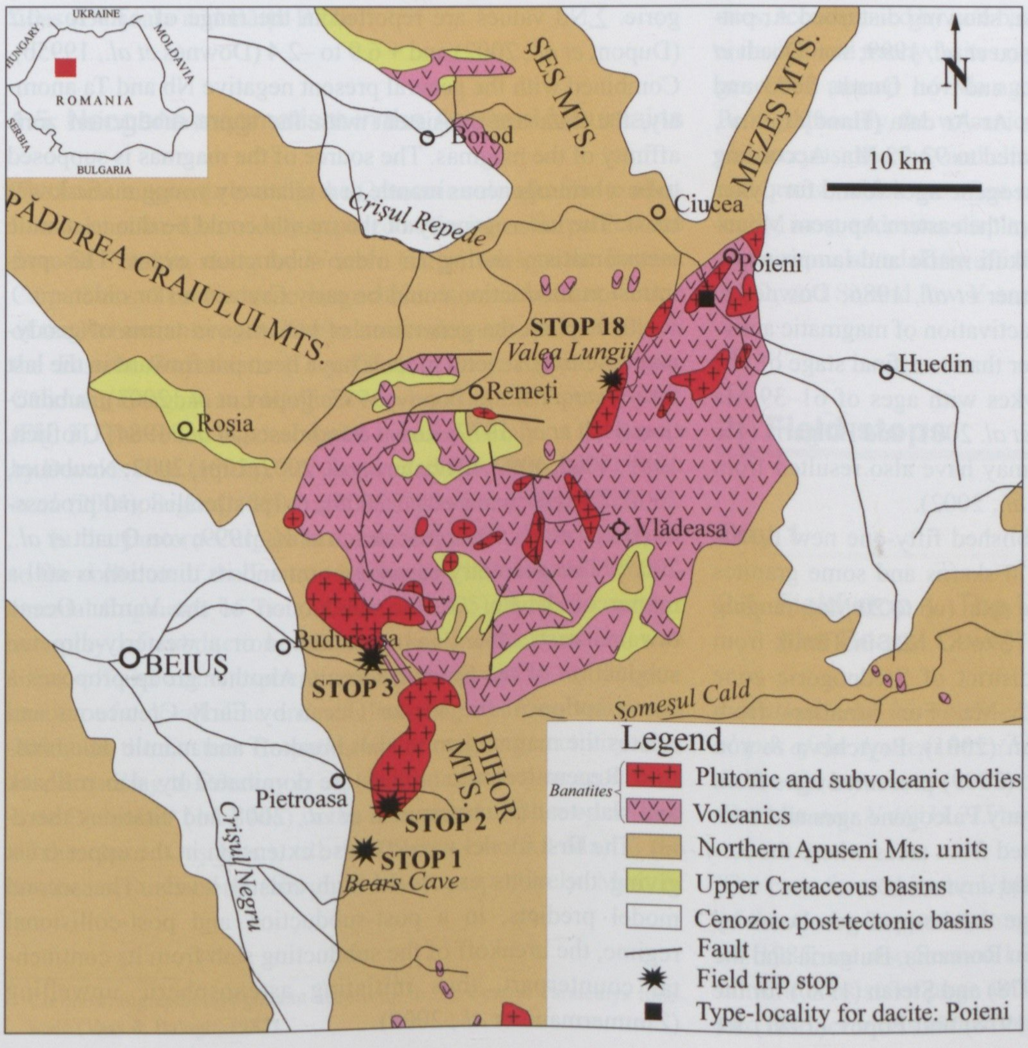

Fig. 13. Distribution of banatites in the Northern Apuseni Mountains, according to Giuşcă *et al.* (1967b,c) and Istrate (1978), simplified. Black stars mark field stops 1, 2, 3 and 18. Black square shows the type locality for dacite, according to Ştefan *et al.* (1996). The position of the map within the Romania is shown in the upper left insert.

distribution discernible, *e.g.* the Northern Apuseni banatites are dominated by dacitic and rhyolitic compositions, those from Bocşa (Russo-Săndulescu *et al.,* 1978) and Surduc in the Banat range from basaltic to trachytic composition, with high-K and a shoshonitic affinity. Banatites from the Southern Carpathians are dominated by medium-K series and basaltic-andesitic to andesitic composition.

Two main igneous stages of banatites are recognized in the Apuseni Mts. (Ştefan *et al.*, 1988, 1992): Phase I with volcanic character, and Phase II with intrusive character. The volcanic phase includes andesites, dacites and rhyolites, developed in time as follows: andesitic lavas at the beginning, followed by dacitic lavas. Rhyolites, sometimes with ignimbritic character, end the volcanic activity. The intrusive phase is represented by small bodies of diorite and quartz diorite as well as by granodiorite (+ granite) plutons (Budureasa, Drăganului Valley and Pietroasa). During the second phase of Ştefan *et al.* (1988, 1992), rhyolite, rhyodacite, aplite, microgranite, porphyrytic microgranite and micropegmatite dykes were emplaced along NW–SE faults. Towards the south, in the Băiţa Bihor area basic rocks (basalts, lamprophyres), probably originating from a deeper source, crosscut the main magmatic suite.

Ciobanu *et al.* (2002) published a complete compilation of age dating in the banatitic belt. The whole range of K-Ar ages spans the time from 110 to 50 Ma, showing disturbed Ar pattern. Relevant U-Pb ages (Nicolescu *et al.,* 1999; von Quadt *et al.*, 2002, 2003, 2005; Peytcheva and von Quadt, 2003 and Georgiev *et al.,* 2009), as well as Ar-Ar data (Handler *et al.,* 2004; Wiesinger, 2006) are restricted to 92-72 Ma. According to Ciobanu *et al.* (2002), the Palaeogene ages found for dykes with intermediate composition from the eastern Apuseni Mountains (Lemne *et al.,* 1983) and alkali mafic and lamprophyre dykes from Poiana Ruscă (Kräutner *et al.,* 1986; Downes *et al.,* 1995b) can be assigned to re-activation of magmatic activity caused by later tectonics, rather than to a final stage of the banatitic magmatism *s.s.* The dykes with ages of 61–39 Ma reported from Serbia (Jovanovic *et al.* 2001) and Bulgaria (45-42 Ma; Harkovska *et al.* 2001), may have also resulted from later tectonic events (Ciobanu *et al.*, 2002).

Zimmerman *et al.* (2008) published fifty-one new Re-Os ages measured on molybdenum in skarns and some granites and could confirm the narrow time span (up to 20 Ma), ranging in the Apuseni–Banat area from 72 to 83 Ma, in Timok from 81–88 Ma, and in the central district of Srednogorie zone (Panagyurishte) from 87 to 92 Ma. For banatites from Srednogorie Mts. Peytcheva *et al.* (2001), Peytcheva & von Quadt (2003) and Georgiev *et al.* (2009) presented ages from 75 to 84 Ma. They did not reveal any Paleogene ages and concluded that Paleogene ages reported from areas where the Re-Os dating was done, were not valid anymore.

Within the last decades a large database of geochemistry was generated by petrologists from Romania, Bulgaria and the former Yugoslavia, *e.g.* Istrate (1978) and Ştefan (1980) for the Apuseni Mts., Boccaletti *et al.* (1978) and Popov (1981) for Srednogorie, or Cioflica & Vlad (1973) and Russo-Săndulescu *et al.* (1984) for the Southern Carpathians. Only major elements and few trace elements were analyzed at that time, modern high-precision data, including isotopes analyses are only available since the 90s (*e.g.* Ştefan *et al.*, 1992; Downes *et al.,* 1995b; Karamata *et al.,* 1997; Dupont *et al.*, 2002).

Berza *et al.* (1998) reviewed the available data and identified four trends: a tholeiitic, a calc-alkaline, a high-K calc-alkaline, including shoshonites, and finally a peralkaline trend, found only in the eastern part of Srednogorie. They assigned the acidic intrusives to A-type granitoids, with an origin in the upper mantle or deeper crust. For the calc-alkaline intrusives, Berza *et al.* (1998) recorded three stages, more primitive: monzodioritic, dioritic and granodioritic evolution, found in the Southern Apuseni, South Banat and Timok, as well as in the central and western Srednogorie. The more evolved granodioritic to granitic trend is recorded in the Northern Apuseni, North Banat and Ridanj–Krepoljin. The alkaline trend is restricted to east and west Srednogorie and western Banat. The evolution of the magmas is mainly attributed to fractional crystallization (Dupont *et al.,* 2002). Modern trace elements and isotopic data confirm these results. The initial $^{87}Sr/^{86}Sr$ isotope ratios are in the range of 0.705 to 0.709 in the Apuseni Mts., 0.703 to 0.706 in the Banat, 0.706-0.710 in Timok and finally, 0.704 to 0.705 in Srednogorie. ΣNd values are reported in the range of +3.9 to –0.2 (Dupont *et al.*, 2002) and + 6.9 to –2.4 (Downes *et al.,* 1995b). Combined with the general present negative Nb and Ta anomaly, the data are consistent with the supra-subduction zone affinity of the magmas. The source of the magmas is supposed to be a heterogeneous mantle or a relatively young mafic lower crust. The heterogeneity of the mantle could be due to mantle metasomatism during an older subduction event. This pre-intrusion subduction could be early Cretaceous or older.

To explain the generation of banatites in terms of geodynamic setting, several models have been put forward in the last forty years: rifting (Popov, 1987; Popov *et al.,* 2003), subduction (Hsü *et al.* 1977; Russo-Săndulescu *et al.,* 1984; Cioflica, 1989; Vlad, 1997; Ciobanu *et al.*, 2002; Lips, 2002; Neubauer, 2002; Zimmermann *et al.*, 2008), or (post)collisional processes (Berza *et al.*, 1998; Nicolescu *et al.,* 1999; von Quadt *et al.*, 2005). The geometry of subduction and its direction is still a matter of debate, *i.e.* the subduction of the Vardar Ocean towards east, others have postulated or a westerly-directed subduction of the Severin Ocean. Another group proposes a consumption of the Vardar Ocean by Early Cretaceous and relates the magmatism to slab-breakoff and mantle delamination. Recent tectonic models are dominated by slab-rollback and slab-tear (Zimmermann *et al.,* 2008 and citations therein). The first model would cause extension in the upper crust giving the melts excess to high crustal levels. The second model predicts, in a post-subduction and post-collisional regime, the breakoff of the subducting slab from its continental counterpart, thus initiating astenospheric upwelling (Zimmermann *et al.*, 2008).

According to paleomagnetic data (Pătraşcu *et al.*, 1992, 1994; Panaiotu, 1998; Roşu *et al.*, 2004), the BMMB should have been straight east-west oriented before the 60° rotation during the Cenozoic deformation. This is consistent with a northward subduction of the Vardar Ocean. Steepening of the subduction zone forced the asthenosphere and mantle lithosphere to flow beneath the overriding plate, causing normal faulting and giving opportunity to the newly created melts to reach the surface. These processess would shift the melt generation towards the south, causing an observed younger age trend from the north to the south (Zimmermann *et al.,* 2008).

The banatitic magmatism is responsible for a widespread metallogeny. It comprises four major types: a) skarns and replacement mineralization found in the Apuseni Mts., Banat Mts., Ridanj–Krepoljin, Timok and Srednogorie; b) porphyry mineralization, common in the Southern Banat, Timok, central and eastern Srednogorie; c) massive sulphides, restricted to Timok and central Srednogorie and d) vein-stockwork mineralizations, in eastern Srednogorie, western Apuseni Mts. and Timok. The main metals are Cu, Au, Mo, Zn, Fe and subordinately Bi, U, Ti, Co and Ni. In the Northern Apuseni Mts. skarn bodies occur around Pietroasa and Budureasa intrusions. The most important mineralization is found in the Gilău–Bihor zone, with mineralized skarns associated with veins and impregnations (Băişoara and Băiţa-Bihor).

1.5 Neogene magmatism in the Apuseni Mountains

The latest magmatic event is the Cenozoic calc-alkaline to alkaline volcanism, also widely exposed at the inner Carpathian arc from SE Austria, along the Western Carpathians, to the Eastern Carpathians. The Neogene magmatism is an important feature of the Romanian orogen. The related volcanics can be found in the Eastern Carpathians (Oaş, Gutâi, Ţibleş, Rodna, Bârgău, Călimani, Gurghiu and Harghita Mts.) and in the Apuseni Mts. (Fig. 3). The following description is based mainly on the papers by Downes *et al.* (1995a), Pécskay *et al.* (1995, 2006, 2009), Roşu *et al.* (2004), Seghedi (2004) and Seghedi *et al.* (2004, 2005).

In the Eastern Carpathians, the magmatic event is related to the collision of the Eastern European Platform and the continental fragments of Tisia (Csontos, 1995). A very complex interplay of subduction, rollback, back arc extension, slab break-off and asthenosphere uprise is responsible for the generation of calc-alkaline and alkaline magmas (Seghedi *et al.*, 2005). In the Eastern Carpathians the magmatic activity covers, according to Pécskay *et al.* (1995), a large time span from Late Badenian[1] (14.3 Ma) in the Romanian north-western end of the chain (Oaş Mts.), till the Pleistocene (0.2 Ma) in the southeasternmost end (Harghita Mts.), respectively. Most of the products are volcanics (extrusives and subvolcanic bodies) but intrusives are found as well (Pécskay *et al.*, 2009).

In the Apuseni Mts., as part of the Tisia block, the development of the Neogene volcanism was closely related to an extensional tectonic stage (Roşu *et al.*, 2004), as a consequence of both the Neogene development of the Pannonian Basin (Fodor *et al.*, 1999) and the translational and rotational movements of the Tisia block (Pătraşcu *et al.,* 1994; Csontos, 1995). Similar to the Eastern Carpathians, the volcanic activity in the Apuseni Mountains covers a time span from Middle Miocene to Pleistocene.

In the Apuseni Mts., the Neogene calc-alkaline and alkaline magmatism mostly occurs in intramountain extensional basins and is found as NNE to SSW trending scattered bodies (Roşu *et al.*, 2004). These small basins *e.g.* Zarand, Brad and Zlatna are aligned from NW to SE and the volcanics are accompanied by Miocene sediments. In several areas, the Neogene magmatic activity is responsible for important metallogenetic processes, in particular hydrothermal Au, Ag and base metal mineralization. The famous Gold Quadrilateral including the gold deposits of Roşia Montană, Baia de Arieş, Săcărâmb and Brad, as well as the porphyry copper deposit of Roşia Poieni are associated with the Neogene volcanism (see also Benea & Tămaş, 2010).

The Neogene volcanics in the Apuseni Mts are mainly andesites, accompanied by rare basaltic andesites and dacites. Trachyandesites and trachydacites represent the youngest eruptive rocks generated around 1.6 Ma ago (*e.g.* Uroi Mountain near Deva). Petrographically, the volcanics are highly porphyritic and display phenocrysts of ortho- and clinopyroxene, amphibole, plagioclase and quartz, which are often corroded by the magma. Geochemically, most andesites display a normal calc-alkaline trend, but some show, an adakitic affinity ("adakite-like rocks" with high Al_2O_3 and Sr, and low Y and HREE – Roşu *et al.,* 2004).

[1] Chronostratigraphic assignment according to the Central Paratethys timescale (Vass & Balogh, 1987).

2. Field stops

Day 1

2.1 Field stop 1. The Urşilor Cave (the Bears' Cave), in Chişcău village

Location: *~80 km south-east of Oradea, on the west-facing slopes of the Apuseni Mts. (Fig. 13), near the village of Chişcău, Bihor County.*

Coordinates: *N 46°33.225' and E 22°34.169'; Elevation: 485 m.*

The cave formed in the Upper Jurassic (Tithonian) limestones of the Codru nappes system (Giuşcă *et al.*, 1967c; Bleahu *et al.,* 1985; see also Section 1.2), in the Northern Apuseni Mountains. It consists of large and narrow passages developed along two distinct levels extending over 1500 m (Figs. 14a,b). Only

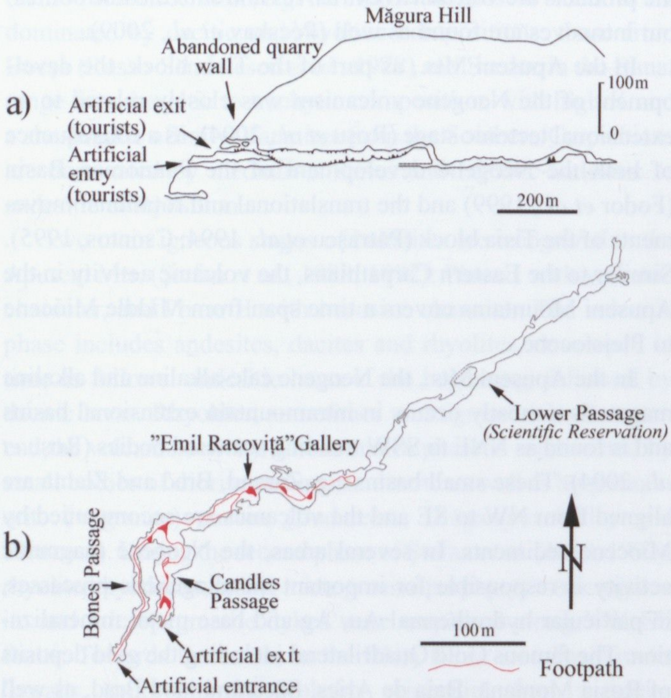

Fig. 14. Urşilor Cave (Bears' Cave), Chişcău: a) Cross-section (from Onac *et al.,* 2002), b) Schematic map showing the main rooms of the cave (based on Rusu, 1981, redrawn) and the route for visitors.

the lower one, carrying an underground stream, is still hydrologically active. It hosts most of the cave bear remains being preserved as a *Scientific Reservation* with a restricted public access (Rusu, 1981; Onac *et al.,* 2002). The upper level is not hydrologically active and contains various and spectacular stalactites, stalagmites, columns and draperies (Figs. 15a,b). Therefore, it is included in the show part of the cave (Rusu, 1981; Onac & Tămaş, 2010). The cave microclimate and the temperature are monitored: the mean annual temperature in the cave is 9.8 °C and the relative humidity varies from 95 to 100% (Racoviță *et al.,* 1999, 2003; Onac *et al.*, 2002).

The natural entrance of the Urşilor Cave was blocked by a rock fall. Subsequently, the cave remained closed and hidden until 1975, when it was accidentally discovered after blasting in a nearby marble quarry. An artificial entrance was opened and the cave was prepared for tourism. Since 1980 it became the most important show cave in Romania (Onac *et al.*, 2002), with more than 200,000 visitors/year. The cave preserves important palaeontological remains (mainly cave bear) and also a great variety of speleothems (Onac & Tămaş, 2010).

The touristic part of the cave includes three main sectors (the Bones Passage, the "Emil Racoviță" Gallery – Figs. 14b, 15a, and the Candles Passage - Figs. 14b, 15b) and four rooms (Candles Room, Spaghetti Room, the "Emil Racoviță" Room and the Bones Room. The tourist tour starts with the Bones' Gallery, where most of the cave bears skeletons and remains were found. Apart from the cave bear, which is the dominat species, over 40 species of mammals *e.g.* ibex, cave lion, and cave hyena have been identified (Terzea, 1971, 1978). In the lower passage an almost undisturbed skeleton of *Ursus spelaeus* was found. Numerous gestation holes and claw marks can be seen in the lower section of the cave.

A high number of cave bear skeletons (estimated from 140 to over 500 individuals) were found. The cave bears from Europe, *i.e. Ursus spelaeus* Rosenmüller & Hainroth, were larger than the recent brown bears from the Carpathians. They were among the most massive mammals during the Ice Age (Christiansen, 1999) and used the caves for living during the winter, for giving birth to the cubs and to die. It is estimated that an *Ursus spelaeus* female had an average body mass of around 180-230 kg, but might have reached up to 350 kg, whereas a male weighed between 350 and 630 kg. The cave bear was similar in size to the polar bear, and could reach almost 3 m at the shoulder level, in upright position. Their life expectance was about 20 years. They probably became extinct in the Pleistocene, approximately 28,000-27,000 BP but survived longer in Southern and South-Eastern Europe (Pacher & Stuart, 2008).

