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UNIVERSITY OF BUCHAREST
FACULTY OF GEOLOGY
AND GEOPHYSICS



ALCAPA II

"Geological evolution of the Alpine -Carpathian -Pannonian system"

FIELD GUIDEBOOK

**SOUTH CARPATHIANS AND
APUSENI MOUNTAINS**

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PREFACE

The second ALCAPA meeting begins with an excursion in the South Carpathians, Apuseni Mts and Transylvanian Basin, from Bucharest to Covasna-Voinești. After the brilliant presentation of the eastern Central Alps in 1992, it is the turn of Romanian Carpathians to be the battle field. It is our pleasure and honour to be guides on a classic tour, including Iron Gates, Jiu Gorges, Mureș and Arieș Valleys, Olt Valley at Racoș, etc. The first three days in the South Carpathians our attention will be focused nappe structure, from the lowest Danubian to the upper Supragetic unit (Marginal, Outer and Median Dacites); several stops will show tectonic contacts and others will present specific basement or cover formations. In the Apuseni Mts, the fourth day of the trip will present Mureș ophiolites and other igneous products from the Transilvanides belt, while the fifth day will introduce basement formations and tectonic contacts in the Apusenides (Internal Dacites). The sixth day is dedicated to the largest intramontane basin of the Carpathians – Transylvanian Basin – in a west-east cross-section.

This guide-book is mainly a product of earth scientists from Geological Institute of Romania and the Geological-Geophysical Department of Bucharest University. They tried to introduce new-comers in the way Romanian school of geology separates, names, studies and interprets geological formations. Many local names may be difficult to retain, but the book will be at hand for everyone and we hope that at least the main divisions will not be forgotten. As regards our interpretations, it is the role of such trips to present outcrops and cross-sections and to wait for a torrent of completely different views. The issue of these discussions will be known after the session in Covasna-Voinești, but even now it is sure that the benefit will be mutual, for guides as well as for visitors.

As editor of this guide, I want to thank to Prof. F. Neubauer from Salzburg for deciding to give us this chance and for the financial support offered, to the Geological Institute of Romania for editing and printing it, and to my colleagues contributors and to the editorial staff for the hard work of this hot summer.

Bucharest, end of August, 1994

T. Berza



OVERVIEW ON ROMANIAN GEOLOGY

by

Mircea Săndulescu

Geological Institute of Romania

Romania's geology is dominated by the alpine Carpathian Folded Belt (Orogen) which covers more than half of the territory. The foreland of this orogen includes several platforms and also the Cimmerian North Dobrogea Orogen (Fig. 1).

Carpathian Foreland

The oldest cratonic unit known in the Carpathian Foreland is the East European Platform, represented by its podolo-moldavian sector (the Moldavian Platform). The Lower Proterozoic mesometamorphic basement of the platform, intruded by gabbros, anorthosites and granites, is covered by sedimentary formations which show several sedimentary cycles (Vendian-Cambrian, Ordovician-Silurian, Devonian, Upper Jurassic-Cretaceous, Eocene and Neogene). The platform is fractured by several trans-crustal faults, those situated on its western border belonging to the Tornquist-Teisseyre Fault Zone (Trans-European Lineament).

The Scythian Platform develops south of the East European one, extending from eastern Romania through the north Crimea, north of the Great Caucasus; toward north-west it prolongates below the East Carpathians flysch nappes. The folded basement is of Caledonian-Hercynian age, involving old metamorphic rocks and Paleozoic ankimetamorphic and sedimentary formations. The Platform cover includes Upper Devonian, Carboniferous, Permian, Triassic and Jurassic rocks. The Neogene formations are common with the whole Carpathian Foreland. The boundary with the East European Platform is, east of the Siret River, a trans-crustal fault (or fault system); toward north-west the contact is more complex, being represented by the Torquist-Teisseyre Zone and also supposed thrustings with eastern vergency.

The Moesian Platform is situated in the southern part of the Carpathian Foreland, separated from the Scythian Platform by the North Dobrogea Orogen. The basement is constituted by Precambrian metamorphic rocks (cata- and/or mesometamorphics of Lower and Middle Proterozoic age, ankimetamorphic of turbiditic type, named the "Green Schists", of Vendian-Lower Cambrian (?) age). The Platform cover is thinner and with more gaps in Dobrogea, but it develops toward west in important sedimentary basins in Paleozoic (Cambrian-Carboniferous) and Mesozoic (Permian+Triassic-Cretaceous), reaching in some depressions more than 6 km in thickness. Within the specific features of the Moesian Basin are to be stressed out: the quartzitic developments in Cambrian and Ordovician, the Graptolithic shales of Silurian, the carbonatic platforms in Upper Devonian (with evaporites also), Middle Triassic and Upper Jurassic-Lower Cretaceous, the Keuper-type development of the Upper Triassic (locally also with salt), the chalk-type development of a part of the Senonian. Taphrogenic bimodal magmatism (mafic and acid lava flows) is known in Permian and Triassic, generated in an extensive tectonic moment. Several important trans-crustal fractures are known. The Peceneaga-Camena Fault, representing the north-eastern border of the Platform, is the southernmost segment of the Tornquist-Teisseyre Lineament. The Capidava-Ovidiu Fault, south of the previous one, separates the Central Dobrogea from the South Dobrogea. The Intra-Moesian Fault divides the Platform into two panels (Dobrogean and Wallachian) which partially drifted independently during the Carpathian deformations.

The North Dobrogea Orogen (NDO) is a segment, the westernmost one, of a Cimmerian Folded Belt which extends eastwards in South Crimea and further joins the Asian Cimmerides. The NDO shows a nappe structure with north-east vergency (toward the Scythian Platform); three nappes are recognised: Măcin, Niculițel and Tulcea (from SW toward the foreland) (Fig. 2). The Măcin and Tulcea nappes involve continental crust (Precambrian and/or Lower Paleozoic metamorphic formations and granites) and Middle Paleozoic-Jurassic sedimentary (or ankimetamorphic) formations, while the Niculițel



Fig. 1

Fig. 1 - Tectonic Sketch of Romania, Carpathian Foreland: 1 - East European Platform, 2 - Scythian (Sy) and Moesian (Mo) platforms; 3 - North Dobrogea Orogen Carpathians; 4 - Inner Dacides, 5 - Transylvanides, 6 - Pienides, 7 - Median Dacides, 8 - Outer Dacides, 9 - Marginal Dacides, 10 - Moldavides, 11 - Post-tectogenic covers, 12 - Neogene Molasse depressions and Foredeep, 13 - Upper Cretaceous - Paleocene magmatic arcs, 14 - Neogene magmatic arcs, 15 - thrust-sheets, 16 - faults.

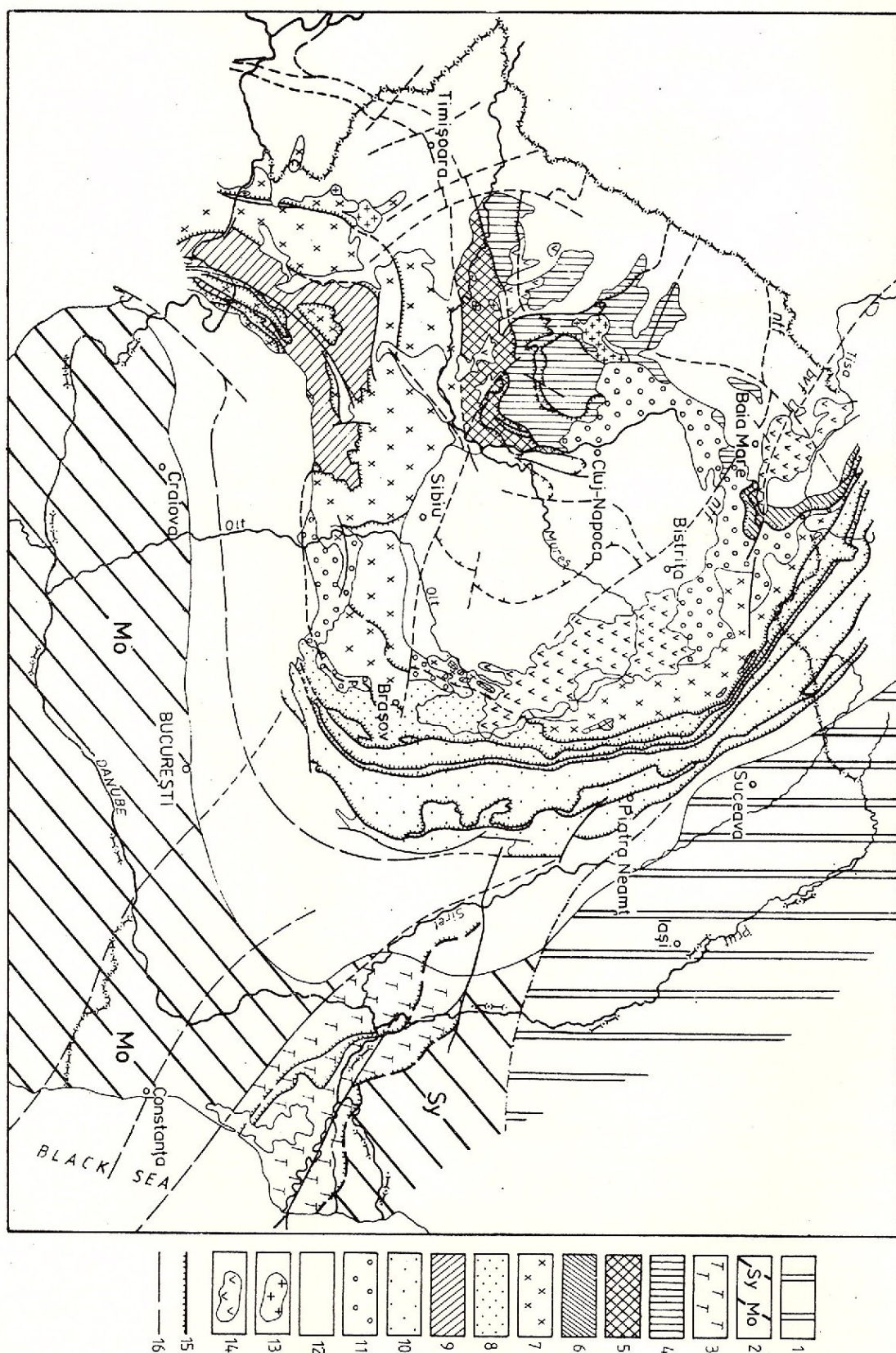


Fig. 2 - Tectonic Sketch of the North Dobrogea Orogen. EE - East European Platform, Sy - Scythian Platform, Mo - Moesian Platform, M+N+T - North Dobrogea Orogen (M - Măcin Nappe, N - Niculițel Nappe, T - Tulcea Nappe), FPC - Peceneaga-Camena Fault.

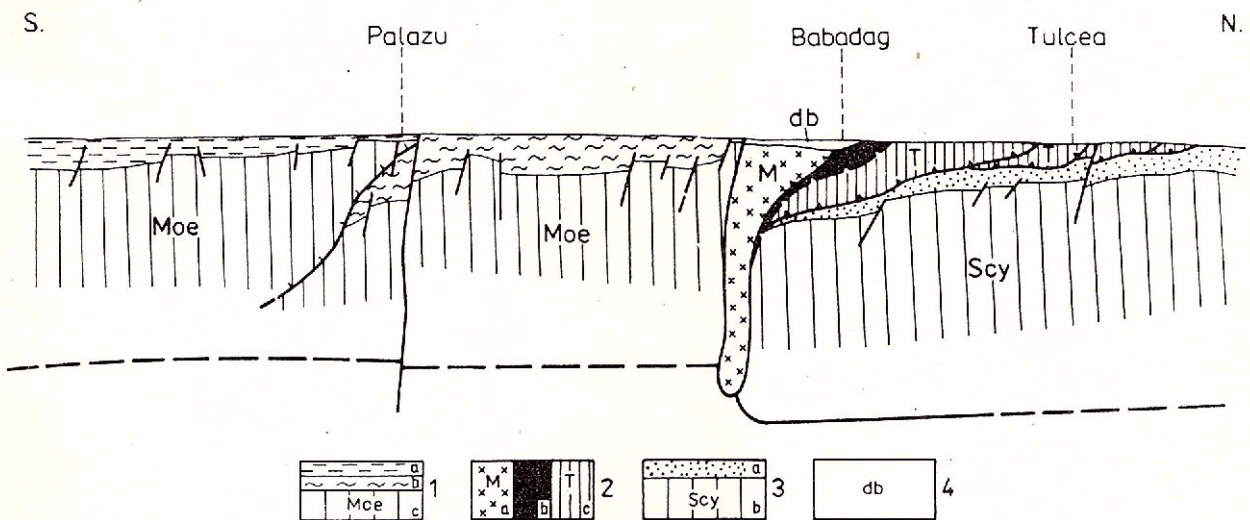
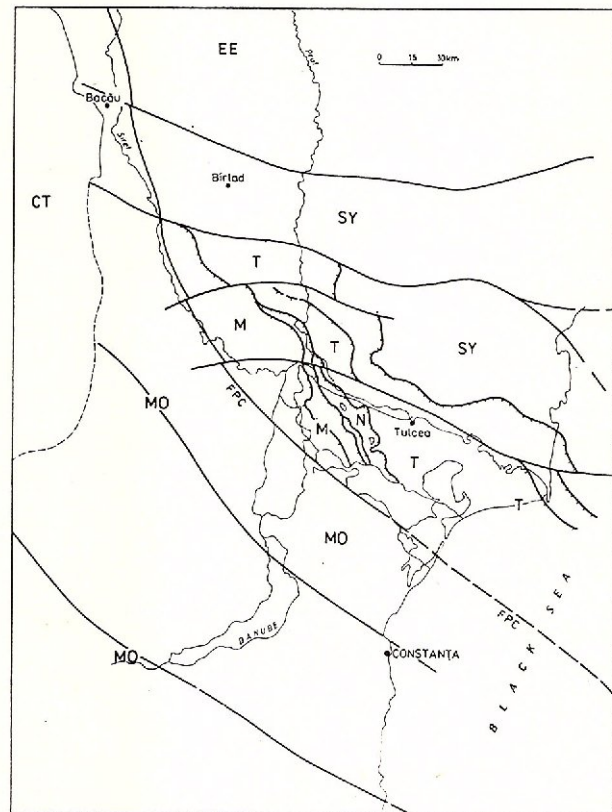


Fig. 3 - Schematic cross-section through Dobrogea (acc. Visarion et al., 1993, simplified), 1 - Moesian Platform (a. basement, b. "Green Schists, c. Paleozoic - Mesozoic formations), 2 - North Dobrogea Orogen (a. Măcin Nappe, b. Niculițel Nappe, c. Tulcea Nappe), 3 - Scythian Platform (a. basement, b. Up. Paleozoic - Mesozoic formations), 4 - Babadag Synclinoria.

Nappe, with a median position between the two others (Fig. 3), plays the role of a "suture", built up of a Spathian-Carnian mafic complex (within-plate basalts and pillow-lavas with pelagic carbonate intercalations), followed by a Norian-Lower Jurassic flysch.

The North Dobrogea Orogen is issued from an Early Mesozoic intra-plate rift which extended from the Early Triassic since the Upper Triassic, allowing in its central part the development of a thinned crust and mafic flows and intrusions. The compressive moments deformed the North Dobrogea Rift (as well as

its correspondent, the South Crimea) in Early Jurassic and in End-Jurassic tectogeneses (Early and Late Cimmerian tectogeneses). An active role in the compressive history of the North Dobrogea Orogen is due to the anticlock rotation of the Moesian Platform in Jurassic time determined by the extension of the Outer Dacidian Rift (see further).

The structure of the NDO prolongates off-shore into the Black Sea continental plateau, as well as the Moesian and Scythian ones. The post-tectogenetic cover starts there with Albian formations followed by Upper Cretaceous ones, the latter cropping out on-land also (Babadag Syncline). North-west of the Danube, the NDO prolongates below a thin Neogene cover; it is known there as the North Dobrogea Promontory. It squeezes out along a major fault – the Trotus Fault – which was probably the fault which limited toward north-west the North Dobrogea Rift during its Triassic extension.

Carpathian Folded Belt (Orogen)

The Carpathian Orogen is a segment of the Tethyan Chain which joins toward west the Alps and toward south the Balkan and Rhodopes. The Carpathian folded area includes deformed remnants of the oceanic crust bearing Tethys and its strongly deformed continental margins. The orogen is the result of several tectogenetic moments in Cretaceous (the inner zones) and Miocene (the outer zones). Post-tectogenetic covers develop above deformed units of the inner zones in Upper Cretaceous and/or Paleogene. Two Neogene molassic depressions (Transylvanian and Pannonian) overlie important parts of the inner zones and their post-tectogenetic covers. A Neogene molassic Foredeep is situated along the outer margin of the Orogen. Two calc-alkaline magmatic arcs, one Senonian-Paleocene, the younger Neogene, are connected with important subduction processes.

The Main Tethyan Suture Zone (MTS) group together tectonic units constituted of ophiolitic complexes and sedimentary formations, which both proceed from the Tethyan oceanic crust bearing area. It prolongates north of the Danube from the Vardar Suture Zone and, passing below the south-east corner of the Pannonian Depression, crops out in the Southern Apusenides ("Metaliferous Mountains"). There, north vergent units are constituted of Jurassic ophiolitic complexes and Upper Jurassic and Cretaceous sedimentary formations issued from a complex paleogeography (calcareous platforms, pelagic rocks, flysch formations). These units prolongate below the Neogene Transylvanian Depression (and some Paleogene post-tectogenetic formations)(Fig. 4). There they represent only a part of the MTS; the other part is represented by the "root zone" of the Transylvanian nappes, obducted from there toward east above the Central East Carpathians units (the proximal European continental margin area in respect with the MTS). The Transylvanian nappes are constituted of ophiolitic complexes (serpentinites, pillow-lavas etc.) of different ages (following the different moments of the Tethyan spreading processes), extending from Middle Triassic to Middle/Upper Jurassic. They are overlain by sedimentary piles (mostly limestones, but starting with radiolarites or cherty limestones). The youngest sedimentary levels known in some Transylvanian nappes are of Barremian (-Lower Aptian ?) age. The MTS units described above constitute the Transylvanides, structured during the Cretaceous. North of the Transylvanian Depression the MTS is represented in outcrops by the Pienides, a group of units which relay "en echelon" the Transylvanides and have recorded Cretaceous, as well as Lower Miocene, tectogeneses. On Romania's territory, the Pienides consist of Botiza Nappe with the Poiana Botizei Klippen Zone in its frontal part (south-east prolongation of the Pieniny Klippen Belt), the Petrova and Leordina nappes (equivalent of the Măgura Nappe). In Poiana Botizei Klippen Zone a Middle Jurassic-Upper Cretaceous pelagic succession (radiolarites/cherty and Calpionella-bearing limestones/dark shales/"couches rouges") is known. In Petrova and Leordina nappes a Maastrichtian-Paleocene flysch develops ("Inoceranian Beds"-type); well developed Paleogene flysch formations are known in all Pienidian nappes. The actual general structural shape of the Pienides is due to the Lower Miocene (Burdigalian) tectogeneses, when the nappes were overthrust above the post-tectogenetic cover (neoautochthon) of the Median Dacides (Central East Carpathians nappes, namely). The traces of the Cretaceous tectogeneses are visible only in the Poiana Botizei Klippen Zone (where formations of this age were preserved). The Lower Miocene transport of the Pienidian nappes was directed and/or accentuated by several important fractures with strike-slip components, mainly the North Transylvanian Fault (ntf) and the Bogdan Vodă Fault (bvf) (Fig. 1). Consequently, escape tectonics processes developed.



Fig. 4

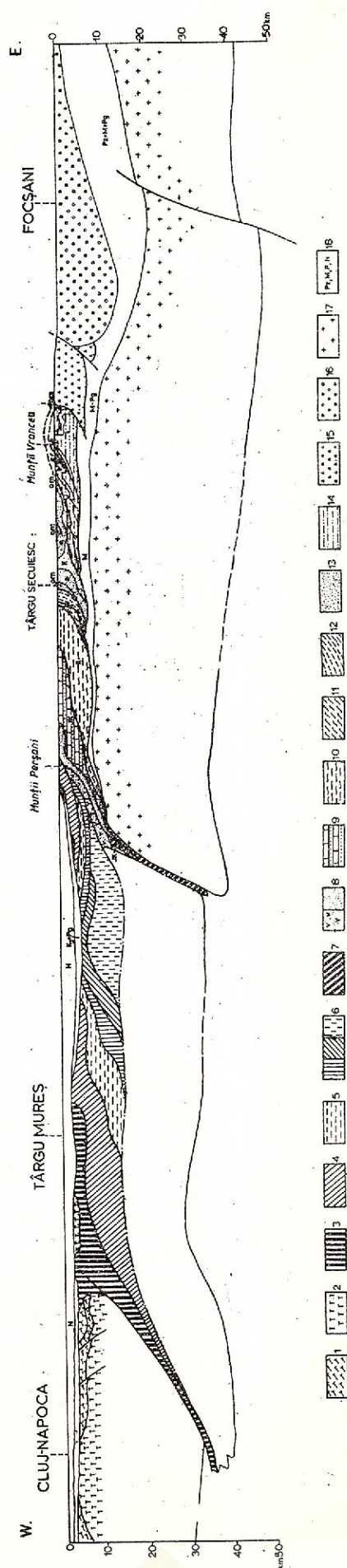


Fig. 4 - General cross-section through the Romanian Carpathians (acc. to Săndulescu, 1984). Inner Dacides (1+2): 1 - Codru-Arșeni nappe system, 2 - Bihor Unit; Main Tethyan Suture: 3 - Transylvanides; Median Dacides (4-6): 4 - Bucovinian Nappe, 5 - Subbucovinian Nappe, 6 - Infrabucovinian nappes; Outer Dacides (8+9): 8 - Black Flysch Nappe, 9 - Ceahlău Nappe; Moldavides (10-15): 10 - Convolute Flysch, 11 - Macia Nappe, 12 - Audia Nappe, 13 - Tarcău Nappe, 14 - Marginal Folds Nappe, 15 - Subcarpathians Nappe; Foredeep: 16 - Focșani Depression; Underthrust elements (17+18): 17 - Crystalline basement, 18 - Sedimentary formations (Pz-Paleozoic, Mz-Mesozoic, Pg-Paleogene).

The Inner Dacides (ID) which are part of the Forcapulian Block (Figs. 4, 5) are situated west and north of the salient bended MTS. In Romania this group of units crops out in the Apuseni Mts (Northern Apusenides); it was also reached by drillings below the Pannonian Depression and the north-western Transylvania. The ID consist of several north and north-east vergent nappes constituted of metamorphic rocks (and granites locally) and sedimentary formations. The metamorphic series are pre-Cambrian and Paleozoic. The sedimentary succession starts (in the southern units) with molassic Upper Carboniferous and, mostly, Permian. The orthoquartzitic Lower Triassic is followed by a carbonatic Triassic sequence, more shallow-water toward north (tectonically deeper units) (where the Upper Triassic is missing) and progressively deeper-water toward south (higher units), the southernmost one showing a Hallstadt-type lithofacies. The Lower Jurassic (paralic in the north, deep-water in the south) follows the model. Middle Jurassic (condensed carbonatic rocks), Upper Jurassic (carbonatic platform) and Cretaceous (pre-Senonian) (bauxites followed by neritic-urgonian-type limestones, marls and turbidites) formations were preserved from erosion only in the northern units; nevertheless, a turbiditic Tithonian-Neocomian sequence is known in the southern one. The overthrusting processes are (Upper ?) Turonian in age. The structured and partly eroded nappes were overlapped by a Senonian post-tectogenetic cover, showing the Gosau lithofacies. With a stratigraphic gap it continues in some areas with Paleogene epicontinental formations.

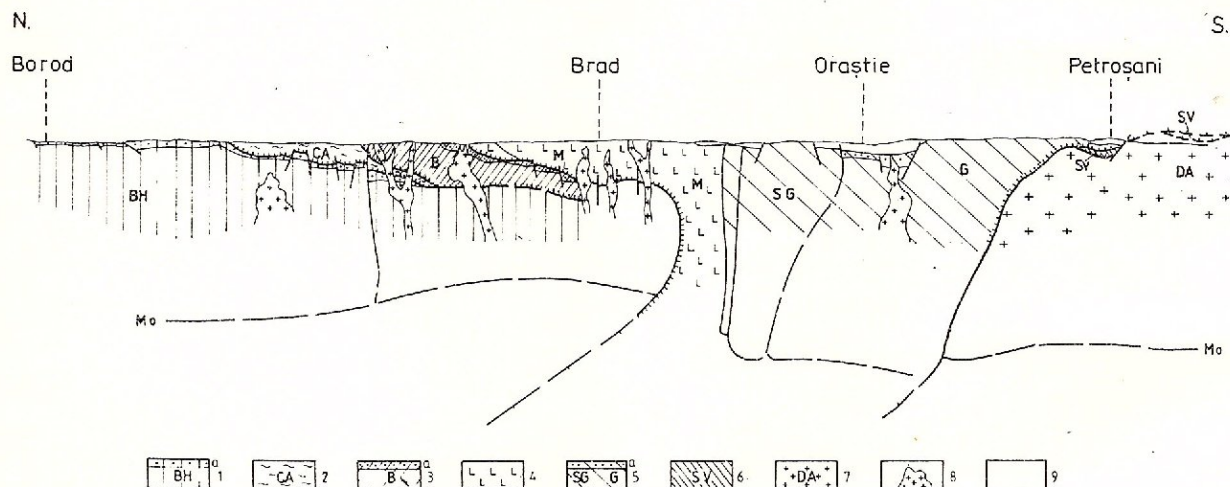


Fig. 5 - Structural cross-section through the Apuseni Mts and the South Carpathians. 1 - Bihor Unit (a - sedimentary formations), 2 - Codru-Arieseni nappe system, 3 - Biharia nappe system (a - Gosau Formation), 4 - Transylvanides (Main Tethyan Suture), 5 - Getic and Supragetic nappes (a - sedimentary formations), 6 - Severin Nappe, 7 - Danubian, 8 - Severin-Palococene calcalcaline intrusions (Banatites), 9 - Post-tectogenetic covers.

The ID/MTS tectonic relationships are complex: while in the southern sector the Transylvanides are "back-thrust", in the Senonian, above the Inner Dacides, in the northern sector the ID thrust the Pienides, in the Lower Miocene. This important changing is determined by the transcrustal fault (ntf) which, crossing the MTS and its continental borders, allowed opposite and composite translations of different panels. The north vergent ID/MTS tectonic relationships are supported by drillings and geophysics in Romania/Ukraine boundary area.

The Median Dacides (MD) are situated on the opposite side of the MTS in respect with the ID. The MD units crop out in the Central East Carpathians and an important part of the South Carpathians; they proceed from the nearest strip, toward the MTS, of the European continental margin. These units are basement-shearing nappes each of them involving metamorphic rocks and their sedimentary envelopes. In the Central East Carpathians the nappes are (upside/downside): Bucovinian, Subbucovinian and Infrabucovinian nappes; the latest correspond to the Getic Nappe (Domain), the former two to the Supragetic nappes, of the South Carpathians. Within the metamorphic formations predominate the mesometamorphic series with a complex premetamorphic composition and a polymetamorphic pre-Cambrian and Paleozoic history. The epimetamorphic series proceeds from terrigenous or volcano-sedimentary formations (Lower and Middle Paleozoic); their metamorphism can be Caledonian and/or Hercynian.

The MD sedimentary formations show several sequences which are more or less expressed sedimentary cycles: Upper Carboniferous and/or Permian molasses (locally developed), quartzitic (late) Lower Triassic

Fig. 6

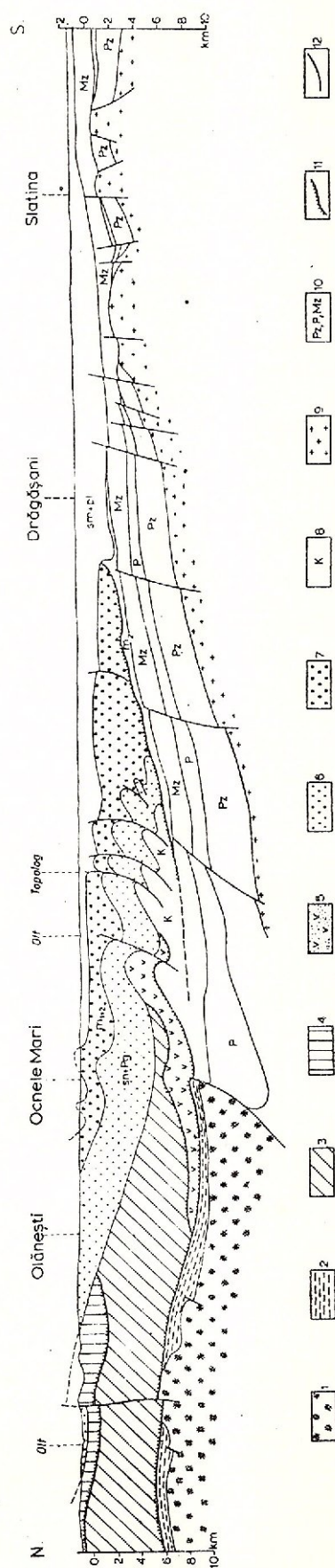


Fig. 6 - Geological cross-section through the South Carpathians and the Moesian Platform (along the Olt Valley).
 1 - Danubian crystalline, 2 - Danubian sedimentary formations, 3 - Getic Nappe, 4 - Supraretic nappes, 5 - Senonian-Paleogene, 6 - Miocene, 7 - undivided Cretaceous, 8 - basement of Moesian Platform, 9 - sedimentary cover of the Moesian Platform, 10 - overthrusts, 11 - thrusts, 12 - Neogene molasses of the Foredeep and Moesian Platform.

followed by carbonatic Middle Triassic, detrital Lower Jurassic, sandy-marl, Middle Jurassic (ending locally with radiolarites), neritic or pelagic calcareous Upper Jurassic-Neocomian. With the Lower Cretaceous (wildflysch in Bucovinian, calcareous in Infrabucovinian nappes) the MD succession ends in the Central East Carpathian nappes, there the Upper Cretaceous (molassic Cenomanian, almost marly Turonian-Senonian) representing the post-tectogenetic (post-nappe) cover. In the South Carpathians the Lower Cretaceous is mostly calcareous, ending with glauconitic Albian; the Upper Cretaceous rocks are molassic in the lower part, followed by marly-sandy formations and ending with turbiditic or volcano-sedimentary sequences. This Upper Cretaceous rocks seal some Mid-Cretaceous compressive structures, but are involved also in End-Cretaceous deformations (Fig. 6).

The post-tectogenetic cover of the East Carpathians MD are preserved in some sunk areas (gulfs) on their western slope, partly being covered by the eastern parts of the Transylvanian Depression and the East Carpathian volcanic arc. The Upper Cretaceous formations are there followed by Lutetian molasses, Priabonian limestones, marls or flysch and Oligocene-Lower Miocene, alternating, pelitic (partly bituminous) and arenitic sequences. The post-tectogenetic cover of the South Carpathians MD starts with Paleogene molasses, followed on their southern slope by Oligocene marly-sandy formations and Lower Miocene conglomerates.

The *O u t e r n D a c i d e s* group together a strip of units which proceed from a Jurassic-Lower Cretaceous paleorift developed within the European continental margin. In the East Carpathians (Black Flysch, Baraolt and Ceahlău nappes) they are built up of Jurassic within-plate basalts and/or Tithonian-Lower Cretaceous flysch formations (locally with conglomerates in the Upper Aptian and Albian); some marly Upper Cretaceous formations are known in the Ceahlău Nappe. In the South Carpathians (Severin Nappe) Jurassic ophiolites are followed by Tithonian-Lower Cretaceous flysch. The Outer Dacidian units (nappes) were twice deformed: during the Mid-Cretaceous and End-Cretaceous tectogenetic moments. In the South Carpathians they are "sandwiched" between the Getic Nappe and the Marginal Dacides (Danubicum); in the East Carpathians they are partly overthrust by the MD and run parallel with the external border of these. The Outer Dacides constitute a satellite suture in respect with the MTS.