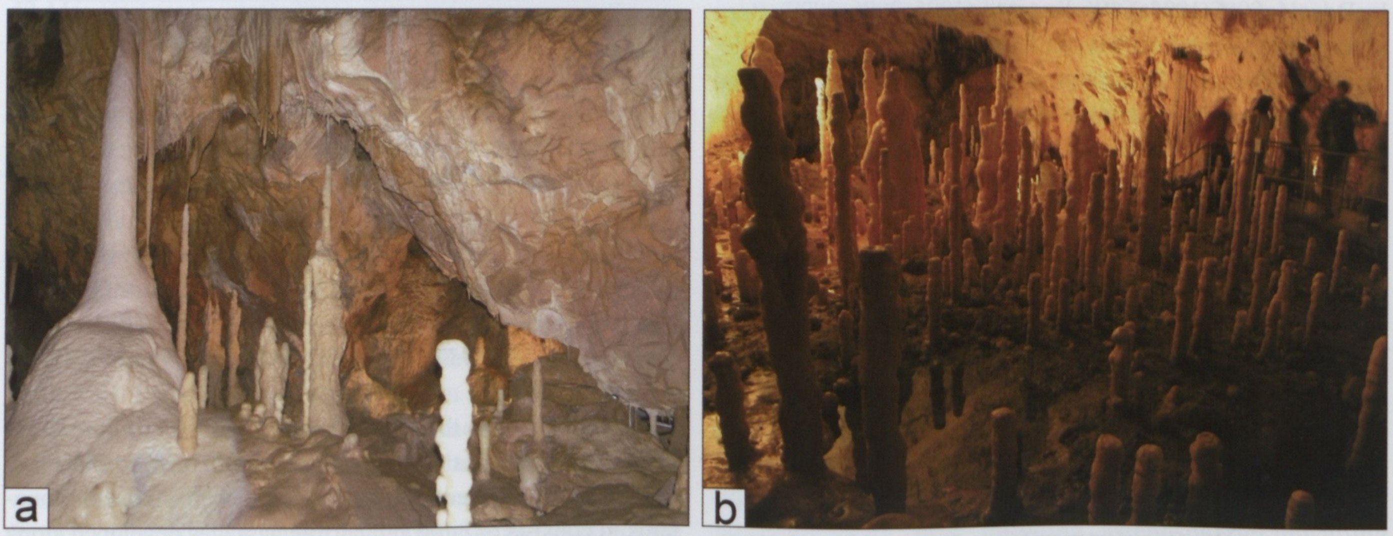

Fig. 15. Urşilor Cave (Bears' Cave), Chişcău: a) "Emil Racoviță" Gallery; b) Candles Passage (Photos by B.P. Onac).

2.2 Field stop 2. Pietroasa Quarry: Upper Cretaceous granodiorites (banatites)

Location: *about 1 km west of the Pietroasa village (Figs. 13 and 16), at the southern slope of the Crişul Pietros valley.*
Coordinates: *N 46°35.022' and E 22°33.871'; Elevation: 385 m.*

The Pietroasa and the Budureasa plutons (Fig. 13) are the largest and the most important intrusives in a large banatitic sequence extending in the Northern Apuseni Mts. (see also Section 1.4) between the Crişul Repede and the Crişul Negru rivers. According to Ştefan *et al.* (1988, 1992), they belong to the second, mainly intrusive phase of the Late Cretaceous igneous activity in the Northern Apuseni Mts. The first cycle would consist mainly of volcanics including andesites, dacites and rhyolites. Age dating (Lemne *et al.*, 1983; Bleahu *et al.*, 1984; Ciobanu *et al.*, 2002) shows more complex relations between volcanics and intrusives on a regional scale (Apuseni–Banat). The Pietroasa and Budureasa plutons intruded within a time range of 70-74 Ma (Bleahu *et al.*, 1984). Late dacite and rhyodacite dykes, sometimes up to 15 km in length, crosscut the intrusions.

Petrographically, the intrusive banatitic rocks in the Budureasa and Pietroasa area (Fig. 17a) range from granites to quartz diorites, with a prevalence of monzogranites and granodiorites (Istrate & Udubaşa, 1981; Ionescu, 1996a; Ionescu & Har, 2001). Quartz monzonites, quartz monzodiorites and diorites occur as well (Istrate & Udubaşa, 1981). Granodiorites are hypidiomorphic-granular, whereas porphyritic textures are restricted to the marginal facies. Granodiorites have a greyish to yellowish or pinkish colour and consist of K-feldspar, quartz, plagioclase (mostly oligoclase or oligoclase/andesine), biotite, magnesiohornblende and rare clinopyroxene. Titanite, zircon, apatite and allanite-(Ce) occur as accessory minerals.

Xenoliths (Fig. 17b) of magmatic, metamorphic and sedimentary origin along the margins of the intrusions are typical features (Stoicovici & Selegean, 1970; Istrate & Udubaşa, 1981; Ştefan *et al.*, 1988; Ionescu, 1996a; Mârza & Ionescu, 1998). The xenoliths occur over large areas and have various sizes,

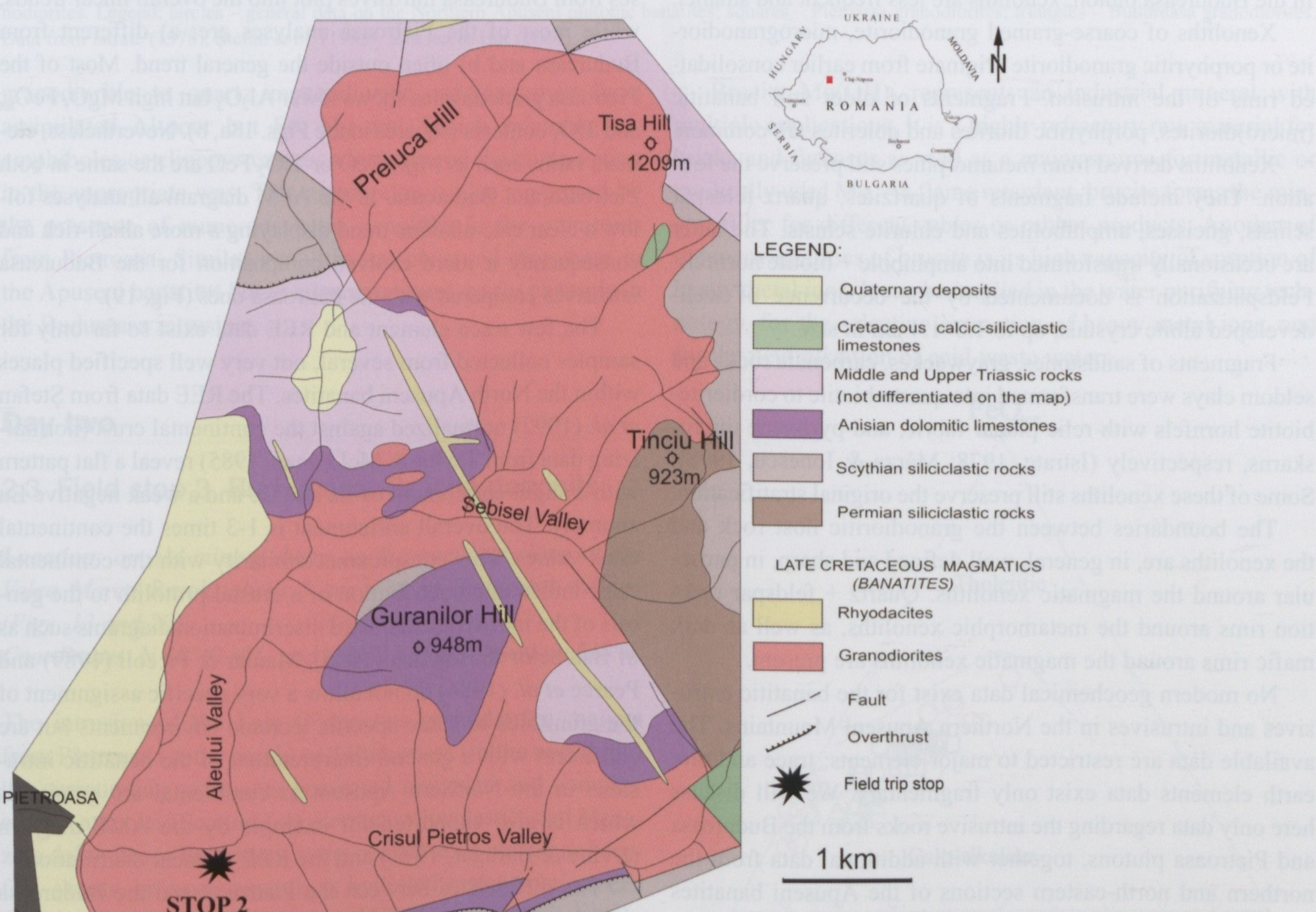

Fig. 16. Simplified geological map of Pietroasa area (from Ionescu & Balaban, 1998; based on Popa *et al.*, 1984 – unpublished report in the archives of Transgex SA, Cluj-Napoca, Bleahu *et al.*, 1985 and the field data of the authors), with the field stop 3. The insert in the upper right shows the position of the map within Romania.

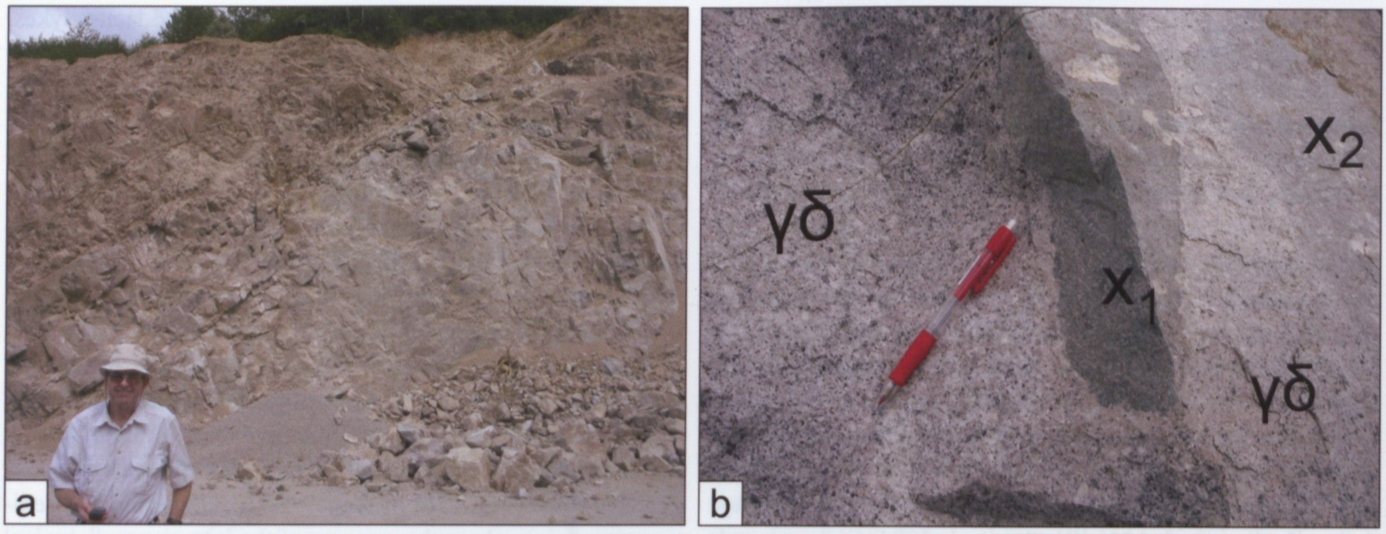

Fig. 17. Pietroasa Quarry: a) General view; b) Various volcanic xenoliths (x_1, x_2) in granodiorite (γδ).

ranging from 1-2 cm to tens of meters (Stoicovici & Selegean, 1970). Their maximum size and number is recorded in the Pietroasa intrusion, along the Crişul Pietros valley (Figs. 17 a,b). In the Budureasa pluton, xenoliths are less frequent and smaller.

Xenoliths of coarse-grained granodiorite, microgranodiorite or porphyritic granodiorite originate from earlier consolidated rims of the intrusion. Fragments of older still banatitic (micro)diorites, porphyritic diorites and dolerites are common.

Xenoliths derived from metamorphics still preserve the foliation. They include fragments of quartzites, quartz-feldspar schists, gneisses, amphibolites and chlorite schists. The latter are occasionally transformed into amphibole + biotite hornfels. Feldspatization is documented by the occurrence of well-developed albite crystals, up to 1.0–1.4 cm in size.

Fragments of sandstones, graywackes, carbonate rocks and seldom clays were transformed into quartz-biotite to cordierite-biotite hornfels with relic planar fabric, and pyroxene micro-skarns, respectively (Istrate, 1978; Mârza & Ionescu, 1998). Some of these xenoliths still preserve the original stratification.

The boundaries between the granodioritic host rock and the xenoliths are, in general, well defined and sharp, in particular around the magmatic xenoliths. Quartz + feldspar reaction rims around the metamorphic xenoliths, as well as dark mafic rims around the magmatic xenoliths are present.

No modern geochemical data exist for the banatitic extrusives and intrusives in the Northern Apuseni Mountains. The available data are restricted to major elements; trace and rare earth elements data exist only fragmentary. We will discuss here only data regarding the intrusive rocks from the Budureasa and Pietroasa plutons, together with additional data from the northern and north-eastern sections of the Apuseni banatites (Istrate, 1978; Ştefan *et al.*, 1992; Ionescu & Har, 2001).

The combined analyses form linear arrays in variation diagrams of the major elements oxides *vs.* SiO_2 and MgO, respectively (Figs. 18a,b). Most oxides correlate negatively with SiO_2 and positively with MgO, except for K_2O wich increases with increasing SiO_2 and decreases consequently with increasing MgO. These well-defined linear trends suggest a continuous evolution from dioritic to granitic rocks due to fractional crystallization mainly of Fe-Mg silicates and plagioclase. The analyses from Budureasa intrusives plot into the overall linear trends, while most of the Pietroasa analyses are: a) different from Budureasa and b) often outside the general trend. Most of the Pietroasa granodiorites shows lower Al_2O_3 but high MgO, FeO_{tot} and TiO_2 contents (squares in the Figs. 18a, b). Nevertheless, element ratios such as MgO/FeO or TiO_2/FeO are the same in both Pietroasa and Budureasa. In the AFM diagram all analyses follow a clear calc-alkaline trend displaying a more alkali-rich and consequently a more evolved composition for the Budureasa intrusives compared with the Pietroasa ones (Fig. 19).

The few trace element and REE data exist so far only for samples collected from several, not very well specified places within the North Apuseni banatites. The REE data from Ştefan *et al.* (1992) normalized against the continental crust (normalizing data from Taylor & McLennan, 1985) reveal a flat pattern with a slight enrichment of La and Ce and a weak negative Eu anomaly. The overall enrichment is 1-3 times the continental crust values. This conspicuous similarity with the continental crust indicates a contribution of a crustal protolith to the genesis of the intrusives. Standard discrimination diagrams such as of Batchelor & Bowden (1985), Maniar & Piccoli (1989) and Pearce *et al.* (1984) do not allow a very specific assignment of the granitoids to some specific tectonic environments but are consistent with a general interpretation of the banatitic intrusions in the Northern Apuseni as continental arc granitoids which is also supported for example by the AMF diagram (Irvine & Baragar, 1971) and the REE element distribution.

The differences between the Pietroasa and the Budureasa intrusives are complex. The more SiO_2-poor composition would suggest an earlier crystallization of the former than the latter within the overall evolution trend. This is not consistent with the relative enrichment of Fe, Mg, Ti and the depletion of Al in the Pietroasa rocks (Figs. 18a,b). This requires, compared with the overall trend, additional processes. Possibly

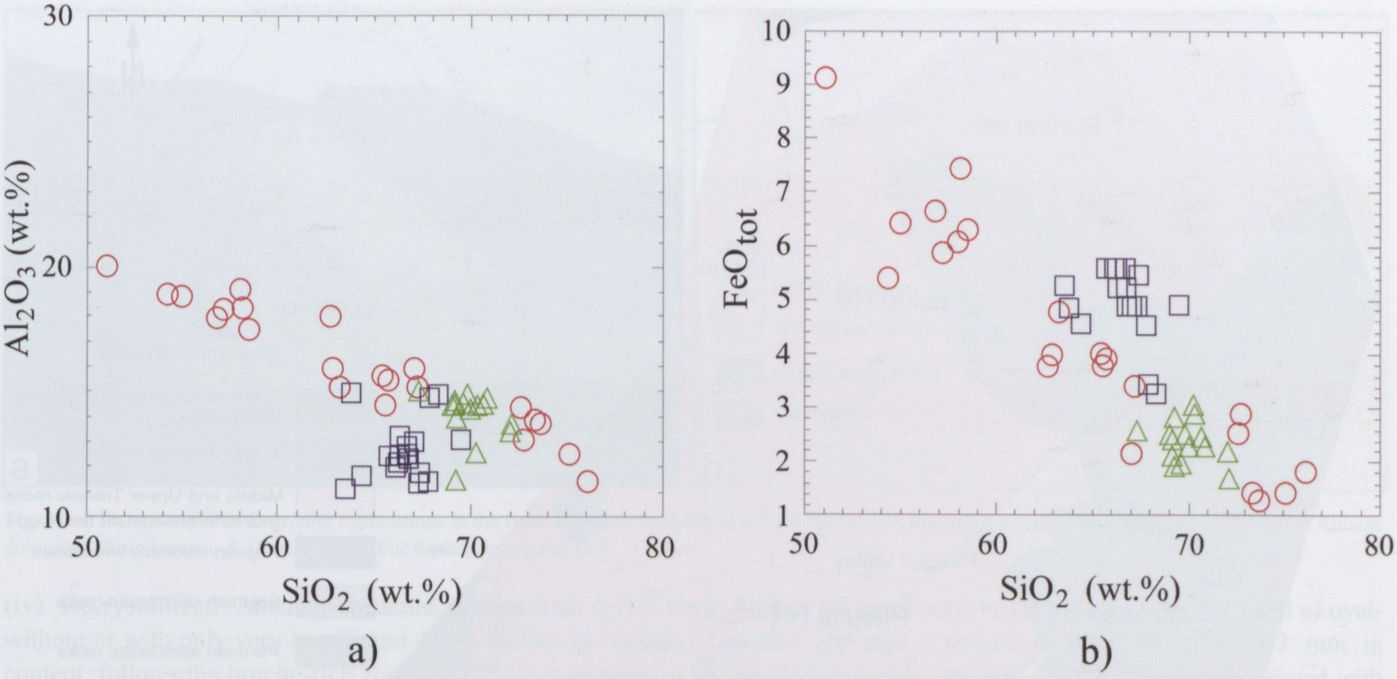

Fig. 18. The (a) SiO_2 *vs.* Al_2O_3 and (b) SiO_2 *vs.* FeO_{tot} diagrams for plutonic banatites from the Northern Apuseni Mts., including Pietroasa and Budureasa granodiorites. Legend: circles – general data on the Northern Apuseni plutonic banatites; squares – Pietroasa granodiorites; triangles – Budureasa granodiorites. Data from Istrate (1978), Ştefan *et al*. (1992), Ionescu & Har (2001).

granodioritic to quartz monzodioritic magmas may have assimilated Al-poor but Fe, Mg and Ti-rich minerals, *i.e.* amphiboles or clinopyroxenes, which altered the composition in the appropriate way. This assumption is also supported by the presence of numerous basic xenoliths in the granitoids from Pietroasa. Similar xenoliths are observed elsewhere in the Apuseni banatites but in a lower amount, as for example in the Budureasa intrusion.

Day two

2.3 Field stop 3. Budureasa: Brucite deposits

***Location**: an old mining dump on the northern slope of the Valea Marea Brook, about 5 km west of Budureasa village (Figs. 13 and 20).*

***Coordinates**: N 46°40.395' and E 22°33.361'; Elevation: 510 m.*

The intrusion of the Late Cretaceous granodioritic magma from Pietroasa (in the south) and Budureasa (in the north) into the surrounding sediments generated extended and complex contact aureoles (Section 1.4). Permian siliciclastics and Mesozoic dolomites to limestones at the contact zone appear as hornfels, skarns, or more general, as hydrated metasomatic rocks (Rafalet, 1963; Istrate & Udubaşa, 1981; Ionescu, 1987, 1996b, 1999; Ştefan *et al*., 1988; Marincea, 1993). Within the Anisian dolomitic and calcareous protoliths large brucite-bearing zones occur in the contact aureoles of the granodioritic intrusions. They form two main deposits: Pietroasa in the south (Fig. 16) and Budureasa in the north (Fig. 20).

Brucite $Mg(OH)_2$ represents an industrial mineral with multiple applications. It is a highly refractory raw material for bricks and furnaces as well as a major source for metallic or medically used Mg. As flame retardant, brucite forms the mineral filler for different cables or rubber products. Another of the many qualities of brucite is its high capacity of sorption of heavy metal ions. It can be implied in the water purifying technology, for the selective extraction of heavy metal ions, and for the neutralization of acid waste waters.