The *M a r g i n a l D a c i d e s* (MAD) (Danubian) crop out in a huge half-window, below the Getic and Severin nappes (Fig. 1). The major part of the Danubian units of basement shearing type, involving crystalline formations and their sedimentary envelope. Pre-Cambrian mesometamorphic rocks (mostly metaclastics and/or amphibolitic, with serpentinite bodies), with important retromorphic processes, are intruded by numerous granitic, granodioritic and dioritic massifs of Latest pre-Cambrian or Lowermost Cambrian age. Near the Danube a gabbro-peridotitic massif (Iuti) may represent a part of a pre-Cambrian or Paleozoic ophiolitic suture, included in the pre-Alpine basement of the MAD. Paleozoic slightly metamorphic quartzose, conglomeratic sandy and shaly rocks are thin and areally discontinuously developed. The "envelope" formations include Permian and Mesozoic rocks. The Permian molasses (with rhyolitic intercalations) are disconformably followed by Lower Jurassic paralic and/or marine deposits (Gresten lithofacies); the Middle Jurassic is sandy-calcareous while, during the Upper Jurassic and a part of the Lower Cretaceous, carbonatic platforms develop in the external parts and basinal formations in the internal ones. The Albian and Cenomanian show pelagic marly formations followed, during the Senonian and Turonian, by a wildflysch (olistostrome) formation (which includes blocks of very different sizes proceeding both from the Danubian, Severin and Getic domains). In the western part of the MAD, below the Getic Nappe and above the internal Danubian units, the Arjana Unit is situated, a complex nappe constituted of sedimentary and volcano-sedimentary formations (quartzitic sandstones and coarse-grained Lowermost Jurassic/alternating shales, limestones, trachytes and alkaline basalts lavas and tuffs of Lower (?) Middle and Upper Jurassic age/Neocomian limestones/Upper Cretaceous wildflysch). It proceeds from the transitional strip between the Danubian and Severin domains (the boundary strip between the MAD and the Outer Dacidian paleorift).

The actual structure of the MAD is the consequence of two main tectogenetic periods, a Mid-Cretaceous (Aptian-Albian) one and an End-Cretaceous (Senonian) one. Anchimetamorphic events occur in connection with both of them. At the end of the Cretaceous the whole MAD area was tectonically covered by the Getic (together with Severin) Nappe. Erosional processes occur since the earliest Cenozoic. The Paleogene post-tectogenetic covers are common for the MD and MAD.

The *M o l d a v i d e s* are the outermost Carpathian units (Fig. 7). They correspond to a major part of the East Carpathian Flysch Zone (excepting the Outer Dacidian nappes). From inside to outside they are: Convolute Flysch, Macla, Audia, Tarcău, Marginal Folds and Subcarpathian nappes. These nappes are sedimentary allochthonous bodies overthrust progressively above foreland elements. The stratigraphic successions, showing different lithofacies, extend from the Lower Cretaceous up to the Lower Miocene.



During this interval the detrital rocks were supplied by two main sources: an external source situated in the foreland and an internal one represented by "cordilleras" or, mostly in the Cenozoic, by the still structured internal units of the East Carpathians. The highly subsiding trough migrated from inside to outside since the Lower Cretaceous (Convolute Flysch area) until the Paleogene (Tarcău area) and even Lower Miocene (Subcarpathian area). The Lower Cretaceous formations show two main lithofacies: the external Black Shales development, covering the Audia, Tarcău and Marginal Folds domains, relatively thin (600-900 m) and dominated by euxinic sedimentation and the inner, Convolute Flysch developments, several kilometres thick (5-6 km), developed in a subsiding flysch trough supplied by an inner source area (the Peri-Moldavian Cordillera). In the Convolute Flysch Trough the turbiditic subsiding sedimentation continues up to the Turonian (even Lower Senonian ?); while in the Black Shales domain a condensed variegated very thin sequence, partly deposited below the CCD level, sedimented in the Upper Vraconian-Lower Senonian time span. In the Senonian time the main flysch lithofacies migrated toward the exterior, the more specific ones being known in the Tarcău domain (calcareous flysch); toward the interior sandy flysch developed (Audia domain) while the external lithofacies are of pelagic nature. The highest subsiding flysch area migrated once more

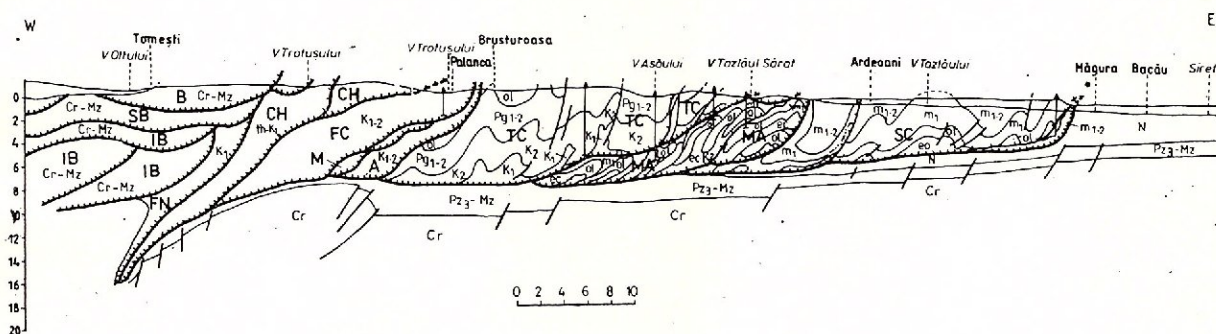


Fig. 7 – Geological cross-section through the East Carpathians Flysch Zone (in central Moldavia). B – Bucovinian Nappe, SB – Subbucovinian Nappe, IB – Infrabucovinian Nappe, FN – Black Flysch Nappe, CH – Ceahlău Nappe, FC – Convolute Flysch Nappe, M – Macia Nappe, A – Audia Nappe, TC – Tarcău Nappe, MA – Marginal Folds Nappe, SC – Subcarpathian Nappe, Cr – crystalline formations, Pz-3 – Upper Paleozoic, Mz – Mesozoic, N – Neogene, TH – Tithonian, k₁ – Lower Cretaceous, k₂ – Upper Cretaceous, Pg₁ – Paleocene, Pg₂ – Eocene, ol – Oligocene, m₁ – Lower Miocene, m₂ – Middle Miocene.

in the Paleogene (Tarcău domain) determining the development of many lithofacies, from the proximal sandy one in the inner part, to the distal shaly-calcareous one in the external part. The double source areas, in foreland and hinterland, were still existing. During the Oligocene and Early Miocene a specific bituminous-quartzitic sandy lithofacies develops in the external part of the Moldavides (external Tarcău, Marginal Folds and Subcarpathian nappes); synchronously a flysch development is known in the inner part of the Tarcău Nappe.

Following an evaporitic event (salt and gypsum) of (Middle) Burdigalian age the Lower and Middle Miocene molassic formations start (Tarcău, Marginal Folds and Subcarpathian nappes), which are included in the thrust-sheets. The post-nappes molasses are developed in the Foredeep.

The main tectogenetic moments which structured the Moldavides are of Burdigalian, Badenian and Sarmatian age. Some precursory foldings are recorded in the Convolute Flysch and Audia nappes during the End-Cretaceous time.

The Foredeep is filled with Upper Miocene-Pliocene-Lowermost Pleistocene molasses entirely supplied by the deformed rising Carpathians. They cover the most external parts of the East and South Carpathians and a part of the neighbouring platforms. In the Carpathian bending area and in the southern Subcarpathians the inner part of the Foredeep is folded (Plio-Quaternary deformations). The most subsiding segment of the Foredeep is situated in front of the Carpathian Bend (the Focşani Depression) where about 10 km of Upper Miocene-Lowermost Pleistocene molasses accumulated.

The Transylvanian and Pannonian, Neogene, molassic depressions develop above the Inner Carpathians deformed units and their post-tectogenetic covers. There are Middle and Upper Miocene

and Pliocene formations, sandy or/and coarse grained, partly in schlier facies, with an evaporitic level in the Lower Badenian (salt layer in the Transylvanian Depression). Volcanic tuffs develop at different levels.

The magmatic activity during the Mesozoic and Cenozoic, in the Carpathian area, may be summarized as follows:

1. Ophiolitic complexes developed in the oceanic crust bearing Tethys (preserved in the Transylvanides=MTS) during the Middle Triassic-Upper Jurassic; Jurassic ophiolites are known in a part of the Outer Dacidian paleorift (preserved in the Severin Nappe), synchronous with within-plate basalts, in the same paleorift (East Carpathians Outer Dacides).
2. Alkaline magmatism of Jurassic age developed in the extensive margins of the Outer Dacidian paleorift (in the frontal part of the Median Dacides and the innermost part of the Danubian (Marginal) Dacides).
3. Calc-alkaline magmatism developed during the compressive history of the Carpathians, connected with subduction paleoplanes. Two main periods were documented: Senonian-Paleocene in the South Carpathians (Getic and Supragetic areas) and the Apuseni Mts (North Apuseni=Inner Dacides) and Neogene in the East Carpathians and Apuseni Mts. The first period is predominantly intrusive with a few extrusives structures preserved; the second one shows an important volcanic arc in the East Carpathians (with some outcropping subvolcanic and/or hypabyssal bodies) and a smaller arc in the Apuseni Mts.
4. Intracontinental basalts of Plio-Quaternary age (Perșani Mts and Mureș Valley) are connected to deep (transcrustal) fractures.

Geotectonic History

The End-Proterozoic (Panafrican) cratonisation is recognised in the whole Carpathians foreland and, as relics, in the whole Carpathian Orogen. This huge cratonic area, preserved actually in the East European Platform was split, south and west of the former, within Paleozoic mobile areas. The Scythian Platform proceeds from one of them, the Paleozoic metamorphic series of the Carpathians from another branch. Within the last ones, remnants of a Paleozoic oceanic crust bearing domain seem to be acceptable. A second large cratonisation occurs after the Lower Carboniferous, including the Carpathian area and its foreland.

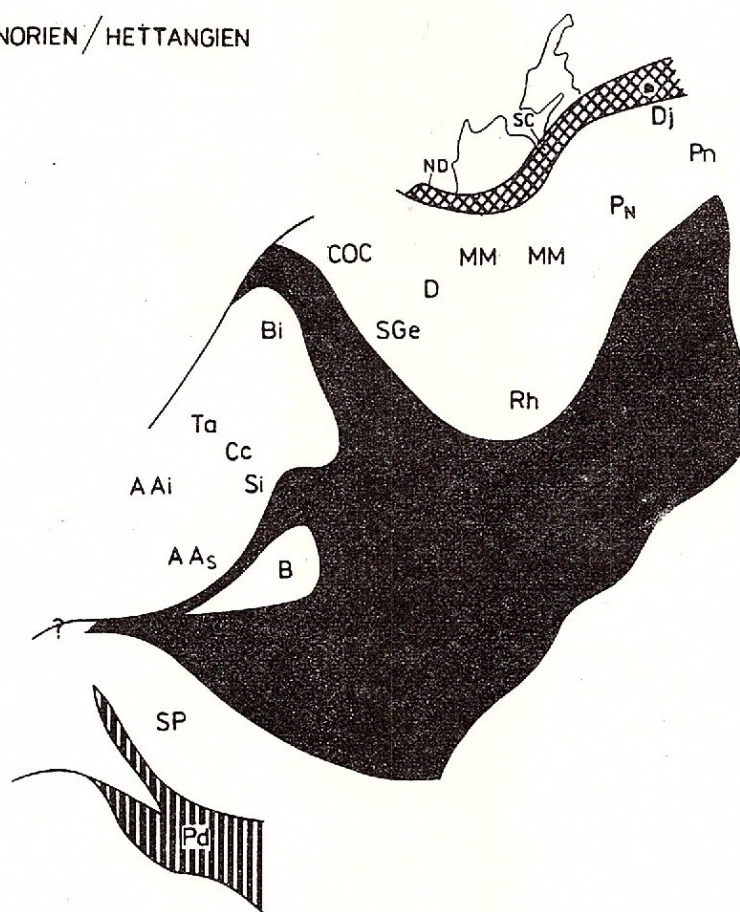
The earliest Mesozoic riftings occur in the North Dobrogea-South Crimea aulacogene, a possible pull-apart extensive structure connected with the strike-slip movements of the Tornquist-Teyssere Lineament. The rifting processes are relevant since the Spathian and continue during the whole Triassic.

The Tethyan oceanic spreading starts in the Middle Triassic (the oldest ophiolites preceeding from the Transylvanian Domain) separating the European continental margin from the Forcapulian Block (Fig. 8). The processes continue during the Jurassic propagating into the Pienidian Domain (Fig. 8). The opening of the Tethys is followed by the compressional period of North Dobrogea, as a result of the north-east rotation movement of the Moesian block. This movement is also connected with the rifting processes in the Outer Dacidian area. At the Middle/Upper Jurassic boundary, the spreading (in the Tethyan oceanic area) and the rifting (in the Outer Dacidian area) reach their maximum sizes. The End-Jurassic compression in the North Dobrogea Orogen is a consequent result.

The earliest important crustal shortenings in the Carpathian area was recorded in latest Tithonian/earliest Berriasian, within the oceanic Tethys. It is contemporary with the overthrust of the Vardarian ophiolites above the Serbo-Pelagonian Massif, but of less amplitude (Fig. 8). It generated subduction to which is connected, above the oceanic crust, a calc-alkaline arc, involved actually in the structure of the Transylvanides (Metalliferous Mts) (Fig. 5). The compressional processes involved the European continental margin (Median Dacides) as well as a part of the oceanic domain (Transylvanides and the obducted Transylvanian nappes), in the Mid-Cretaceous time (Fig. 9). Since the Lowermost Cretaceous the North Dobrogea Orogen was integrated in the stable (cratonic) Carpathian Foreland. End-Turonian (Pre-Gosau) compressional events affected the Apulian Block (Inner Dacides) generating their actual structure. The End-Cretaceous determined the final closing of the oceanic Tethys in the Transylvanian sector while only partial shortening in the Pienidian one. End-Cretaceous deformations were recorded in the Median (South Carpathians), Outer and Marginal Dacides (Fig. 9).



NORIEN / HETTANGIEN



TITHONIQUE / NÉOCOMIEN

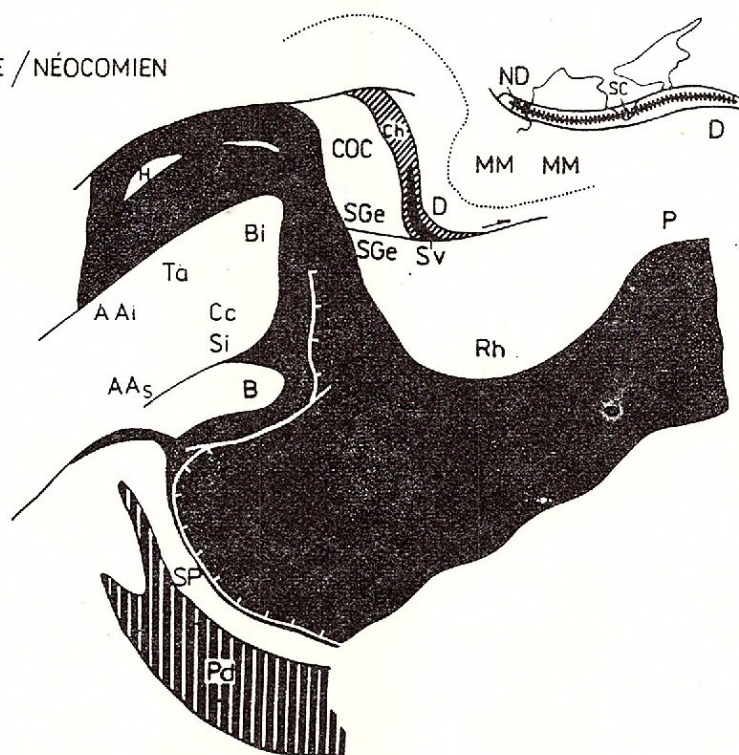
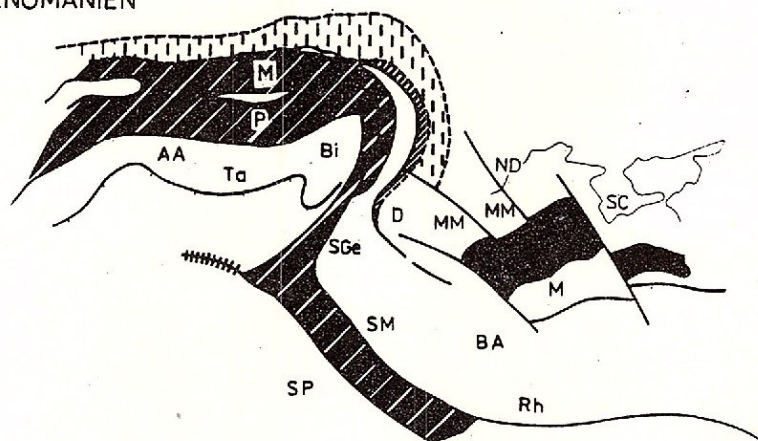
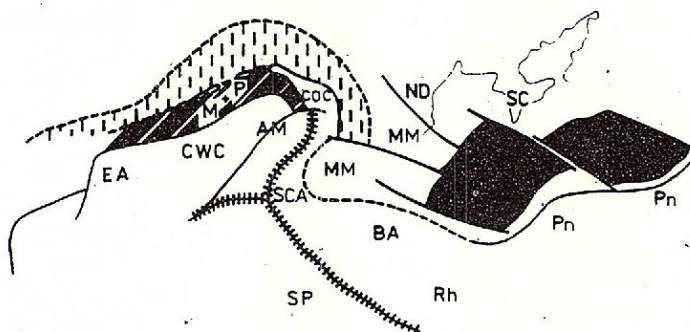


Fig. 8 - Palinspastic sketches of the Carpathians and their Foreland in Triassic and Jurassic (legend similar with the Fig. 9).

CÉNOMANIEN



PALÉOGÈNE



OLIGOCÈNE

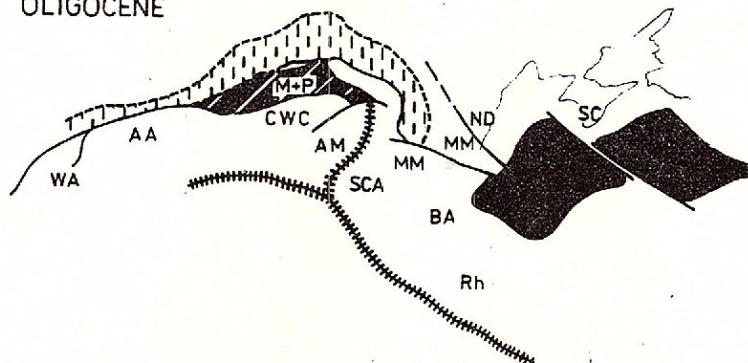


Fig. 9 - Palinspastic sketches of the Carpathians and their Foreland in Cretaceous and Paleogene. 1 - Tethyan oceanic crust, 2 - Thinned and/or oceanic crust (Pindus), 3 - Thinned and oceanic (a) crust (Outer Dacides), 4 - North Dobrogea - South Crimea Cimmerian rift, 5 - Continental crust, 6 - Deformed oceanic crust, 7 - Thinned crust (Moldavides). AA - Austroalpin, AAi - Lower Austroalpin, AAs - Upper Austroalpin, AM - Apuseni Mts, B - Bükk, BA - Balkanides, Bi - Bihor, Cc - Choč, Ch - Ceahlău, COC - Central East Carpathians, CWC - Central West Carpathians, D - Danubian, Dj - Djirula, EA - Eastern Alps, MM - Moesia, M - Măgura, ND - North Dobrogea, P - Pieniny Klippen, Pd - Pindus, Pn - Pontides, Rh - Rhodopes, SC - South Crimea, SCA - South Carpathians, SGe - Getic and Supragetic, Si - Silicicum, Sv - Severin, SP - Serbo-Pelagonian, Ta - Tatrides, WA - Western Alps.

Starting with the Early Paleogene the imobile areas, receiving important flysch sedimentation, remain the Pienidian and the Moldavidian domains (Fig. 9). The Pienidian mobile area was limited toward south by the North Transylvanian Fault which separated it from the yet sutured Transylvanides (Fig. 9). The Moldavides, with important flysch subsidence developed above thinned continental crust. The south-western end of the Moldavidian flysch troughs is connected with the Intra-Moesian Fault; west of it the Paleogene formations represent a small (proto)foredeep of the South Carpathians showing lateral lithofacial changes. The Pienides and the Moldavides will be deformed during the Lower Miocene (Pienides and Inner Moldavides) and the Middle-Upper Miocene (Outer Moldavides). In the inner parts of the Carpathians, deformed during the Cretaceous, develop in Paleogene (mostly Eocene and Oligocene) extensional basins; in some parts (western part of the Central East Carpathians) they start in the Upper Cretaceous (following the Mid-Cretaceous tectogenesis). There post-tectogenetic covers accumulated.

The consumption of the Tethyan oceanic crust as well as that of the Outer Dacidian rift (oceanic or thinned) and Moldavidian thinned crust generated several calc-alkaline arcs of Senonian-Paleocene age (in the North Apuseni and the South Carpathians) and Neogene age (in the Apuseni Mts and along the inner part of the East Carpathians).

The actual twice-bended shape of the Carpathians is the result of a mutual interaction during the Cretaceous and Miocene deformations, of the Moesian block, showing western translations and the Fore-Apulian block, with eastward translations and clockwise rotations. The western wandering of the Moesian Platform is in connection with the "opening" of the western Black Sea, which starts in the Albian and continues during the Upper Cretaceous and in some periods of the Cenozoic.

The recent history of the Carpathians is dominated by the development of the Neogene molasse basins: the Transylvanian and the Pannonian above the inner parts of the still structured Carpathian and the Foredeep along the outer border of the chain and, partially, above the foreland. The Transylvanian and Pannonian depressions develop above two tectonic groups: the strongly deformed Carpathian units and the slight or non-deformed post-tectogenetic basins.

The youngest deformations recorded in the Romanian Carpathians are situated to the outer part of the Carpathian Bend. It is of Lower Pleistocene age (the Wallachian "Phase"). The deformed area is limited by two important transcrustal faults: the Intra-Moesian Fault (with left-lateral Neogene translations) in the west and the Peceneaga-Camena Fault (right-lateral translations) in the north-east. The panel thus delimited moved towards the Carpathian Bend generating the Pleistocene deformations. The high seismic area of Vrancea is situated within this panel.



GEOPHYSICAL FEATURES OF THE ROMANIAN TERRITORY

by

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All kind of available information was taken into account in order to point out the main geophysical characteristics of the Romanian territory.

Gravity

The new Bouguer map of Romania (Fig. 1) offers a nice opportunity for geological connections and advancing in clearing up different problems of the crustal structure of Romania.

Platform regions. The gravity anomaly of Moesian Platform is clearly connected with its structure. The different upheavals/sinkings of the Moesian basement are visible in the gravity by alternance of maximum anomalies, corresponding to lifted blocks (e.g. Balota, Turnu Măgurele, Balș-Craiova) with minimum anomalies, generated by subsided blocks (e.g. Calafat, Alexandria-București). The Moesian Platform is clearly bounded by the isome of 25 mgal.

A particular structure of the gravity field is characteristic of its eastern sector, Central and Southern Dobrogea. The main crustal faults which are recognised into the most geophysical maps, Peceneaga-Camena and Capidava-Ovidiu are easily observable and change the gravity patterns of the two Dobrogean sectors. The subsided Southern Dobrogea is pointed out by values up to 0 mgal, while lifted Central Dobrogea, with the greenschists basement, is characterised by high values up to + 28 mgal. Other geophysical information is going to suggest its prolongation to the NNE but from the Bouguer data it could be considered as ending in front of the Carpathian arc. Northern Dobrogean orogen is shaped by alternancies of maximum and minimum anomalies due to the intricate structure. As an overview, the tectonic lines are concordant with the Dobrogean faults (NNW-SSE).

The Moldavian Platform, also bounded by an isoline (+ 25 mgal), has a complicated structure with a probable basement subsidence around Dolhasca-Hârlău.

Alpine-Carpathian folded regions. The main feature of the folded structures seems to be the very large and extended anomaly of minimum-minimorum, placed more or less over the Carpathian foredeep. The intense negative values (down to -126 mgal) are superposed to whole foredeep, in front of both Eastern and Southern Carpathians. The only break is along the Trotuș Valley, where a jump of the minimum Bouguer anomaly of 30-60 km suggests the importance of the Trotuș line, also clear in some other geophysical data and absolute opaque for geological investigations. It seems to represent the south-western termination of the mobile margin of the East European Platform. The negative isogals are also going to shape the Neogene volcanic chain, from the Oaș to the Harghita Mts.

The Apuseni Mountains, although presenting a NNE-SSW orientated minimum anomaly at their border with Transylvanian Basin, have not a clear foredeep as the Carpathians. The structure of the Apuseni Mts is dominated by the minima of Muntele Mare and Bihor and maxima probably generated by the ophiolitic chain which can be followed on more than 600 km, along the Mures River and under the Transylvanian Basin.

A very complicated distribution of the gravity field seems to be on the Romanian part of Pannonian basin, with a very strong alternancy of maxima and minima gravity anomalies.



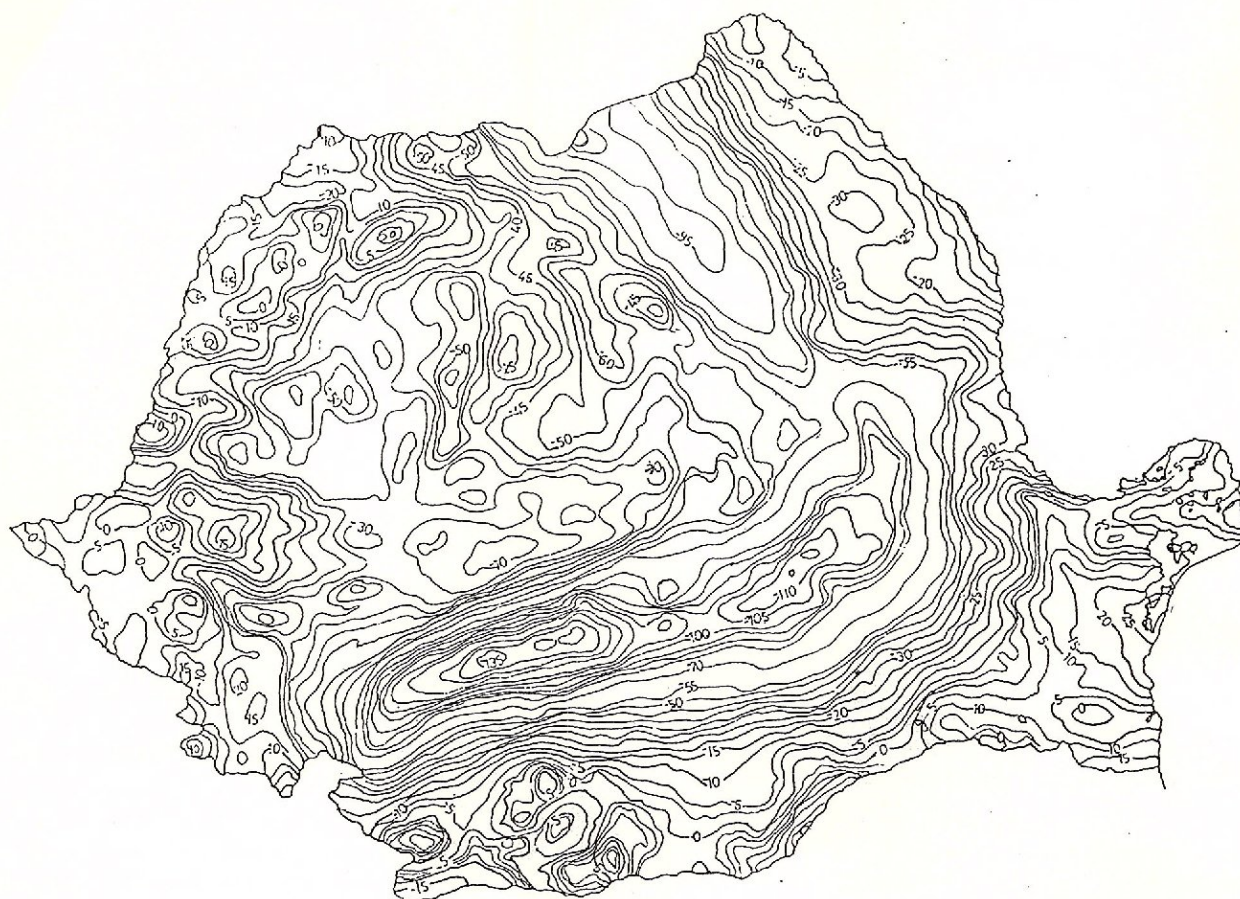


Fig. 1 - Romanian Territory. Scheme of Gravity Field.

Magnetics

Unfortunately, the airmagnetic map of Romania can be used only for limited regions of interest. So, for the general geomagnetic overview only the distribution of ΔZ and ΔZ_a can be considered in order to point out some regional features of the magnetic field.

Platform regions. There are six regional magnetic anomalies, corresponding to metamorphosed basements. There are different both from petrographical and age viewpoint: the magnetic anomalies of the Moesian Platform (Prebaikalian basement), the anomaly of metamorphic complex which also contains rocks of Baikalian age (Krivoi Rog type), the greenschists anomaly that is prolonged to the NW, between the Siret River and the Carpathian flysch (Prebaikalian basement probably rejuvenated during Caledonian), The North Dobrogea anomaly, extended to the South Moldavia, until Birlad Depressin (Hercynian basement), the anomaly of the Moldavian Platform between Siret and Prut rivers (Baikalian basement) and the anomaly of the East European Platform basement, extended on north-east Moldavia and east of the Prut River (Prebaikalian basement).

It is to be pointed out that the Prebaikalian and greenschists basements correspond to areas of dominant negative magnetic values; the Baikalian and Hercynian basements, to areas of dominant positive values.

Most of subregional platform magnetic anomalies were considered as being generated by intrusive mass, mainly intracrystalline and only very few to volumes of metamorphic rocks of high transformation/metamorphism. A particular characteristics is the alignment of subregional anomalies that suggests the position of crustal faults along which the magna advanced to the surface or where important petrological/dynamical local processes were located. Such deep faults are Peceneaga-Camena, Capidava-Ovidiu, Intramoesian Oravița etc.

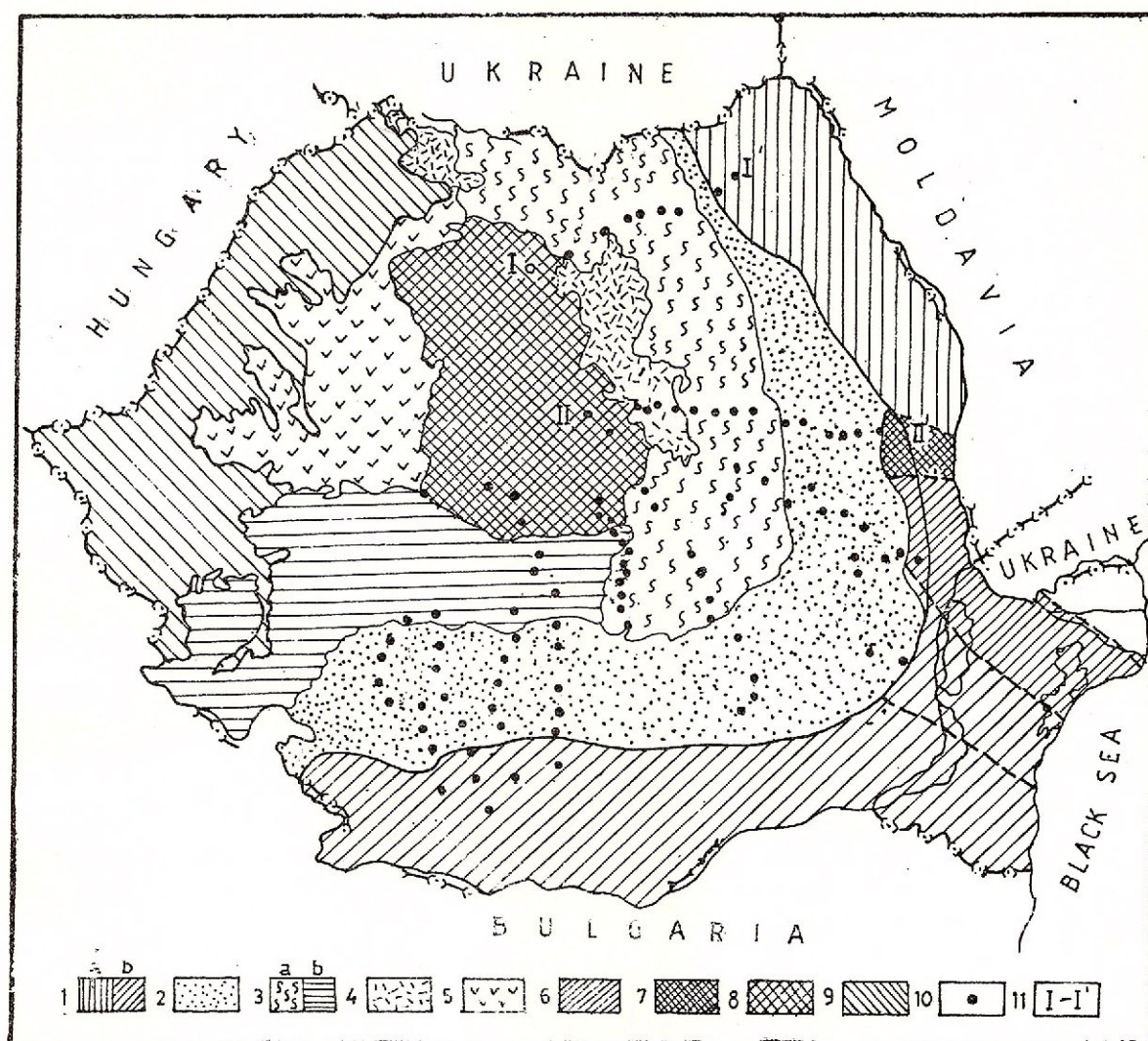


Fig. 2 Romanian Territory. MTS locations. 1a, Moldavian Platform; 1b, Moesian Platform; 2, Carpathian Foredeep; 3a, Eastern Carpathians; 3b, Southern Carpathians; 4, Neogene volcanic rocks; 5, Apuseni Mts.; 6, North Dobrujean Orogenic; 7, Bârlad depression; 8, Transylvanian Depression; 9, Pannonian basin; 10, MTS locations; 11, MTS profiles location.

Local anomalies correspond to sources placed either inside the crystalline basement (as metamorphic facies of high ferromagnetic content) or into the sedimentary cover (layers of andesitic ash, as those from Casin-Oituz or flows of Mesozoic lava, at Bals-Craiova for example).

Alpine-Carpathian folded regions. The metamorphosed basement of this kind of regions represent 2/3 from the whole Romanian territory. It is differentially rejuvenated on the orogenic phases which affected the folded system. As a result, it is strongly broken up and heterogeneous from age and petrographic composition viewpoint. It is considered to be superposed over a unique regional magnetic anomaly, separated by the last one through a only magnetic limit placed near the boundary between the non-rejuvenated extra-Carpathian and the rejuvenated intra-Carpathian basements. The extension of this unique regional magnetic anomaly is dominated by negative magnetic values. Large areas of this values cover most of the Eastern Carpathians, Southern Carpathians and Apuseni Mountains, a large belt on the border of the Transylvanian Basin and the northern part of the Romanian sector of the Pannonian Basin.

One important semi-regional magnetic anomalies is to be mentioned. It is placed into the central part of the Transylvanian Basin. It is the most important subregional magnetic anomaly over the Romanian territory and has a complex morphology of local anomalies as an effect of intrusive mass of different age (Triassic, Jurassic,

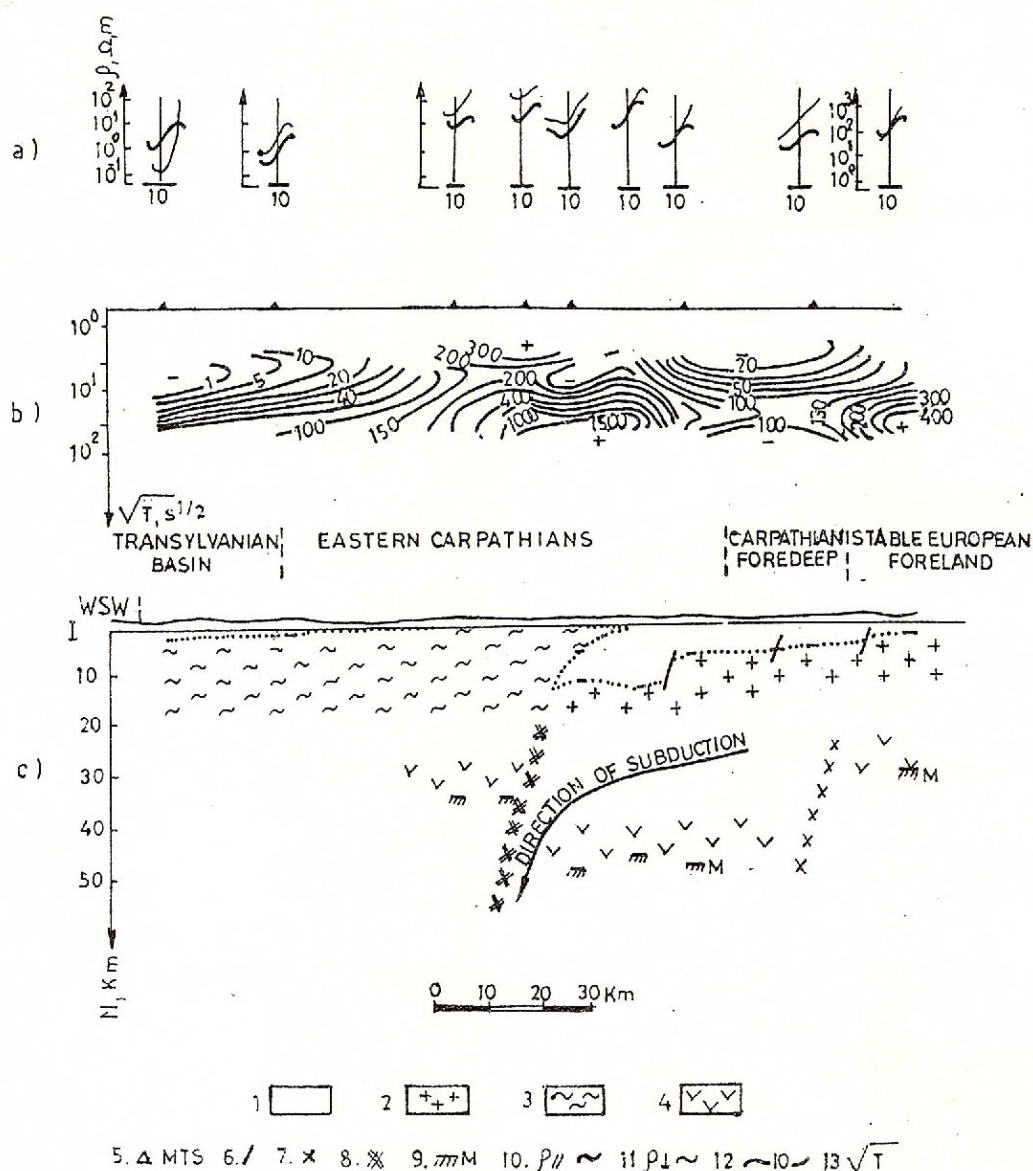


Fig. 3 MTS profile I - P. a) Experimental MTS curves; b) Resistivity pseudosection; c) Deep geological profile: 1, sedimentary cover; 2, Platform type basement; 3, Central-East Carpathian Nappe System; 4, Crust (basalt part); 5, MTS locations; 6, Basement fractures; 7, Crustal fractures; 8, Deep dislocations; 9, M discontinuity; 10, ρ_{\parallel} , E-polarized curves; 11, H-polarized curves; 12, ρ_{\perp} isolines; 13, square root of period (after Stănică et al., 1986).

Paleogene and Neogene). They are placed along dominant alpine or ophiolitic eruption directions from the Apuseni Mountains. It has two main branches to the west: the southern one, along the Mureș River and the northern one, on the direction Cluj Napoca-Huedin-Beiuș. Both lines consist of many local anomalies, as effects of Mesozoic eruptive (on the southern line) and banatitic mass (on the northern line).

Another important semiregional anomaly is placed in Banat. It is also strongly broken up, consisting of very many local anomalies as an effect of Mesozoic, banatitic and Neogene eruptions.

Around both semiregional anomalies, a great number of local magnetic anomalies is distributed, as effect of ophiolitic (Apuseni Mountains and Banat), banatitic (Banat, Apuseni Mountains, Southern Carpathians) and Neogene (Apuseni Mountains, Romanian sector of the Pannonian Basin and Neogene volcanic chain) eruptions.

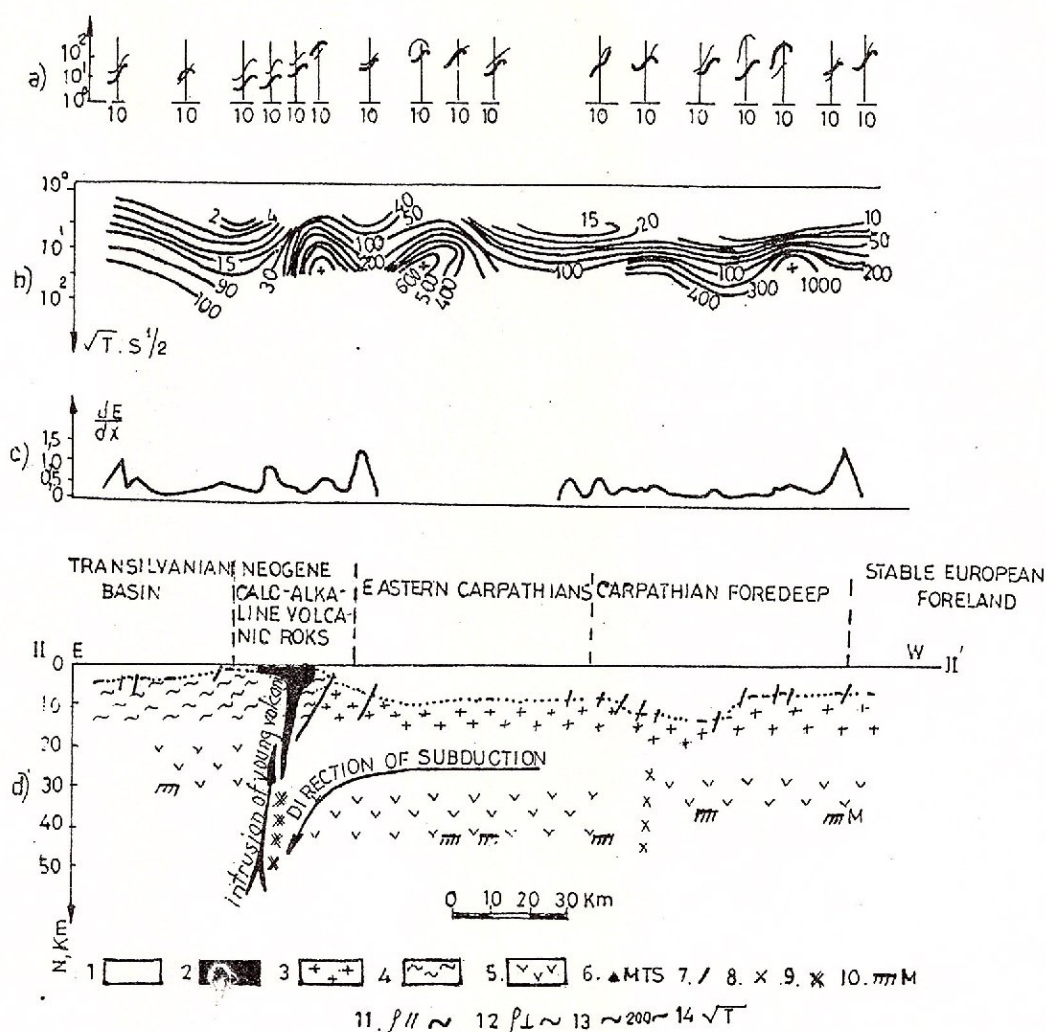


Fig. 4 - MTS profile II - II'. a) Experimental MTS curves; b) Resistivity pseudosection; c) Horizontal gradient of the telluric field; d) Deep geological profile: 1, sedimentary cover; 2, Neogene Volcanic rocks; 3, Platform type basement; 4, Central-East Carpathian Nappe System; 5, Crust (basalt part); 6, MTS locations; 7, Basement fractures; 8, Crustal fractures; 9, Deep dislocations; 10, M discontinuity; 11, $\rho_{II} \sim$, E-polarized curves; 12, $\rho_{\perp} \sim$, H-polarized curves; 13, ρ_{II} isolines; 14, square root of period (after Stănică et al., 1986).

Most of the local anomalies are placed along aphiolitic, banatitic and Neogene eruption alignments. Only very few (as those from the crystalline domain) are emplaced along the mineralization alignments. The most local anomalies from the Neogene volcanic chain are anomalies of minimum. By comparison with the maximum anomalies, they suggest to be a the result of Neogene magnetic reversals. The local magnetic anomalies from the Southern Carpathians correspond to Mesozoic eruptions, where having character of maximum and reverse magnetization, were of minimum. Local anomalies from the Romanian sector of Pannonian Basin are emplaced along the direction of Neogene eruptions from the Apuseni Mountains and the Neogene volcanic chain.

Magnetotellurics

Advanced magnetotelluric studies (Stănică et al., 1981, 1984, 1986) were carried out along certain long profiles crossing the Carpathian chain (Fig. 2). Some of the most important results is extension the known depths to the crystalline basement which was defined by boreholes or other geophysical surveys. The material of the basement is, on the average, more resistant than the sedimentary cover, so that an increased contrast of

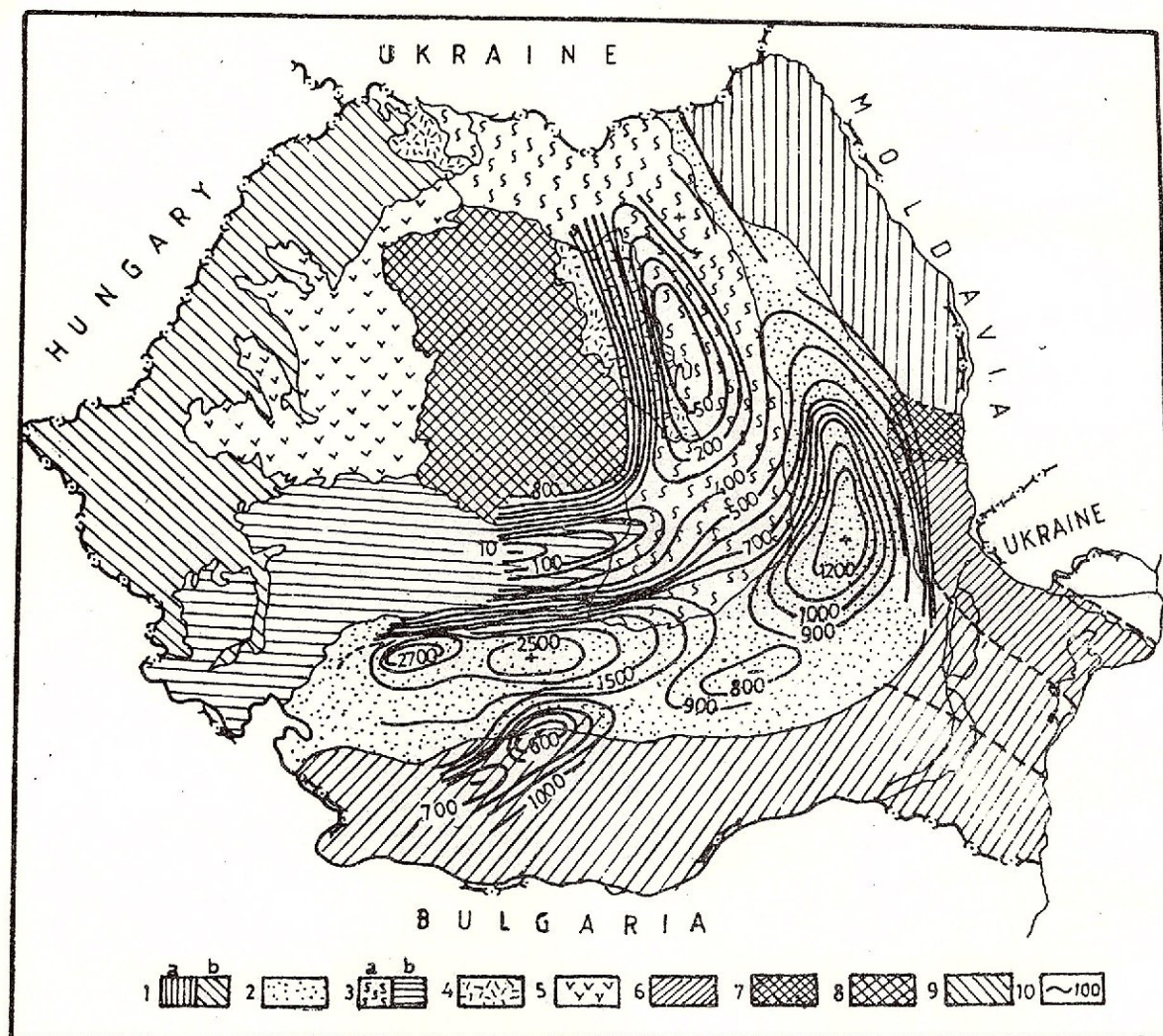


Fig. 5 - Romanian territory. Distribution of Effective Conductance. 1a, Moldavian Platform; 1b, Moesian Platform; 2, Carpathian Foredeep; 3a, Southern Carpathians; 3b, Southern Carpathians; 4, Neogene volcanic rocks; 5, Apuseni Mts.; 6, North Dobrudgean orogene; 7, Bârlad depression; 8, Transylvanian Depression; 9, Pannonian basin; 10, S_{ef} isolines (mho) (after Stănică et al., 1986).

resistivity occurs at the base of sedimentary cover. Another task would be finding possible connections between the edge of the first conductivity layer in the upper mantle and the asthenosphere.

At the transition between Moldavian Platform and the Eastern Carpathians, the depth of the resistive horizon, computed on the basis of 1-D master curves interpretation, has a gentle sloping from east to west, with the maximum depth of 14 km in Câmpulung Moldovenesc area (Fig. 3) and also corresponding to the position of minimum gravity anomaly. The overthrusting of Central-East Carpathian Nappe System on the flysch sedimentary cover is clearly pointed out by resistivity pseudosections, confirming the nappe character of the crystalline mountain chain. A deep dislocation was presented as the subduction scar or the Alpine consumption paleo-plane of the oceanic and thinned crust.

To the west, the MT data suggest the thicknesses of the Pliocene andesitic volcanites, overlapping the younger sediments of the Transylvanian basin, are about 500–1,000 m (around the main tectonomagmatic alignment).

The Alpine paleo-plane of consumption was drawn on the basis of a very strong gradient of the resistivity pseudo-section (Fig. 4), superposing the gravimetric, magnetic and MT information about the deep dislocation along which the intrusion of young volcanic lavas reached the surface and the presence of high heat-flow anomaly overlapping the zone.

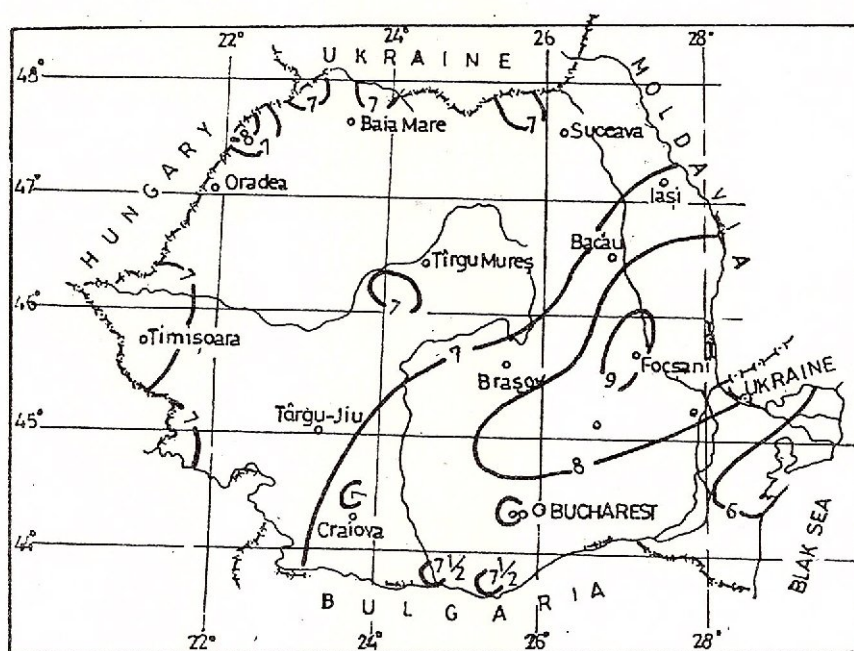


Fig. 6 - Romanian territory. Seismic Zoning by March 4, 1977 Earthquake. 1, isoseists; 2, Făgăraș earthquakes; 3, Banat seisms; 4, Oradea seisms; 5, Sf. Gheorghe seisms (after Cornea and Lăzărescu, 1980).

The distribution of conductance values (Fig. 5) is pointing out: (a) a large area of high conductances with axis corresponding to the maximum thickening of the sedimentary cover on the Carpathian Foredeep, (b) a clear dislocation along the Trotuș line which is going to suggest the mobile margin of the East European Platform.

The presence of the first conductivity layer at a depth of 80–90 km, corresponding to the Low Velocity Layer (LVL) in the Upper Mantle on the top of the asthenosphere is confirmed by seismological and heat flow data.

Seismicity

Romania is episodically affected by earthquakes with Vrancea as main epicentral region. Here, subcrustal earthquakes at the depth of 50–220 km are generated. They have high energy, affecting very large areas. Twice-three time a century, such disasters of 7–7.5 magnitude, 1022–1023 erg, of destroying character are appearing. The last two were generated on November 10, 1940 and March 4, 1977. The most seisms are generated by shearings generated through compression. Sometimes, they generate multishocks, repeated at several seconds, as beats of a unitary fracturing process, as for example the March 4, 1977 earthquake (Fig. 6). The compression appears into a lithospheric slab, supposed to be subducted under the Bending Zone of the Eastern Carpathians (Fig. 7) since Cretaceous. The focal mechanism, very complex and the influence of crustal structure crossed by seismic waves generate a special shape of isoseists. They are elongated on NE-SW direction. Here it is why some earthquakes are very strong affecting Moesian Platform and/or the East European Platform.

Normal earthquakes are produced around other regions of Romania. They are normal (intracrustal) earthquakes (foci placed at 5–30 km depth), of low energy and intensity. Sometimes they are polycyclic earthquakes (many rejoinders after the main shock). They are located at the intersection of some important faults, generating: Făgăraș earthquakes, at the junction between the prolongation of Intramoesian and South Transylvanian faults (important event in 1916), fractures between the Southern Carpathians and the Pannonian Basin, active around Timișoara and Vinga areas (Banat seisms, with important events in 1879 and 1990), the fault system around Oradea (with a local event of 7–8 intensity), Sf. Gheorghe fault from the northern boundary of North Dobrogea (event of magnitude 5.2 on November 1981). Some earthquakes generated in the neighbouring countries are also important for the Romanian territory, as those from Yugoslavia (Morava valley), Bulgaria (Sablă and Maritza valley) etc.



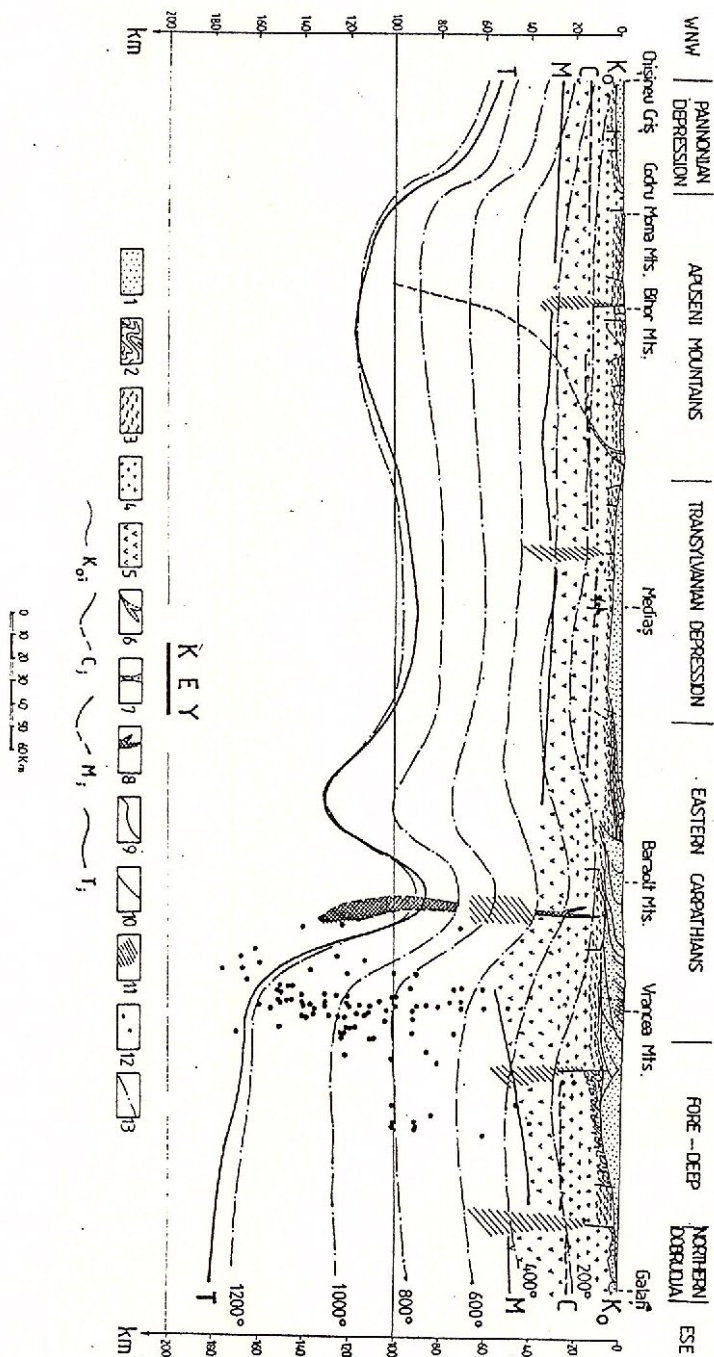


Fig. 7 – Lithospheric cross-section on the Galați-Chişineu Criș profile. 1, sedimentary cover; 2, green schists; 3, crystalline basement; 4, granitic layer; 5, basaltic layer; 6, ophiolites; 7, banatites; 8, Neogene volcanic rocks; 9, overthrusting line; 10, fault; 11, deep crustal fault; 12, hypocenters; 13, isotherms; K₀, sedimentary cover/crystalline basement boundary; C, Conrad boundary; M, Moho boundary; T, lithosphere/asthenosphere thermal boundary.

Moesian Platform
24 CPD Seismic Line across the Central Uplift and the
Roşiori-Alexandria Depression

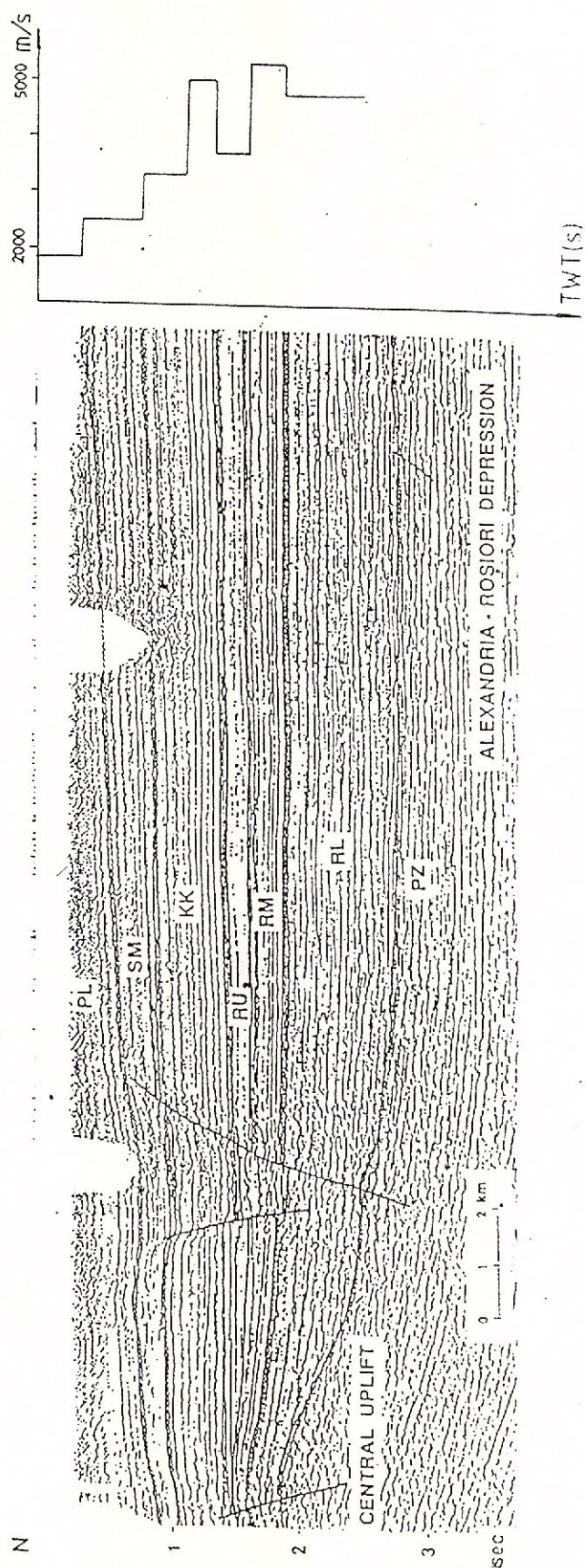


Fig. 8

Moesian Platform
Deep Seismic line - Drăgășani

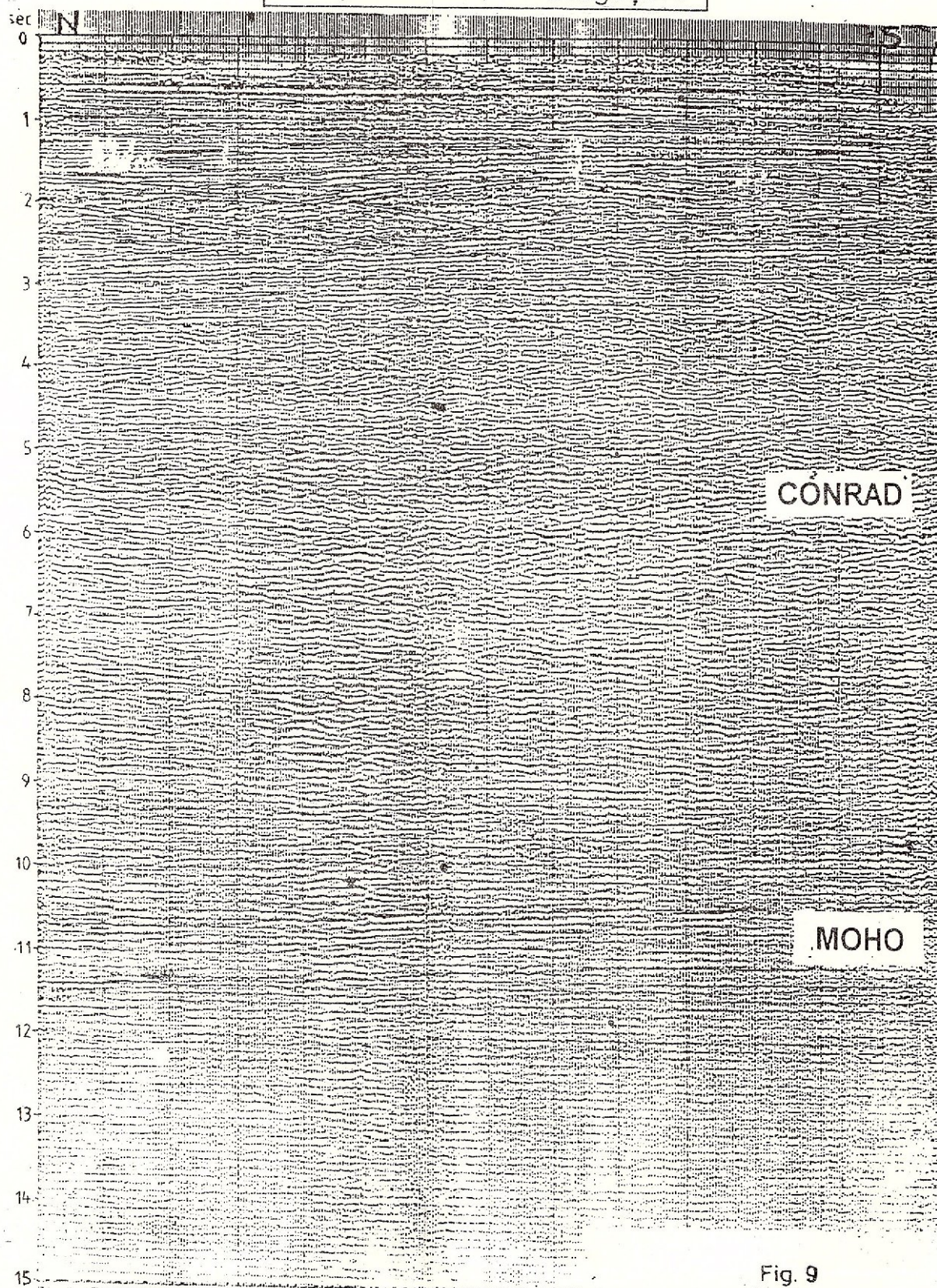


Fig. 9

East Carpathians

Seismic Line across the frontal part of Subcarpathian Nappe (Trotuș Valley)