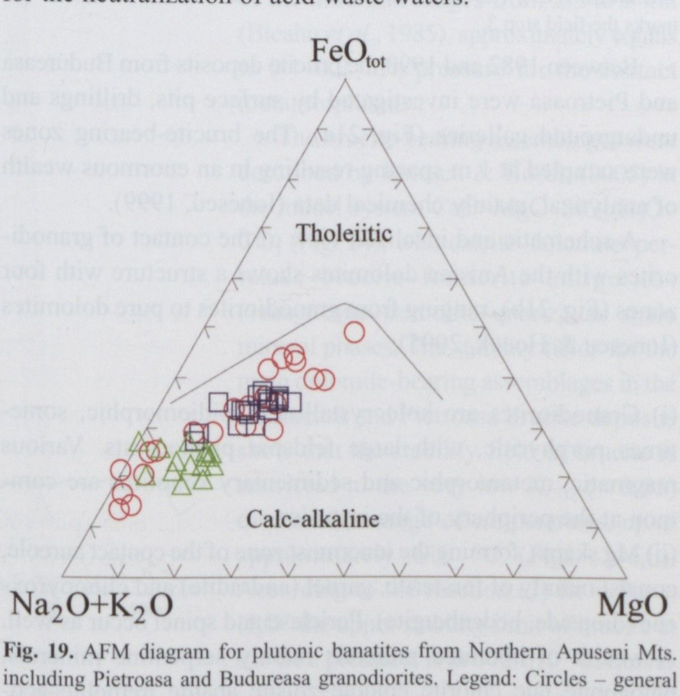

Fig. 19. AFM diagram for plutonic banatites from Northern Apuseni Mts. including Pietroasa and Budureasa granodiorites. Legend: Circles – general data on the Northern Apuseni plutonic banatites, squares – Pietroasa granodiorites, triangles – Budureasa granodiorites. Data from Istrate (1978), Ştefan *et al*. (1992), Ionescu & Har (2001).

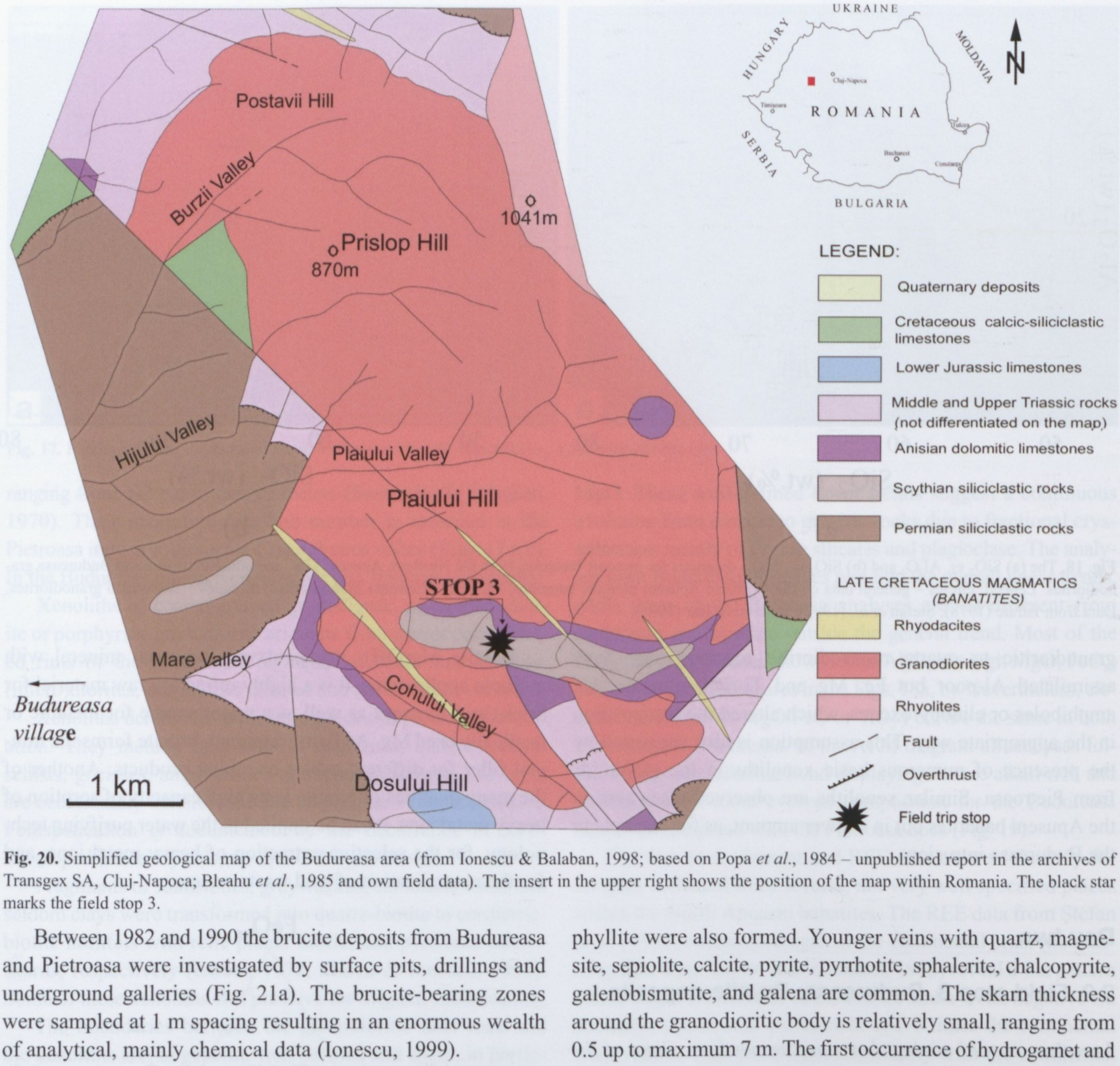

Fig. 20. Simplified geological map of the Budureasa area (from Ionescu & Balaban, 1998; based on Popa *et al.*, 1984 – unpublished report in the archives of Transgex SA, Cluj-Napoca; Bleahu *et al.*, 1985 and own field data). The insert in the upper right shows the position of the map within Romania. The black star marks the field stop 3.

Between 1982 and 1990 the brucite deposits from Budureasa and Pietroasa were investigated by surface pits, drillings and underground galleries (Fig. 21a). The brucite-bearing zones were sampled at 1 m spacing resulting in an enormous wealth of analytical, mainly chemical data (Ionescu, 1999).

A schematic and idealized view of the contact of granodiorites with the Anisian dolomites shows a structure with four zones (Fig. 21b), ranging from granodiorites to pure dolomites (Ionescu & Hoeck, 2005):

(i) Granodiorites are holocrystalline hypidiomorphic, sometimes porphyritic, with large feldspar phenocrysts. Various magmatic, metamorphic and sedimentary xenoliths are common at the periphery of the intrusion.

(ii) Mg skarns, forming the innermost zone of the contact aureole, consist mainly of forsterite, garnet (andradite) and clinopyroxene (diopside, hedenbergite). Periclase and spinel occur as well. A wealth of hydrated minerals mainly serpentine minerals, phlogopite, talc, chlorite, epidote–zoisite, apatite, tremolite–actinolite and subordinately hydrogarnet, vesuvianite, chondrodite, clinohumite, hydrotalcite, brucite, hydromagnesite and pyrophyllite were also formed. Younger veins with quartz, magnesite, sepiolite, calcite, pyrite, pyrrhotite, sphalerite, chalcopyrite, galenobismutite, and galena are common. The skarn thickness around the granodioritic body is relatively small, ranging from 0.5 up to maximum 7 m. The first occurrence of hydrogarnet and magnesioferrite in Romania was described from the Budureasa area by Ghergari & Ionescu (2000).

(iii) Brucite-bearing zones occur only at some distances (0.5 to 7 m) from the contact (Figs. 22a,b,c). The irregular, sometimes lens-shaped brucite-bearing zones range from several metres up to tens of metres in width and from tens to several hundreds of metres in length, respectively. The thickness of the brucite-bearing zone can significantly vary within the short distance. The variation of brucite content across the contact aureola around the granodioritic intrusion is highly inhomogeneous, ranging from brucite-rich, with up to 40 wt% brucite to brucite-poor domains, with less than 5 wt% brucite (Figs. 22a,b,c). The average content of brucite is around 10.5 wt% in the Budureasa area and 7.5 wt% in the Pietroasa area, respectively (Ionescu, 1999).

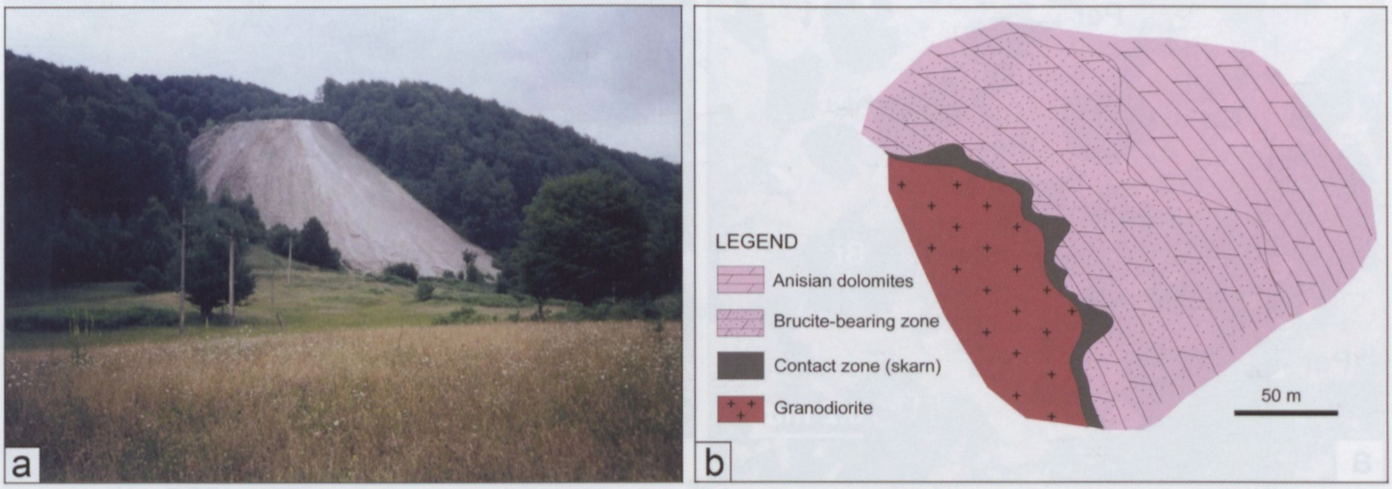

Fig. 21. a) Dumps of the underground explorations in the Budureasa brucite deposit; b) Idealized sketch through the contact between granodiorites and Anisian dolomites (from Ionescu & Hoeck, 2005). For further details see text.

(iv) Recrystallized Anisian dolomite, without or with only very low Si and Al content, follows the brucite-rich zones. Brucite forms small lamellae of 20×20×2 µm up to 80×50×6 µm (length, width, thickness). Large individual lamellae, over 1 mm in length are only exceptionally found. Brucite lamellae group in clusters of various shapes and sizes (in average from 0.05 to 1.3 mm). Fillings of small veinlets or isolated lamellae are rare. Three main types of brucite clusters could be identified: a) small, isometric clusters, about 0.05 mm in diameter, rarely containing relics of periclase (Fig. 23 a); b) large, irregular or rhombohedral-shaped clusters, often containing carbonate relics; the diameter of these clusters ranges from 0.5 up to 1.6-1.8 mm (Fig. 23b) and c) oval-shaped clusters, about 0.1×0.3 mm in average, showing brucite associated with forsterite relics and serpentine minerals.

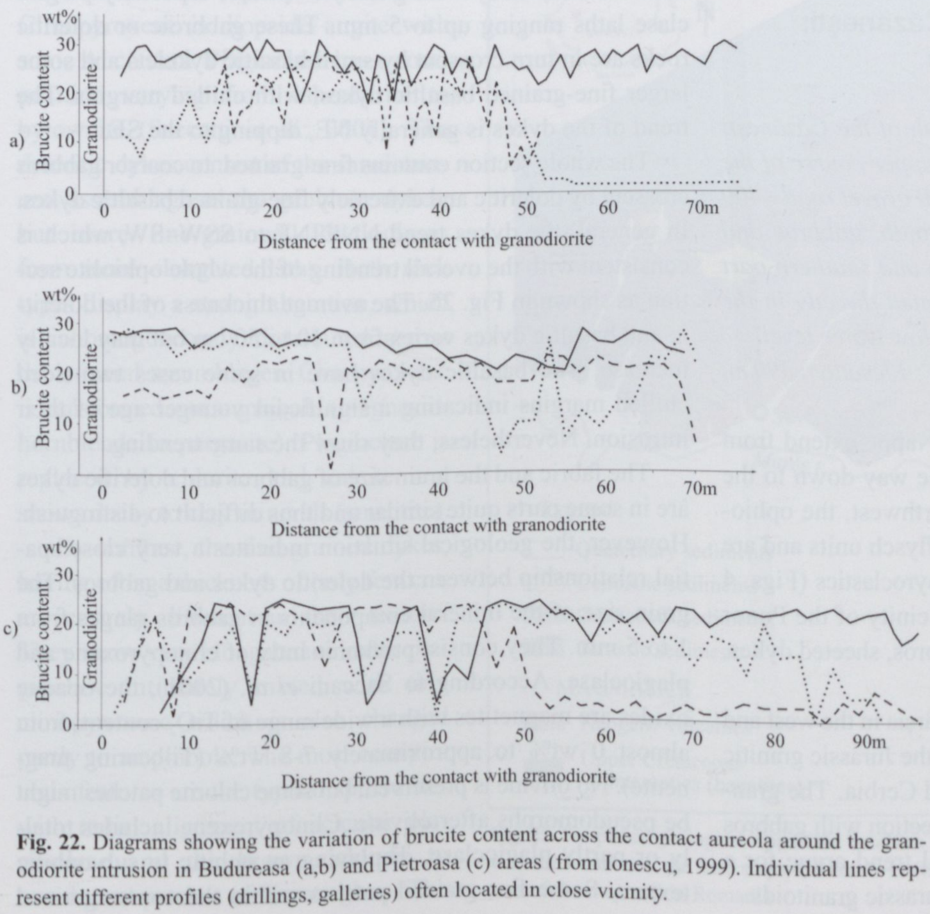

Fig. 22. Diagrams showing the variation of brucite content across the contact aureola around the granodiorite intrusion in Budureasa (a,b) and Pietroasa (c) areas (from Ionescu, 1999). Individual lines represent different profiles (drillings, galleries) often located in close vicinity.

The studies carried out on the mineralogy, petrology and genesis of the brucite deposits reveal a model of heating and cooling sequence under conditions of very low X_{CO2} during the contact metamorphism (Ionescu & Hoeck, 2005). The pressure estimates for the heating and cooling paths can be based on field relations. The stratigraphic column of sediments covering granodiorites at the time of the intrusion ranges from 2.5 to 4 km (Bleahu *et al.*, 1985), approximately equals to 0.1–0.2 GPa pressure for the contact metamorphism.

The brucite-bearing assemblages were described by Ionescu & Hoeck (2005) in the model system $CaO–MgO–SiO_2–H_2O–CO_2$ (Fig. 24), with calcite–dolomite–periclase–brucite–forsterite–antigorite–(+magnesite–tremolite–quartz) as main mineral phases. The stability fields for the main dolomite-bearing assemblages in the Budureasa and Pietroasa brucite deposits show that the stability field of brucite is restricted to the very low X_{CO2} (< 0.05) over a wide range of temperatures, up to approximately 610 °C (Figs. 24a,b). According to the reaction (1) Br = Per + H_2O* the upper stability limit of brucite at 0.1 GPa pressure, is at 600-610 °C. Lower

* Mineral names abbreviation according to Kretz (1983).

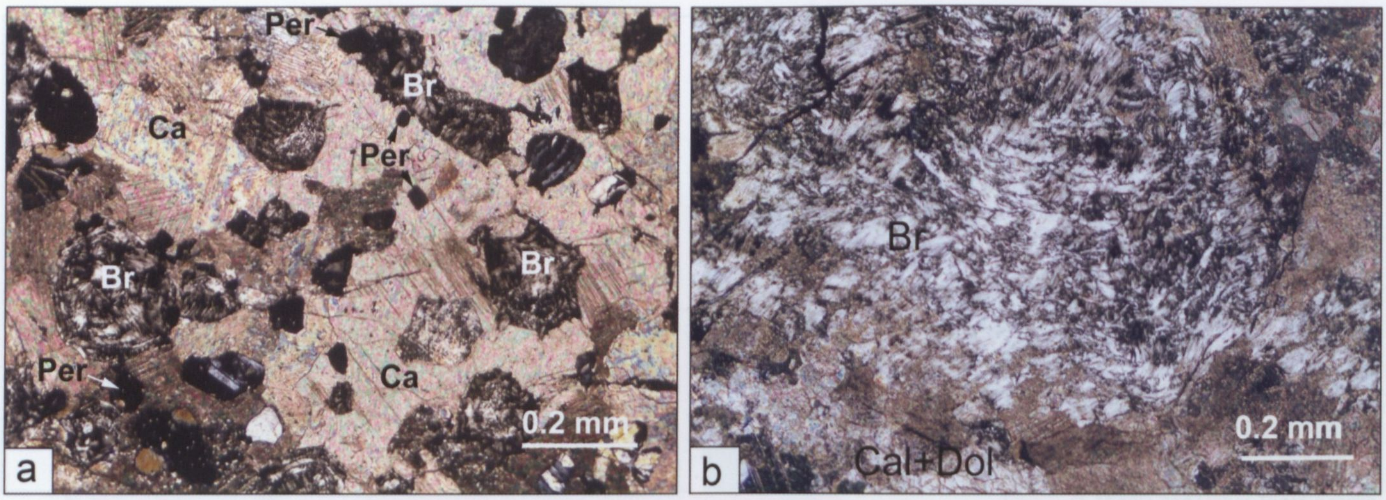

Fig. 23. a) Brucite clusters (Br) formed on the expense of periclase (Per) in a carbonate (Ca) matrix; b) Large and irregular-shaped brucite (Br) cluster, with internal onion-skin texture, in a carbonate matrix (Cal + Dol) (from Ionescu & Hoeck, 2005). Budureasa deposit. Crossed polarizers.

temperatures, below 400 °C, can be estimated from the reaction (4) 20Br + Atg = 34Fo + 51H_2O. The reaction (10) Dol + H_2O = Br + Cc + CO_2 can take place over a wide range of temperatures.

Based on field observations and theoretical considerations it is concluded that brucite formed by three main processes: a) a prograde reaction, which generated periclase in the close vicinity of the intrusion, followed by a retrograde reaction forming brucite in the presence of a vapour phase; b) directly from dolomite at lower temperatures and c) by decomposition of forsterite.

2.4 Field stop 4. Ponor valley at Căzăneşti: Jurassic gabbro–dyke section

***Location**: ~12 km south-west of Vaţa, south of the Căzăneşti village, close to a western affluent on the upper course of the Ponor valley (Fig. 25). At a bend of a small gravel road ~400 m upstream the Ponor valley from the mouth, gabbros and various dykes are exposed on the northern and southern part of the road. The southern exposure is located directly in the stream, being accessible only at relatively low water level.*

***Coordinates**: N 46°07.387' and E 22°30.159'; Elevation: 390 m.*

The ophiolites of the Căpâlnaş–Techereu Nappe extend from the entry of the Ponor valley at Vaţa all the way down to the Mureş River in the south. Towards the northwest, the ophiolites are thrust over the Lower Cretaceous flysch units and are in turn overlain by Neogene lava flows or pyroclastics (Figs. 4 and 25). In the north-eastern part, in the vicinity of the Ponor valley, the ophiolites consist mainly of gabbros, sheeted dykes, massive lava flows and pillow lavas.

The small granitic intrusion between Obârşia in the west and Căzăneşti in the east is most likely a part of the Jurassic granitic intrusions, such as those from Săvârşin and Cerbia. The granitoids are not dated yet, but their close connection with gabbros and their elongation parallel to the general trend argue for a Jurassic age. Both features are typical for Jurassic granitoids.

The general strike in the western part of the ophiolites runs from SW to NE with a certain variation. The sequence from plutonic to volcanic rocks, *i.e.* from gabbros and dykes to lava flows is several times repeated. This is best recognisable in the distribution of gabbros from W to E (Fig. 25): near Juliţa, around the Săvârşin granitoids and in the vicinity of Cerbia gabbros.