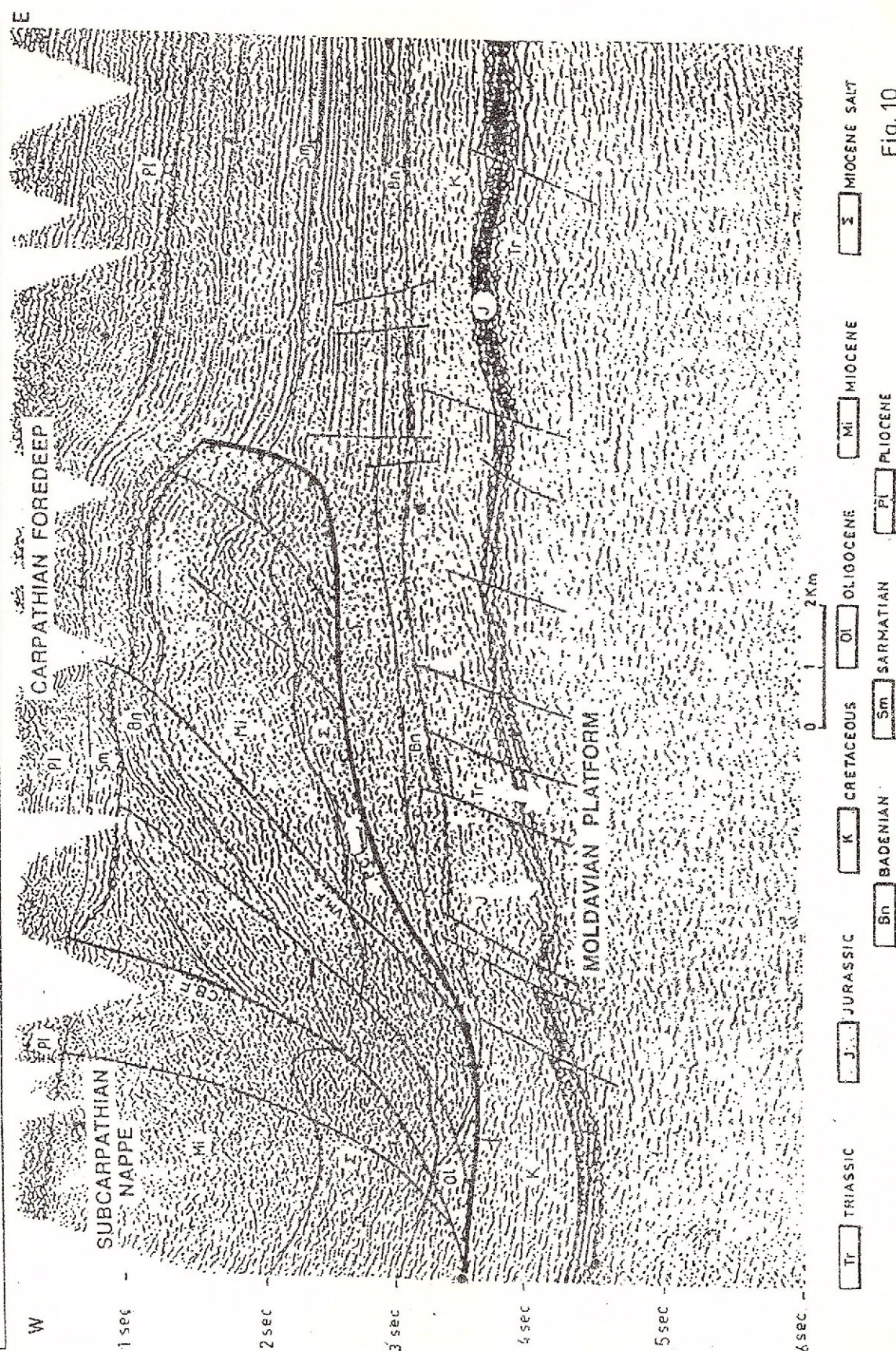


Fig. 10

Focșani Depression

Seismic Line across Roșioru and Ghergheasa gas fields

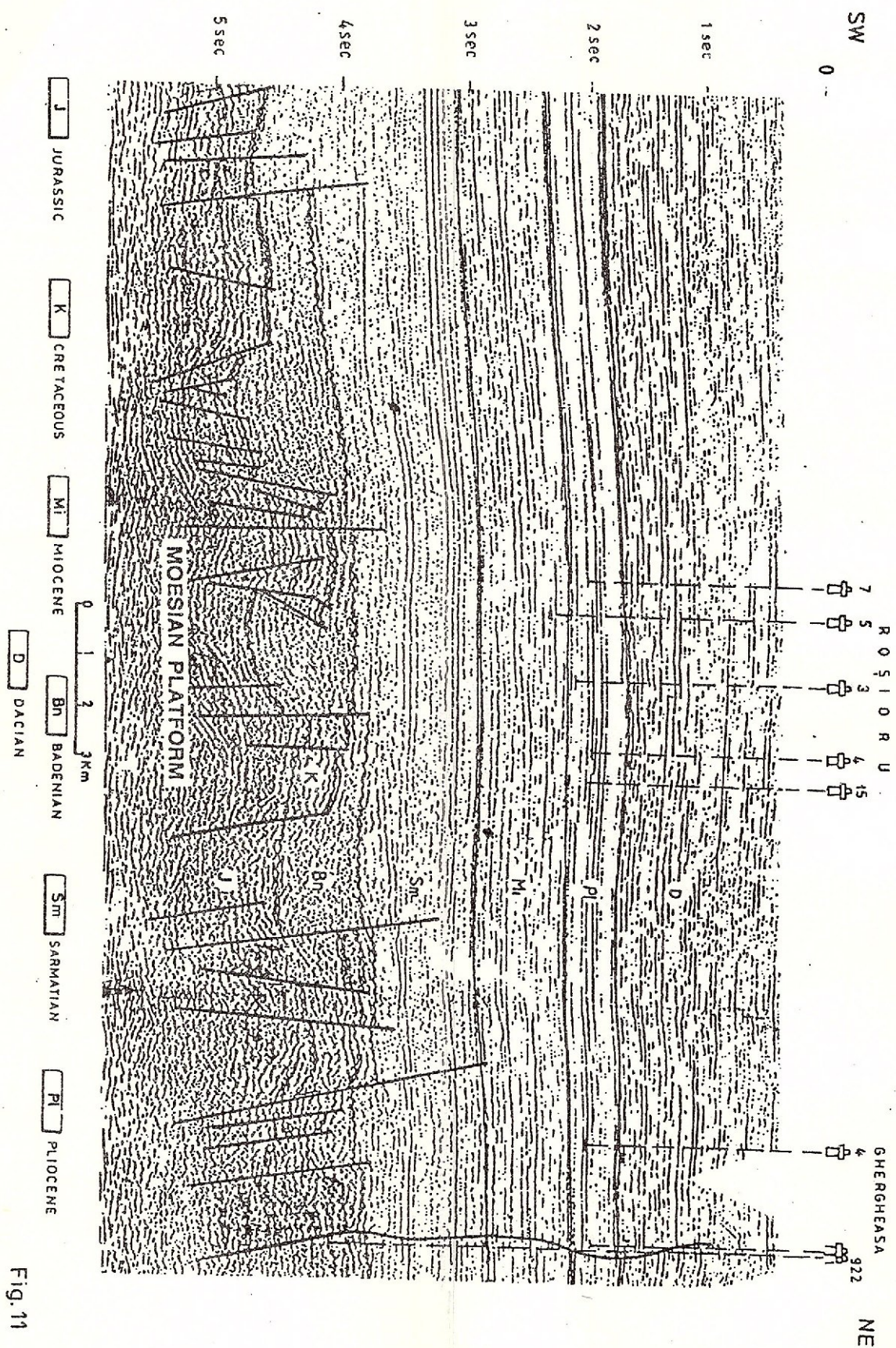


Fig. 11

East Carpathians

Seismic Line across Bucșani and Mănești Fields (Diapiric Folds Zone)

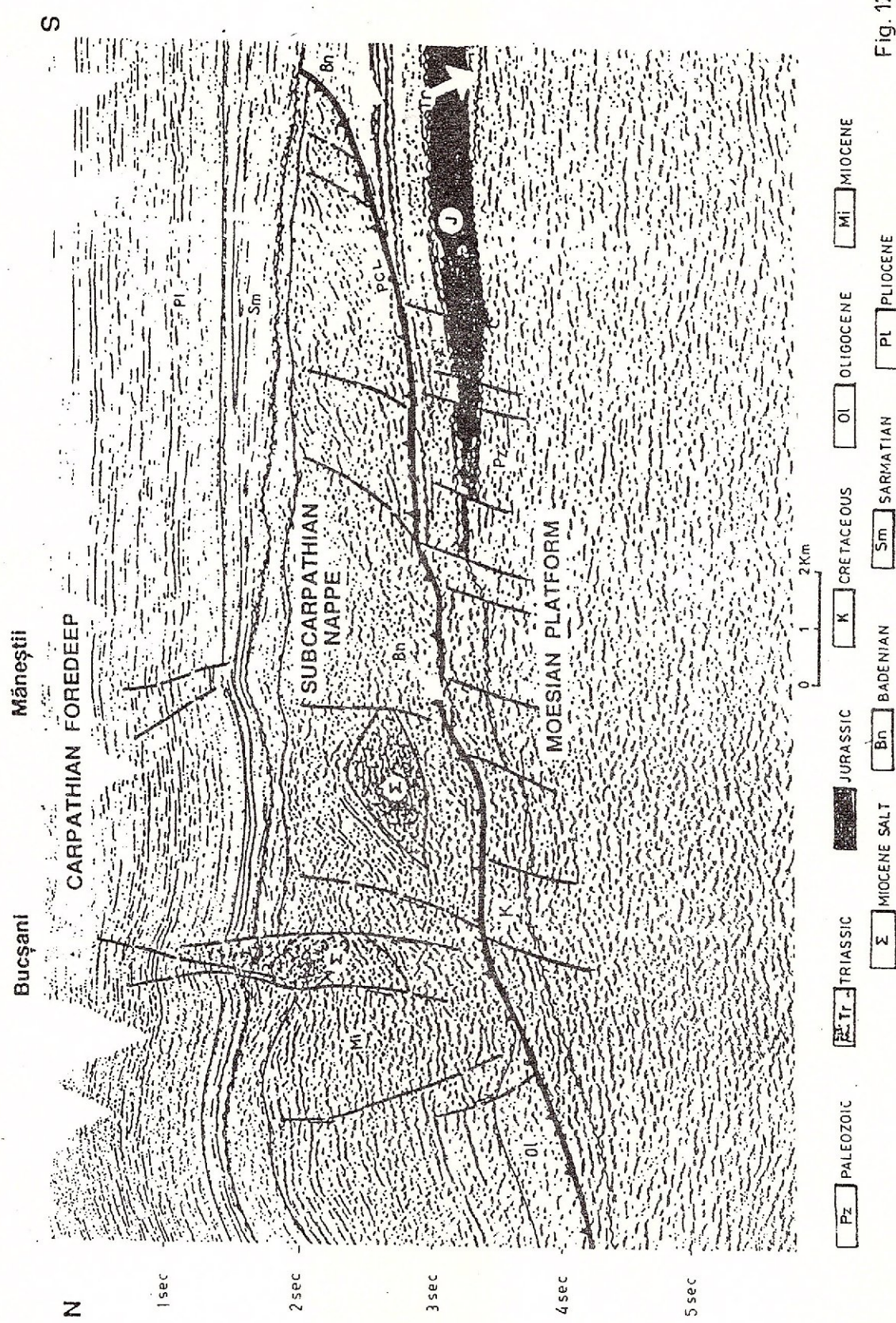


Fig. 12

Carpathian Foredeep - Getic Depression
Deep Seismic line - N Strehaiia

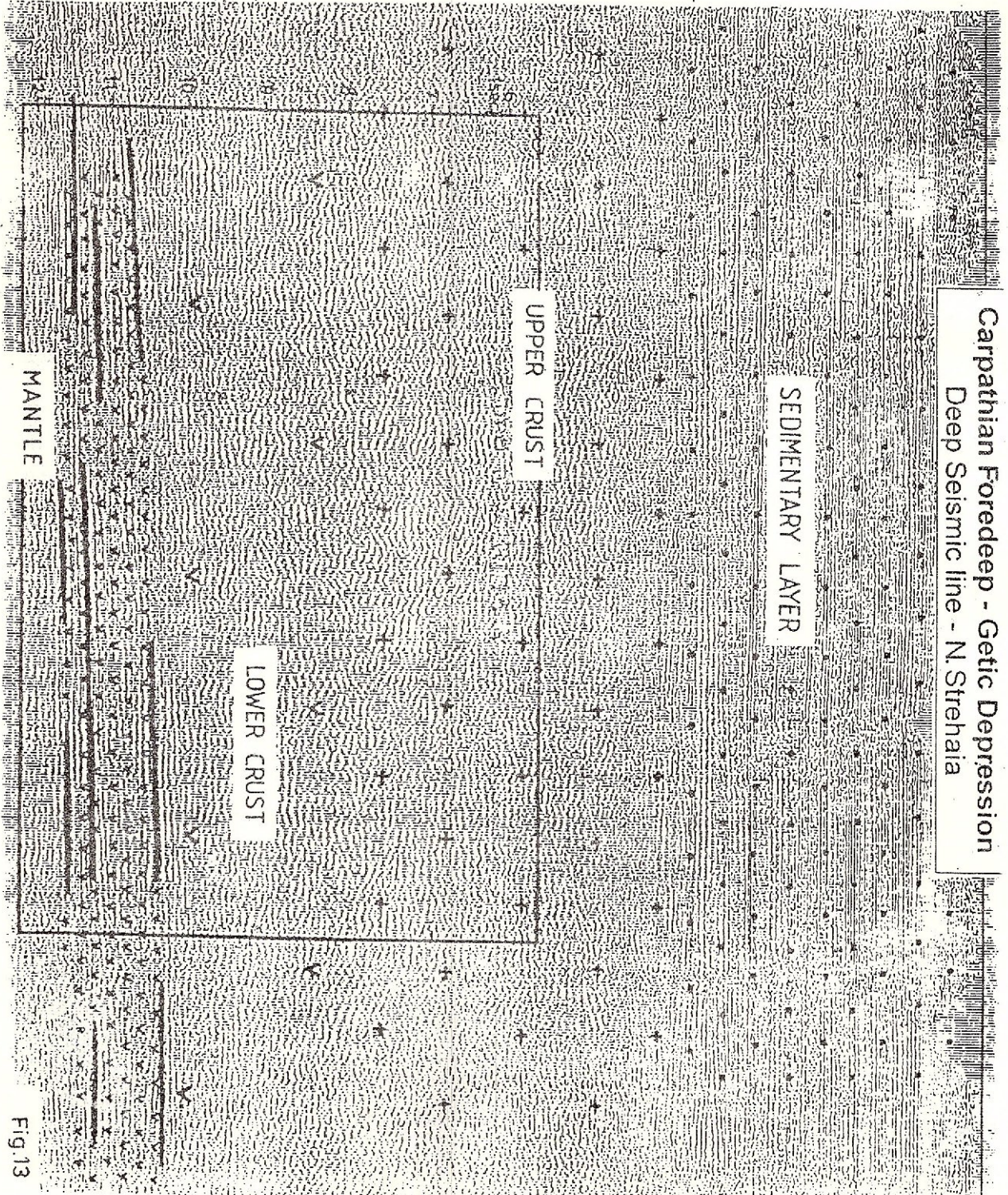
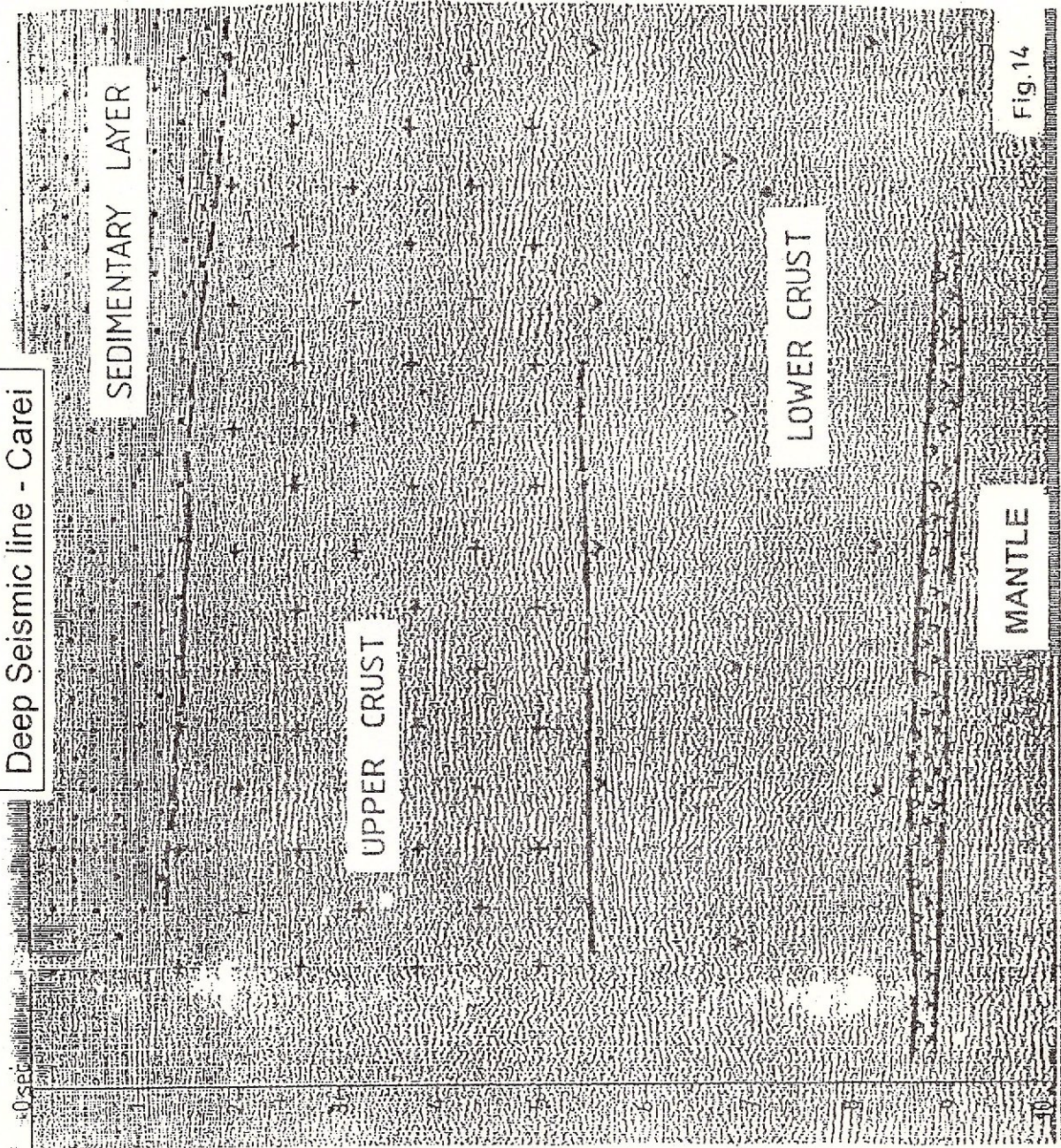


Fig.13

**Pannonian Depression
Deep Seismic line - Carei**



Panonian Depression 24 CPD Seismic Line

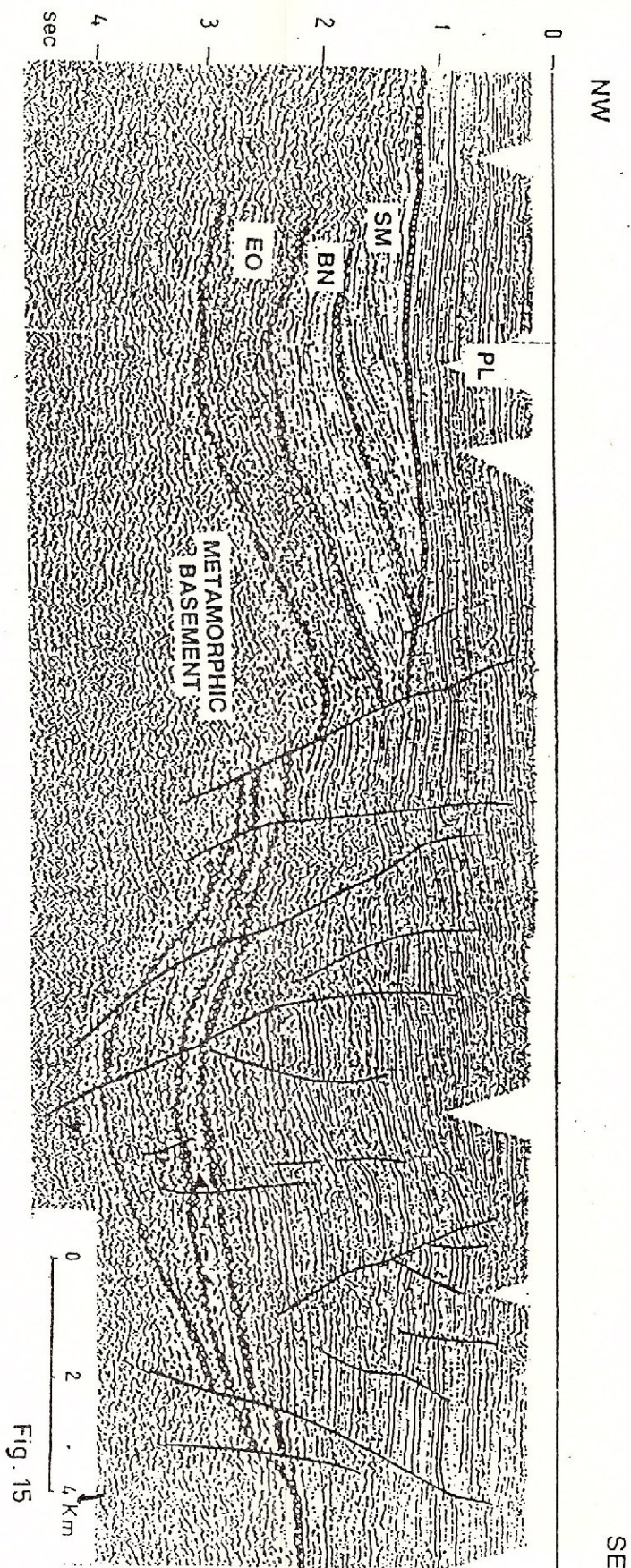


Fig. 15

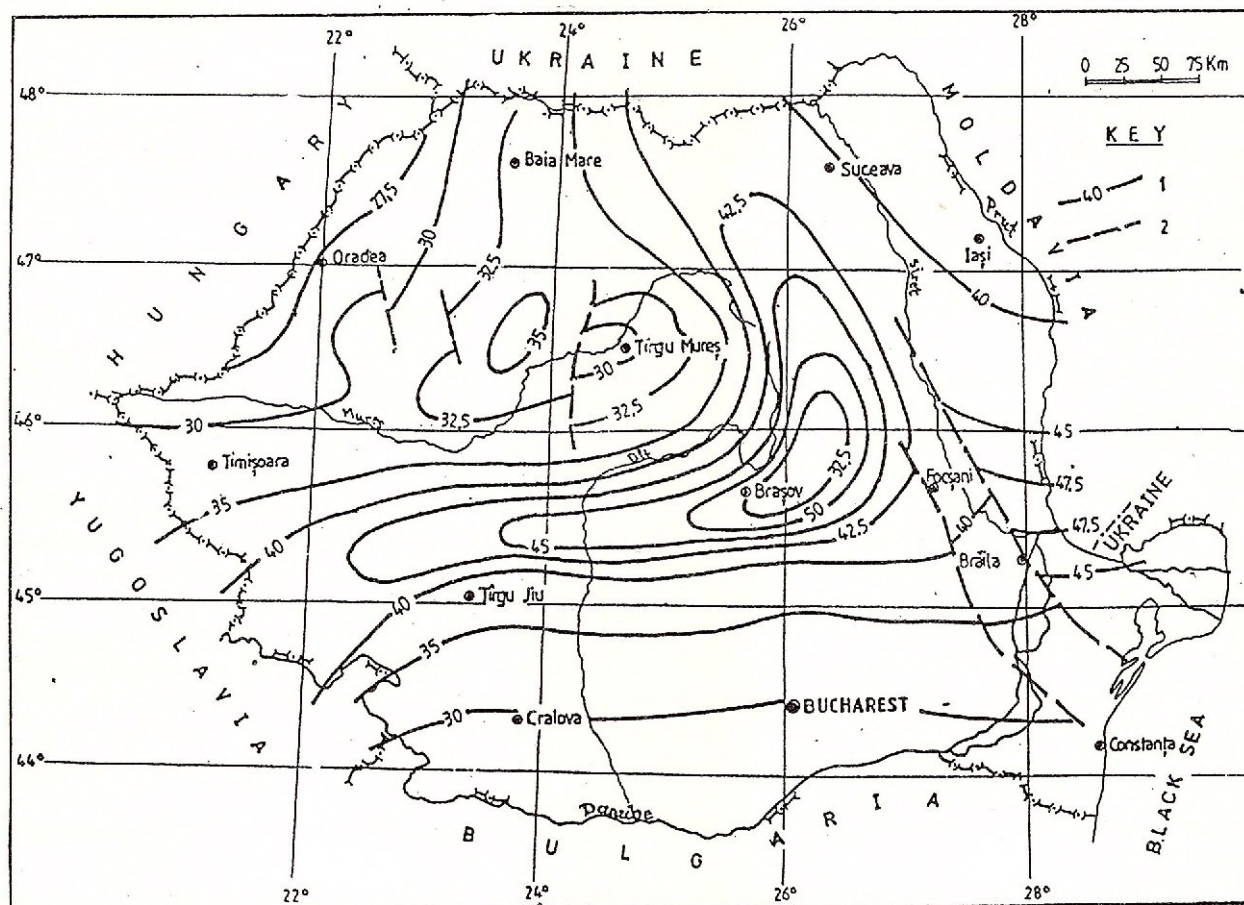


Fig 16 – Romanian territory. Crustal thickness.

Seismics

Only accessible terrains were studied by seismics. The first proposed seismic profile to cross the Carpathians and to be carried out by modern means will hopefully become reality under EUROPROBE umbrella.

Moesian Platform. The sedimentary cover has a thickness of 2–7 km and seismic velocities of 1.7–5.7 km/s. There are two main seismic sequences: the upper one, down to the base of Neogene formations, with parallel and continuous reflections and the lower one, including Mesozoic and Paleozoic formations, with dipping reflectors containing discontinuities, amplitude variations, lower frequencies and diffractions. The characteristic reflectors are: top Jurassic–Lower Cretaceous limestone formation, top of Middle Triassic limestone and dolomite formation (T3/T2 boundary), top of Paleozoic limestone formation, top of crystalline basement (Fig. 8). The consolidated crust (Fig. 9) has a thickness of 25 km (south) to 30 km (north) and seismic velocities of 5.8–6.4 km/s in the upper part, 7 km/s around the Conrad discontinuity and 7.5–7.7 km/s in the lower part, down to Moho discontinuity. The characteristic reflectors are: Conrad discontinuity, at 16–18 km depth (5–6 s TWT), well pointed out as strong head waves on reflection profiles; Moho discontinuity at 30–35 km depth (10.5–11.5 s TWT), with continuous and strong reflectors and a multiple character (4–6 km thickness), representing the transition from Crust to Upper Mantle.

Moldavian Platform. The sedimentary cover has a thickness of 1–3 km and seismic velocities of 1.5–5.3 km/s. There are two main seismic sequences: the upper one (down to the base of the Neogene formations, on



large scale correlability - the most prominent reflections belonging to the anhydrite horizon of Middle Badenian) and the lower one (top of head waves from the Paleozoic limestone). The sedimentary cover/crystalline basement is a strong seismic marker of 0.22 reflection coefficient. The consolidated crust has a thickness of about 40 km, thicker to the Ukrainian shield and Eastern Carpathian orogen, being the oldest tectonic unit of the Romanian territory: Proterozoic.

Carpathian Foredeep. The Neogene seismic sequence has average seismic velocities of 1.7–2.5 km/s and different seismic behavior. Inside the Subcarpathian nappes there are: folds, thrusts faults, diapiric structures (Figs. 10, 11, 12); external to the Pericarpethian line, there are (sub)horizontal prominent seismic reflections into the Upper Miocene deposits; common Pliocene deposits covering both internal and external zones, there are strong and continuous reflections.

The Paleozoic-Mesozoic seismic sequence belongs to the surrounding platform sequences, so it has the same seismic behaviour.

Consolidated crust does not generate prominent reflections along the whole foredeep (Fig. 13). The Upper Crust appears to be almost transparent. The Lower Crust has increasing reflectivity by depth increasing, with bends of subhorizontal reflections. The Crust/Mantle transition has a 2–3 km thickness at 9–11.5 s TWT. The general thickness of the Crust is 34–37 km and much more, up to 42 km, in the Focșani depression.

Transylvanian Depression. There are two main seismic sequences, separated by the Badenian tuff complex as the strongest seismic marker around the region: the upper one (Upper Badenian-Pannonian age sediments, with pronounced reflections of high frequencies, well correlated with various tuff levels) and the lower one (Upper Cretaceous-Burdigalian age sediments, with only several reflections suggesting the discontinuous areal development of different sedimentary formations). The Mesozoic folded basement includes crystalline formations, ophiolites and Paleozoic-Lower Cretaceous sedimentary deposits, folded and thrust. They have strong seismic reflections corresponding to the T/J limestone formation (velocity near those of crystalline basement: 5.6–5.8 km/s). The consolidated crust has two sectors: the Upper Crust is almost transparent, with a local seismic marker at 3.8–4.0 s TWT and a strong regional reflection at 5.5–6.5 s TWT, corresponding to the Upper/Lower consolidated Crust (Conrad discontinuity). The Lower Crust has short and dense subhorizontal reflections. Moho discontinuity is placed at 11 s TWT, not very well distinguished on the reflection seismic sections, corresponding to 30–32 km depth.

Pannonian Basin, as a Neogene extensional basin, consists of the sedimentary cover and the consolidated Crust.

The sedimentary cover presents two main seismic sedimentary sequences: the upper one (consisting of Pannonian deposits, with very well individualised reflections) and the lower one (consisting of Paleogene-Miocene deposits, with discontinuous reflections, due to non-completed development of sedimentary formation generating the morpho-structural palcorelief). The boundary between sedimentary cover and metamorphic basement is clear especially in the area of uplifted blocks (Fig. 14). The consolidated crust is generally transparent with local subhorizontal reflections at 5–6 s TWT, probably corresponding to Upper Crust / Lower Crust limit (Fig. 15). A laminated zone of 3–4 km thickness, at Crust/Mantle boundary, of ductile behaviour, probably because of the high heat flow is also presented. The Crust is thin (24–28 km, corresponding to 8.5–9 s TWT), a transition to the very thin crust from the central part of the Pannonian Basin (Fig. 16).

Recent crustal movements

The recent crustal movements obtained by high accuracy repeated levelling (Fig. 17) point out some very interesting information possible to be correlated with other classic geophysical information, as gravity, geomagnetics, heat flow, magnetotellurics.

Moesian Platform. The general shape of the isolines suggests the general structure of the platform crystalline basement, as from gravity and some magnetotelluric data. It is possible to separate different uplifted blocks (Bals-Craiova, Balota, SE Bucharest) and subsided blocks (Alexandria, NW Bucharest etc.).

Moldavian Platform. The northern part is clearly affected by the very active uplifting of the Eastern Carpathians. At the parallel of the Trotuș River, the movement character is going to be changed and an active subsidence of -1.4 mm/y clearly separates the Bârlad Depression.

Transylvanian Depression is very well individualised by its quiet realm from the recent crustal movement viewpoint. The most part of the depression is stable (0 mm/y), suggesting the present day stability of its basement.



Pannonian Basin. As from other geophysical data, it is clearly divided into different blocks, orientated NNW-SSE and affected by low scale subsidence (as along the Mureş and Bega rivers) or uplifting (Someş River

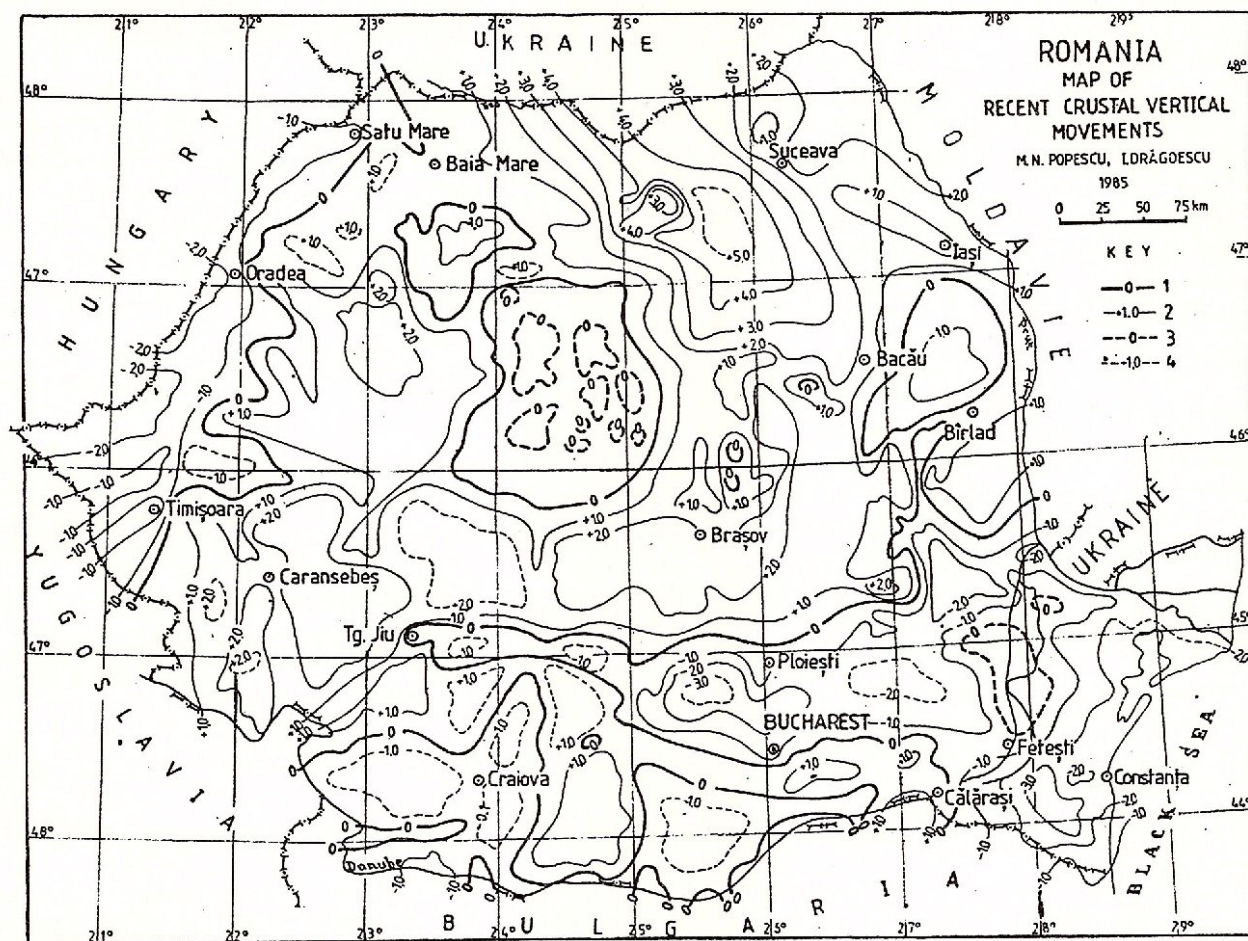


Fig. 17 Romanian territory. Recent crustal movements (after Popescu, Drăgoescu, 1986).

at Satu Mare).

Alpine Carpathian Folded Region. The main feature is represented by active uplifting of the whole Carpathian chain, with velocities up to $+5.5$ mm/y, in the northern part of the Romanian sector of the Eastern Carpathians. This active movement seems to affect the westernmost part of the East European Platform, suggesting a possible suture between the Alpine Europe and the Moldavian Platform. To the south, an important change of crustal movements (subsidence down to -1.3 mm/y on Bârlad depression) supports the previous idea that the Trotuș Fault represents an important geotectonic boundary affecting the Eastern Carpathians. The southern sector of Eastern Carpathians, Southern Carpathians and Apuseni Mts have an active uplifting of about $+2.5$ mm/y, as the most alpine structures around the world.

Heat flow

Generally, the heat flow information is well superposed and clearly controlled by the general tectonic structure of the Romanian territory.

Moesian Platform. The local maximum of 60 mW/m² corresponds to the Alexandria sinking and shape a hot region of the platform. In Dobrogea, the heat flow lines are parallel to the Peceneaga Camena and Capidava Ovidiu crustal faults.

Moldavian Platform. The values of heat flow are low and the whole platform is well shaped by the isolines of 50 and 40 mW/m². The southern part of the Moldavian Platform has a different aspect, the heat flow isolines looking like the tectonic structure.



Pannonian Basin. The hot behaviour of the Pannonian Basin is clearly stopped at the eastern tectonic boundary of the basin. It is possible to separate detailed geothermal fields which are important from different economical point of view, as thermal water for housing and spa. Transylvanian Depression. It is very well shaped by the 70 mW/m² isoline. The most interesting thing is that the whole depression is a very cold province, the heat flow values decreasing down to 38 mW/m².

Alpine Carpathian Folded Region. The Eastern Carpathians, which have the "maximum maximorum" recent upheaval, appear just a little bit different in the heat flow image. The Călimani-Gurghiu-Harghita Neogene volcanic chain represents a separate maximum heat flow anomaly (up to 100 mW/m²). The Southern Carpathians appear with a different behaviour, the heat flow isolines being perpendicular on the tectonic structure and probably suggesting some very deep access ways for the hot/cold fluids. The Apuseni Mts are not clearly separated into the heat flow information, because of not enough data. It is also possible that the very different characteristics of the Pannonian (hot) and Transylvanian (cold) sector might rule the general heat flow information.

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SOUTH CARPATHIANS

by

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I. INTRODUCTION

As area, elevation and geological variety, the South Carpathians represent the main segment of the Carpathian fold and thrust belt, and one of the important areas of the Alpine orogen running in Southern Europe, from Spain to Turkey. Between Prahova valley in the East, where the northward bending to the East Carpathians begins, and Timok valley to the West, where they pass to the Balkans, the South Carpathians crop out over a length of 500 km, attaining 100-150 km in width and covering an area around 700 km². In South-Western Romania (Banat) and Eastern Serbia, the South Carpathians describe a loop, curving around the Moesian Platform, from E-W to N-S, and again to E-W directions. As a mountain belt, the South Carpathians represent a girdle of individual massifs, separated by intramontane basins or deep valleys. The main morphology is a high plateau (2000-2300 m), but some rocky summits reach 2500 m. They are bordered southwards by the Getic Basin, westwards by the Pannonian Basin and northwards by the Southern Apuseni Mountains and the Transylvanian Basin. As a fold and thrust belt, the South Carpathians occupy a larger area beneath the Cenozoic basins, neighbouring northwards and westwards with the South Apuseni - Vardar ophiolitic belt and southwards with the Moesian Platform. The northern margin is represented by the steeply dipping South Transylvanian Fault (Săndulescu, 1984), but the southern border is a low-angle north-dipping overthrust (Ștefănescu et al., 1988). Since geological presentations of the South Carpathians for various meetings and excursions are included in the guide-books of Codarcea et al. (1961, 1968), Kräutner et al. (1981), Năstăseanu et al. (1981) and Balintoni et al. (1989), in this text we will focus on the nappe structure and post-nappe activity in this Alpine fold and thrust belt.

II. THE NAPPE STRUCTURE OF THE SOUTH CARPATHIANS

1. General structure

Since Murgoci (1905 a,b,c; 1912), an upper "Getic Nappe" and lower "Autochthonous Massifs", separated by a sheet of Alpine sedimentary cover, were generally accepted for the South Carpathians. In the Olt and Banat regions, at the top of the Getic Nappe, Streckeisen (1934) introduced the "Superior Nappes", separated later by Codarcea et al. (1967) as the "Supragetic Nappe". The number, areal development, internal constitution and age of the Getic-Supragetic tectonic units was a controversial subject for the last 20 years, each author favouring its own proposal (Maier, 1974; Dimitrescu, 1978; Năstăseanu, 1978; Săndulescu, 1975, 1984; Iancu, 1986; Balintoni et al., 1986; Kräutner et al., 1981, 1988; Gherasi et al., 1986; Hann, 1994). Here we will present the proposal of Balintoni et al. (1989) both for the map (Pl. I) and for



the correlation table 1 (Balintoni et al., 1989).

Table 1

CORRELATION ATTEMPT OF THE AUSTRIAN NAPPE INCLUDED IN THE LARAMIAN NAPPE OF THE GETIC-SUPRAGETIC DOMAIN.	
LOTRU - BISTRA LARAMIAN NAPPE	TIMIŞ - BOIA LARAMIAN NAPPE
Austriides	
Valea lui Stan (Căineni) Nappe	Strâmba Nappe
Bocşa Nappe = Uria Nappe	Moldoveanu Nappe Pârâul Moaşă Nappe
Tălva Drenii Nappe = Măgura Nappe	Argeş Nappe
Maniam Nappe	
Sasca-Gornjak Nappe = Reşiţa Nappe	Getic Nappe
Getic Nappe	
Iloviţa-Borăscu Nappe	Holbav Nappe

At the scale of the Carpathian-Balkan fold and thrust belt, the Getic-Supragetic units correspond to the Median Dacides, correlated to the North with the Central East Carpathian Nappes and to the South with the Inner Balkanides, eastwards of the Timok River (Săndulescu, 1975, 1984). They represent slices of the Getic Plate, a continental fragment of the European margin between the (external) Outer Dacides and the Transylvanian Thetys.

The ideas concerning the rocks below the sole of the Getic Nappe underwent a complex evolution. Firstly, Popescu-Voiteşti (1923) and Codarcea (1940) have separated, at the top of the underlying stack, the Urdele, Severin, Arjana and Cerna nappes. For the Severin Nappe, Codarcea (1940) also used the term "Severin Para-autochthonous", considering it as originating from a different area than the Danubian Domain - the palaeogeographic area of the "Danubian Autochthonous". Berza et al. (1983) introduced the concept of two systems of mainly basement nappes - the Upper Danubian Units and the Lower Danubian Units. The constitution and map pattern of the Danubian Nappes were slightly modified by Iancu et al. (1990), Kräutner et al. (1988), Balintoni et al. (1989) and Berza, Iancu (1994), but the main divisions are almost generally accepted in the form shown in figure 1 and plate 1. The Danubian Nappes represent the Marginal Dacides; they have no corresponding units in the East Carpathians, but in the Balkans their equivalents are the Outer Balkanides (Săndulescu, 1984).

Timing and emplacement of these nappes is (partly) controversial. While the presence of Upper Cretaceous turbidites at the top of most Danubian Nappes, as well as the age of the first post-nappe covers, ascribes them to the Laramian tectogenesis, the absence of such deposits above the Severin Nappe determined Codarcea (1940) to claim an earlier, Mid-Cretaceous (Austrian), age for the Getic/Severin overthrust fault. At the eastern margin of the South Carpathians, the Getic Nappe overthrusts the Ceahlău and Baraolt nappes in the Middle Aptian ("the first Getic phase", Săndulescu, 1984). The latter nappes consist of Sinaia Beds but lack ophiolites; however this is still an indirect evidence for the age of the Severin Nappe in the western part of the South Carpathians. In our opinion the Mid-Cretaceous age of the Severin Nappe has still to be demonstrated, but if the duplex nature of the thrust systems underlying (as roof thrust) the Getic Nappe (Seghedi, Berza, 1994) is real, then its older age is even compulsory. The reverse concerns the

lowest thrust plane known in the South Carpathians (i.e. separating the Lainici Nappe from the underlying Schela Nappe), a plane which has to be the youngest of the exposed thrusts. Stănoiu et al. (1992) advocated a Liassic (Old Kimmeric) tectogenesis for this thrust, because the only sedimentary formation involved is the Lower Jurassic Schela Formation. The 70 Ma K-Ar age (Grünenfelder et al., 1983), yielded by ultramylonites from the Șușița (Precambrian) granitoid body within the thrust zone, testify for the Laramian age of this thrust also.

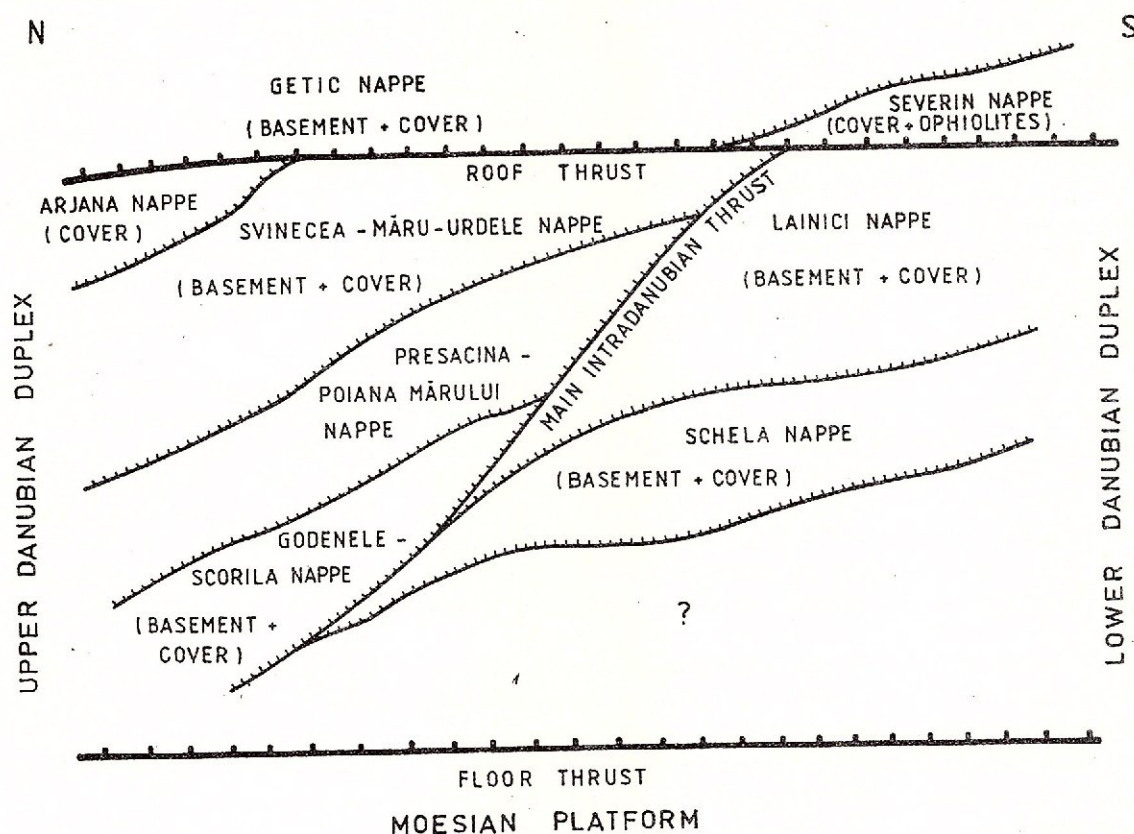


Fig. 1 - Schematic N-S distribution of the Danubian Nappes.

If the Severin Nappe, consisting of Jurassic-Lower Cretaceous turbidites and ophiolites, is a sliver from the Outer Dacidian Rift (Săndulescu, 1984), a Valais type secondary rift in the European margin (Săndulescu, 1994), the Danubian Nappes represent slices from the Euxinic Plate, a microplate separated from the East European Plate by the Peceneaga-Camena dextral strike-slip fault. For Săndulescu (1984), the Danubian Domain represents, as the Prebalkans in Bulgaria, exposed formations of the Moesian Platform. For us, assigning to an initial Laramian phase the thrust over the Moesian Platform pictured by Ștefănescu et al. (1988), the Danubian Nappes represent a different paleogeographic domain. But we do not agree with the proposal of Savu et al. (1991) of the emplacement of the Severin (Jurassic) rift between the Danubian and the Moesian Platform.

Due to the Carpathian bend, in the South Carpathians the array of the thrusts is at present opposite to that of the Alps, even if the Europe directed vergence is the same: here the European margin (the Moesian Platform) is situated South of the fold and thrust belt. By the N-S bending in the Banat-Eastern Serbia area, and then the curving to a W-E trend in Bulgaria, the normal array is recovered and continues eastwards, across the Black Sea, to the Greater Caucasus (Săndulescu, 1984).

2. The Getic-Supragetic thrust-sheets

These tectonic units (Getides-Supragetides) represent in the South Carpathians the Median Dacides, slices from the Getic Microplate; the latter is a fragment of the European margin, bordered northwards by

the Thetyan oceanic realm and southwards by the oceanic/thinned continental realm of the Outer (External) Dacides (Săndulescu, 1984, 1994). As opinions concerning the present structure of the Getic-Supragetic Thrusts vary considerably, we will present here the ideas of Balintoni, Iancu and Hann (in Balintoni et al., 1989). According to them, the Mid-Cretaceous shortening resulted in a stack of several basement, basement-cover or cover nappes, all of them showing strong similarities in the lithology of their pre-Alpine basements and Alpine covers. The main unit is the lowest thrust sheet and corresponds to the Getic (*sensu stricto*) Nappe; the overlying units correspond to the Supragetic Nappes. This shortening event must be considered Austrian (Mid-Cretaceous), similarly to the main tectogenesis affecting the Central East Carpathian Nappes (Săndulescu, 1984), since the youngest formations involved in thrusting are Aptian and the youngest post-nappe covers belong to the Albian-Cenomanian. The present geometry of the upper structural compartment of the South Carpathians (Getides-Supragetides) was created during the Laramian shortening of the Getic-Supragetic (Austrian) nappe pile, when the northern Timiș-Boia Nappe was emplaced onto the southern Lotru-Bistra Nappe (table 1).

2.1. The Timiș-Boia (Laramian) Nappe

This nappe crops out in the north-western and eastern parts of the South Carpathians. On the attached map (Pl. 1) it is labelled Tm-B (NE-SW hatches) and the constituting Austrian thrusts are omitted in the western part, due to the scarcity of reliable available data (but an attempt to separate them was made by Balintoni, Iancu (1986); at the eastern part of the South Carpathians, the Austrian thrust-sheets are outlined.

The western outcrop area of the Timiș-Boia Nappe consists only of basement formations. They probably belong to the Upper Precambrian (Buziaș Hills; Tincova-Rușchita-Alunu-Ghelar area in Poiana Ruscă Mts), but also to a (pre-Upper Carboniferous) Paleozoic sequence (the northern part of the Poiana Ruscă Mts). Kräutner et al. (1981, 1988) considered all the formations Paleozoic, showing a prograde Variscan metamorphism, with increasing metamorphic grade from east (chlorite zone) to west (staurolite zone). Balintoni, Iancu (1986) interpreted the eastern, greenschist looking areas as retrogressed rocks of the Precambrian Făgăraș Series. East of Deva, across the Mureș River, these formations are exposed in the Southern Apuseni Mountains, and are surrounded by Upper Cretaceous and Miocene deposits or pierced by Neogene volcanics.

At the eastern end of the South Carpathians, in the Olt Valley and Făgăraș-Leaota mountains, Balintoni et al. (1986, 1989) and Hann (1994) proposed the following pre-Laramian nappes (Austrian, according to the quoted authors) (from top to bottom):

- Strâmba Nappe (Proterozoic metamorphic basement);
- Moldoveanu nappe (Proterozoic metamorphic basement, Permo-Mesozoic cover);
- Pârâul Moașa Nappe (Proterozoic metamorphic basement);
- Argeș Nappe (Proterozoic metamorphic basement, Permo-Mesozoic cover);
- Getic Nappe (Proterozoic metamorphic basement, Permo-Mesozoic cover).

According to Balintoni et al. (1989), in the Timiș-Boia (Laramian) Nappe, the basements of the Austrian nappes show the following constitution:

- Strâmba (litho) Zone of Proterozoic age, in the Strâmba Nappe, represented by garnet bearing paragneisses and micaschists and quartzo-feldspathic rocks;
- Proterozoic Făgăraș (litho) Group in the Moldoveanu Nappe, with a lower Suru (litho) Zone (represented by carbonate rocks, amphibolites, paragneisses, micaschists, quartzites and graphitic schists) and an upper Sâmbăta (litho) Zone (dominated by quartzo-feldspathic rocks);
- Proterozoic Moașa (litho) Group in the Pârâul Moașa Nappe, with a lower Scara (litho) Zone (carbonate rocks, amphibolites and staurolite-kyanite bearing paragneisses) and an upper Serbota (litho) Zone (staurolite-kyanite bearing micaschists and paragneisses), showing intense retrogression at the top;
- Proterozoic Cumpăna (litho) Group in the Argeș Nappe, including: a lower Vâlsan complex, characterised by exotic rocks (mainly metaultramafites) and by the large development of ophiolitic migmatites ("Cozia Gneiss"); an upper Măgura Căinenilor (litho) Zone, represented by staurolite and kyanite-bearing micaschists, with the quartzitic+amphibolitic Vemeșoia (litho) Zone at the top.

— Iezer and Leaota (litho) Groups of Proterozoic age in the Getic Nappe; the age for the upper part of the Leaota (litho) Group is controversial, being also considered Early Paleozoic (Dimitrescu, 1978; Kräutner, 1980; Iancu, Mărunțiu, 1994), or even Late Paleozoic (Dimitrescu et al., 1990); it is represented by retrogressed paragneisses (with garnet porphyroclasts) and new albite, chlorite and muscovite (Lerești Formation); the upper, controversial, Călușu Formation consists of metabasites and metaterigenous schists with

greenschist appearance.

At the western tip of the South Carpathians, in the Poiana Ruscă Mountains, traditional lithostratigraphy (Kräutner et al., 1981, 1988) uses the names Bătrâna Series (Ordovician?), Găvojdia Series (Middle-Upper Devonian) and Padeș-Nădrag Series (Lower Carboniferous). According to Balintoni, Iancu (1986), the Bătrâna, Găvojdia, Ghelar and Nădrag formations represent highly retrogressed medium-grade Proterozoic sequences; the Padeș Group is considered Lower Paleozoic and the Hunedoara-Luncani Formation is Middle Paleozoic, both showing only a low-grade Paleozoic metamorphism.

The covers found in some of the Austrian nappes constituting the Timiș-Boia (Laramian) Nappe show a minor development compared to their pre-Alpine metamorphic basements:

- in the Modoveanu Nappe, polymictic Permian conglomerates and, possibly Triassic, carbonate rocks (Balintoni et al., 1986);

- in the Argeș Nappe, Lower Triassic quartitic sandstones, Middle-Upper Triassic carbonate rocks and Aptian reef limestones (Urgonian) (Balintoni et al., 1986);

- in the Getic Nappe, different interpretations were proposed by Săndulescu (1975, 1984, 1994) and Balintoni (1986, 1994). For Săndulescu, the Triassic-Jurassic deposits in the Brașov area represent members of the Alpine cover of the Getic Nappe, while for Balintoni et al. (1986, 1989), they belong to distinct Austrian Nappe (Holbav, etc.); Lower and Middle Triassic detrital and carbonate rocks (Guttenstein facies) are unconformably overlain by Lower Jurassic terrigenous sequences, Middle Jurassic pelagic limestones and siliceous deposits, Upper Jurassic carbonate rocks (Stramberg facies) and Lower Cretaceous neritic and pelagic carbonate rocks; in the Dâmbovicioara area, Lower Cretaceous marly sequences are included to the Getic Nappe both by Săndulescu (1975, 1984, 1994) and Balintoni (1994);

- within a distinct unit near Holbav, Lower Jurassic (Gresten facies) deposits with alkaline volcano-sedimentary sequences crop out (Săndulescu, 1975).

A common feature for these Alpine covers is the absence of metamorphism and pervasive deformation, their sedimentary and paleontological records being well preserved.

The post-Austrian cover of the Timiș-Boia (Laramian) nappe is exposed in Hia, Deva, Cîsnădie areas and Postăvaru and Bucegi Mountains and consists of coarse detrital Upper Cretaceous deposits (Raciu Breccia, Postăvaru and Bucegi conglomerates showing Vraconian-Albian ages; in the western Poiana Ruscă Mountains, volcanoclastic deposits are known in the Nădrag area).

2.2. The Lotru-Bistra (Laramian) Nappe

This is the main Laramian body of the Getic-Supragetic realm. On the attached map (Pl. 1), it is figured with NW-SE hatches and the constituent Austrian nappes and Upper Cretaceous cover are separated on the whole outcrop area. According to Balintoni et al. (1989), the internal structure of the Lotru-Bistra (Laramian) Nappe involves:

- an Upper Cretaceous cover (sedimentary and volcanic deposits), common to all Austrian nappes and exposed in the basins of Rusca Montană, Densuș and Pui;

- Valea lui Stan-Căineni Nappe (Proterozoic metamorphic basement and Upper Paleozoic-Mesozoic cover);

- Bocșa-Uria Nappe (Proterozoic and Paleozoic metamorphic basement and Upper Paleozoic-Mesozoic cover);

- Tâlva Drenii-Măgura Nappe (Proterozoic metamorphic basement and Upper Paleozoic-Mesozoic cover);

- Moniom Nappe (Paleozoic metamorphic basement and Mesozoic cover);

- Sasca-Gornjak-Reșița Nappe (only Upper Paleozoic and Mesozoic formations);

- Getic Nappe (Proterozoic and Paleozoic metamorphic basement and Upper Paleozoic-Mesozoic cover);

- Borăscu-Ilovița Nappe (Proterozoic metamorphic basement and Upper Paleozoic-Mesozoic cover).

Basement stratigraphy of the Austrian Nappes making-up the Lotru-Bistra (Laramian) Nappe will be presented according to Balintoni et al. (1989):

- Cumpăna (litho) Group of Proterozoic age in the Valea lui Stan-Căineni Nappe, represented by paragneisses and micaschists with kyanite and staurolite, amphibolites and Cozia (augen) Gneisses (Hann, 1994);

- Bocșița-Drimoxa and Sibișel (litho) Groups of Proterozoic age in the Bocșa-Uria Nappe (garnet-bearing paragneisses and amphibolites in Bocșița-Drimoxa and garnet-staurolite mica-gneisses, carbonate, graphitic and amphibolitic rocks for Sibișel, both strongly retrogressed) and Lower Paleozoic low grade metamorphic Caraș Group, represented by a bimodal volcano-sedimentary formation, overlain by black slates and quartzites (Iancu, 1984; Hann, 1994).



— Proterozoic Cumpăna Group in the Tâlva Drenii-Măgura Nappe, consisting of augen gneisses (Cozia Gneiss), paragneisses and amphibolites (Iancu, 1984, Hann, 1994); -

— Upper Devonian-Lower Carboniferous Moniom Group in the Moniom Nappe, represented by very low-grade basic tuffs, carbonatic and pelitic rocks and conglomerates (Iancu, 1985);

— Proterozoic Sebeș-Lotru (litho) Group in the Getic Nappe, the most extended metamorphic group in the South Carpathians, represented usually by a lower migmatitic complex and upper, kyanite-staurolite bearing micaschists; manganese silicate rocks and tectonic inclusions of anisofacial rocks, as metaultramafites, eclogites and granulites are hosted (Iancu, Mărunțiu, 1994); pegmatitic dykes are also typical, in places with economic importance (Lotrului Mts and Banat area); low-grade Lower Paleozoic Miniș Group (metaterigenous) and Buceava Group (including strongly broken metabasalts in black shales and carbonate rocks) crop out in an antiform refolding an older thrust-fault with Sebeș-Lotru Group rocks in the hangingwall; the age of this thrust plane predates the intrusion of the Variscan Ponișca granitoid (Iancu et al., 1988).

— Proterozoic Sebeș-Lotru (litho) Group in the Borăscu-Ilovita Nappe, which is a local slice from the Getic Nappe, exposed at the sole of the Godeanu outlier and East of Orșova.

The covers of the Austrian nappes making up the (Laramian) Lotru-Bistra Nappe form minor volumes compared to their pre-Alpine metamorphic basements, with the exception of the Sasca-Gornjak-Reșița Nappe which is a pure cover nappe. The Upper Paleozoic-Mesozoic formations are (Codarcea et al., 1961; Sădulescu, 1984):

— in the Valea lui Stan - Căineni Nappe, Permian conglomerates and Triassic marls ;

— in the Bocșa - Uria Nappe , Middle Jurassic sandstones , Lower Cretaceous (Urgonian) limestones (exposed only in the Bocșa area);

— in the Tâlva Drenii- Măgura Nappe, Upper Carboniferous conglomerates with volcanic sequences, Middle Jurassic limestones, Lower Cretaceous (Urgonian) limestones ;

— in the Moniom Nappe, Middle - Upper Jurassic sandstones and limestones ;

— in the Sasca-Gornjak-Reșița Nappe , Upper Carboniferous conglomerates and sandstones , Permian black schists and red sandstones, Lower - Middle Triassic conglomerates followed by limestones Lower Jurassic detrital rocks , Middle Jurassic marls and Upper Jurassic limestones;

— in the huge outcrop area of the Getic Nappe, several zones are known which preserved Alpine covers, from Reșița - Moldova Nouă in the West , to the western slope of Sebeș Mts and the Vânturarița Mt. in the East: Upper Carboniferous coal-bearing conglomerates and sandstones, followed by Permian black schists and red sandstones (Reșița Zone), Lower Jurassic detrital sequences (coal bearing in Reșița Zone), Middle Jurassic marls or sandstones, Upper Jurassic limestones, Lower Cretaceous marls and (Urgonian) limestones represent the dominant lithologies;

— in the Borăscu-Ilovita Nappe the cover is represented by Permian (?) red conglomerates and Lower Cretaceous marls and sandstones.

Upper Cretaceous deposits represent the post-tectogenetic cover common to the Austrian Nappes making up the Lotru - Bistra (Laramian) Nappe. They are exposed as red Senonian marls in Pogănișului Valley or vulcano-sedimentary formation in the Rusca Montană and Densuș areas, as sandstones followed by red clays with Dinosaurians in Hațeg area in the center, and as conglomerates, sandstones and marls in the Vânturarița - Băile Olanesti area.

Similar to the Timiș - Boia (Laramian) Nappe , the Upper Paleozoic-Mesozoic covers of the Austrian Nappes belonging to the Lotru - Bistra (Laramian) Nappe and the common post-Austrian cover show no metamorphism and strong deformation , their sedimentary and paleontological features being well preserved.