The outcrops of the gabbro–dyke section in the stream of the upper Ponor valley at the southern side of the road can only be studied when the water and the weather allow it. Few metres north of the stream, close to the road is an outcrop of a dark intrusive body with gabbroic texture, especially plagioclase laths ranging up to 5 mm. These gabbroic or doleritic rocks are in turn crosscut by small basaltic dykelets and some larger fine-grained basaltic dykes, with chilled margins. The trend of the dykes is generally NE, dipping to the SE.

The whole section contains fine-grained to coarser gabbros crosscut by doleritic and extremely fine-grained basaltic dykes. In general, the dykes trend NNE–NE to SSW–SW, which is consistent with the overall trending of the whole ophiolite section as shown in Fig. 25. The average thickness of the doleritic and basaltic dykes varies from 10 to 50 cm but may locally increase. The basaltic dykes have in some cases two-sided chilled margins indicating a significant younger age of their intrusion. Nevertheless, they show the same trending.

The fabric and the grain size of gabbros and doleritic dykes are in some parts quite similar and thus difficult to distinguish. However, the geological situation indicates a very close spatial relationship between the doleritic dykes and gabbros. The grain size of the mineral components in gabbros ranges from 1 to 5 mm. They consist predominantly of clinopyroxene and plagioclase. According to Saccani *et al.* (2001), the opaque oxides are magnetites with a wide range of TiO_2 content, from almost 0 wt% to approximately 7-8 wt% (Ti-bearing magnetite). No olivine is preserved, but some chlorite patches might be pseudomorphs after olivine. Clinopyroxene includes totally or partly plagioclase, displaying an ophitic to sub-ophitic texture. Some fine-grained patches within the coarse-grained

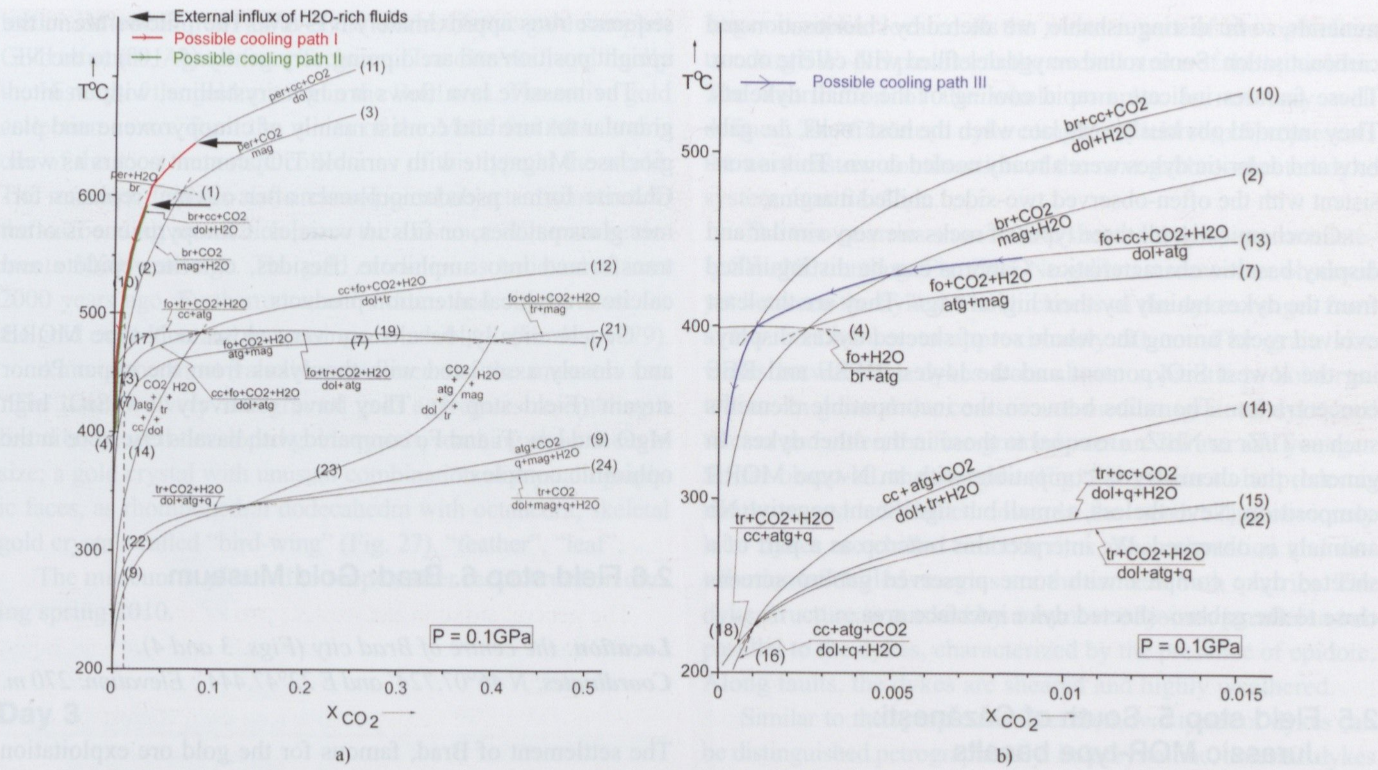

Fig. 24. Phase diagrams for brucite calculated with Thermocalc by Powell & Holland (2001) for a) 200–700 °C and 0–0.5 X_{CO_2} and b) detail for 200–550 °C and 0–0.015 X_{CO_2}. The diagrams show possible cooling paths and genesis of brucite: from periclase (reaction 1), dolomite (reaction 10) and forsterite (reactions 4 and 7). Diagrams from Ionescu & Hoeck (2005). Mineral names abbreviation according to Kretz (1983).

gabbro could be interpreted as former melt inclusions, rapidly crystallized. Clinopyroxene is diopside–augite with some Al content. Plagioclase has a composition varying from labradorite to bytownite (Saccani *et al.*, 2001). The doleritic dykes consist of the same mineral assemblage as the gabbroic dykes, but more fine-grained. Plagioclases form more elongated laths. Both rock types display a strong alteration, which is inhomogeneously distributed across the mineral assemblages. Clinopyroxene is transformed into amphibole (magnesiohornblende to pargasite). Plagioclase is replaced by chlorite and calcite and shows decay to clay minerals (smectite).

By contrast, the fine-grained dykes, having often a thickness of only few cm, are entirely different in their texture. They show very elongated acicular plagioclase intimately mixed with tiny clinopyroxene which is also changed partly to amphibole. The most conspicuous features are the typical spherulites with radial alignment of acicular plagioclase and the brown, cloudy to translucent patches of former glass. The

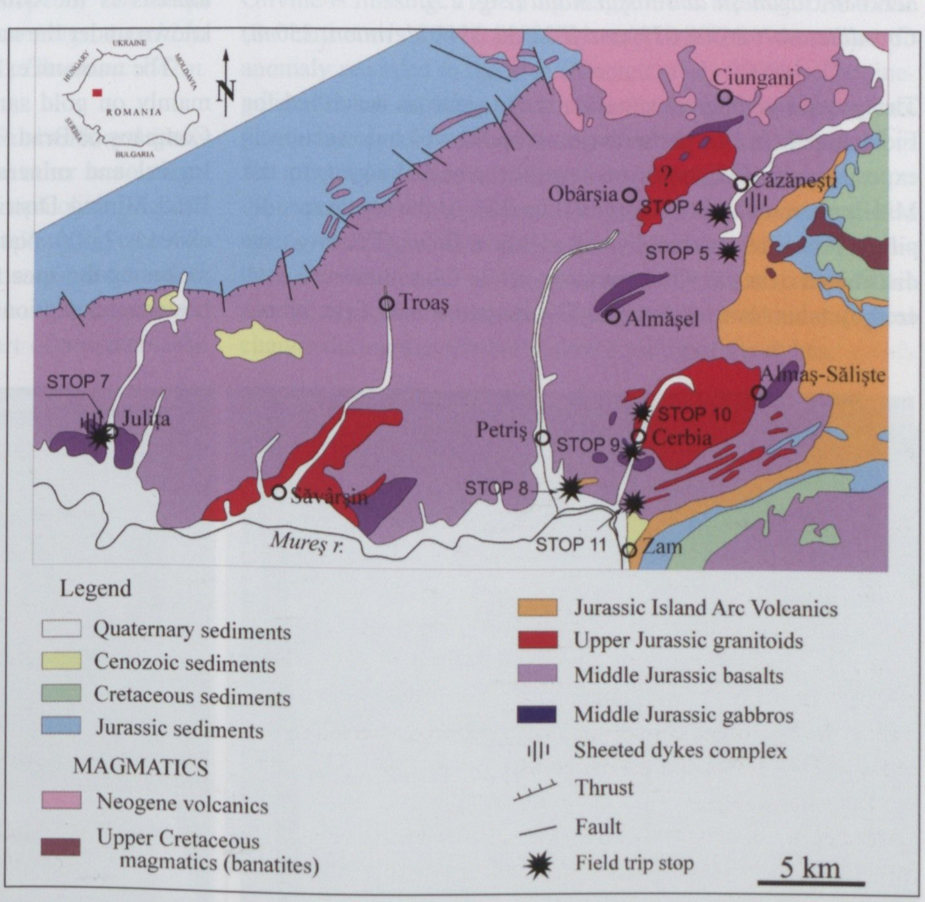

Fig. 25. Simplified geological map of the Southern Apuseni ophiolites (according to Giuşcă *et al.*, 1967c), with the location of the field stops 4, 5 and 7–11 (black stars). The insert in the upper left shows the position of the map within Romania.

minerals, so far distinguishable, are altered by chloritisation and carbonatisation. Some round amygdales filled with calcite occur. These features indicate a rapid cooling of the small dykelets. They intruded obviously very late when the host rocks, *i.e.* gabbros and doleritic dykes were already cooled down. This is consistent with the often-observed two-sided chilled margins.

Geochemically, all three types of rocks are very similar and display basaltic characteristics. Gabbros can be distinguished from the dykes mainly by their higher Mg#. They are the least evolved rocks among the whole set of sheeted dykes displaying the lowest SiO_2 content and the lowest HFSE and REE concentration. The ratios between the incompatible elements such as Ti/Zr or Nb/Zr are equal to those in the other dykes. In general, the chemistry is compatible with an N-type MORB composition. Nevertheless, a small but significant negative Nb anomaly is observed. We interpret this outcrop as a part of a sheeted dyke complex with some preserved gabbro screens close to the gabbro–sheeted dyke interface area.

2.5 Field stop 5. South of Căzăneşti: Jurassic MOR-type basalts

Location: *an abandoned small quarry ~9 km west of Vaţa village, on the western slopes of the Ponor valley, on the road between Căzăneşti and Roşia Nouă (Fig. 25).*
Coordinates: *N 46°08.073' and E 22°30.264'; elevation: 330 m.*

The general geological situation is the same as described for Field stop 4. In this particular outcrop the volcanic section is exposed, with two types of basaltic lavas belonging to the Mid-Jurassic ophiolite complex (Fig. 26). At the southern side, pillow lavas are overlain by massive lava flows. The lavas are distributed over the whole outcrop, while the pillows are covered by talus towards the NE. The overall trend of the whole sequence runs approximately NE–SW. The pillows are in the upright position and are dipping very gently (5–10°) to the NE.

The massive lava flows are holocrystalline, with an intergranular texture and consist mainly of clinopyroxene and plagioclase. Magnetite with variable TiO_2 content occurs as well. Chlorite forms pseudomorphoses after olivine, replaces former glass patches, or fills in vesicles. Clinopyroxene is often transformed into amphibole. Besides, chlorite, epidote and calcite are typical alteration products.

Geochemically, basalts are very close to N-type MORB and closely associated with the dykes from the upper Ponor stream (Field stop 4). They have relatively low SiO_2, high MgO and low Ti and Fe compared with basalts elsewere in the ophiolitic complex.

2.6 Field stop 6. Brad: Gold Museum

Location*: the centre of Brad city (Figs. 3 and 4).*
Coordinates*: N 46°07.724' and E 22°47.444'; Elevation: 270 m.*

The settlement of Brad, famous for the gold ore exploitation in the surroundings, was mentioned in documents for the first time in 1445. It was one of the main centers of Revolution of 1848-1849 in Transylvania. The most attractive objective for tourists is the Mineral collection in the Museum of Brad known under the name of "Gold Museum".

The museum exhibits a large mineralogical collection based mainly on gold samples (Fig. 27). It belongs to the Mining Company of Brad. The museum started around 1896, as a geological and mineralogical collection, representative for the Brad Mining District. Nowadays the number of the samples exceeds 2,500. Native gold samples as well as gold minerals makes up the specific character of this museum. The museum has several sections (Verdeş, 1979): Systematic mineral col-

Fig. 26. Outcrop of MOR-type basaltic pillow lavas (Ponor stream, near Căzăneşti).

Fig. 27. Dendritic gold called the "bird-wing" (length of the sample: 6 cm) from the Gold Museum, Brad (from Ionescu *et al.*, 2009a; Photo F. Forray).

lection, Minerals from all over the world, Native gold samples, Gold minerals, Agates from Techereu (Trascău Mts.) and finally the History of the gold mining in the Brad area. The native gold collection comes from the mines of the Metaliferi Mts. and is one of the most important collections of this type in the world. The museum displays also archaeological objects discovered in the surrounding area which prove the existence of human settlements 5000 years ago. The gold extraction is believed to start 2000 years ago. Furthermore, old utilities and objects used for the gold extraction and processing are exhibited (Verdeş, 1979).

Among the most valuable specimens of the museum are: "The lizard" – very fine-grained gold aggregates resembling a lizard body; well-developed gold crystals, reaching about 1 cm in size; a gold crystal with unusual combination of crystallographic faces, as rhombohedral dodecahedra with octahedra; skeletal gold crystals called "bird-wing" (Fig. 27), "feather", "leaf".

The museum is planned to reopen after reconstruction during spring 2010.

Day 3

2.7 Field stop 7. Julița Quarry: Jurassic sheeted dykes

***Location**: two quarries on the western slope of the Julița stream – a northern affluent of the Mureş River (Figs. 4 and 25).*
***Coordinates**: N 46°03.611' and E 22°08.344'; Elevation: 190 m.*

A sheeted dyke complex is exposed in two abandoned quarries, one marked as northern quarry and the other as southern quarry. They are situated in the SW part of the ophiolites (Fig. 25) in the close vicinity of a larger gabbro complex (Savu, 1982; Savu *et al.*, 1979a,b). Due to the poor exposures, the connection between gabbros and the sheeted dyke complex cannot be studied in detail. The sheeted dykes are part of a presumably larger complex located at the western part of the ophiolites trending SW–NE parallel to the overall strike of the ophiolites. The distribution of the whole complex is not very well defined. The Căzăneşti dyke complex (Field stop 5) represents the northeastern continuation of the next gabbro–sheeted dyke system near Săvârşin.

The two quarries show 100% sheeted dykes, approximately SW–NE trending (Fig. 28a). No gabbro screens were observed yet between the dykes. The thickness of the dykes ranges from a few centimetres to approximately 50 cm. The grain size varies between the dykes from almost aphanitic to holocrystalline. In general the coarse- and medium-grained dykes are thicker and older; the fine-grained ones are thinner and younger. So far observed, the margins (Fig. 28b) seem to be predominantly one-sided chilled towards the north. The obviously later dykes are chilled on both rims. The orientation and distribution of chilled margins are the same in both quarries. The dyke structure is masked by a number of joints, parallel to sub parallel to the dykes, characterized by the presence of epidote. Along faults, the dykes are sheared and highly weathered.

Similar to the complex at Căzăneşti, two types of dykes can be distinguished petrographically: holocrystalline, doleritic dykes and fine-grained dykes, respectively. The former are coarse grained, comprising mainly plagioclase and clinopyroxene, displaying an intersertal texture. No large phenocrysts are visible. Olivine is missing. Ti-bearing magnetite is the major Fe oxide (Saccani *et al.*, 2001), probably responsible for the magnetic anomaly recorded in the area (Ionescu *et al.,* 2009b). The fine-grained dykes show the same composition, with clinopyroxene, plagioclase and Fe oxides. As expected, the dykes are altered in the greenschist facies, with numerous amphiboles grown on the expense of clinopyroxene. The amphiboles in turn are altered during cooling, at least partly, into chlorite. Additionally, epidote occurs. Similar to the complex at Căzăneşti, amphiboles are the typical minerals indicating the thermal mineralogical change during the relatively slow cooling of the dykes.

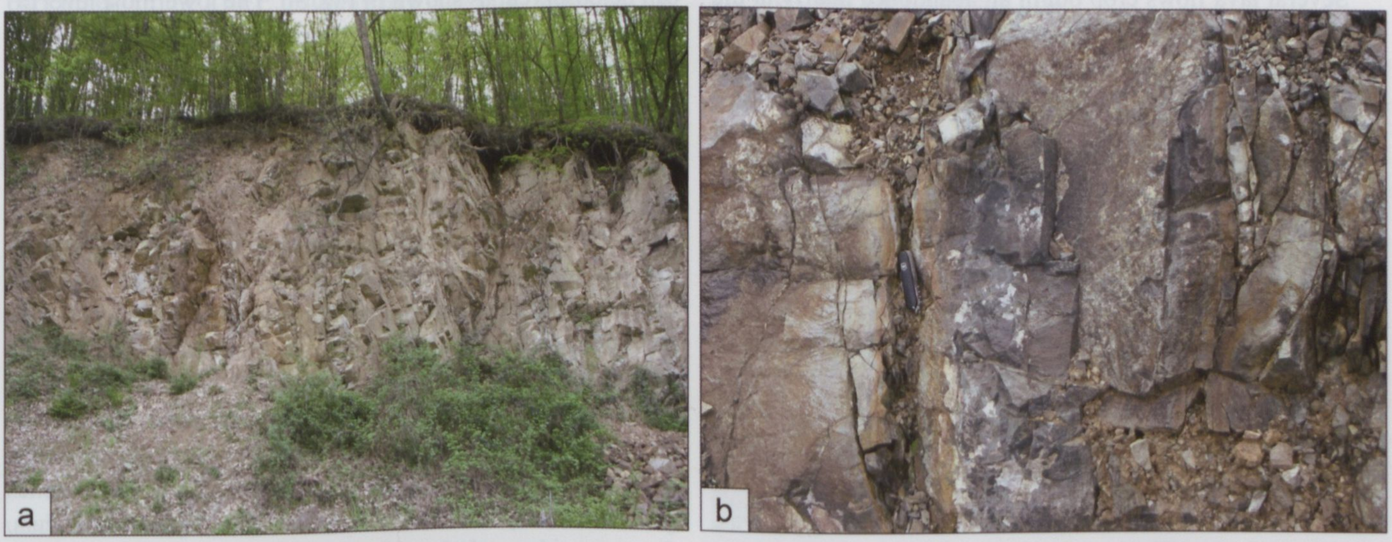

Fig. 28. Sheeted dyke complex in the Julița Quarry:
a) General view; b) Detail of dykes.

According to Saccani *et al.* (2001) plagioclase is normally zoned and has an anorthite content between An_{50} and An_{80}. Clinopyroxene is diopside to augite, with a high Mg#, ranging from 0.75 to ~0.85 and a considerable amount of Ti, Na and Cr.

Geochemically, the dykes have a basaltic to basaltic-andesitic composition. In particular the fine-grained dykes are more evolved. Generally, the Juliţa dykes contain the highest concentration of Fe, Ti, P and the HFSE. At the same time, the Mg and Al contents are low. In this respect, they contrast with the dykes in the Ponor Căzăneşti area, which have low concentration of these elements. This is indicated by the high Fe, Ti, P and the HFSE. At the same time, they show the lowest concentrations of Mg and Al. The overall enrichment of the REE (up to 60 times chondrite) is the highest among the dykes. A weak negative Eu anomaly is common, indicating plagioclase fractionation. The dykes consist of enriched MOR-type basalts.

Lavas are generally less enriched in incompatible elements and REE compared with the Juliţa dykes. By contrast, the former are more depleted in REE and HFSE but match well with other less evolved dykes as for example with those from Căzăneşti. Otherwise, the dykes and lavas are in the same range of element concentrations.

2.8 Field stop 8. Petriş Quarry: Jurassic lavas

Location: *the northern side of the highway from Deva to Săvârşin, between Zam and road junction to Petriş (Figs. 25 and 29).*
Coordinates: *N 46°01.156' and E 22°24.575'; Elevation: 185 m.*

The intense mapping of the ophiolite complex by Savu *et al.* (1979a,b) showed that they are occasionally intruded by dykes or overlain by lava flows of the island arc volcanics. Their occurrences are all relatively small, frequently being found east of Juliţa. One of the examples of the island arc sequence on top of the ophiolites is exposed close to Petriş.