3. The Severin Nappe

The Severin Nappe represents in the South Carpathians the Outer (External) Dacides, obducted slices from a rift with oceanic/thinned continental crust formed in Jurassic on the European margin (Sădulescu, 1975, 1984, 1994). According to Codarcea (1940) and Sădulescu (1984), the Severin Nappe was overthrust by the Getic Nappe (in fact by the Getic Nappe s.str. from the bottom of the Laramian Lotru - Bistra Nappe) in Middle Cretaceous (first "Getic phase"); in block, at the end of the Cretaceous, both nappes were overthrust onto the Danubian (second "Getic phase"). The Severin Nappe, defined by Codarcea (1940), includes Upper Jurassic ophiolites and pelagic deposits (red or green marls and siliceous rocks associated to basalts - Azuga Beds) and Lower Cretaceous flysch deposits - Sinaia and Comarnic Beds.

The ophiolitic sequence, known as the Obarșia Complex (Stănoiu, in Bercia et al., 1977) is exposed mainly in the Mehedinți Mountains (discontinuously bordering the Balna outlier) and to a smaller extent



in the northern Parâng Mountains (Stănoiu et al., 1992). The Obârșia Complex is an ophiolitic mélange (Mărunțiu, 1987), consisting of highly broken oceanic plate rocks - basalts, gabbros, harzburgitic ultramafites and siliceous pelagic deposits. Geochemistry of basaltic rocks indicates ocean floor tholeiites (Cioflica et al., 1980), interpreted as an obducted marginal basin crust (Mărunțiu, 1987), or as a transitional ridge segment (T-type MORB) (Savu et al., 1991). Basalts show a subduction related low temperature metamorphism in prehnite-pumpellyite facies (Seghedi et al., 1994).

The Sinaia Beds (Upper Tithonian-Lower Vallanginian - Stănoiu, 1978a) represent distal limestone turbidites, dominated by facies D with thinly bedded T_{cd} and T_{de} Bouma divisions. Pelagic limestone interbeds yielded a microfauna rich in *Calpionella alpina*, *Calpionella eliptica* and *Tintinnopsella carpathica*. Pelitic lithofacies are rich in trace fossils, mainly Chondrites and Helminthoidea, suggesting a deep water depositional environment.

4. The Danubian Thrust Sheets

The Danubian thrust sheets (Danubides, Danubicum) represent in the South Carpathians the Marginal Dacides. They are slices from the continental crust of the European margin, bordered northwards by the Outer (External) Dacides trough and pass southwards to the Moesian Platform (Săndulescu, 1984, 1994). They are recognised in Bulgaria in the Prebalkans and the Stara Planina (Săndulescu, 1984), exposing pre-Alpine granitised basements and Permian-Eocene covers, the latter showing strong affinities both with the sedimentary sequence of the Moesian Platform (thick Permian and Triassic German-type sequences) and with the Danubian Jurassic-Cretaceous deposits. Most of the Danubian covers involve Upper Cretaceous sequences, indicating that Danubian Nappes were emplaced during the same shortening episode (the "second Getic phase" - Săndulescu, 1984) as their hangingwall units, the Getic and Severin Nappes, in the Laramian tectogenesis. Excepting the major division between Upper and Lower Danubian Nappes (Berza et al., 1983), known also as the External and Internal Danubicum, the names, areal extent, stratigraphy and number of thrust sheets differ largely from one author to another (Codarcea, 1940, Codarcea et al., 1961, 1968; Stănoiu, 1973; Kräutner et al., 1981, 1988; Berza et al., 1983, 1986b, 1988a, 1988b; Gherasi et al., 1986; Iancu et al., 1990; Năstăsescu et al., 1988a, 1988b; Pop, 1988; Balintoni et al., 1989). A simplified version according to Balintoni et al. (1989) will be presented here (PL. II).

4.1 The Upper Danubian Nappes

These nappes are exposed in the western and northern parts of the Danubian Window, between the contact with the Getic Nappe as roof thrust and the Lainici (Lower Danubian) Nappe as floor thrust; in Bulgaria their equivalents crop out in the Stara Planina (Săndulescu, 1984). From top to bottom, the Upper Danubian Nappes show the following geometry (Fig. 1):

- Arjana Nappe (Mesozoic cover deposits);
- Urdele-Măru-Svinecea Nappe (Proterozoic basement and Mesozoic cover);
- Poiana Mărului-Cornereva Nappe (Proterozoic basement and Paleozoic-Mesozoic cover);
- Godenele-Scorila Nappe (Proterozoic basement and Mesozoic cover).

In contrast with the basement of the Getic-Supragetic thrusts, the metamorphic basement of the Danubian thrusts was for long time regarded as lower-grade in fact, the green-schist appearance is the effect of penetrative greenschist facies mylonitisation, connected with both pre-Alpine and Alpine deformation. The Upper Proterozoic age of the early, amphibolite facies metamorphism of the basement formations is supported by geochronological evidence in the basement rocks of the Lower Danubian Nappes, which yielded radiometric ages ranging between 650-550 Ma (Grünenfelder et al., 1983; Liégeois, unpublished data; Dallmeyer et al., 1994). The Upper Danubian Nappes show the following basement stratigraphy:

- Drăgșan (Măru, Zeicani) and Lainici-Păiuș (Măgura Marga) Groups of Proterozoic ages in Urdele-Măru-Svinecea Nappe;
- Poiana Mraconia, Lainici-Păiuș (Neamțu), Drăgșan (Zeicani, Bărnita) Groups of Proterozoic age in Poiana Mărului-Cornereva Nappe;
- Drăgșan (Godenele) Group in Godenele Nappe.

The names Drăgșan and Lainici-Păiuș (previously used only for the Lower Danubian Nappes) were introduced by correlation for the basements of the Upper Danubian Nappes (Berza, Seghedi, 1983; Iancu, 1983), in an attempt to replace local terminology. Just like in the type-area, Drăgșan Group is a prevailing amphibolitic sequence, with associated quartz-plagioclase gneisses (leptynites) and kyanite-staurolite bearing muscovite gneisses and micaschists (Savu et al., 1984; Mărunțiu, 1987). The metaterigenous Lainici-Păiuș



Group is a highly migmatized, prevailing quartzitic sequence, with biotite gneisses, amphibolites, marbles and calcisilicate rocks (Gherasi et al., 1968; Anastasiu, 1976; Dimitrescu et al., 1994). Drăgșan and Lainici-Păiuș Groups are both intruded by granitoid plutons (Muntele Mic, Sfârdinu, Cherbezeu, Ogradena, Ștevia), and a pre-Alpine overthrust of the former onto the latter was supposed in Țarcu Mountains (Săndulescu, 1984) and Parâng Mountains (Berza et al., 1986b). The Poiana Mraconia Group is constituted mainly of mica gneisses, amphibolites, quartzites and migmatites; before the Upper Carboniferous, it was overthrust onto the Iuți ultramafic complex and onto the Cherbezeu granitoid pluton (Bercia, Bercia, 1980).

The covers of the Upper (and also of the Lower) Danubian Nappes have two characteristic features, contrasting with the Getic-Supragetic covers: (1) a very low-grade Alpine metamorphism and (2) the presence of Lower Paleozoic formations, displaying paragenetic (and sometimes, structural) conformity with the Upper Paleozoic-Mesozoic formations. The Upper Danubian covers show distinct stratigraphy in each unit:

— Lower Jurassic Armeniș Formation, consisting of sandstones, siltstones and shales, with basaltic tuffs and basalts; Middle-Upper Jurassic volcano-sedimentary Țarcu Formation with basic and alkaline lava flows, pyroclastic and epiclastic deposits, associated to stromatic limestones and red shales (Russo-Săndulescu, in Kräutner et al., 1990); Upper Cretaceous Arjana flysch (Codarcea, 1940) - in the Arjana Nappe;

— Devonian sandstones and slates of the Răul Rece and Drencova Formations; Lower Carboniferous Ideg Limestones and Sevastru Slates, followed by Permian conglomerates; Lower Jurassic Schela Formation, dominated by black conglomerates, sandstones and slates, with chloritoid and pyrophyllite; Middle-Upper Jurassic-Lower Cretaceous Lupeni Limestones and Upper Cretaceous Iseroni flysch (Năstăsescu, 1979; Iancu et al., 1990; Berza et al., 1986b, 1988a, 1988b; Codarcea, 1940) - in the Urdele-Măru-Svinecea Nappe;

— Lower Paleozoic Nijudin Formation with metaterigenous and metabasic phyllites; Silurian (?) Brustur Formation (including Baicu Conglomerates which rework basic-ultrabasic rocks of the Iuți complex); Upper Carboniferous coal bearing conglomerates, sandstones and slates; Permian red-beds with rhyolitic volcanic and volcanoclastic rocks; Lower Jurassic coal bearing Bogăltin and Ohaba Formations with conglomerates, sandstones and slates (Presacina facies) and Lower Jurassic oolitic limestones, coal bearing conglomerates and shales, siliceous and carbonatic sandstones (Svinița Zone); Middle Jurassic marls and iron oolitic limestones (Svinița Zone); Upper Jurassic limestones and Lower Cretaceous marls - in Poiana Mărlui-Cornereva Nappe (Codarcea et al., 1961; Năstăsescu, 1979; Iancu et al., 1990);

— Lower Jurassic detrital deposits in Godenele-Scărița Nappe.

Intensity of Alpine metamorphism in each of the covers varies between high-grade diagenesis and low-grade metamorphism, with an eastward increase in PT conditions, most obvious in the Urdele-Măru-Svinecea Nappe.

4.2 Lower Danubian Nappes

These nappes are exposed in the central and southern parts of the Danubian Window, covered by the Severin Nappe in the SW, by the Upper Danubian Nappes in the North, or directly by the Getic Nappe in the East; in Bulgaria, their correspondents crop out in the Prebalkans. From top to bottom, their geometry was described by Berza (in Balintoni et al., 1989):

— Lainici Nappe (Proterozoic metamorphic-granitic basement and Paleozoic-Mesozoic cover);

— Schela-Petrcanu Nappe (Proterozoic metamorphic-granitic basement and Mesozoic cover);

— one or several unknown thrust sheets, until the floor thrust of the Danubian duplex (Seghedi, Berza, 1994), the thrust onto the Moesian Platform.

The basement of the Lower Danubian Units is represented by two main sequences - the Lainici-Păiuș and Drăgșan Groups, typical for the entire Danubian. The Lainici-Păiuș Group is mostly metaterigenous, with a lower Carbonatic-Graphitic Formation (Mg-silicates bearing marbles and cordierite-andalusite-sillimanite bearing graphitic schists) and an upper Quartzitic Formation; low P-high T metamorphism, large scale migmatization and intrusion of granitoid batholiths occurred 550-650 Ma ago, as indicated by U-Pb, K-Ar and Ar-Ar ages in granitoids (Grünnfelder et al., 1983; Dallmeyer et al., 1994; Liégeois, unpublished data). The leptino-amphibolitic Drăgșan Group, with a lower Augen Gneiss, a main Amphibolitic Formation and an upper, kyanite-staurolite mica-gneiss Formation shows no migmatization, but is intruded by several granitic plutons. K-Ar and Ar-Ar ages of these plutons do not exceed 300-400 Ma, but Rb-Sr and Sm-Nd isotopes indicate 600 Ma (Liégeois, unpublished data).

The basement of the Lainici Nappe is the classical area for recognising the Variscan overthrust of Drăgșan onto Lainici-Păiuș Group. Because a Devonian cover (Stănoiu, 1982) is exposed in places in the footwall and the thrust is sealed by Permo-Mesozoic deposits, the thrusting was timed as Variscan. Dynamic deformation is most obvious in the footwall, where the Lainici-Păiuș rocks underwent greenschist facies mylonitization



over a great thickness (one km or more) (Berza et al., 1983; Berza, Iancu, 1994). Alpine tectonics also contributes to widespread greenschist facies retrogression above and below the thrust planes.

The basement of the Schela-Peteanu Nappe is mostly of Lainici-Păiuș type, with important graphite ores in the Carbonatic-Graphitic Formation from the Parâng Mountains; the Drăgăsan Group is not present in this nappe. In the Petreanu Mountains, different sequences are exposed below the Lainici-Păiuș Group (locally named Bodu) and its granitoids and migmatites, downwards from a mylonitic zone with upper greenschist facies mylonites: a sheet of Furcătura Gneiss, then a sequence of leptyno-amphibolites showing biotite zone retrogression, overlying micaschists with large garnet porphyroblasts (Rof Formation), the latter very unusual for the Danubian metamorphic basement (Kräutner et al., 1988).

The Lower Danubian covers show different stratigraphy in each of the Lower Danubian Nappes:

— Upper Ordovician-Lower Silurian Valea Izvorului Formation, with quartzites and slates; (Ordovician?) Coarnele Formation with quartzites; Devonian Valea de Brazi Formation, with conglomerates, sandstones, slates and rhyolites; Permian red-beds; Lower Jurassic Gresten facies deposits (conglomerates, sandstones and slates), called Baia de Aramă and Schela Formations; Middle Jurassic sandstones with carbonate cement; Upper Jurassic-Lower Cretaceous limestones, Middle Cretaceous marls (Nadanova Beds) and Upper Cretaceous turbidites - in the Lainici Nappe (Pop, 1973; Stănoiu, 1973, 1982; Drăgănescu, in Berza et al., 1986b, 1989b).

— only Lower Jurassic terigenous deposits (Gresten facies), known as the Schela Formation in Schela-Peteanu Nappe (Drăgănescu, in Berza et al., 1986b, 1989b; Stănoiu et al., 1992).

There is no striking deformational or metamorphic contrast between the Paleozoic and Mesozoic cover members. For the Lainici Nappe, most important are the cover thickness (increasing southwards) and the latitudinal position, sedimentary and paleontological evidence being preserved only in the South. A typical example is the case of the Upper Jurassic-Lower Cretaceous limestone sequence, showing northward increase in degree of recrystallisation and shearing (loosing gradually all primary sedimentary and diagenetic features), as it is perfectly documented along the eastern border of the Danubian Window (Polovragi-Funicelu saddle area). Index minerals occur in appropriate lithological compositions: chloritoid-pyrophyllite in Lower Jurassic quartzitic sandstones and slates, rich in organic debris (Schela Formation) and prehnite-pumpellyite in Upper Cretaceous volcanoclastic sandstones (Seghedi, Iancu, this volume). Illite crystallinity index in pelitic compositions varies considerably between diagenesis and low-grade metamorphism, the Alpine metamorphism of Mesozoic cover rocks being of very low-grade type and only in places reaching the chlorite isograde (Iancu et al., 1984; Seghedi et al., 1994).

III POST-NAPPE GEOLOGY

During Cenozoic times, the Cretaceous nappe stack underwent not only elevation, exhumation and (partial) erosion, but also important magmatic, tectonic and sedimentary activity.

1. Magmatic activity

In the western South Carpathians, a belt of calc-alkaline to subalkaline intrusions emplaced into the Getic-Supragetic Nappes are known since more than a century as "banatites". On Romanian territory, they form two lineaments: a western one, with important plutons, and an eastern one, with dyke-swarms (Russo-Săndulescu, Berza, 1979). The plutonic belt follows the Carpathian bend, passing from a N-S to an E-W direction in the southern Poiana Ruscă Mountains; in the easternmost outcrop area, banatites cut the Upper Cretaceous deposits South of Bara (16 km west of Petroșani). Petrological and geochronological studies of the banatites from the Romanian South Carpathians were published by Russo-Săndulescu et al., 1986), revealing both distinct differentiation trends and ages between the older, subalkaline western intrusions (85–75 Ma) and the younger, eastern intrusions (65–55 Ma) of the plutonic zone. The dyke-swarms show only structural differences (hypabissic porphyritic rocks) compared to the eastern plutonic intrusions. Important Fe, Cu, Mo and base metal deposits of skarn or porphyry-copper type are connected to the banatites. The banatites cut through the thrusts between Austrian nappes of the Lotru-Bistra (Laramian) Nappe: Bocșa, Tâlva Drenii-Măgura, Moniom, Sasca-Gornjak-Reșița and Getic; at Tincova, a pluton seals to the West even the Laramian thrust between the Timiș-Boia and Lotru-Bistra Nappes. The banatitic magma was generated by consumption of the subducted oceanic lithosphere of the Outer Dacides Rift, and subsequently raised through the hangingwall to give the plutons cutting the Getic and Supragetic nappes (Rădulescu, Săndulescu, 1973).



2. Tectonic activity

Even if the Southern Carpathians do not show (as the Eastern Carpathians) an outer belt with Neogene deformational history, in the Cainozoic they suffered important vertical and horizontal displacements related to several extensional and compressional events. An early extension was responsible for normal faulting and basin formation in Hațeg or Titești Basins, with Eocene deposits. Then pre-Chattian dextral slip faulting developed along several lineaments, the most obvious being the Cerna-Jiu Fault system with a displacement of 30-40 km (Berza, Drăgănescu, 1988) and the initiation of Petroșani basin, a pull-apart half-graben. The Balta-Baia de Aramă graben, with Miocene infill, might also be initiated this way. Sarmatian compressions were responsible for the final overthrust of the South Carpathians onto the Moesian Platform, as well as for reverse faulting occurred west of Lupeni along the northern border of the Petroșani Basin. Late E-W extension produced N-S trending normal faults, some even noticeable in the morphology, like those located at the western margin of the Parâng crest (Sadu-Petrila). A kinematic analysis in the Central South Carpathians (Ratschbacher et al., 1993) proposed a somewhat different structural history: ductile-brittle dextral wrenching, E-W compression and formation of the Petroșani Basin along the Cerna-Jiu Fault system during the Paleogene: large-scale Miocene dextral wrenching along the northern margin of Moesia; probably Pliocene-Early Pleistocene N-S compression connected with sinistral wrenching.

3. Intramontane basins

Despite the stiffness of the chain, the South Carpathians include several intramontane basins: Titești, Vidra, Petroșani, Hațeg-Strei, Caransebeș-Lugoș, Bozovici, Bahna. If most of them represent only Neogene molassic depressions, Brezoi and Hațeg basins also involve Paleogene deposits and are not easy to separate from the sedimentary covers of the underlying Laramian units.

Deposits of the Hațeg Basin overlie basement and cover formations of the Getic (Austrian) Nappe of the Lotru-Bistra (Laramian) Nappe (Sebeș-Lotru metamorphics and Lower Jurassic-Aptian cover) and of the own cover (Cenomanian-Turonian) of the Lotru-Bistra Nappe. Maastrichtian-Paleocene red beds (conglomerates, sands and clays with Dinosaurs) represent the first post-Laramian cover member. The basin fill continues with Rupelian clays and sandstones, Aquitanian green marls, Langhian sandstones and conglomerates and Sarmatian clays and sands (Popescu, in Berza et al., 1986a, 1989a).

The Titești Basin seals the Laramian thrust, emplacing Timiș-Boia Nappe on top of the Lotru-Bistra Nappe. Paleocene conglomerates of this basin are overlain by Ypresian silty marls and sandstones, Eocene marls and clays, Lower Miocene conglomerates, sandstones and tuffs (Ștefănescu et al., 1982).

The southern border of the Petroșani basin lays on Sebeș-Lotru metamorphic rocks from the Getic Nappe (Austrian) of the Lotru-Bistra (Laramian) Nappe; the northern border of this basin is a reverse fault of the Cerna-Jiu system (Berza, Drăgănescu, 1988). Basin sequences include: lower Rupelian red deposits (sandstones, conglomerates) and fresh-water limestones; Middle-Chattian coal bearing formation (clays, sandstones); Upper Chattian detrital sequence (sandstones, conglomerates) (Moisescu, in Berza et al., 1986a); at the eastern margin of the basin, younger gravels (Miocene and Pannonian) are exposed, reaching heights 700 m. In the Lotru valley, at 1300 m altitude, the small Vidra Basin with Tortonian clays and sands rests on Sebeș-Lotru basement rocks of the same unit as that underlying the Petroșani Basin.

The western basins (Strei, Caransebeș-Lugoș, Bozovici, Bahna and the eastern bays of the Pannonian) rest on both Timiș-Boia and Lotru-Bistra Laramian Nappes and consist only of Neogene (Miocene and Pannonian) formations.

Even if the source area of the Maastrichtian, Paleogene and Neogene formations from the intramontane basins is undoubtedly Carpathian (mostly of Getic-Supragetic provenance), geological and geodetic data indicate that the upraisal of the present South Carpathian mountain belt is very young and even in progress.

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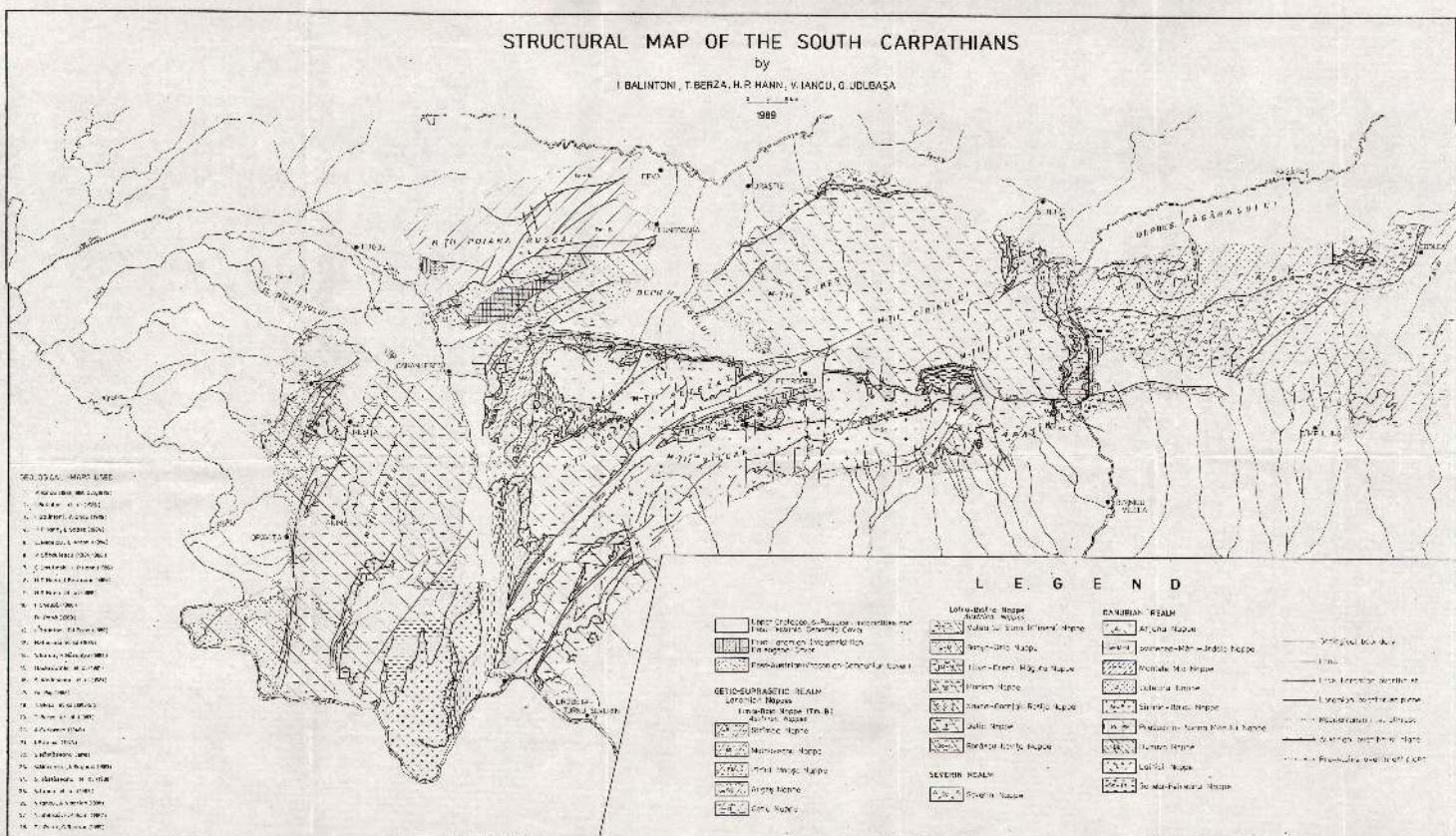
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STRUCTURE OF THE APUSENI MOUNTAINS

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I. General Geological Framework

The Apuseni Mountains represent an isolated massif of the Carpathian arc, located at its interior. They are separated from the Transylvanian basin and the South Carpathians by the Mureş zone, where ophiolite tectonic units crop out. These are known as the western Transylvanides and represent the expression of the major Tethyan suture (e.g. Debelmas, Săndulescu, 1987). The south-western prolongation of the Apuseni Mountains (Voivodina) is separated from the Dinarides by the south-Pannonian suture (e.g. Săndulescu, 1984). The major Tethyan suture, together with its north-western prolongation (the Pieniny Klippen belt) and the south Pannonian suture, confine a realm with a specific Triassic development (increasingly south Alpine in character in a southern direction), a specific paroxysmal tectogenesis (pre-Gosauian or Mediterranean) and a specific post-tectogenetic sedimentary cover (the Upper Cretaceous Gosau facies sediments) (e.g. Andrusov et al., 1973; Săndulescu, 1975; Kovacs, 1982; Săndulescu, 1984). This realm can be considered the continental crust of one single plate – the Austro-Bihorean block (Săndulescu, 1984) or the pre-Apulian plate (Săndulescu, 1994) – or of two microplates, the Tisia to the east (Kovacs, 1982) and the Carnic to the west (Dewey et al., 1973). If we consider the common pre-Gosau tectogenesis and Gosau post-tectogenetic cover for both the Western Carpathians and the Apuseni Mountains, then one can begin to move the two microplates only at late Turonian times. Thus, the intra-Pannonian mobile belt is not older than the Senonian and that explains the post-Santonian age of the Szolnok flysch (e.g. Nagymarosy, Baldi-Beke, 1993). The intra-Pannonian mobile belt is the result of some transcurrent faulting (Patruşiu et al., 1971). This mobile belt has been connected with the Pieniny Klippen belt until the Burdigalian time, when they have been deformed together (Nagymarosy, Baldi-Beke, 1993).

The pre-Gosau tectogenesis generated the basement nappes of the Apuseni Mountains – the Apusenides (e.g. Săndulescu, 1984) – by shearing off the continental crust of the pre-Apulian plate. Most correlations done up to date (e.g. Săndulescu, 1975, 1984; Debelmas, Săndulescu, 1987) connect the Transylvanides with the Pienides and the Apusenides with the Western Carpathians.

II. The Structure and Lithostratigraphy of the Apuseni Mountains

1. General problems. The Apuseni Mountains nappes have been emplaced during three tectogeneses: the first mid-Cretaceous, the second pre-Gosau and the third Laramian. Considering Royden's (1993) geotectonic criteria for all the contractional period mentioned above, the tectonic regime at the convergent boundaries between the pre-Apulian and Getic plates show advancing subduction boundaries, because: (1) the nappes are antithetical; (2) the crystalline basement is strongly deformed; (3) the post-collisional convergence lasted over 25 Ma; (4) the foredeep basins practically do not exist. During the Middle Tertiary, the tectonic regime shows retreating subduction boundaries, but related with convergence in the external Dacian rift, because the plate boundaries in the Mureş zone were blocked after late-Cretaceous times. The main expressions for this regime are the appearance of some extensional basins and crustal thinning within the Pannonian area (e.g. Royden, 1988). The responses within the Apuseni Mountains are similar: the appearance of the extensional basin, striking in the same direction as the Pannonian basins, and development of the Tertiary calc-alkaline magmatism, running in the prolongation of the basins as far as the south-eastern boundary of the pre-Apulian plate. The mid-Cretaceous tectogenesis can be recognized within the Transylvanides, the pre-Gosau events created the Apusenide edifice and the Laramian events affected



the Transilvanides again. Reactivation of the Transilvanides during the Laramian tectogenesis reflects the gradual immobilization of the plate boundaries along the Mureş zone.

Plate motion data (Dercourt et al., 1986, fide Ratschbacher et al., 1993) show a constant NE to E displacement of the Getic plate since mid-Cretaceous. But because the Mureş zone curvature parallels the bend of the South Carpathians (Braşov), and both of them have been formed in the same time span, it may be inferred that the eastern part of the pre-Apulian plate followed a similar path. The displacement was sensibly faster between 110–80 Ma.

2. The Transilvanides. The Transilvanides proceed from the major Tethyan lithosphere. According to Nicolae et al. (1992) the pre-collisional tectonic setting was that of an island arc, accompanied by a marginal basin. The subduction was directed beneath the pre-Apulian plate. The magmatites are represented by a tholeiite series, a calcalkaline series and some spilites – Middle Jurassic – Lower Cretaceous in age – as well as Early Cretaceous granitoids, intruded in the arc magmatites (e.g. Ştefan, 1986).

2.1. The Austrian Transilvanides. The first geologist who mapped a mid-Cretaceous nappe in the Trascău Mountains was Ilie (1936), who named it the Metaliferi Mountains Nappe. He attributed the Tithonian limestones of the Bedeleu summit to the main body of this nappe. The idea was re-emphasized by ophiolite obduction models (Ianovici et al., 1976 and Bleahu et al., 1981), especially due to the Albian wildflysch outcrops.

In the Trascău Mountains, several Austrian nappes have been mapped (Balintoni, Iancu, 1986). The separation of these nappes was possible only in places, not only because the overlying Cretaceous cover hides contacts, but also because of the subsequent reactivation of the Austrian thrusts during the late-Cretaceous (Laramian) events.

Balintoni, Iancu (1986) have separated the following Austrian tectonic units within the Trascău Mountains: the Izvoarele Nappe (Aptychus beds); the Valea Muntelui Nappe (ophiolites and associated jaspers and limy rocks); the Colţu Trascăului Nappe (detrital Permian rocks and the Paleozoic low-grade Trascău series); the Bedeleu Nappe (the massive Stramberg limestones from Bedeleu summit: Tithonian-Neocomian) and the associated ophiolitic rocks at their bottom. We accept the same age to the Ardeu Unit separated by Mantea (in Borcos et al., 1981), and for the Căbeşti and Bejan Units considered by Bleahu et al. (1981) Laramian. The Căbeşti Unit is the lower part of the ophiolitic Laramian Căpâlnaş-Techereu Unit, while the Bejan Unit is probably the single southern Transilvanide situated in the same position as the eastern Transilvanides, above the Median Dacides (or Getides) (Săndulescu, 1984). The Transylvanian Tethys has closed during the Austrian tectogenesis and did not reopen afterwards. However, the suture remained mobile until the Laramian tectogenesis.

2.2. The Laramian Transilvanides. Bleahu et al. (1981) distinguished the following Laramian Transilvanides: the Bucium Unit; the Groşi Unit; the Criş Nappe; the Feneş Nappe; the Frasin Nappe; the Vulcan Nappe; the Căbeşti Unit; the Bejan Unit; the Valea Mică-Galda Nappe; the Bozeş Nappe; the "Bedeleu Nappe System", having mid-Cretaceous internal relationships. In accordance with the tectonic sketch (Fig.) we accept the following Laramian Transilvanides: the Bucium Unit; the Groşi Unit; the Criş Nappe; the Feneş Nappe; the Frasin Nappe; the Vulcan Nappe; the Căbeşti Unit; the Bejan Unit; the Valea Mică-Galda Nappe; the Bozeş Nappe; the "Bedeleu Nappe System". In this scheme the Valea Mică-Galda Nappe is attached to the Căpâlnaş-Techereu Nappe, conform to the Zlatna sheet of the 1:50000 scale map (Borcos et al., 1981); the "Bedeleu Nappe System" is attached to the Feneş Nappe; the Căbeşti Unit is considered the lower part of the Căpâlnaş-Techereu Nappe, and the Bozeş Nappe is in fact the Senonian cover of the Rapolt crystalline metamorphic inlier, which in turn belongs to the South Carpathians. This final observation has a great significance, because no post-Senonian transcurrent faults can be traced between the Apuseni Mountains and the South Carpathians. Thus, the Laramian boundary between the Apuseni Mountains and the South Carpathians plus Transylvanian basin runs along the northern tectonic margins of the Bejan Unit and Bozeş Beds.

In the mid-Cretaceous time, the Bedeleu Nappe, the Căpâlnaş-Techereu Nappe, the Căbeşti Unit, the Vulcan Nappe and the Ardeu Nappe probably formed a single tectonic body, consisting mainly of island arc ophiolites and Stramberg limestones. The Frasin Nappe was part of the Feneş Nappe, and the Colţu Trascăului nappe involves the continental crust of an island. In accordance with Nicolae et al. (1992) the Feneş and the Criş Nappes include marginal basin ophiolites which were parts of the same basin during the mid-Cretaceous times.



The Laramian fault structures are initiated as reverse faults in the northern zone, where they strike north-southward. In the southern zone they gradually increase in amplitude and take a WSW strike. The inflexion takes place in the Zlatna-Alba Iulia zone and reflects the clockwise rotation of the ensemble formed of the pre-Apulian and Getic plates around the Moesian (Euxinic) corner (Stage two, Fig. 12, Ratschbacher et al., 1993). It is worth mentioning that the Getic and pre-Apulian rotations around the Moesian (Euxinic) corner began late in the Early Cretaceous, because the transform fault between the Moesian (Euxinic) plate and the East-European plate (the Peceneagă-Camena fault) is sealed by the "Babadag basin", where sedimentation started in the Vraconian (e.g. Săndulescu, 1984). Otherwise, the Laramian nappes could not be recognized anymore on the northern boundary of Transylvanian basin.

As it was earlier emphasized, along the segment between Căpâlnaş and Alba-Iulia, the border between the Laramian Transylvanides and the South Carpathian Laramides is represented by the South Transylvanian Fault (Visarion, Săndulescu, 1979). This fault represents a complex convergent - transform boundary between the pre-Apulian and Getic plates, allowing repeated dextral slips. Its importance is emphasized at the western end of the fault by the alpine low grade metamorphism, of the Căbeşti beds (Dinică et al., 1994). The fault is sealed by the Laramian magmatites from the Mureş valley (Roşu, verbal communication, 1994). This fault also represents the axis of vergency change of the Laramian overthrusts (northern vergency in the Metaliferi Mountains, southern vergency in the South Carpathians). Westwards of the Rapolt crystalline unit the fault branches, one branch separating the pre-Apulian and Getic plates, the other one the Getic and Moesian (Euxinic) plates. The Getic plate is completely sheared out along the segment discussed in this text, enabling the underplated northern boundary of the Moesian (Euxinic) plate to come into contact with the pre-Apulian plate.

3. The Apusenides. The Apusenides are basement and cover nappes localized outside of the Transylvanides. They have been emplaced during the pre-Gosauian tectogenesis. General descriptions of the Apusenides are given in Ianovici et al. (1976), Bleahu et al. (1981) and Săndulescu (1984). Depending on the presence or absence of sedimentary covers and of their facies, the Apusenides have been divided into two groups: the Codru Nappe System, built up of basement units and their covers, or only of cover units, and the Biharia Nappe System, made up of mainly or only basement units. Considering the specific features of the basement and the facies of the Permo-Triassic cover, the Arieşeni Nappe of the Codru Nappe System should be included to the lower part of the Biharia System.

3.1. The Bihor Unit. We mention first the Bihor Unit for it forms the foundation of all the Apusenides. The Bihor Unit can be accepted as autochthonous, being composed of the medium grade polymetamorphic Someş Series, the Muntele Mare intrusive granitoids and a Permo-Mesozoic cover, that reaches up until the Turonian (e.g. Bleahu et al., 1981). The sedimentary facies of the Mesozoic sequence is the most distal one comparatively with the South Alpine realm.

3.2. The Codru Nappe System. It follows above the Bihor Unit. We adopt here the classification scheme proposed by Bleahu et al. (1981), and modified by Săndulescu (1984). The essential feature is, excluding the Finiş (-Ferice-Gârda-Următ) Nappe, that it develops only westwards of the Muntele Mare. It includes the following units (from the lower part to the upper part, or from outside to inside): (a) the Valani Nappe; (b) the Finiş Nappe; (c) the Dieva-Bătrâncu (-Vetre) Nappe; (d) the Moma Nappe; (e) the Vaşcău-Coleşti Nappe.

(a) The Valani Nappe is a cover nappe built up of Permian to Albian deposits, similar to the cover of the Bihor Unit.

(b) The Finiş Nappe is a major unit of the Codru Nappe System. It has a basement formed of the Codru medium-grade polymetamorphic Series, pierced by the venitic Codru migmatites in the eastern part, and by well individualized granitoid bodies in the Codru-Moma and Highiş massifs. The Codru migmatites probably belong to the (M2) late-Proterozoic metamorphic event, while the western granitoids, surrounded by hornfels, appear to be much younger possibly of Variscan age (Stan, 1989). The Finiş Nappe cover is composed of a sequence beginning with a rhyolite-bearing Permian and ending with a flysch type Neocomian pile. The cover is developed only in the western part of the Finiş nappe.

(c) The Dieva-Bătrâncu(-Vetre) Nappe. These are three discontinuous cover units, similar in sedimentary facies. Sedimentation may begin with a thick Permian sequence, characterized by a bimodal rhyolite-basalt volcanism, and ends in the Triassic, which is mainly limy in composition.

(d) The Moma Nappe is also a cover nappe which begins with the Permian containing the bimodal



volcanics and continues with a dominantly limy Triassic.

(c) The Vascau-Colești Nappe is situated at the top of the Codru Nappe System. It is formed by a Triassic-Lower Jurassic limestone sequence, identical with the upper Austro-Alpine facies.

Reminding that this short description of the Codru Nappe System was in a north-south facies order, the following correlations with the Western Carpathians are possibly: the Bihor Unit and the Tatrîdes on one hand, and the Codru Nappe System with the Veporides plus the Gemerides, on the other hand (e.g. Săndulescu, 1975, 1984; Kovacs, 1982).

3.3. *The Biharia Nappe System* consists of three major basement units: (a) the Arieșeni Nappe; (b) the Biharia Nappe; (c) the Baia de Arieș Nappe.

(a) The Arieșeni Nappe, the lower unit of the Biharia Nappe System, is formed by metabasites and a Permian rhyolite-bearing detrital sequence, overlain by a limy Triassic sequence. As a very interesting fact, the sedimentary sequence of the Arieșeni Nappe can be correlated with that of the Finiș Nappe of the Codru Nappe System (Kovacs, 1982), although the Arieșeni Nappe lies over the whole Codru Nappe System. Also, between Scărișoara and Garda village the transition from the Biharia Series to the "Arieșeni Series" is exposed, due to the development of a penetrative mylonitic foliation, affecting the Biharia Series metabasites and the Lunca Largă type granitoids, associated to the metabasites. Consequently, the "Arieșeni Series" represent a part of the Biharia Series which underwent a strong deformation, probably during the Upper Paleozoic, because the Permian-Mesozoic cover appears less deformed than the basement. The original position of the "Arieșeni Series" was next to the Biharia Series and its sedimentary cover was near the Finiș Nappe cover and not south of the Vascau-Colești Nappe within the South-Alpine domain. Near the Arieșeni village, in the Cobliș brook a conglomerate intercalation crops out. Until now there are no pertinent structural studies, to demonstrate if these rocks are deformed metaconglomerates or tectonic pseudo-conglomerates.

(b) The Biharia Nappe. The Biharia Nappe has a comparable development as the Finiș Nappe, covering initially the Bihor Unit, because it crops out from the eastern part of the Apuseni Mountains, near Huedin and as far as their south-western extremities, near Lipova. The Biharia Nappe has a complex constitution. As an alpine nappe, it is built up by the following tectono-stratigraphic entities: the Biharia Series and the conformable plagio-granitoids of Lunca Largă type, in accordance to Balintoni (1985, 1986); the Poiana Series, separated in the Bihor massif by Bordea et al. (1988); the Păiușeni Series, mapped in the Bihor massif by Bordea et al. (1988); the Cladova formation and the Păiușeni Series from the Highiș-Drocea massif, as are they sketched by Balintoni (1986); the Biharia and the Păiușeni Series of the Piatra Grăitoare slice, drawn by Bordea et al. (1988); the intrusive Upper Paleozoic Highiș granites (Balintoni, 1986); the Băișoara (Permian ?) conglomerates (according to Hărtopanu et al., 1982). The Alpine Biharia Nappe is formed of the following pre-Alpine units: the Lipova Nappe (Balintoni, 1986), consisting of the Biharia Series, Lunca Largă granitoids, the overlying Păiușeni Series in the Bihor Mountains and the Băișoara conglomerates; the Piatra Grăitoare Slice; the Highiș-Poiana Nappe, consisting of the Păiușeni Series in the Highiș-Drocea massif and of the Poiana Series in the Bihor massif. In the western part of the Highiș massif Dimitrescu (1967) figured for the first time a Saalic overthrust within the Păiușeni Series. Balintoni (1986) extended the Biharia Series in the Highiș-Drocea massif and traced that overthrust along the whole chain, between the Biharia Series and Cladova Formation, and considered a pre-Gosau tectonic contact between the Cladova Formation and the Păiușeni Series. Pană and Ricman (1988) showed that the Cladova Formation cannot be genetically separated from the Păiușeni Series. They considered the whole crystalline Highiș-Drocea massif as an Alpine shear zone affecting a bimodal magmatic association of the Paleozoic age, attributing this association to the Biharia Series.

(c) Geotectonic signification of the Biharia, Arieșeni and Păiușeni Series. All these tectono-stratigraphic units and associated granitoids represent elements of a major Paleozoic ophiolite suture, that is of oceanic lithosphere which has been obducted during the formation of the Variscan mobile belt. In this context, the Biharia Series can be considered a remnant of the ocean floor created during spreading, and the plagio-granitoids as the acid differentiates of the mid-ocean ridge magmatism. The Arieșeni Series comprises a part of the Biharia Series, much more intensively deformed and foliated during the period of convergence and consumption of the Biharia oceanic lithosphere. The Păiușeni and Poiana Series were probably deposited during the convergent movements in an island arc setting, the Highiș granitoids being the island arc acid intrusives. It is difficult to explain the first metamorphic event which affected the Biharia Series, because all the rocks are foliated, rotated albite porphyroblasts frequently occur, and the entire rock pile exhibits strong ductile flattening. All these tectono-stratigraphic entities underwent together an Upper Paleozoic mainly dynamic metamorphism which produced a horizontal axial-plane foliation. The bottom of the Biharia Series



was also strongly mylonized during the Mediterranean tectogenesis. The Biharia suture can be correlated to the Rakovec suture described in the Central Western Carpathians (e.g. Plasienka, 1993). It preserves the remnants of a southern branch of the Variscan ocean; the subduction was directed northwards and the nappes had a southward vergence.

(d) The Baia de Arieș Nappe is the uppermost unit of the Biharia Nappe System and it covers a similar area as the Biharia Nappe. This nappe has also a complex constitution and is formed by the following tectonostratigraphic entities: the Baia de Arieș Series which constitutes the main nappe body, including the Sohodol marbles developed south of the Arieșul Mic river; the Biharia Series from the Muncel slice; the metamorphosed association of the calcareous and quartzitic rocks, which sometimes have a conglomerate resemblance, known as the Vulturese-Belioara Series (e.g. Ianovici et al., 1976); the pre-Austrian cover of the Baia de Arieș Series. We admit a pre-Alpine contact between the Biharia Series of the Muncel slice, and the Baia de Arieș Series. The Vulturese-Belioara Series can be considered either a strongly deformed sequence of the Baia de Arieș Series or a part of the other lithostratigraphic unit transported at the base of the Baia de Arieș Nappe. As for the Păiușeni Series meta- or pseudo-metaconglomerates for the time being there are no suitable structural studies to solve the question.

The Baia de Arieș Series is a medium grade polymetamorphic sequence, similar from our point of view to the Someș and Codru Series. However, it has quite a different lithology: marbles, graphite quartzites, micaschists and paragneisses, metaporphyrroids and metagranites, as well as the Vinta intrusive granitoids. If the Someș and Codru Series formed the southern margin of the Biharia ocean, the Baia de Arieș Series made up its northern margin. On the other hand, the Baia de Arieș Series built up the basement of the Transylvanides, that is the southern margin of the Transylvanian Tethys. This is an important fact in understanding the emplacement of the Apusenides.

(e) The problem of the Apusenide emplacement. Nappe correlations may be done according to their origin, or along the strike of the tectonic bodies. Confusions appear if the author does not specify the way of correlation. For instance, the Transylvanides are Austrian and Laramian antithetic thrusts, and the Pienides are Laramian and intra-Burdigalian (Karpatian) synthetic tectonic units.

Although they originate from the lithosphere of the same ocean – the main Tethys – (e.g. Săndulescu, 1984; Debelmas, Săndulescu, 1987), they cannot be correlated unit by unit. At the same time they can obviously be correlated only along the northern margin of the pre-Apulian continental crust.

The Apusenides cannot be correlated with the Central Western Carpathian nappes following the northern margin of the pre-Apulian continental crust because: (1) both the Apusenides and the Central Western Carpathian Nappes are antithetic tectonic units with northern vergency; (2) the Tatrîdes have a northern vergency in relation to the Pienides, but this relationship is lacking between the Apusenides and Transylvanides; (3) a Permo-Mesozoic cover is missing or has a German type facies on the northern margin of the pre-Apulian plate (Andrusov et al., 1973; Ianovici et al., 1976; Plasienka, 1993); (4) the Permo-Mesozoic cover is thicker and of more south-Alpine character towards the south along the strike (Andrusov et al., 1973; Săndulescu, 1975; Kovacs, 1982); (5) this lateral north-south succession in sedimentary facies is found in the vertical sequence of the pre-Gosauian nappes, the uppermost nappes showing the southernmost facies (Andrusov et al., 1973; Săndulescu, 1975; Bleahu et al., 1981).

We can explain the relationships between the pre-Gosauian nappes of the Central Western Carpathians and the Apusen Mountains easier if we consider the following observations:

1. Both the Apusenides and the Transylvanides terminate against the North-Transylvanian fault;
2. Except the Finiș Nappe, the nappes of the Codru System do not surpass the central meridian line of the Apusen Mountains, toward the east.
3. The same location has the Biharian Arieșeni Nappe.
4. The Biharia and Baia de Arieș Nappes practically have almost no sedimentary covers.
5. The Finiș Nappe has a sedimentary cover only in its western part.
6. The Moesian (Euxinic) plate was in its current position at late Early Cretaceous times, because the Peceneaga-Camena transform fault, its northern boundary against North Dobrogea, is sealed by the "Babadag basin", which begins with the Vraconian (e. g. Săndulescu, 1984).

The above mentioned comments and observations encourage the following inferences:

1. The Codru Nappe System correlates with the Fatrîdes, Veporides and Gemerides and the Biharia Nappe System with the Tatrîdes, just as the Bihor unit.
2. Because the Tatrîdes occupy the lowermost position and the Biharia Nappe System the uppermost position, the latter must have been emplaced in a special tectonic setting.



3. This special tectonic setting has been brought about by the corner effect of the Moesian (Euxinic) plate, which forced the clockwise rotation of shears at the eastern end of the pre-Apulian plate, during its north-eastern movement.

4. The spatial propagation of the Moesian (Euxinic) plate corner effect was dependent on its size. That is why the Apusenides, as well as the Transylvanides, did not extend beyond the north-Transylvanian fault.

5. The Moesian (Euxinic) plate corner effect determined the advancing subduction features at the boundary between the pre-Apulian and Getic plates, too.

6. The mid-Hungarian dextral wrench faults, generated during the Senonian time, must be attributed to the blocking of the eastern end of the pre-Apulian plus the Transylvanian Getic fragment, east of the western termination of the corner of the Moesian (Euxinic) plate.

7. The relationships between the Codru and Biharia Nappe Systems have been conditioned by the style of rifting of the pre-Apulian plate, during the Jurassic time. The rift cuts obliquely the facies zones.

8. As between the Transylvanides and the Pienides, the correlation between the Apusenides and the Central Western Carpathian nappes is not possible unit by unit. The Apusenides are the result of the collision between the pre-Apulian and Getic plates, while the Central Western Carpathian nappes resulted from the collision between the Dinarides and the pre-Apulian plate.

III. Magmatism and Structure

In the Apuseni Mountains a Jurassic-Eocretaceous island arc magmatism, a collisional-subductional banatitic late-Cretaceous magmatism and an extensional Tertiary magmatism are manifested.

1. The island arc magmatism. It is represented by the arc rocks and the marginal basin rocks, the subduction being directed beneath the pre-Apulian plate (e. g. Nicolae et al., 1992).

2. The banatitic magmatism. The banatitic magmatism parallels the Transylvanian suture. It is a subduction related magmatism of continental margin type, and occurred only when the compressive tectonics began to relax, that is around the boundary between Cretaceous and Tertiary. The banatitic magmatism of the Apuseni Mountains is associated with the subduction front of the Getic plate and it terminates northerly where the pre-Gosau and Laramian overthrusts end.

3. The Tertiary magmatism. This is an extensional type magmatism associated to the Lower Miocene graben formation in the Pannonian area, and western part of the Apuseni Mountains. The arrangement of the Miocene magmatites follows evidently the strike of these grabens, whose distension was determined by the retreating subduction boundary from the external Dacidian basin (Royden, 1988, 1993). The Miocene age of the magmatites corresponds to the timing of extension in the Pannonian area. The strikes of the intrusions, volcanoes and their products are perpendicular to the Transylvanian suture. They do not extend beyond this suture. The Tertiary volcanism is not alkaline, because it is situated within the area of the banatitic intrusions and it derives from the same source. The banatitic magmatism destroyed the sub-continental pre-Apulian mantle and the temperatures remained high enough for a long time near the base of the crust.

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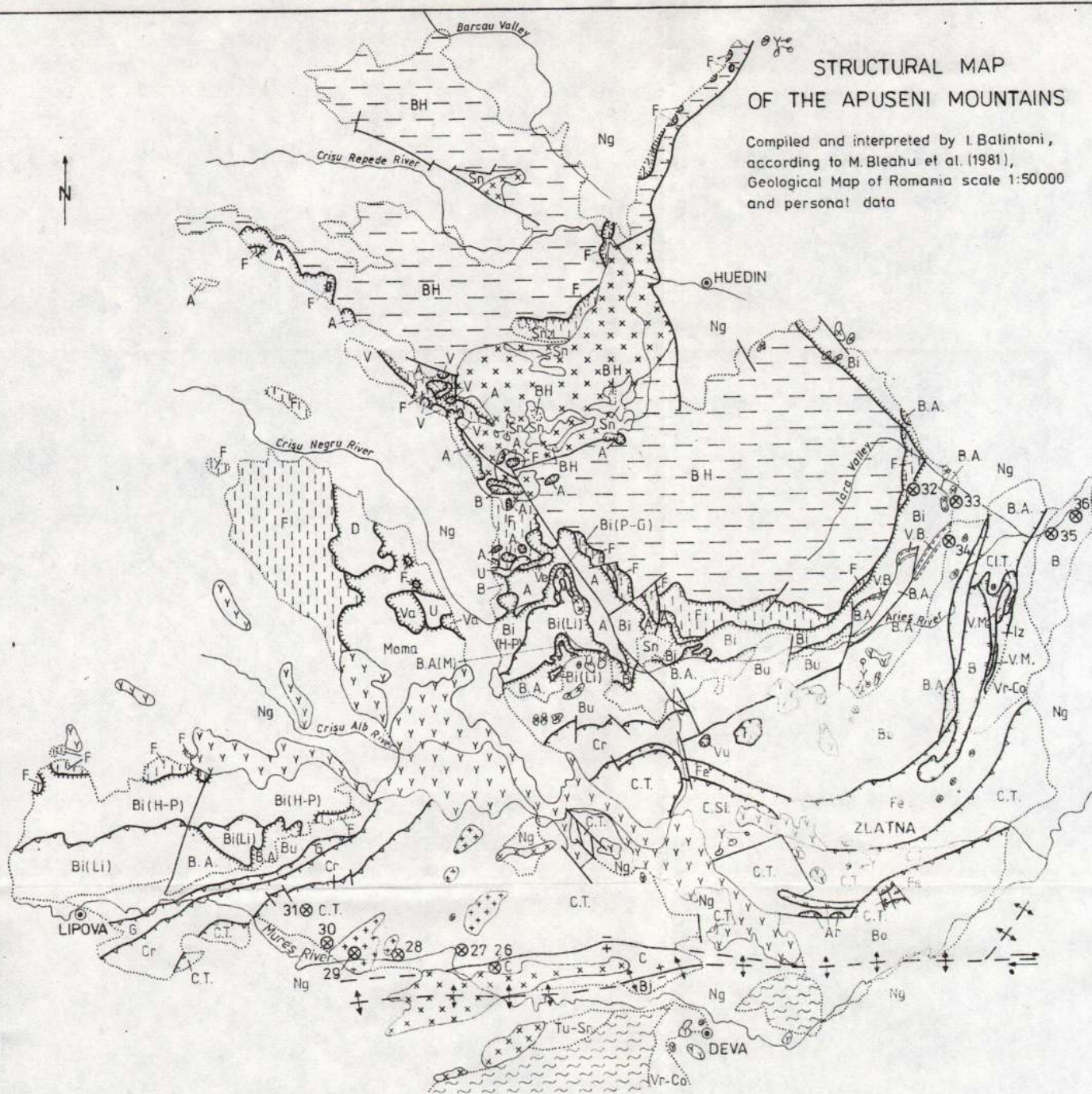
Legend of Plate

1, Neogene; 2, Senonian; 3, Turonian-Senonian; 4, Vraconian-Coniacian; 5, Neogene magmatites; 6, Banatitic magmatites; 7, Eocretaceous magmatites (granites). **Laramian Transilvanides:** 8, Căpâlnaş-Techereu Nappe; 9, Curechiu-Stănişia Nappe; 10, Feneş Nappe; Criş Nappe; 12, Groşi Unit; 13, Vulcan Nappe; 14, Frasin Nappe; Bucium Unit. **Meso-Cretaceous Transilvanides:** 16, Bedeleu Nappe; 17, Colţu Trascăului Nappe; 18, Valca Muntelui Nappe; 19, Izvoarele Nappe; 20, Ardeu Unit; 21, Căbeşti Unit; 22, Bejan Unit. **Apusenides (Biharia Nappe System):** 23, Baia de Arieş Nappe; (M), Muncelu Scale; (V.B.), Vulturese-Belioara Series; 24, Biharia Nappe; (Li), Lipova Nappe; (H-P), Hghiş-Poiana Nappe; (P-G), Piatra Grăitoare Scale; 25, Arieşeni Nappe. **Apusenides (Codru Nappe System):** 26, Vaşcău-Coleşti Nappe; 27, Moma Nappe; 28, Dieva-Bătrânescu-Vetre Nappe; 29, Feniş Nappe (-Ferice-Gârda-Următ); 30, Vâlcani Nappe; 31, Bihor Unit; 32, South Carpathian Crystalline; 33, Fault; 34, Reverse Fault; 35, Laramian overthrust; 36, Pre-Gosau overthrust; 37, Meso-Cretaceous overthrust; 38, Variscan overthrust; 39, Transgression; 40, Magmatite boundary; 41, Transform fault, plate boundary; 42, Wrench-fault movement; 43, Fault compartments; 44, Unspecified tectonic contact; 45, Stop; 46, Bozeş beds.



STRUCTURAL MAP OF THE APUSENI MOUNTAINS

Compiled and interpreted by I. Balintoni,
according to M. Bleahu et al. (1981),
Geological Map of Romania scale 1:50000
and personal data



0 6 12 18 24 30 km

Ng	1	Sn	2	Tu-Sn	3	Vr-Co	4	Y _Y Y _Y Y _Y	5	X _X X _X X _X	6	+ ₊ + ₊ + ₊	7	C.T.	8	C.St.	9	Fe	10	Cr	11
G	12	Vu	13	Fr	14	Bu	15	B	16	Cl.T.	17	V.M.	18	Iz	19	Ar	20	C	21	Bj	22
B.A(M)(V.B.)	23	Bi(Li)(H-P)(P-G)	24	A	25	Va	26	Moma	27	D.B.Ve.	28	F.U.	29	V	30	B.H.	31	~ ~ ~	32		
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THE TRANSYLVANIAN BASIN AND ITS UPPER CRETACEOUS SUBSTRATUM

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The Transylvanian Basin, with its present-day roughly circular shape, lies in a very complex and interesting position in the middle of the Carpathians. The basin is surrounded by the East and South Carpathians and by the Apuseni Mountains (Plate). The Carpathians are a Cretaceous to Neogene mainly external-vergent fold-and-thrust belt. The Apuseni Mountains, on the contrary, have acquired their main tectonic structures during the Cretaceous. In the eastern part of the basin, the Neogene volcanic chain of andesitic affinity is found.

From a geophysical point of view, the Transylvanian Basin shows a number of features which differ from those observed in many sedimentary basins. With values of ca. 40 mW/m², the heat flow in the basin is almost three times less than in some parts the Pannonian Basin and is also less than in the surrounding mountainous regions (Fig. 1). The Moho beneath the basin lies between 30 km in the center and 34 km in the eastern sectors and is thereby shallower than the surrounding chains (Rădulescu et al., 1991). Geodetic measurements have shown a marked contrast between strongly uplifting areas in the south (up to 3 mm/yr in the Southern Carpathians) and stable to gently subsiding domains in the central part of the basin (Fig. 2). A strong positive magnetic anomaly is present under the central part of the Transylvanian Basin and has been attributed to the presence at depth of ophiolitic rocks, similar to those found in the Apuseni Mountains (Botezatu, 1982). The Transylvanian Basin has presently quite a strong relief and lies at an average altitude of 600 m, with a maximum elevation of 800 m. This is in striking contrast with most of the Pannonian Basin, where average elevations are around 100 meters.

In this chapter of the guidebook we will make a brief summary of the existing knowledge and ideas about the sedimentary evolution of the basin as published in several articles (e.g. Ciupagea et al., 1970, Royden et al., 1988). We will also present the preliminary results of the work we are carrying out, which mainly addresses the problem of the structural and tectonic evolution of the basin and its relations with the surrounding orogen. The ultimate goal of the project, which involves researchers from the Universities of Bucharest and Amsterdam, is a quantitative understanding of the tectonic processes implicated in the formation of a subsiding area within a contractional domain. In the literature, the Transylvanian Basin is treated as an essentially Neogene feature. The basin, however, is partly underlain by thick sequences of Paleogene and Upper Cretaceous sediments (post orogenic cover in literature; Săndulescu, 1984). The tectonics which controlled the sedimentation of these successions form an interesting and poorly known topic in itself and bear important consequences on the evolution of the basin. We will therefore include also this time span in our discussion.

The evolution of the Transylvanian Basin and its Upper Cretaceous-Paleogene substratum

Late Cretaceous to Paleogene

The oldest sediments unconformably overlying the pre-Albian nappe stack presently found beneath the Transylvanian Basin are Upper Cretaceous. They mainly consist of terrigenous turbidites. In the area occupied by the present-day basin, the Upper Cretaceous pile reaches a maximum thickness of about 1500 m along a roughly N-S oriented trough in the central part of the basin (thickness values here and in the following are not corrected for compaction). Further to the west and to the south, the Upper Cretaceous is cropping out in the eastern part of the Apuseni Mountains and on the northern slope of the Southern Carpathians. In these



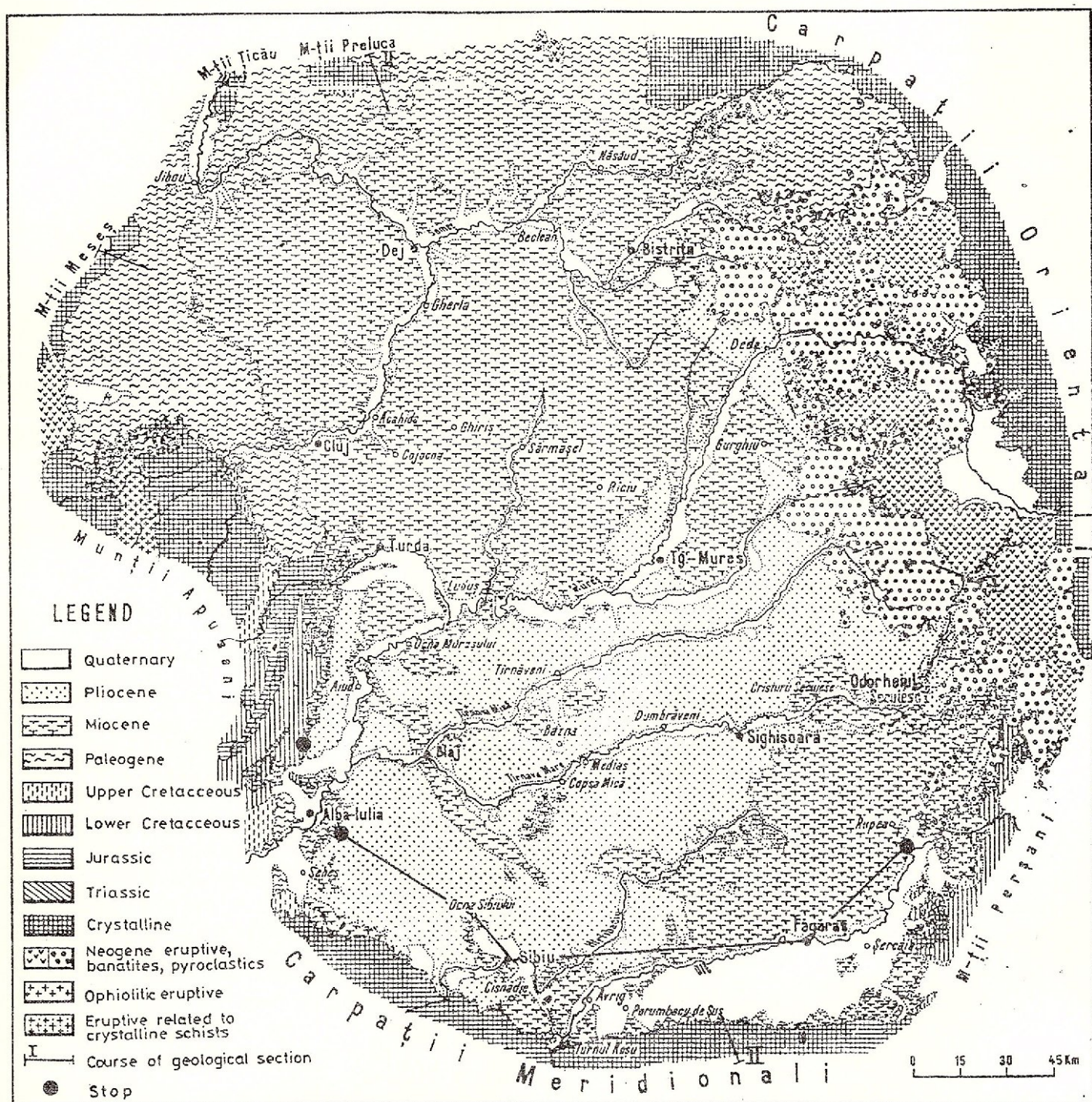


Fig. 1 - Geological Map of the Transylvanian Basin

(Ciupagea et al., 1970).