Several lava flows crop out in an abandoned quarry at the southern end of the Mureş Nappe. A small layer of Upper Jurassic island arc volcanics covers the ophiolitic pillow lavas (Fig. 29). The calc-alkaline lavas form an isolated layer extending approximately 1.5 to 2 km and trending ENE–WSW. The hill in which the quarry is located forms a ridge inclined towards N to NW. In the lower part of this ridge, basaltic rocks, mostly as massive dolerites, are exposed. They are in turn overlain by calc-alkaline rocks, building up the entire hill until the Petriş quarry. The geometry of the contacts suggests that the calc-alkaline rocks overlay the tholeiitic basalts. E to NE of the Petriş Quarry, a small valley trends to the NE following over a short distance the strike of the calc-alkaline rocks.

The ridge towards the south of this valley is built up also by calc-alkaline rocks, mainly highly oxidized massive to pillow lavas and more rarely highly vesicular plagioclase-phyric lavas. At the southern flank of this ridge, just NE of the main road, greenish massive lavas appear again. Further to the west, across

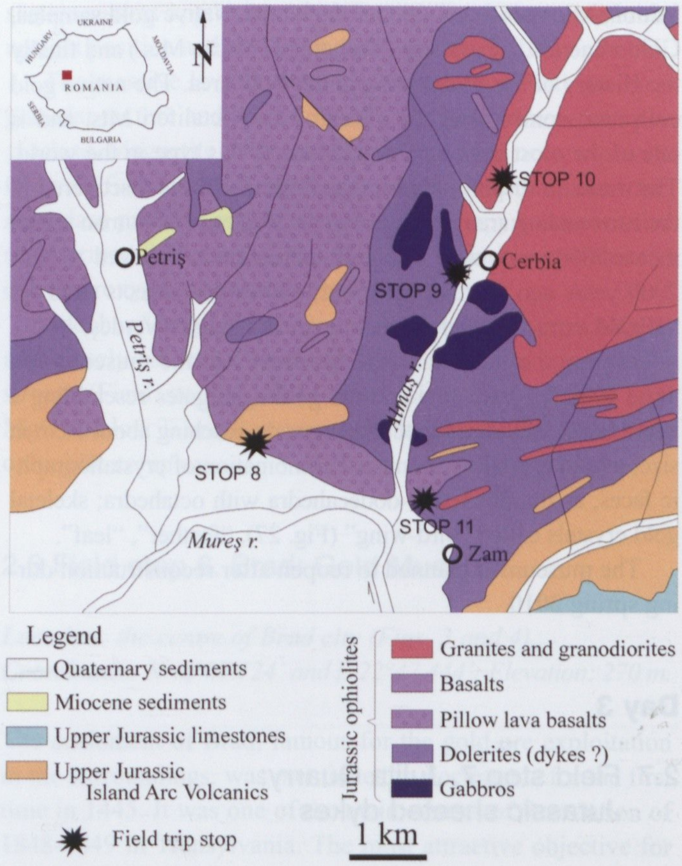

Fig. 29. Simplified geological map of the Petriş–Zam–Cerbia area (according to Savu *et al.*, 1979b), with the location of field stops 8, 9, 10 and 11. The insert in the upper left shows the position of the map within Romania.

the Petriş stream, near the village Sălişte, a similar situation was mapped. The pillow lavas are overlain by calc-alkaline lavas and pyroclastic rock of andesitic to dacitic composition. The regional geological situation indicates that in this area the ophiolites form the oceanic basement of the island arc volcanics.

In the Petriş quarry, four distinct volcanic events (a to d) can be observed, as shown in Fig. 30. The lowermost level (a) consists of pillow lava with pillows of metre size. The pillow rims are completely chloritized, and contain a high amount of iron oxides and hydroxides, indicating a high activity of water. Above the pillow lavas an yellow highly altered porphyritic layer (b), with plagioclase phenocrysts, follows. The lava stream shows columnar joints perpendicular to the cooling surfaces. The thickness of this layer is about 6-7 m. It is in turn overlain by a pillow breccia (c) with 3-4 m thickness and with reddish to purple colours. The highest layer (d) is more acidic and has a thickness of ~3 m.

The pillow lavas are very fine grained, aphyric and characterized by highly acicular plagioclase in a sheaf-like appearance. These features and swallow textures of plagioclase laths suggest a rapid cooling. Plagioclase exists mainly as form relics and is widely replaced by chlorite. Rarely, clinopyroxene occurs. Opaque oxides and glass altered to chlorite, form the groundmass together with plagioclase. Polymictic breccias can be

Fig. 30. Four layers of basaltic rocks in the Petriş Quarry (from bottom to top): *a) Pillow lavas; b) Massive lava flows; c) Pillow lavas; d) Massive lava flows.*

found nearby. They comprise clasts of various basalts in a groundmass of volcanic ash consisting of quartz, clinopyroxene, plagioclase, glass and zeolites.

The Petriş lavas have a SiO_2 content of approximately 52 wt% and high $Na_2O + K_2O$, thus classifying as basaltic trachyandesites in the TAS diagram. While K_2O is relatively low, Na_2O is very high, being combined with low CaO. This argues for strong alteration, which is recorded by the petrography. Originally the volcanics were presumably basaltic andesites with high Al_2O_3 and low TiO_2. They are depleted in HFSE and exhibit a clear Nb and Ta anomaly. With 10-15 times chondrites, the overall REE enrichment is, lower than in ophiolitic basalts and dykes. These features and the spider pattern indicate volcanic arc tholeiites. They might be considered as early eruptions of the IAV sequence directly overlaying the ophiolitic basalts.

The geochemistry of the Petriş lavas show some similarities to those from Poiana Aiudului (Field stop 15), with chondrite-normalized HREE around or below 1, and a slight enrichment of the LREE resulting in a $(La/Yb)_N$ ratio of 1.1 to 1.3 (Fig. 11). Similarly, the N-type MORB normalized HFSE are relatively depleted in both locations, LIL elements are enriched and Nb shows a negative anomaly. For further discussion see Section 2.15.

2.9 Field stop 9. Cerbia: Jurassic gabbros

Location: *at 3.1 km north from the junction with the road Deva-Arad, on the western slope of the Almaş brook, near Cerbia village (Figs. 25 and 29).*
Coordinates: *N 46°02.207' and E 22°26.784'; Elevation: 195 m.*

As seen in Fig. 29, gabbros together with dolerites, are distributed over the whole ophiolite area in three SW–NE trending zones. Most likely they are, at least to a large part, remnants of a sheeted-dyke complex. The westernmost zone includes gabbros and sheeted dykes near Julița (Field stop 7), the central one extends from Săvârşin via Almăşel to Căzăneşti (Field stop 5) and the third one includes the Almaş Valley between Zam and Almaş-Sălişte (Field stop 9). Gabbros cropping out SSW of Cerbia are part of the third easternmost gabbro–dyke complex and are exposed on a fresh cut (Fig. 31) which was opened recently during the reconstruction of a road into Almaş valley (Fig. 29). Gabbros form relativelly small, elongated bodies about 1 to 1.5 km long and a few hundred meters wide. They are surrounded, According to Savu *et al.* (1979b), they are surrounded by dolerites, most likely sheeted dykes.

Gabbros are dark green, isotropic, medium to coarse grained, with a grain size varying from 1 to 10 mm. Mineralogically, they consist predominantly of clinopyroxene and plagioclase. The former is preserved only as relic. Most of clinopyroxenes are changed into amphiboles. Partly, only amphibole gabbros are preserved. Magnetite with variable Ti content (Saccani *et al.*, 2001) are ubiquitous but may be missing in other gabbros. Fresh olivine was not observed in the isotropic gabbros, however some patchy pseudomorphs of serpentine minerals might have been formed after olivine. Amphiboles occur in two forms: a) as large crystals, representing pseudomorphs after clinopyroxene and b) as very fine-grained matrix between the large grains of amphiboles. No detailed chemical analyses are available on amphiboles so far. Their optical features suggest the presence of magnesio hornblende to pargasitic amphiboles. According to Saccani *et al.* (2001) pyroxene is diopside to augite. Their Mg# and Ti content are in the range of clinopyroxene from the sheeted dyke complex in Julița. Magnetite, the main carrier of Ti, is partly replaced by titanite, due to high-*T* ocean floor metamorphism which also changed clinopyroxene to amphibole and olivine to serpentine. During further cooling, calcite, chlorite and probably glauconite-celadonite formed as alteration products.

Geochemically, the isotropic gabbros are similar to MOR-type magmas, with similar major elements, REE and trace elements content. Apart from these, ratios such as Nb/U and Zr/Nb are high, and the LREE are low.

Fig. 31. Gabbros outcropping south of Cerbia village.

Saccani *et al.* (2001) mentioned gabbros dominated by plagioclase, which show sometimes cumulus texture and have a different geochemistry, with high LREE, low V/Ti and high $(La/Yb)_N$ ratios.

2.10 Field stop 10. Cerbia: Jurassic granodiorites

Location: *at 4.7 km north from the junction with the road Deva-Arad, on the western slope of the Almaş brook, near Cerbia village (Figs. 25 and 29).*

Coordinates: *N 46°02.911' and E 22°27.233'; Elevation: 212 m.*

Within the Mureş Zone ophiolites, a number of granitoid intrusions occur at various places and in different sizes, such as Săvârşin, Cerbia and possibly Obârşia. The granitoid intrusions are elongated in the same direction as the general trend of gabbros and sheeted dykes *i.e.* SW–NE (Fig. 25), showing thin contact metamorphic aureoles (Savu *et al.*, 1979a,b). Previously, these granitoids were believed to be of Late Cretaceous-Paleogene (Giuşcă *et al.*, 1967c), or, based on K-Ar, Early Cretaceous in age (Ştefan, 1986; Ştefan *et al.*, 1992). Recently, based on U/Pb isotopes, Pană *et al.* (2002) dated the Săvârşin intrusion as "Oxfordian– Kimmeridgian" (152-156 Ma). Additionally, Re/Os dating on molybdenite from granitoid near Cerbia (Zimmermann *et al.*, 2008) revealed an age of 159.1±0.5 Ma (Oxfordian).

In the outcrops opened along the new road to the villages of Cerbia and Almaş-Sălişte, light pink granitoids are exposed (Fig. 32a). It is one of the best outcrops of the Jurassic granitoids in the Southern Apuseni Mts. The Cerbia granitoid intrusion shows variable compositions and comprises diorites, quartz diorites, granodiorites and granites, according to Savu *et al.* (1979b).

The granitoid rocks outcropping near Cerbia are light pink granodiorites consisting of quartz, plagioclase, K-feldspar and biotite. The grain size, with 0.2 to 0.5 mm, is small. Occasionally, larger crystals of K-feldspar and plagioclase may occur. Plagioclases are normally zoned, whereas the K-feldspars exhibit a "chessboard-type" albitisation. Granodiorite is slightly deformed, without showing conspicuous shearing. Feldspars, in particular K-feldpars are in general highly altered to clay minerals. Biotites are yellowish-brown to reddish-brown and are partly chloritized.

Geochemically, granitoids from Cerbia are similar to those from Săvârşin. They are calc-alkaline and characterized by SiO_2 between 66 and 72 wt%, low TiO_2 and a low Mg#. Nb/U and Ti/Zr ratios are low, the LIL elements such as Ba, Rb, Th, and K are strongly enriched. The REE patterns normalized against continental crust (normalizing values from Sun & McDonough, 1989) are flat, with LILE, Yb and Lu around unity and a slight depletion of MREE (Fig. 33a). Spider diagrams for trace elements, also normalized against continental crust (Sun & McDonough, 1989), reveal a distinct negative anomaly for Nb and Ti and less significant for Ta. Elements such as K, Rb, Ba Th, Zr and Hf are slightly enriched (Fig. 33b). These features argue for a contribution of continental crust during the magma evolution.

Granodiorites are cut by fine-grained dark grey and dark pink dykes. The dark grey dykes (Fig. 34) are microgranodiorites and consist of quartz, plagioclase, K-feldspar, biotite and magnetite. Microscopically, a flow texture, parallel/sub-parallel to the boundary with granodiorites is obvious. Compared with the pinkish granodiorite, the dyke material contains more biotite and magnetite, responsible for the darker colour. Along the contact with the host granodiorite, microgranodiorite shows enrichment in biotite in very narrow zones. The dyke contains conspicuous clusters of pinkish crystals, in millimetre size, which are sometimes surrounded by a dark rim rich in biotite. They consist of either K-feldspar or plagioclase, occasionally accompanied by quartz. The clusters seem to be rather fragments of the host granodiorite, which were not digested by

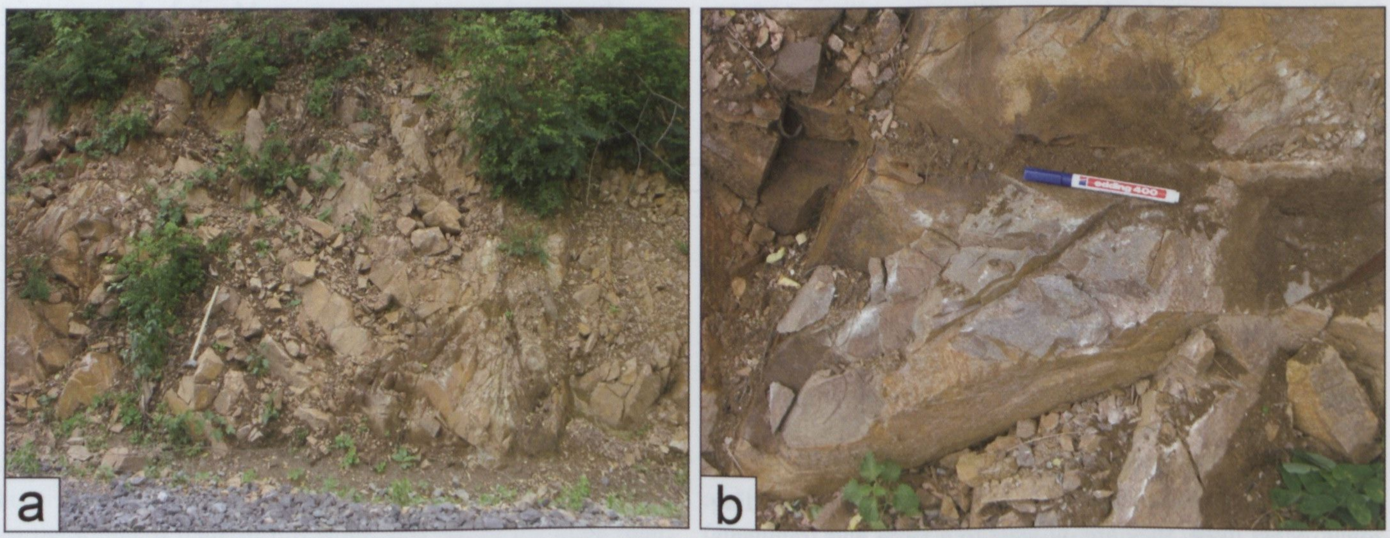

Fig. 32. Granodiorites cropping out south of Cerbia. a) General view, b) Granodiorite (yellowish) with late andesitic to rhyolitic (dark grey) dykes.

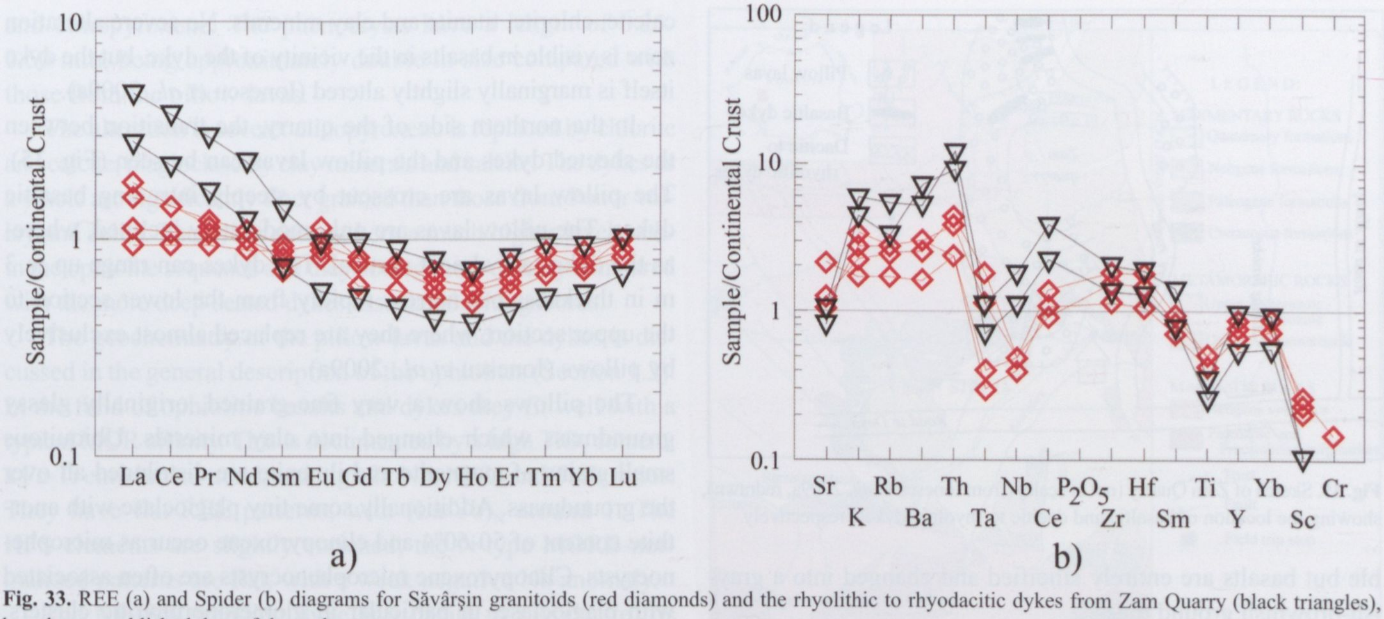

Fig. 33. REE (a) and Spider (b) diagrams for Săvârşin granitoids (red diamonds) and the rhyolithic to rhyodacitic dykes from Zam Quarry (black triangles), based on unpublished data of the authors.

the dyke magma. This is also supported by the finding of cm-sized angular xenoliths, which resemble the composition of the host granodiorite (Fig. 34).

The dark pink dykes are narrow, only few cm across. They are rhyodacites, have a porphyritic texture and consist of quartz, K-feldspar, plagioclase and rarer biotite phenocrysts (between 0.5 to 2 mm in size) in a microcrystalline quartz-feldspar groundmass. Additionally, there are numerous smaller phenocrysts, consisting of mainly quartz and seldom feldspars, up to 0.1 mm in size.

2.11 Field stop 11. Zam Quarry: Jurassic dyke–pillow transition

Location: *on the northern slope of the Mureş valley, approximately 45 km west of the city of Deva (Figs. 25 and 29).*
Coordinates: *N 46°00.834' and E 22°26.285'; Elevation: 175 m.*

The main part of the Zam Quarry is covered by pillow lavas of the ophiolite complex. They are present in all the ophiolite area and follow the spatial distribution of the gabbro–dyke sequences. They form the front of the Mureş Nappe in the W to NW part of the ophiolites and are most abundant west and east of the Săvârşin–Căzăneşti gabbro–dyke zone. In the SE part of the ophiolit zone they can be followed beneath the island arc volcanics (Savu *et al.,* 1979a,b). The Zam Quarry is positioned at the northern border of a small E–W trending and S dipping pillow lava zone overlaying dolerites (Fig. 29).

The quarry is not operating anymore. One of the largest in the area, it extends over several hundred metres and displays pillow lavas with dacitic-rhyolitic dykes, and towards the north a dyke - pillow transition (Fig. 35). By far the most widespread feature is marked by pillow lavas (Fig. 36a) alternating with massive lava flows and sometimes intercalated with basaltic breccias and conglomerates. The pillows trend NNE–SSW and dip gently towards S. They have sizes of decimetres to metres in diameter (Ionescu *et al.,* 2009a). In the center they are massive, in the outer parts highly vesicular. The vesicules are filled with calcite (± hematite), epidote and clay minerals. In the southern part of the quarry the pillows are crosscut by two dacitic to rhyolitic, but different dykes (Figs. 35, 36b).

The first dyke is located at the southern side of the pillows, NW–SE trending and steeply dipping to the SW. The dyke is a highly plagioclase-phyric dacite to rhyolite, highly altered, with visible yellowish-brownish spots of the plagioclase phenocrysts. Further phenocrysts are K-feldspar and quartz. Muscovite occurs in addition. The groundmass consists of quartz, feldspars, muscovite and Fe oxides. This dyke is accompanied by a silicification zone, which affects the joining pillow lavas (Ionescu *et al.,* 2009a). The pillow structure is still recogniza-

Fig. 34. Macroscopic image of the contact between the host granodiorite (left, pink) and the microgranodiorite dyke (right, dark grey) in the Cerbia outcrop (Field stop 10).

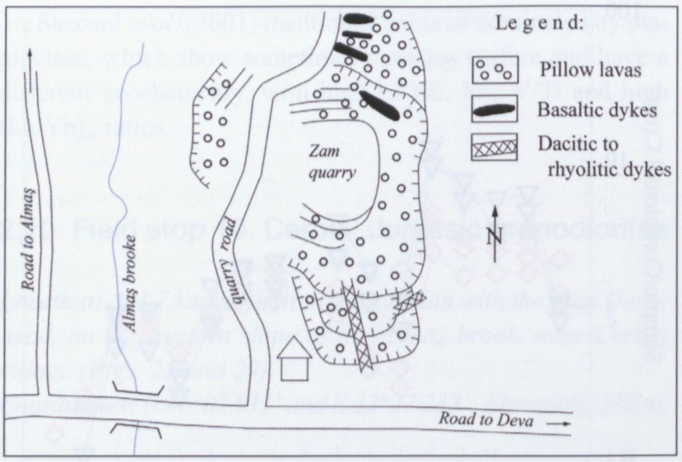

Fig. 35. Sketch of Zam Quarry (not at scale) (from Ionescu *et al.*, 2009a, redrawn), showing the location of basaltic and dacitic to rhyolitic dykes, respectively.

ble but basalts are entirely silicified and changed into a grayish-brownish ground mass.

The second dacitic to rhyolitic dyke trends WSW–ENE and dips steeply to the NNW. It is fairly fresh and highly porphyritic, with phenocrysts of plagioclase, K-feldspar, quartz and Fe oxides in a very fine greyish and dense groundmass. Quartz is highly resorbed; plagioclase, in glomeroporphyritic clusters, is often completely included in K-feldspar. Alteration products are calcite, chlorite, titanite and clay minerals. No severe alteration zone is visible in basalts in the vicinity of the dyke, but the dyke itself is marginally slightly altered (Ionescu *et al.*, 2009a).

In the northern side of the quarry, the transition between the sheeted dykes and the pillow lavas can be seen (Fig. 35). The pillow lavas are crosscut by steeply intruding basaltic dykes. The pillow lavas are only moderately inclined, whereas the dykes are almost vertical. The dykes can range up to 3 m in thickness but narrow rapidly from the lower section to the upper section, where they are replaced almost exclusively by pillows (Ionescu *et al.*, 2009a).

The pillows show a very fine-grained, originally glassy groundmass which changed into clay minerals. Ubiquitous small grains of magnetite and ilmenite are distributed all over the groundmass. Additionally some tiny plagioclase with anorthite content of 50-60% and clinopyroxene occur as microphenocrysts. Clinopyroxene microphenocrysts are often associated with plagioclase, in particular in glomeroporphyritic clusters, where clinopyroxene is best preserved. Otherwise, it is changed into fibrous amphibole (actinolite–tremolite), chlorite and occasionally epidote.

The basaltic dykes intersecting the pillow lavas in the northern part of the quarry are holocrystalline and more coarse grained. The groundmass consists of fine-grained plagioclase

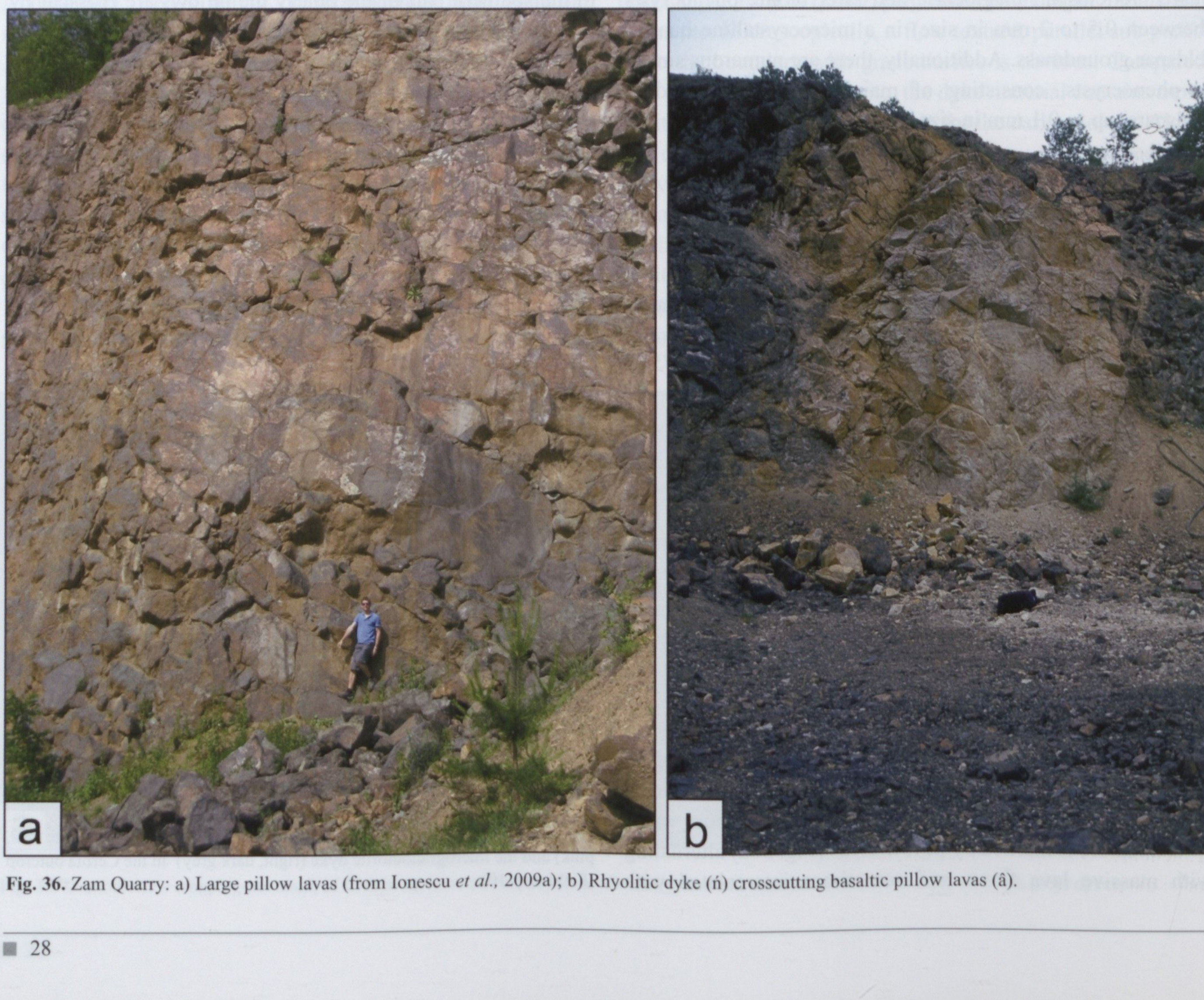

Fig. 36. Zam Quarry: a) Large pillow lavas (from Ionescu *et al.*, 2009a); b) Rhyolitic dyke (ń) crosscutting basaltic pillow lavas (â).

and clinopyroxene. The phenocrysts have a length of 0.25-0.50 mm, being approximately double in size compared with those from the pillow lavas.

The alteration is severe: clinopyroxene is replaced by chlorite and calcite, plagioclase by clay minerals and calcite. The dykes as a whole are significantly finer grained than those from Ponor valley and Juliţa. The dykes present here a much higher level within the ophiolite sequence and they cooled more quickly compared with the more deep-seated dykes associated with gabbros.

The geochemistry of the pillow lavas and the dykes is discussed in the general description of the ophiolites (Section 1.3). In the field of ophiolitic basalts and dykes they fit well with a typical MOR affinity. This is documented by a high TiO_2 content, a Fe-Ti enrichment trend, and high Zr/Nb, Nb/U and Ti/Zr ratios. They have flat REE patterns, with $(La/Yb)_N$ around 1. The HFS elements are slightly enriched; the N-type MORB-normalized patterns are flat, with a weak negative Nb anomaly.

The NE–SW trending dyke is a dacite to rhyodacite. Its geochemistry has some similarity with the Jurassic granitoids from Săvârşin and Cerbia (Fig. 33a,b). Compared with the granitoids, dacites and rhyodacites are slightly more enriched in the LREE and depleted in the MREE and partly also in the HREE, displaying a spoon-formed shape. Likewise, the LILE and additionally Zr, Hf and Nb are high. Such a geochemistry is not compatible with those of andesites in the IAVs sequence but has some affinity to the Jurassic granitoid rocks. A comparison with the banatitic extrusives shows that the geochemistry is not compatible with that of the dyke. Based on the presently available data, the dyke it is best interpreted as being genetically related to the Jurassic granitoids.

Summarizing, the Zam Quarry is an excellent example of the upper oceanic crustal part of an ophiolite including the dyke/pillow transition, the pillow lavas and some later acidic dykes.

Day 4

2.12 Field stop 12. Hunedoara: Matthias Corvinus Castle

Location: *at 20 km from Deva, within Hunedoara (Fig. 37).*
Coordinates: *N 45°44.952' and E 22°53.302'; Elevation: 250 m.*

The Corvins' Castle is a beautiful Middle Age fortress and displays a combination of various architectonic styles, mainly Gothic but also Renaissance (Figs. 38a,b). Ioan (Iancu) of Hunedoara, Governor of Transylvania, owned the castle between 1419 and 1456. As one of the most important political and military leader of the 15th century Europe, he was known for his military campaigns against the Ottomans.

The castle had, with time, a number of owners (see Velescu, 1961), among which the most important were Matthias Corvinus, King of Hungary between 1458 and 1490, son of Ioan of Hunedoara. During his times the kingdom had its widest territorial extent.

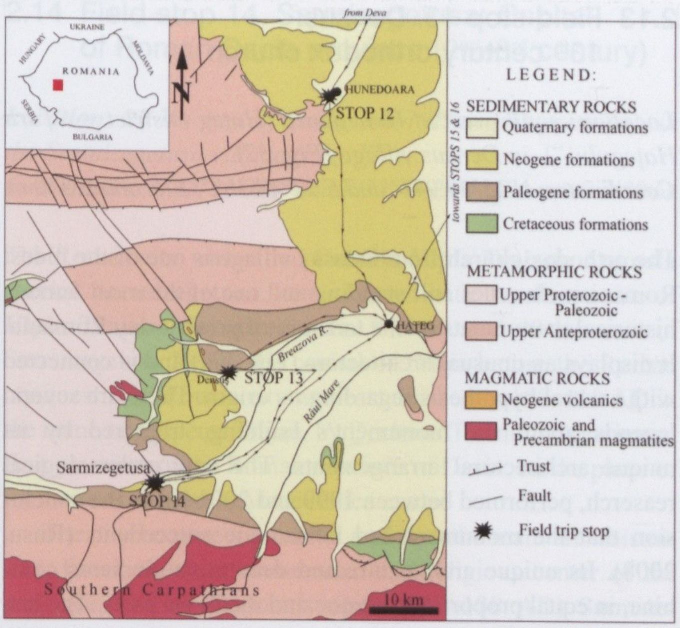

Fig. 37. Simplified geological map of the Hunedoara–Sarmizegetusa–Haţeg area in the Southern Carpathians (according to Codarcea *et al.*, 1967), with the location of field stops 12, 13 and 14. The insert in the upper left shows the position of the map within Romania.

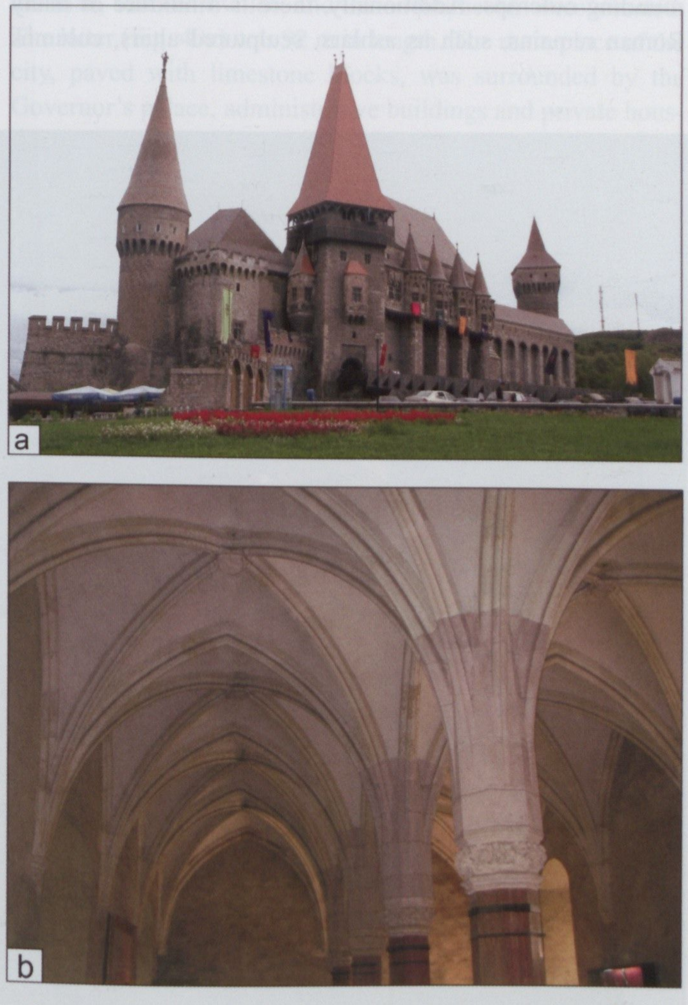

Fig. 38. The Corvins' Castle in Hunedoara: a) Overall view [from Ionescu *et al.* (2009a); Photo: F. Forray]; b) Knights' Hall.

2.13 Field stop 13. Densuş: 13th century orthodox church

Location: *south-west of Hunedoara (Haţeg District = "Ţara Haţegului"), in Densus village (Fig. 37).*
Coordinates: *N 45°34.941' and E 22°48.303'; Elevation: 390 m.*

The orthodox church from Densuş village is one of the oldest Romanian churches still standing and one of the most famous historical monuments on the territory of present-day Romania. It displays an unusual architecture (Fig. 39a) and is connected with several hypothesis regarding its origin. There are several legends about the monument's building, triggered by its unique architectural arrangements. The new archaeological reaserch, performed between 1999 and 2001 led to the conclusion that the monument had no antique antecedents (Rusu, 2008). Its unique architecture and decorative elements combine, in equal proportion, antique and medieval parts. The late Romanesque and early Gothic architecture dates the church in the second half of the 13th century (Rusu, 1997, 2008; Rusu & Burnichioiu, 2008).

The building materials include alluvial pebbles, limestones, volcanics and various metamorphics, originating from the surrounding outcrops. Additionally, there is a mixture of many Roman remains, such as ashlars, sculptured altars, columns and blocks with inscriptions, bricks from a nearby ruined building (Fig. 39b). The walls inside the church are partly covered with paintings.

According to Rusu (2008), the church was founded collectively by several local Romanian nobles with common ancestors Among them, the members of the Muşina family are best known. In the end of the 16th century, the nobles transformed the church into a Calvinist temple and only at the beginning of the 18th century when the village community, which adopted Greek Catholicism, retrieved the church.

Archaeological investigations on the church's architectural elements proved that the altar, nave, and diaconicon were built in tight succession. After a short interval the southern side chapel was erected as well. It underwent some modifications in time, today being severely ruined. New evidence attested the existence of an older iconostasis and an inner scaffold before the middle of the 16th century. A western extension can probably be dated in the end of the 15th century. The bell tower was the last element to be added to the ecclesiastical complex. It had a wooden phase dated no later than the 18th century.

A number of 349 tombs were found in the surroundings of the church. Most of them lack elements for dating. Ca. 240 tombs had no funerary inventory and others contained highly corroded coins or other items which could not be dated. Only two sepulchers could be dated in the 15th century with all certain-

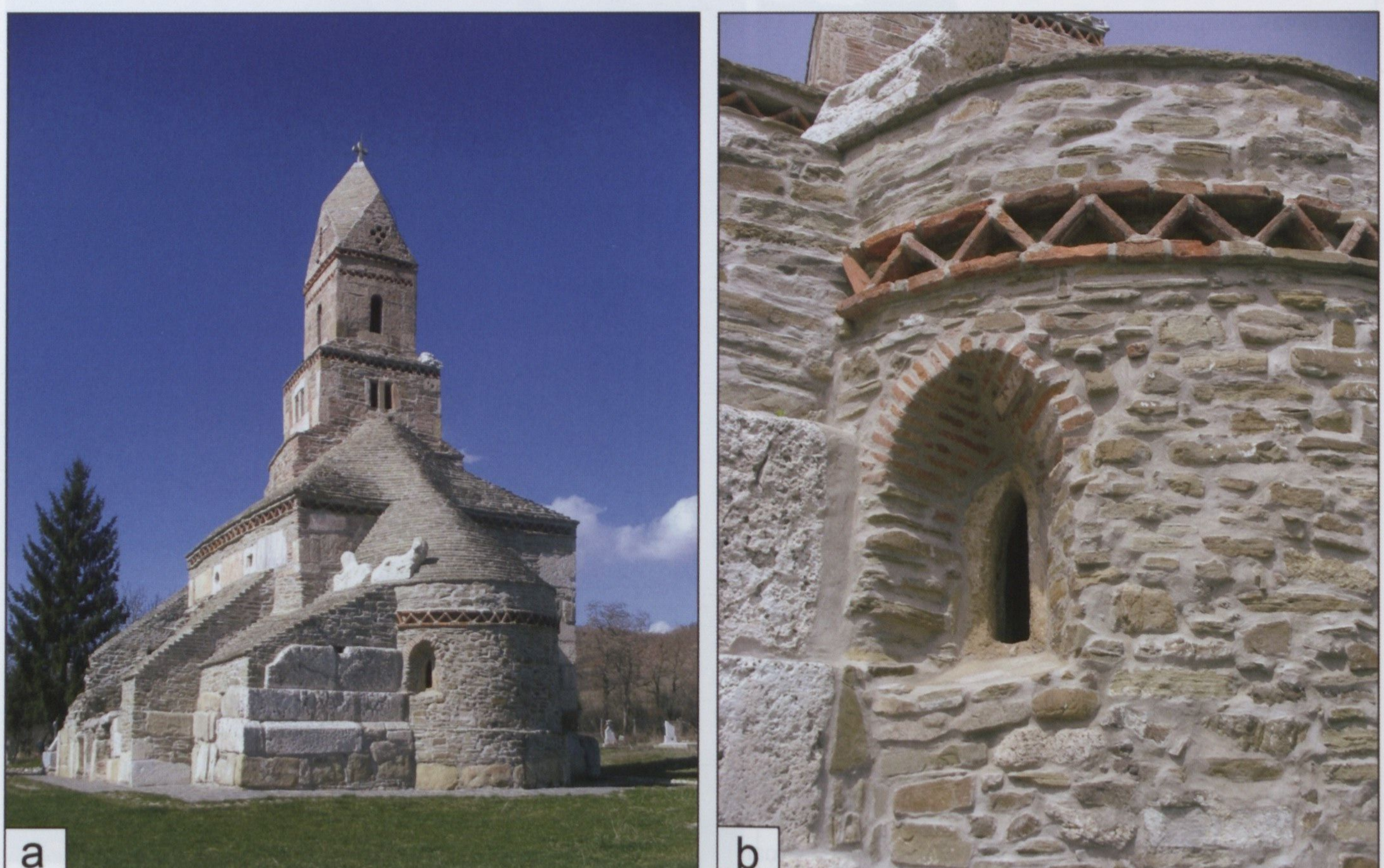

Fig. 39. The Densuş Church: a) General view; b) Detail of the walls.

ty. Other tombs were dated based on coins at the end of the 15th and the beginning of the 16th century. The dating of the church differs from that of the tombs around it (Rusu, 2008).

The cemetery grew especially during the 18th and 19th century when the village seems to have reached a demographic peak. There is evidence that old funerary stones were re-used for graves. Besides the entirely preserved one placed above the entrance door of the church, another fragment was discovered during the research. The re-use of pieces of Roman marble is dated at the end of the 13th century, the probable time when the church was built (Rusu, 2008).

Nearby the church, a Middle Age noble residence, including a 5×8.5 m building was found. The absence of collapsed walls indicates that the building was built out of wood above the ground level. Only the cellar, accessed on a slope made of numerous Roman spolia, preserved the traces of three renovations. The research of the cellar also showed that the building was abandoned peacefully. The historical context and certain archaeological materials, especially stove tiles, allowed to date the residence in the final quarter of the 15th century (Rusu, 2008).

According to the same author, the archaeological materials discovered during the archaeological research on the church site are: items of non-ferrous metals (rings, hair pins, earrings), dress accessories (buttons, clasps, appliqués, buckles), coffin decorations, coins, iron items such as utensils (scissors, knives, frames of wooden shovels), instruments, weapons (arrowheads), accessories such as pins, construction items (cramps, locks, nails). Stone items, either from prehistory (a piece of silex) or the Roman period (inscriptions fragments), manufactured glass objects such as tableware and beads, fragments of wall painting, medieval and Roman building ceramics (stove tiles and bricks, roof and floor tiles respectively), and tableware were also found.

2.14 Field stop 14. Sarmizegetusa: Capital of Roman Dacia province (2nd–3rd century)

Location: *20 km east of Haţeg town (Haţeg Depression), in the Sarmizegetusa village (Fig. 37).*
Coordinates: *N 45°31.035' and E 22°47.205'; Elevation: 480 m.*

The Roman city was called **Colonia Ulpia Traiana Augusta Dacica Sarmizegetusa**, better known with the short name of **Ulpia Traiana Sarmizegetusa.** It is located 40 km east of Sarmizegetusa Regia, the capital of Dacia before the Roman conquest. The Roman city was eastablished in 108–110 AD in the name of Emperor Trajan, by Decimus Terentius Scaurianus, first governor of the Roman province of Dacia, on the place of the famous Legion V Macedonica camp.

The city became the capital and thus the main administrative, fiscal, military, economical and religious center of the province. During Hadrian's times (117–138 AD) Sarmizegetusa got the name of Colonia Ulpia Traiana Augusta Dacica Sarmizegetusa. With time, the name simplified and became Colonia Sarmizegetusa. The city lasted till the 4th century. The ruins of the Roman city extend on approximately 1.5 km^2. The first fortification wall was 5 m in height. Outside the wall, thermae, suburban villae, and the amphitheater were situated. The latter (Fig. 40a) was 92 m in length. The main place of the city, paved with limestone blocks, was surrounded by the Governor's palace, administrative buildings and private houses (Fig. 40b). Remnants of ovens for bricks and a necropolis were found nearby the city. The population was estimated to 20,000/25,000 inhabitants.
(http://cimec.ro/Arheologie/UlpiaTraiana/ descriere/sit.html)

Fig. 40. Ruins of Sarmizegetusa: a) Amphitheater (from Ionescu *et al.*, 2009a); b) Forum.

2.15 Field stop 15. Poiana Aidului: Jurassic basalts with dykes (island arc volcanics)

***Location:** in the village of Poiana Aiudului, on the eastern slope of the Rachiș Valley, at 200 m from the road between Aiud and Rimetea (Fig. 41).*

***Coordinates:** N 46°22.130' and E 23°36.239'; Elevation: 390 m.*

Whereas the ophiolites and the few island arc volcanics occurring NW of Deva are part of the Căpâlnaș–Techereu Nappe, the rocks from the Poiana Aiudului belong to the Rimetea–Bedeleu Nappe in the Trascău Mts. (Figs. 4, 41), which comprises mostly island arc volcanics (Bleahu *et al.,* 1981; Balintoni & Iancu, 1987; Balintoni, 2003; Nicolae & Saccani, 2003). They extend from Cheile Turzii in the NNE to Poiana Aiudului in the south. At some places (Fig. 4; not shown in Fig. 41) the IAVs of the Rimetea–Bedeleu Nappe are underlain at their eastern side by ophiolitic rocks, displaying a depositional contact (Gandrabura, 1981; Savu & Udrescu, 1992).

Similar to the tectonic situation found near Petriș and Zam, such a relation suggests an oceanic setting of the island arc sequence of the Rimetea–Bedeleu Nappe. However, the island arc volcanics are overlain at the western side of the Rimetea–Bedeleu nappe by Upper Jurassic limestones (Săsaran & Bucur, 2006) with a facies close to the Tithonian "Stramberg limestone". The outcrops in Poiana Aiudului represent isolated occurrences separated from the main island arc volcanics towards the north by Upper Cretaceous sandstones. Along the small stream of Rachiș, just north of Poiana Aiudului, several large outcrops can be studied. They consist of pillow lavas cut by a number of sub-parallel oriented dykes. The thickness of the dykes varies from few decimetres to approximately 1 m. They trend SSW–NNE, plunging to ESE.

The pillows are brecciated, vary in pillow size from few centimetres to 1 m, are highly chloritized/celadonitized and have often glassy rims. The pillows are in upright position, tilted slightly towards NW. Under the microscope, they are highly porphyritic, with plagioclase as predominat phenocrysts, and exhibit sometimes a glomeroporphyritic texture. The An content of plagioclase ranges from 70 to 80%, with a weak zoning. Microphenocrysts of clinopyroxene occur as well. The groundmass consists of plagioclase, clinopyroxene und glass, which is partly devitrified and partly changed into chlorite. Occasionally, pyroxene crystals, possibly xenocrysts, occur up to 1 cm in size. Vesicles are filled with carbonate, which also appears on small veinlets. The amount of Fe oxides varies. In samples with higher amount of Fe oxides, chlorites are also Fe-rich.

The dykes are by contrast partly aphyric, partly sparsely phyric; the texture of the groundmass is intergranular. Quartz is common. Clinopyroxene and plagioclase microphenocrysts are altered alike. The former is chloritized, the latter is replaced by carbonate.

Geochemically, the Poiana Aiudului pillow lavas can be classified as trachybasalts and basaltic trachyandesites. The

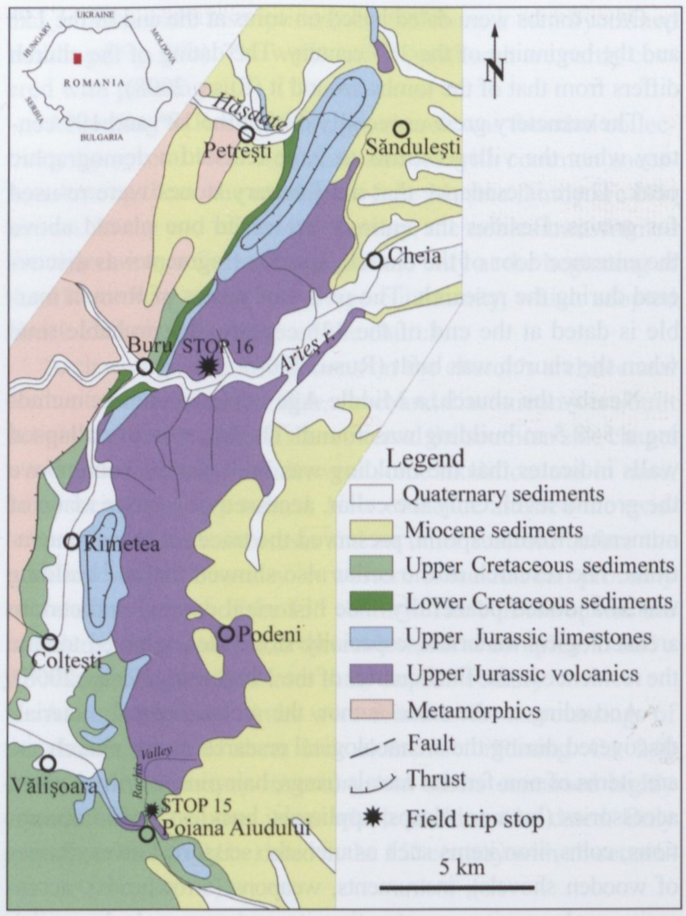

Fig. 41. Simplified geological map of the Poiana Aiudului–Buru area, in the Trascău Mountains (according to Giușcă *et al.*, 1967a), with the location of field stops 15 and 16. The insert in the upper left shows the position of the map within Romania.

high total alkali content is due to Na_2O. The K_2O content is low. At least partly, the Na_2O enrichment is probably due to alteration processes (see also Field stop 16). The Poiana Aiudului pillows and dykes are geochemically different from N-type MORBs and from andesites found mainly in the Rimetea–Bedeleu Nappe. They have, for example, high Ti/Zr and low Nb/Zr or Zr/Y ratios. The HFS elements are lower than in N-type MORB. The LIL elements such as Rb and Ba are in the MORB range. The Poiana Aiudului lavas have a strong negative Nb and Ta anomaly. The LREE normalized against chondrites are slightly enriched. The difference to the common island arc andesites is demonstrated by lower contents of Hf, Zr, Sm, P, LREE as well as LILE values. It should be remembered that the IAVs of Petriș (Field stop 8) have a resemblance to those in Poiana Aiudului. As dicussed already at Field stop 8, this regards in particular the chondrite-normalized HREE around or below 1 and a slight enrichment of the LREE resulting in a $(La/Yb)_N$ ratio of 1.1 to 1.3. Similarly, the N-type MORB normalized HFSE are relatively depleted in both locations. A systematic enrichment of the LILE and a significant negative Nb anomaly is observed (see also Fig. 11). The overall picture of the Poiana Aiudului trachybasalts resemble more the tholeiitic island arc volcanics than the calc-alkaline basalts.

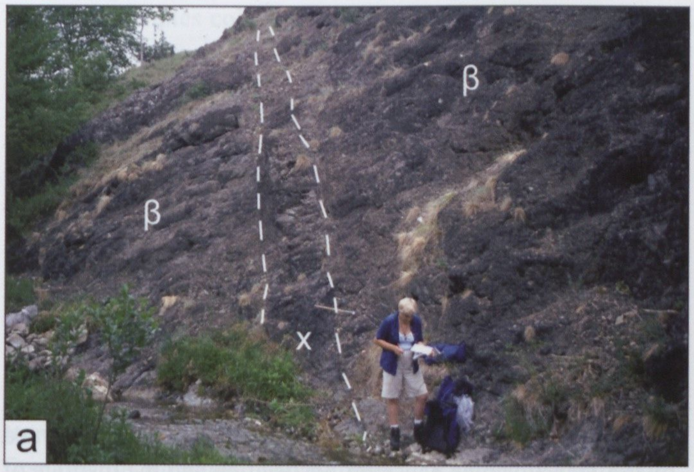

Fig. 42. Mesozoic volcanics in Poiana Aiudului, Rachiş valley: a) Doleritic dyke (x) crosscutting basaltic pillow lavas (β); b) Basaltic and doleritic dykes (in relief) in between pillow lavas (dark brown, eroded).

2.16 Field stop 16. Buru: Jurassic island arc volcanics

Location: *at 15 km west of Turda city, on the road to Abrud, on the northern slope of the Arieş river (Fig. 41).*
Coordinates: *N 46°30.495' and E 22°38.461'; Elevation: 385 m.*

The oucrops along the road from Buru to Cheile Turzii (Fig. 43) represent a cross-section through the northern part of the Rimetea–Bedeleu Nappe. The geological situation of the nappe is already described at Field stop 15, Poiana Aiudului. The Arieş river crosses E of Buru the island arc volcanics along a distance of several kms. The best exposures are found near Cabana Buru and continue about 1 km further to the east. The rocks are mainly massive lava flows, breccias and pyroclastics with the chemistry of basaltic andesites to andesites. Pillow lavas occur as well. Occasionally, rhyodacitic and rhyolitic dykes crosscut the volcanic edifice. No stratification can be observed. However, in the eastern part, a steeply W-dipping layering was observed.

The texture varies from highly phyric to almost aphyric, with only few phenocrysts. Plagioclase predominates over clinopyroxene and amphibole. In more evolved types, quartz phenocrysts occur rarely. The groundmass consists of fine-grained plagioclase, clinopyroxene, chlorite and fine-grained Fe oxides. Some samples exhibit a pronounced flow texture. The anorthite content of plagioclase varies from An_{70} to An_{30}. Almost pure K-feldspars, alkali-feldspars with Ab_{40-60} and ternary feldspars occur additionally. The latter ones seem to have crystallized as individual crystals within the groundmass, the former are mostly alteration products.

Clinopyroxene is SiO_2-rich, with low contents of TiO_2 and Na_2O. They can be classified as augite, in agreement with the results of Nicolae & Saccani (2003). In various diagrams, using the clinopyroxene composition for tectonic discrimination (Nisbet & Pearce, 1977; Beccaluva *et al.*, 1989), they plot in the field of island arc volcanics. Regarding pyroxenes, only clinopyroxene was identified around Buru, but orthopyroxene occurs elsewhere in the island arc volcanics (Nicolae & Saccani, 2003).

Zeolites as alteration product are obviously restricted to the island arc volcanics, where they appear frequently. They have not described yet from the ophiolites. The most common zeolites are laumontite, heulandite, Ca-rich mordenite, more rare are leonhardite (?) and clinoptilolite (our own EMPA data, *unpublished*). They replace together with other clay minerals phenocrysts of plagioclase and clinopyroxene. Minerals formed dur-

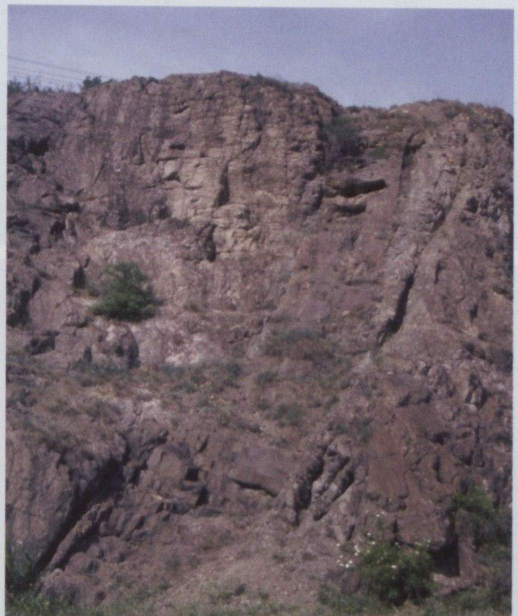

Fig. 43. Island Arc Volcanics in the Buru area: andesites and basaltic andesites lava flows.

ing alteration processes such as calcite, chlorite, celadonite, glauconite and illite occur additionally.

The island arc volcanics in the vicinity of Buru range from basaltic andesites to rhyolites, The more SiO_2-rich compositions show low CaO but partly very high Na_2O, resulting in a low CaO/Na_2O ratio, due to subsolidus alteration reactions with seawater. While other oxides such as MgO, FeO, Al_2O_3 or TiO_2 show a strong and well defined negative correlation with SiO_2, CaO exhibits a wide scatter in composition, but still a poor negative correlation is discernible. K_2O is not correlated with SiO_2. This distribution of CaO and K_2O, the low content of the former combined with the high content of Na_2O testify for a relatively strong alteration.

The REE pattern shows an enrichment of LRE elements, with $(La/Yb)_N$ ratio of around 2. Normalized against N-type MORB, they show a strong increase from the HFSE towards the LILE and a negative Nb and Ta and also a negative Ti anomaly.

Day 5

2.17 Field stop 17. Cluj-Napoca: Mineralogical Museum

Location: *center of Cluj-Napoca (Fig. 1), in the main building of the "Babeş–Bolyai" University.*
Coordinates: *N 46°46.048' and E 23°35.481'; Elevation: 350 m.*

The Mineralogical Museum contains a rich collection of mineral specimens, accumulated during its long history. The museum was established as purely academic collection in 1919, on the basis of the mineral and rock collection earlier jointly owned by the Transylvanian Museum Association and the Institute of Mineralogy and Geology of the "Ferenc József" University. Subsequently, the collections were reorganized, continuously and substantially enriched by donations, exchanges, acquisi-

Fig. 44. Images from the Mineralogical Museum (from Ionescu *et al.*, 2009a): a) Dendritic gold crystal from Roşia Montană (length of the specimen 2 cm, b) The Mociu (Mocs) chondritic meteorite (width of the specimen 35 cm, c) Systematic exhibition of minerals, d) Bucium pyrite (length of the specimen 66 cm).

tions and samples collected by the professors and students in geology (Ionescu & Tămaş, 2003; Ionescu *et al.*, 2009a). The Chair of Mineralogy scientifically coordinates the museum. Rare specimens (very large crystals of pyrite, tetrahedrite, aragonite, halite, quartz and gypsum), as well as rare mineral species (native tellurium, tellurides, some silicates), the cut gemstones and the meteorites are among the main attractions in the museum.

The museum houses about 25,000 samples, grouped in the following main collections: "Native gold" (450 samples), "Meteorites" (210 samples), "Systematic collection" (about 10,000 samples), "Regional collection" (about 1,500 samples), "Cut gemstones" (250 cut gems), "Romanian gemstones" (about 3,500 samples), "Minerals discovered for the first time in Romania" (40 samples), "Radioactive minerals" (about 200 samples), "Ore minerals" (5,000 samples), "The crystallographic collection" (1,500 samples) and "The petrographic collection" (1,000 rock samples).

The *Native gold* collection, with 450 samples (Fig. 44a), represents the second largest gold collection, whereas the collection of 10 meteorites is unique in Romania. Fragments from five Romanian meteorites are displayed, including the famous Mociu (Mocs) meteorite. The largest piece, weighting 35.7 kg (Fig. 44 b), fell down in 1882 at 40 km east of Cluj-Napoca.

The systematic collection contains about 10,000 samples from the world's most important occurrences, illustrating over 800 mineral species ordered according to the Strunz classification (Figs. 44c,d). This collection represents the greatest number of different minerals species gathered in a single museum, in Romania. *The regional collection* consists of 1,500 mineral samples collected from over 70 mines, outcrops and quarries from Romania.

The cut gems collection (250 cut gems), the richest in Romania, is displayed in an original cupboard from the XIX[th] century. It contains the main gems, such as diamond, corundum (with both ruby and sapphire varieties), beryl (emerald, aquamarine), garnets, tourmaline, quartz (amethyst, citrine, rock crystal, and agate), opals (precious opal, fire opal), spinels, zircon, turquoise, as well as pearls and corals. Synthetic counterparts of the main gemstones can also be seen.

The collection of the Romanian gemstones consists of 3,500 samples (Ionescu *et al.*, 2009a), from which only 1,200 cabochon-cut gemstones, mainly quartz, chalcedonies, agates, opals, jaspers, are displayed. This collection was set up in 1987, based on the specimens collected from 90 occurrences by the professors of the Geology Department.

The collection of the minerals discovered for the first time in Romania contains 18 mineral species, *e.g.* native tellurium, sylvanite, nagyagite, krennerite, petzite, fizélyite, fülöppite, semseyite, andorite, tellurite *a.o. The collection of radioactive minerals* (about 200 samples) is not exhibited publicly and is used only for scientific purposes. *The ore minerals collection* contains about 5,000 samples originating from Romania and foreign countries, genetically grouped in: orthomagmatic, pegmatitic, skarn, hydrothermal, metamorphic and sedimentary deposits. *The crystallographic collection* (about 1,500 samples) consists of 750 natural crystals as well as casts. *The petrographic collection* (1,000 rock samples from Europe) is located in the Microscopy room and is used mainly for teaching.

2.18 Field stop 18. Valea Lungii Quarry: Upper Cretaceous granodiorites (banatites)

Location: *Vlădeasa Massif, on the Valea Lungii brook, 1.5 km from the confluence with the Valea Drăganului River and 13 km south of the road between Cluj-Napoca and Oradea. Cluj-Napoca is situated at 80 km ESE (Fig. 13).*

Coordinates: *N 46°53.224' and E 22°48.920'; Elevation: 625 m.*

The porphyritic granodiorite in the Valea Lungii Quarry (Fig. 45) is part of the Vlădeasa Massif, situated at the northernmost part of the banatitic belt (see Section 1.4). The Vlădeasa Massif consists of a sequence of intermediate to acidic volcanics and intrusions. The volcanics constitute by far the largest part of the massif, the intrusives are considerably smaller. The Budureasa and Pietroasa intrusives, occurring SW of Vlădeasa, are the largest in the Northern Apuseni banatites. By contrast, in the Vlădeasa area only small but numerous instrusives occur (Fig. 13). The age relations among the volcanics and the intrusives were already shortly discussed at Field stop 2 (Pietroasa Quarry).

The igneous massif of Vlădeasa is situated at the very limit between the Bihor Mts. and the Pădurea Craiului Mts. (Fig. 3). The regional basement is represented by predominantly Variscan metamorphic complexes overlain by Permo-Mesozoic sedimentary deposits, including Permian rhyolitic flows (Ianovici *et al.*, 1976; Istrate, 1978; Ştefan, 1980; Bleahu *et al.*, 1981). In the NW, the basement is part of the "Bihor Autochtonous Unit". Only in the SW the banatitic massif crosscuts the Alpine nappes (Codru nappe system, in Fig. 3).

The mineralogical, petrographic and geochemical features of the magmatic rocks from Vlădeasa show significant differ-

Fig. 45. Granodiorites in the Valea Lungii Quarry.

ences between the rhyolitic rocks, representing the main part of the massif and the more basic rest of the banatitic rocks which occur subordinately (Istrate, 1978; Istrate & Bratosin, 1976; Ştefan, 1980). Rhyolites form extended flows mainly covering the Cretaceous sedimentary rocks. In some places remnants of Upper Cretaceous sediments are still "floating" above rhyolites. The dioritic and granodioritic apophyses, which penetrate the metamorphic complex and the sedimentary deposits as well as the rhyolite mass, point to a large intrusion in the depth (Istrate, 1978).

The Valea Lungii Quarry is opened in porphyritic granodiorites, which contain numerous and various xenoliths of magmatic, metamorphic and sedimentary origin. The xenoliths are similar to those found in the Pietroasa granodiorite. Petrographically, the Valea Lungii granodiorite consists of a medium-grained holocrystalline groundmass and larger plagioclase phenocrysts. The groundmass is made of plagioclase, K-feldspar, biotite, amphibole, Ti-bearing magnetite to ilmenite and quartz as major constituents. Chlorite, calcite and epidote are typically low-*T* alteration products. The groundmass plagioclases are homogeneous and show an andesine composition. The plagioclase phenocrysts show conspicuous oscillatory zoning, compositionally ranging from An_{50} to An_{30} (Hoeck *et al.*, 2010). The alkali feldspar is very K-rich, with an Ab content of maximum 10%, and displays little variation (Fig. 46a). Biotite has an intermediate Fe#, calculated based on Fe_{TOT}, and is relatively poor in Al^{IV} (Fig. 46b). Amphibole is magnesiohornblende. A preliminary application of the Al-in-hornblende geobarometer based on several calibrations (Hammarstrom & Zen, 1986; Johnsonn & Rutherford, 1989; Schmidt, 1992) but without temperature estimates (Anderson & Smith, 1995), yielded low pressures, between 0.15 and 0.25 Gpa. This is consistent with earlier pressure estimations for Vlădeasa granodiorites (Istrate, 1978) and for the contact metamorphosed Anisian marbles around the Budureasa and Pietroasa massifs (Ionescu & Hoeck, 2005).

The petrographic composition, including the high magnetite content classifies granodiorites as I-types and as part of the magnetite series granitoids (Ishihara, 1977), respectively.

The xenoliths of magmatic rocks represent the majority and consist of various andesites, quartz andesites, and mainly dacites (Fig. 47a), all of them being consanguineous with the host rock. The contact between the andesite xenoliths and granodiorites is generally clear, but recrystallization or silicification of the groundmass is sometimes visible.

The close genetic relation among the host granodiorites on one side and the volcanics and intrusives xenoliths, on the other side is underlined by the similar mineralogy including amphibole, biotite, plagioclase, K-feldspar and magnetite. Subordinately, ilmenite occurs as well. The composition of most of the minerals is very close to those in granodiorite, except for K-feldspar, which shows a wide range from Na-rich to K-rich compositions. In many xenoliths, the same type of phenocrysts as in the host granodiorite is common.

Dacitic xenoliths are often large and, in turn, include fragments of older dacites or andesites (Fig. 47a). The contact of dacites with the granodioritic host rock is either sharp or gradual. Feldspar-rich rims are common at the contact.

Seldom, fragments of Permian rhyolites with fluidal structure occur as xenoliths. They reveal the influence of the host

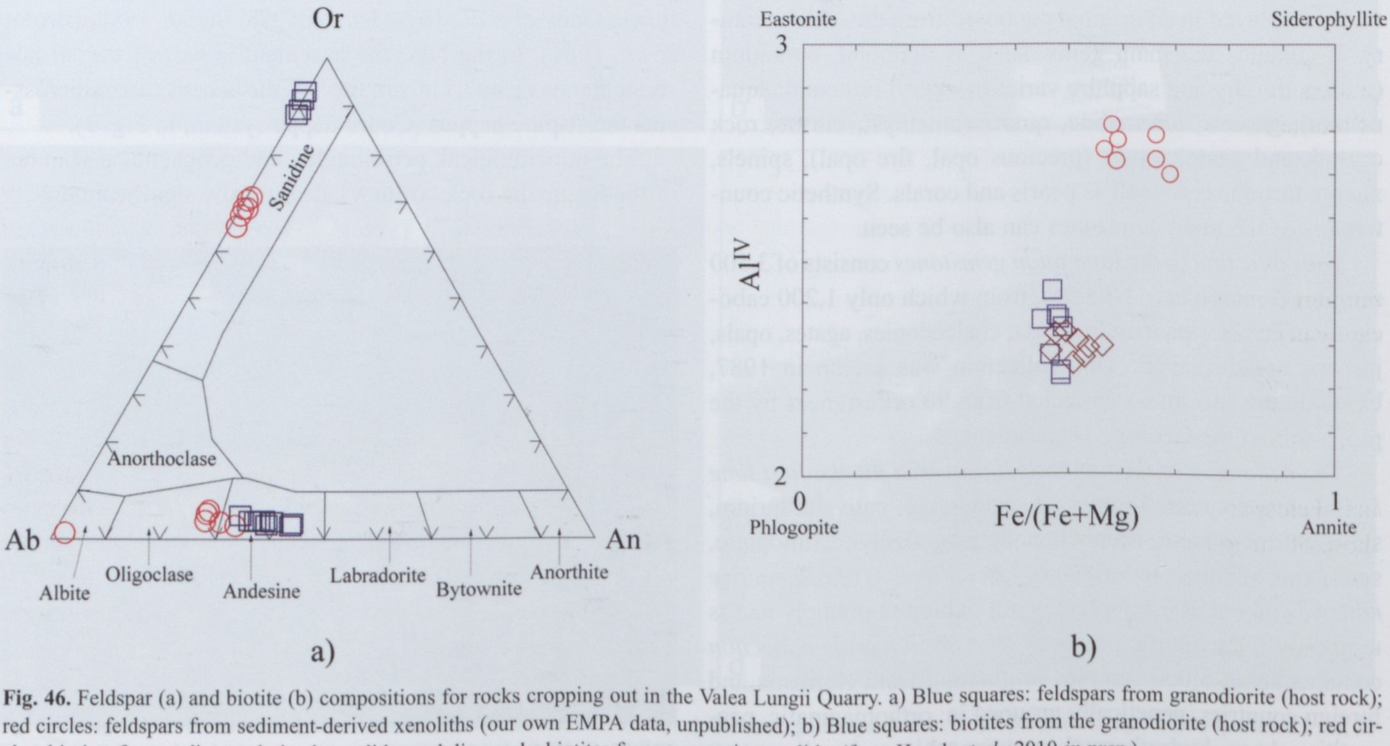

Fig. 46. Feldspar (a) and biotite (b) compositions for rocks cropping out in the Valea Lungii Quarry. a) Blue squares: feldspars from granodiorite (host rock); red circles: feldspars from sediment-derived xenoliths (our own EMPA data, unpublished); b) Blue squares: biotites from the granodiorite (host rock); red circles: biotites from sediment-derived xenoliths; red diamonds: biotites from magmatic xenoliths (from Hoeck *et al.*, 2010 *in prep.*).

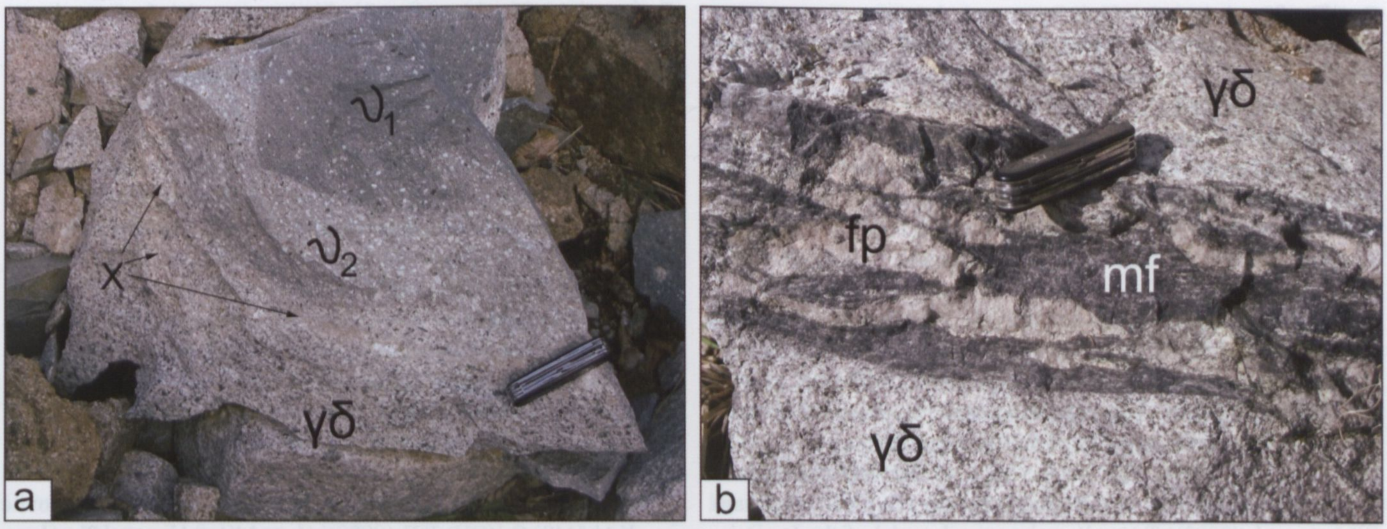

Fig. 47. Xenoliths in the Valea Lungii granodiorite (γδ): a) Dacite(υ_1)–in-dacite(υ_2) xenolith. The latter shows feldspathic reaction zone (x) at the contact with granodiorite; b) Metamorphic xenolith (mf) showing feldspathisation (fp) along foliation.

rock, as shown by the recrystallization of the glassy groundmass and the formation of the quartz + orthoclase spherulitic textures (Ghergari & Ionescu, 2001).

The xenoliths of metamorphic rocks are various in size, from few centimeters to tens of centimeters, being less frequent than the magmatic xenoliths. The metamorphic rocks belong to the crystalline basement being represented by quartzites, amphibolites, amphibole schists, biotite micaschists, quartzofeldspathic schists and gneisses (see also Har & Năstase, 2003). In general, the contact of the granodioritic magma with the metamorphic xenoliths, except for quartzite, show more or less intense metasomatic and thermal processes.

Amphibolite xenoliths are relatively frequent. Their contact with granodiorite is variable, ranging from no reaction to a fine quartzofeldspathic border or to intense feldspathization process throughout the whole xenolith (Fig. 47b). The contact of biotite micaschist xenoliths with granodiorites is generally without any reaction processes, except for few cases of small-scale biotitization of granodiorites as well as the formation of a silicate melt at the contact. From this melt, isometric quartz crystals were formed. Inside the xenolith mass, andalusite and cordierite may occur. Gneisses and quartzofeldspathic rocks rarely occur as xenoliths.

The sedimentary xenoliths are less frequent than the magmatic ones. Common minerals of the sedimentary-derived xenoliths include biotite, plagioclase, K-feldspar, cordierite, andalusite, corundum, magnetite, spinel and tremolite–actinolite. Feldspars and biotites are significantly different form those of the granodiorites. Plagioclase composition is around the oligoclase–andesine boundary, alkali feldspars contain up to 30% albite component (Fig. 46a). Biotites are more Al-rich and have higher Fe content (Fig. 46b). Quartz and occasionally almandine-rich garnet and cordierite were also reported. Muscovite exists mainly as relic in K-feldspar. Chlorite, epidote and calcite are due to hydrothermal alteration. The presence of K-feldspar and corundum in some samples indicates the decomposition reaction of muscovite to K-feldspar + corundum + vapour in a silica-undersaturated environment, taking place in the andalusite field. According to the experimental data obtained by Chatterjee & Johannes (1974), the temperature for this reaction would be between 650 and 700 °C and pressure between 0.15 and 0.3 GPa. This is consistent with the first pressure estimation, based on the Al-in-hornblende geobarometer (see above).

In general, the metasedimentary xenoliths show a sort of layering, represented by layers of various mineral associations. No foliation is visible. However, in few samples old folds can be recognized but they are completely overgrown by the new, contact minerals.

Summarizing, the mineralogy of granodiorites and xenoliths is consistent with a shallow intrusion depth and an intense contact metamorphism. It is visible in the mineralogy of the xenoliths in Valea Lungii and Pietroasa (Field stop 3), and in Anisian dolomitic limestones surrounding the intrusions *e.g.* in Budureasa (Field stop 2).

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Appendix – Itinerary for IMA2010 RO2 Field trip

Saturday, August 28, 2010 (Day 1). Travel from Budapest to Beiuş

08.00–12.00	Travel from Budapest to Beiuş
12.00–14.00	Lunch break in Beiuş
14.00–15.00	Travel to Pietroasa
15.00–16.30	Field stop 1: The Urşilor Cave (The Bears' Cave) near Chişcău village
16.30–18.00	Field stop 2: Pietroasa Quarry: Upper Cretaceous granodiorites (banatites)
18.00–19.00	Travel to Beiuş
19.00–21.00	Dinner in Beiuş
	Accommodation in Beiuş

Sunday, August 29, 2010 (Day 2). Travel from Beiuş to Deva

08.00–09.00	Travel to Budureasa
09.00–10.00	Field stop 3: Budureasa: Brucite deposits
10.00–13.00	Travel to Vaţa
13.00–14.00	Lunch break in Vaţa
14.00–14.45	Field stop 4: Ponor Valley at Căzăneşti: Jurassic gabbro–dyke section
14.45–15.30	Field stop 5: South of Căzăneşti: Jurassic MOR-type basalts
15.30–16.00	Travel to Brad
16.00–17.30	Field stop 6: Brad: Gold Museum
17.30–18.30	Travel to Deva
19.00–21.00	Dinner in Deva
	Accommodation in Deva

Monday, August 30, 2010 (Day 3). Travel from Deva to Juliţa and back, along the Mureş Valley

08.00–10.00	Travel to Juliţa
10.00–12.00	Field stop 7: Juliţa Quarry: Jurassic sheeted dykes
12.00–13.00	Travel to Petriş
13.00–14.00	Lunch break near Petriş
14.00–14.30	Field stop 8: Petriş Quarry: Jurassic lavas
14.30–14.45	Travel to Cerbia
14.45–15.15	Field stop 9: Cerbia: Jurassic gabbros
15.15–16.00	Field stop 10: Cerbia: Jurassic granodiorites
16.00–16.30	Travel to Zam
16.30–18.00	Field stop 11: Zam Quarry: Jurassic dyke–pillow transition
18.00–19.00	Travel to Deva
19.00–21.00	Dinner in Deva
	Accommodation in Deva

Tuesday, August 31, 2010 (Day 4). Travel from Deva to Cluj-Napoca

08.30–09.00	Travel to Hunedoara
09.00–10.00	Field stop 12: Hunedoara: Mathias Corvinus Castle
10.00–11.00	Travel to Densuș
11.00–11.30	Field stop 13: Densuș: XIII[th] century Orthodox Church
11.30–12.00	Travel to Sarmizegetusa
12.00–13.00	Field stop 14: Sarmizegetusa: capital of Roman Dacia province (II[nd]–III[rd] century)
13.00–14.00	Lunch break
14.00–16.00	Travel to Poiana Aiudului
16.00–17.00	Field stop 15: Poiana Aiudului: Jurassic basalts with dykes (Island Arc Volcanics)
17.00–18.00	Travel to Buru
18.00–18.30	Field stop 16: Buru: Jurassic Island Arc Volcanics
18.30–19.00	Travel to Cluj-Napoca
19.00–21.00	Dinner in Cluj-Napoca
	Accommodation in Cluj-Napoca

Wednesday, September 1, 2010 (Day 5). Travel from Cluj-Napoca to Budapest

08.00–10.00	Field stop 17: Cluj-Napoca: Mineralogical Museum
10.00–12.00	Travel to Valea Drăganului
12.00–13.30	Field stop 18: Valea Lungii Quarry: Upper Cretaceous granodiorites (banatites)
13.30–14.00	Lunch break
14.00–19.00	Travel to Budapest

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Rosso, K.M., Bodnar, R.J. (1995): Microthermometric and Raman spectroscopic detection limits of CO_2 in fluid inclusions and the Raman spectroscopic characterization of CO_2. Geochimica et Cosmochimica Acta, **59**, 3961-3975.

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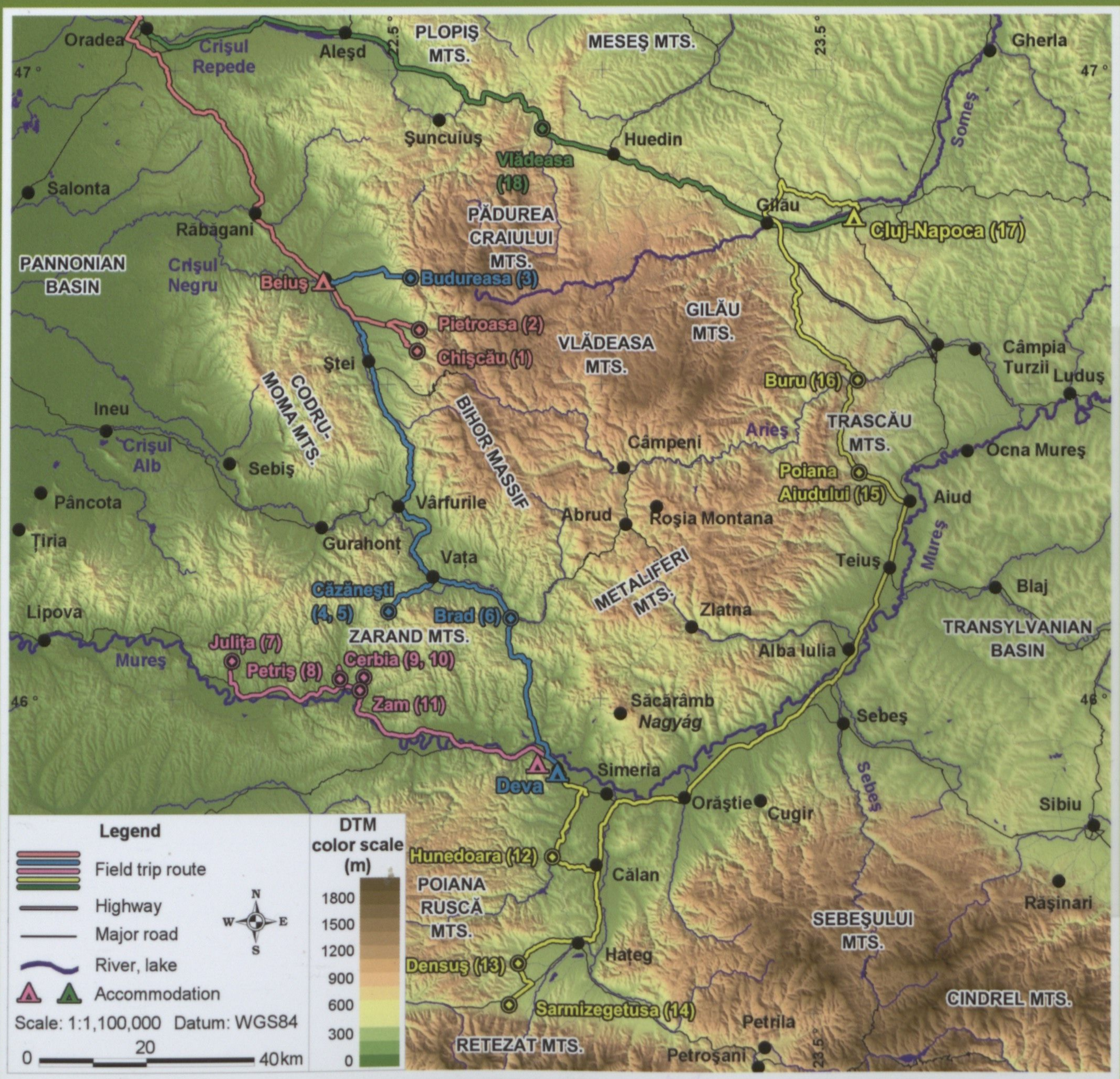

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