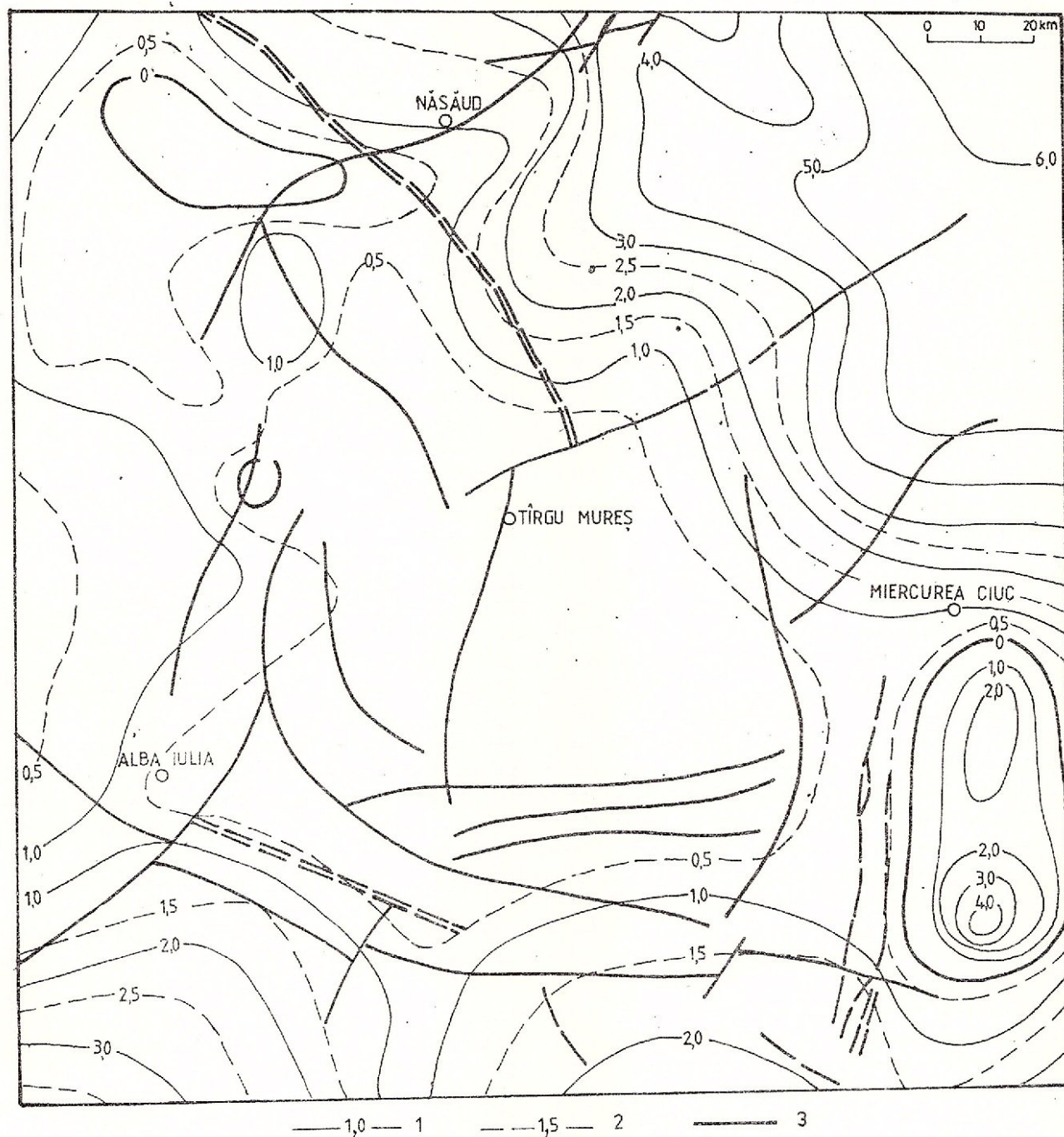


Fig. 2 - Map of Recent Crustal Vertical Movements in the Transylvania Basin

(after M. Visarion, M. Săndulescu, 1977).

regions, they mainly consist of coarse-grained sandstones to conglomerates; in the Southern Apuseni, these can contain also ophiolitic pebbles (Antonescu, 1973).

The Paleogene deposits have their most complete and thickest succession in the northern part of the basin where they are up to 3000 m thick. While the presence of the Eocene and of the Oligocene has been stratigraphically proven, this is not true for the Paleocene. In any case sedimentation was probably non-continuous throughout the Paleogene. Paleogene deposits are mostly shallow water and only in the Oligocene they show deep water facies. The present-day distribution of the Paleogene sediments recalls that of the Upper Cretaceous ones, thinning out towards the margins of the Transylvanian Basin. Sediments of Paleogene age are found along the eastern slope of the Apuseni Mountains and, to the northwest, in the Meseş Mountains showing that the Paleogene basin was larger than the present day outcrops. On the base of palaeontological investigations Bombitã et al. (1971) come to the same conclusion for southern part of the basin.

The Late Cretaceous to Paleogene tectonic history of the basin is still far from being well documented and understood, so that only preliminary results will be given here. Further ongoing field investigations and the release of seismic sections will greatly improve our knowledge.

The Upper Cretaceous sediments cropping out on the eastern slope of the Apuseni Mountains are strongly folded and steepened. The intensity of strain seems to decrease towards the east where seismic sections show much less deformed beds in the basement of the Transylvanian Basin. The intense deformation of the Upper Cretaceous deposits is in apparent contrast with the much less structured aspect of the Paleogene sediments, so that a pre-Paleogene age can be assumed for this phase of deformation ("Laramian phase" in literature). Sedimentation in the basin probably continued during tectonic activity.

The Neogene

With the Early Miocene, a completely different evolutionary stage began marked by a major unconformity of Burdigalian age. Beside Burdigalian deposits which are found only in the northern part of the basin, the most widespread sediments are of Badenian (=Middle Miocene) age. These are found over the entire basin and start with a tuff layer (Dej tuff) followed by a succession of marls and clays with some important salt intercalations. The salt commonly forms diapires, some of each are exploited. In the depocentral areas the thickness of the Badenian can be estimated at 3000 m. Sarmatian and Pannonian s.str. (=Late Miocene) sediments are up to 2000 m thick and were deposited under very shallow water conditions. They are rich in coal intercalations and show repeated tuffitic horizons which have a more basic character in the Sarmatian. The youngest sediments of the Transylvanian Basin consist of shallow water deposits of Middle Meotian (=latest Miocene) to Pliocene age. Tuff layers are found in the Meotian sediments. The sediments above the salt form gas reservoirs, reservoirs which are found in several parts of the basin.

The Neogene is traditionally considered a tectonically very quiet time with deformation being merely a consequence of diapiric movements of the salt layers. If it is true that Miocene sediments show both in seismic sections and in the field a much more regular geometry and are therefore much less deformed, there is ample evidence of substantial tectonic activity which accompanied the subsidence and further evolution of the basin.

Paleostress analysis and structural investigations in the Neogene outcrops along the border of the Apuseni Mountains have led to the recognition of two Badenian or younger phases of shortening. One had a N-S oriented shortening axis and caused widespread folding from the northern part of the basin to the Alba Iulia region. Probably the most spectacular of these folds are those found in the northern part of the basin, in the Dej region, where they are likely to be ramp-folds. Similar folds are observed NW of Alba Iulia (Geoagiu Valley) where they deform the Badenian limestones. The other phase of shortening caused WNW-ESE directed shortening, which determined widespread folding and faulting. The folds seem to be associated with deeper thrusts. One of the major structures associated with this phase of deformation is the well-known Meseş thrust, which is considered by Ciupagea (Ciupagea et al., 1970) to be of Pliocene age. Very similar thrusts, although of smaller magnitude, are found to the south-east of Alba Iulia. All the mentioned structures cause substantial shortening and are not associated with salt bodies.

Scattered extensional structures and paleostress determinations give indication for a phase of NE-SW extension which affected at least the western part of the basin. If these features can be correlated with the kinematically similar grabens found west of the Apuseni Mountains (e.g. Beiuş basin) their age is Badenian.

Some important, roughly N-S striking normal faults are found in the SE corner of the Transylvanian Basin (for instance in the Racoş region). A Pliocene to Quaternary age for these faults can be deduced from published



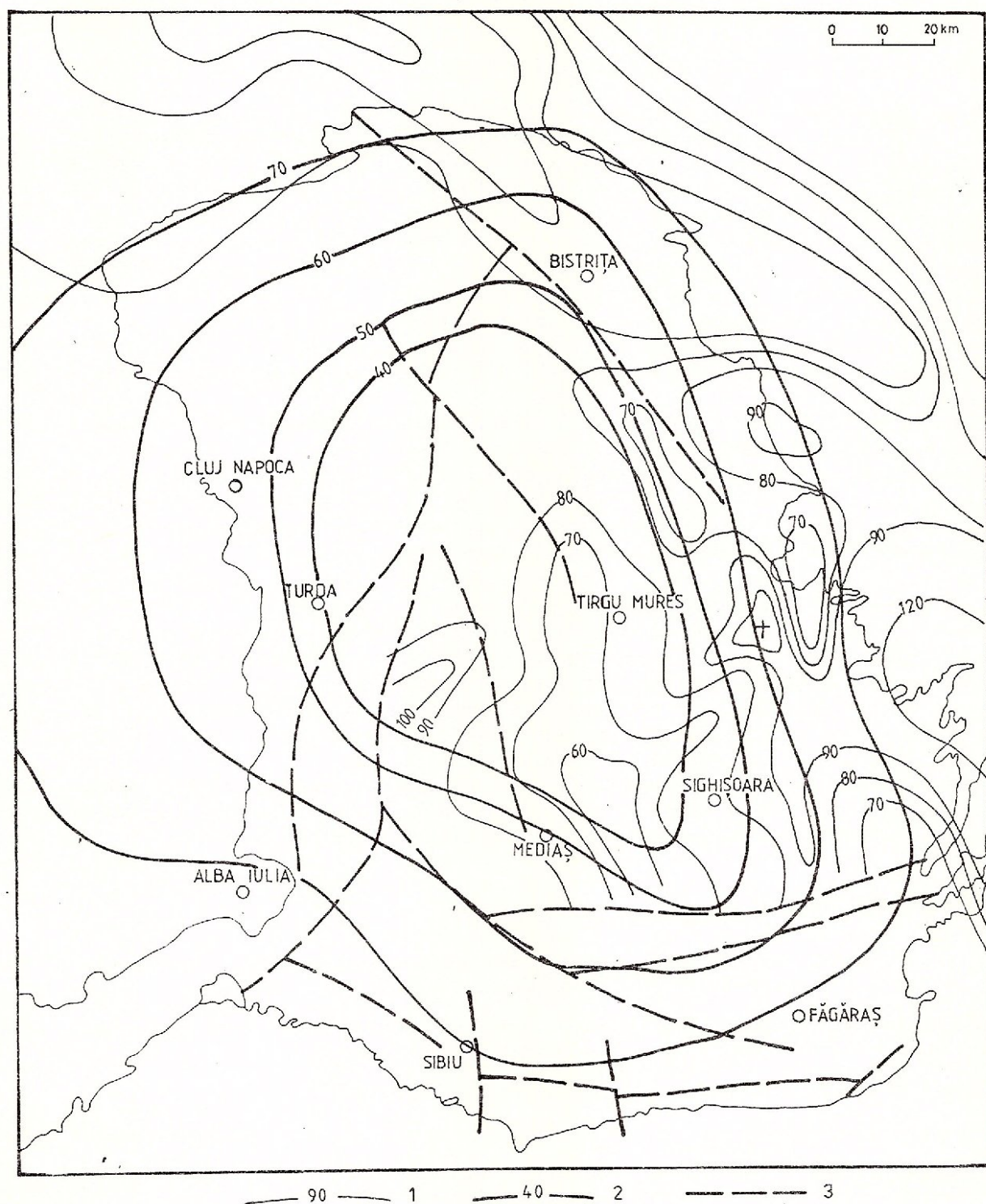


Fig. 3 – Geothermal Map of the Transylvania Basin
(after M. Visarion, A. Butza, C. Georgescu, 1985).

geologic maps and sections. No sign of similar features is found in the seismic sections crossing the Basin.

The tectonic evolution of the Transylvanian Basin

Very little is known until now about the tectonic processes involved in the formation of the Transylvanian Basin. Traditionally, the Late Cretaceous to Paleogene history of the basin has been considered as somehow related to the last stages of evolution of the Apuseni Mts and of the internal part of the Eastern Carpathians. The Neogene evolution is usually seen as the passive filling of a pre-existing morphology with some limited deformation caused by salt movements. It is, however, clear that the overall substantial subsidence throughout the Late Cretaceous to Miocene which allowed for the deposition of more than 6000 meters of sediments clearly calls for a tectonic motor driving the evolution of the basin. Our preliminary structural investigations go in the same direction demonstrating the existence of substantial deformation during basin formation which cannot be attributed to salt tectonics. In fact, the abundance of tectonic structures not associated with any salt body, as well as the amount of shortening accommodated, the alignment of many salt bodies and their asymmetry in cross-sections, suggest that the salt played a passive role, intruding tectonically-driven structures rather than viceversa.

After the formation of the Austrian nappe pile, Upper Cretaceous sediments were deposited in roughly N-S striking basins which we interpret as grabens associated with E-W extension.

Substantial shortening followed causing folding and steeping of the Upper Cretaceous beds. We do not know much about the tectonic context in which shortening took place; however, it ended before the Eocene. Tectonic activity during the Eocene and the Oligocene is demonstrated by seismic cross sections in the basin (Ciupagea et al., 1970) and is reflected in the repeated stratigraphic gaps.

Two phases of shortening affected the area during the Neogene, possibly separated by an extension event. These deformational phases took place in an extremely complex and not well understood overall context: to the north and north-west, important extension was affecting the Pannonian Basin; to the east and to the south substantial shortening was going on in the Carpathians. The relations of these phenomena with the evolution of the Transylvanian Basin are not clear. The extensional structures similar to those detected in the Alba Iulia region are possibly important, but in no way they can accommodate stretching factors large enough to explain the Neogene subsidence of the Transylvanian Basin. In fact, the Neogene history of the Transylvanian Basin is, on the whole, one of shortening induced deformation.

Tectonic activity in the Transylvanian Basin and the surrounding regions continued until quite recent times and is possibly still going on, as shown by the presence in the Southern Carpathians of Badenian marine deposits at 1300 m of elevation and by the high uplift rates found in the Făgăraș Mountains.

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VARISCAN VS. ALPINE TECTONOTHERMAL EVOLUTION WITHIN THE APUSENI MOUNTAINS, ROMANIA: EVIDENCE FROM $^{40}\text{Ar}/^{39}\text{Ar}$ MINERAL AGES

by

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Introduction

A collaborative field and $^{40}\text{Ar}/^{39}\text{Ar}$ geochronological research program is currently being carried out along systematic traverses across the Apuseni Mountains. Preliminary results are available from several areas to be examined during this excursion. Because they provide important constraints for distinction between Alpine and pre-Alpine tectonothermal events the preliminary data are briefly discussed herein.

Geologic Setting

The Apuseni Mountains are comprised of a series of variably metamorphosed nappe complexes (for reviews, see Balintoni, this volume, and Săndulescu, 1984). These were derived from a pre-Alpine palinspastic location situated between the Tethys and Meliata Oceanic Domains. Their overall palinspastic setting was similar to that of structural units within the Eastern Alps and Western Carpathians. Results of recent detailed field work and collaborative $^{40}\text{Ar}/^{39}\text{Ar}$ mineral dating have enabled calibration of a complex tectonothermal evolution that included late Variscan and polyphase Alpine events. These new data require major revision of traditional interpretations of the geologic history which included metamorphic events in the middle Proterozoic, Caledonian and Variscan orogenic cycles followed by middle-late Cretaceous (Alpine) thrusting (for reviews, see Balintoni, this volume; Dimitrescu, 1988; Săndulescu, 1984).

Basement rocks exposed within the Apuseni Mountains comprise four major tectonostratigraphic units. Structurally upward these include (Fig. 1): 1) the Bihor autochthon/parautochthon (Arada and Someș Series); 2) granite intruded amphibolite and gabbro-diorite of the Codru Nappe Complex; 3) a generally low-grade, polydeformed greenstone-granite terrane of the internally imbricated Biharia Nappe Complex (Păiușeni, Bihor and Arieșeni Series); and, 4) an internally imbricated association of carbonate and lenses bearing gneisses (Baia de Arieș Nappe Complex).

Analytical Methods

The techniques used during $^{40}\text{Ar}/^{39}\text{Ar}$ analysis of mineral concentrates generally followed those described by Dallmeyer and Gil-Ibarguchi (1990). Optically pure (*leftarrow* 99 %) mineral concentrates and sized whole-rock powders were wrapped in aluminium foil packets, encapsulated in sealed quartz vials, and irradiated in the TRIGA Reactor at the U. S. Geological Survey in Denver. Variations in the flux of neutrons along the length of the irradiation assembly were monitored with several mineral standards, including MMhb-1 (Samson, Alexander, 1987). The samples were incrementally heated until fusion in a double-vacuum, resistance heated furnace following methods described by Dallmeyer and Gil-Ibarguchi (1990). Measured isotopic ratios were corrected for total system blanks and the effects of mass discrimination. Interfering isotopes produced during irradiation were corrected using factors reported by Dalrymple et al. (1981). Apparent $^{40}\text{Ar}/^{39}\text{Ar}$ ages were



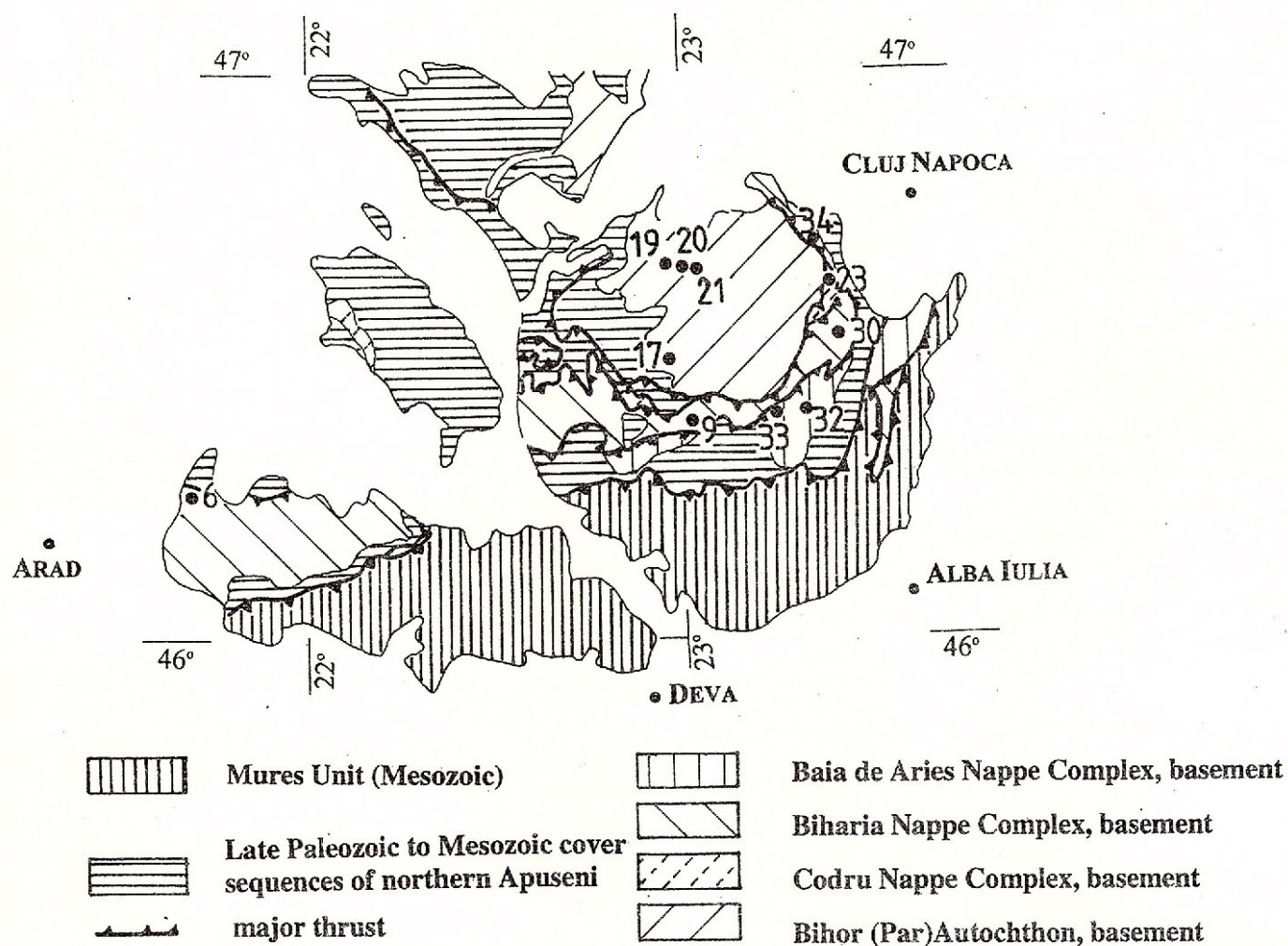


Fig. 1 - Generalized tectonic map of the Apuseni Mountains showing locations for which $^{40}\text{Ar}/^{39}\text{Ar}$ mineral ages have been determined.

calculated from corrected isotopic ratios using the decay constants and isotopic abundance ratios listed by Steiger and Jäger (1977).

Intralaboratory uncertainties have been calculated by statistical propagation of uncertainties with measurement of each isotopic ratio (at two standard deviations of the mean) through the age equation. Interlaboratory uncertainties are ca. $\pm 1.25 - 1.5$ % of the quoted age. Total-gas ages have been computed for each sample by appropriate weighing of the age and percent ^{39}Ar released within each temperature increment. A "plateau" is considered to be defined in the ages recorded by two more contiguous gas fractions (with similar apparent K/Ca ratios) each representing $\rightarrow 1$ % of the total ^{39}Ar evolved (and together constituting $\rightarrow 50$ % of the total quantity of ^{39}Ar evolved) are mutually similar within a ± 1 % intralaboratory uncertainty. Analyses of the MMhb-1 monitor indicate that apparent K/Ca ratios may be calculated through the relationships $0.518 (\pm 0.00095) \times (^{39}\text{Ar}/^{37}\text{Ar})$ corrected for the TRIGA reactor and $0.505 (\pm 0.003) \times (^{39}\text{Ar}/^{37}\text{Ar})$ corrected for the Ford Reactor.

Plateau portions of the analyses have been plotted on $^{36}\text{Ar}/^{40}\text{Ar}$ isotope correlation diagrams. Regression techniques followed the methods of York (1969). A mean square of the weighted deviates (MSWD) has been used to evaluate isotopic correlations.

Results

Bihor Autochthon/Parautochthon

A hornblende concentrate was prepared from a sample of amphibolite collected within the Someş Series of the Bihor autochthon/parautochthon near Ciurtuci (DA-ROM 19-93; Fig. 1). It is characterized by an internally discordant age spectrum which is matched by fluctuations in apparent K/Ca ratios (Fig. 2) which suggests experimental evolution of argon occurred from compositionally distinct, relatively non-retentive phases. These could be represented by: 1) very minor, optically undetectable mineralogical contaminants in the concentrate; 2) petrographically unresolvable exsolution or compositional zonation within constituent amphibole grains; 3) minor chloritic replacement of amphibole; and/or 4) intracrystalline inclusions. The intermediate- and high-temperature gas fractions display little variation in apparent K/Ca ratios suggesting experimental evolution of gas occurred from populations of compositionally uniform intracrystalline site. The intermediate- and high-temperature gas fractions record similar apparent ages which define a plateau of 306.1 ± 0.8 Ma. An $^{36}\text{Ar}/^{40}\text{Ar}$ vs. $^{39}\text{Ar}/^{40}\text{Ar}$ isotope correlations of the plateau data is well-defined (MSWD = 1.96) with an inverse ordinate intercept ($^{40}\text{Ar}/^{36}\text{Ar}$ ratio) of 376.5 ± 12.1 . This is larger than the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio in the present-day atmosphere, and suggests slight intracrystalline contamination with extraneous argon components. Using the inverse abscissa intercept ($^{40}\text{Ar}/^{39}\text{Ar}$ ratio) in the age equation yields a plateau isotope correlation age of 300.4 ± 0.5 Ma. Because calculation of isotope correlation ages do not require assumptions of $^{40}\text{Ar}/^{36}\text{Ar}$ ratios they are more reliable than ages calculated directly from the analytical data. The 300 Ma plateau isotope correlation age is considered geologically significant, and is interpreted to date the last cooling through temperatures required for intracrystalline retention of argon within constituent grains in the amphibole concentrate. Harrison (1981) suggested that temperatures of ca. 500°C are appropriate for argon retention within amphibole at cooling rates which likely characterize most regional metamorphic settings. A hornblende concentrate was prepared from another Someş amphibolite collected near Belslu (DA-ROM21A-93; Fig. 1). It displays a slightly discordant age spectrum (Fig. 2) which records a well-defined intermediate- and high-temperature plateau corresponding to an age of 316.7 ± 0.5 Ma. The plateau data define an isotope correlation age of 313.6 ± 0.5 Ma suggesting only minor intracrystalline contamination with extraneous argon components.

A muscovite concentrate was prepared from a sample of retrogressed felsic gneiss collected within the Someş Series exposed in northwestern sectors of the autochthon/parautochthon near Ciurtuci (DA-ROM20-93; Fig. 1). This displays an internally discordant $^{40}\text{Ar}/^{39}\text{Ar}$ age spectrum in which apparent ages systematically increase throughout low-temperature portions of the analysis (Fig. 3). The intermediate- and high-temperature increments define a plateau age of 313.6 ± 0.2 Ma. Apparent K/Ca ratios are very large with considerable associated uncertainties. Consequently, they are not shown in Figure 3. However, the ratios display no significant or systematic variations suggesting that experimental evolution of gas occurred from compositionally uniform populations of intracrystalline site. The character of spectra discordance displayed by the muscovite concentrate is identical to that which has been described from partially rejuvenated muscovite in other polymetamorphic settings (e.g. Dallmeyer and Takasu, 1992). Comparison of the present discordant spectrum from



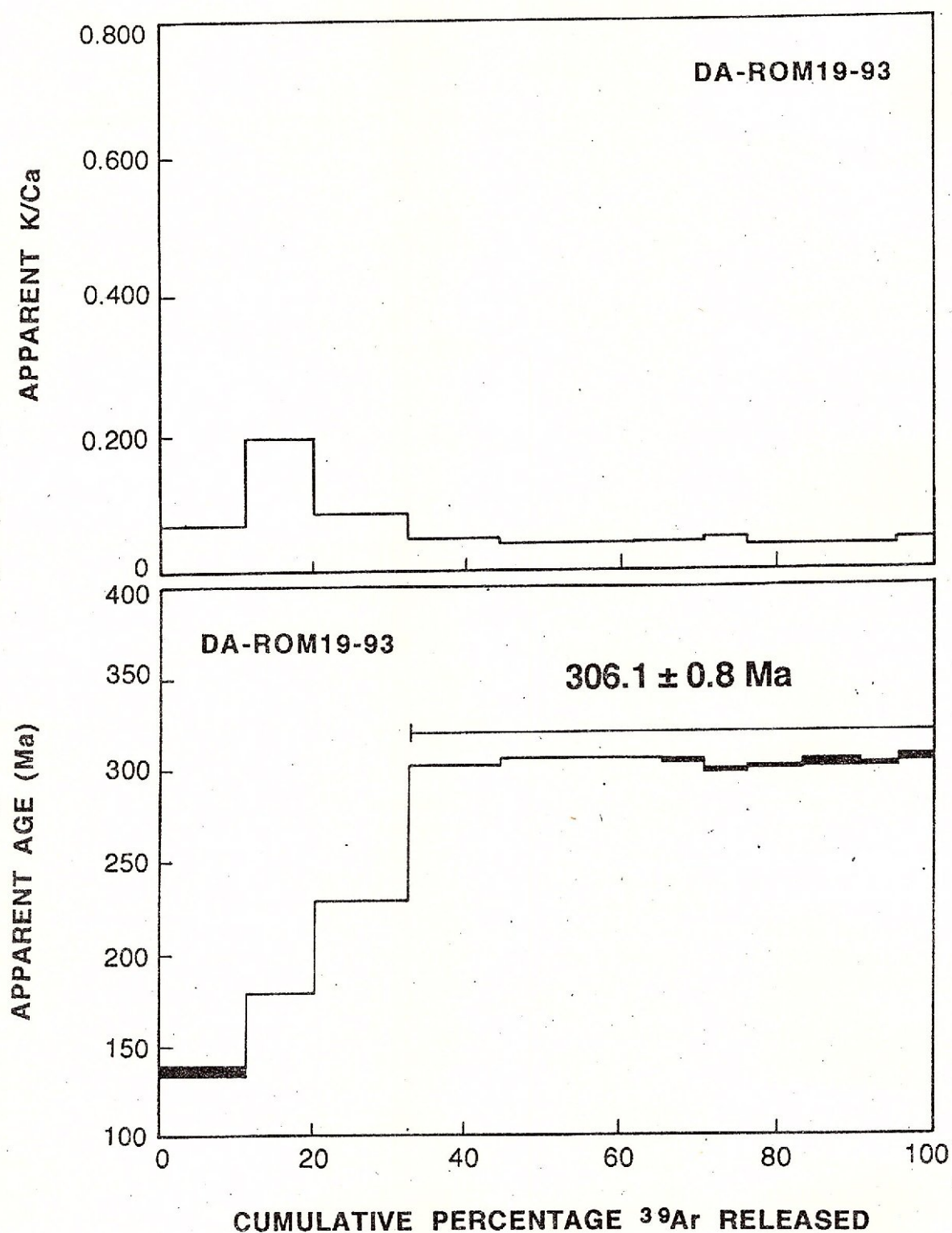


Fig. 2 (part 1) - $^{40}\text{Ar}/^{39}\text{Ar}$ apparent age and apparent K/Ca spectra of hornblende concentrates from amphibolite of the Semeş Series, Bihor autochthon/parautochthon.

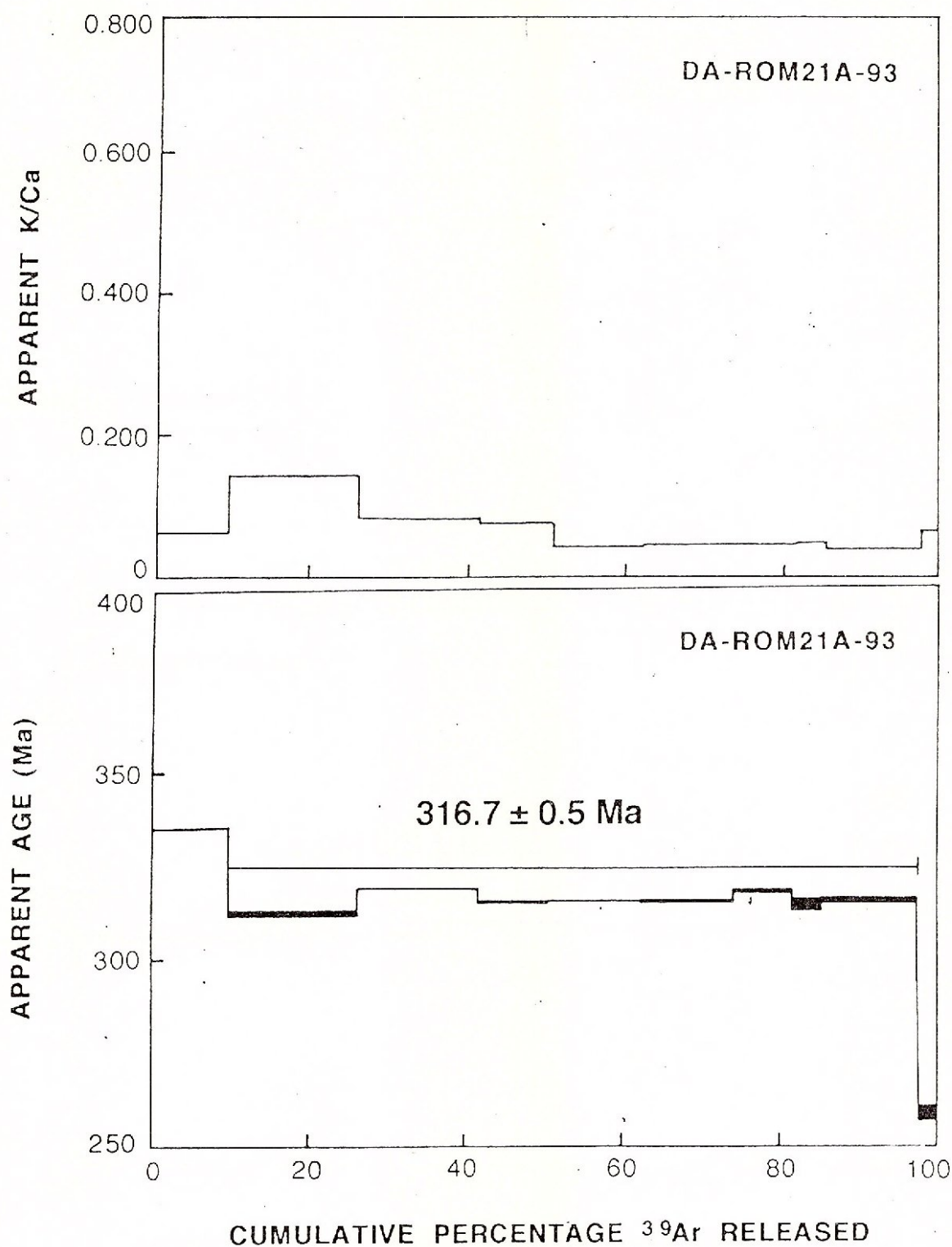


Fig. 2 (part 2) – Analytical uncertainties (two sigma, interlaboratory) represented by vertical width of bars. Experimental temperatures increase from left to right. Total-gas and plateau ages are shown.

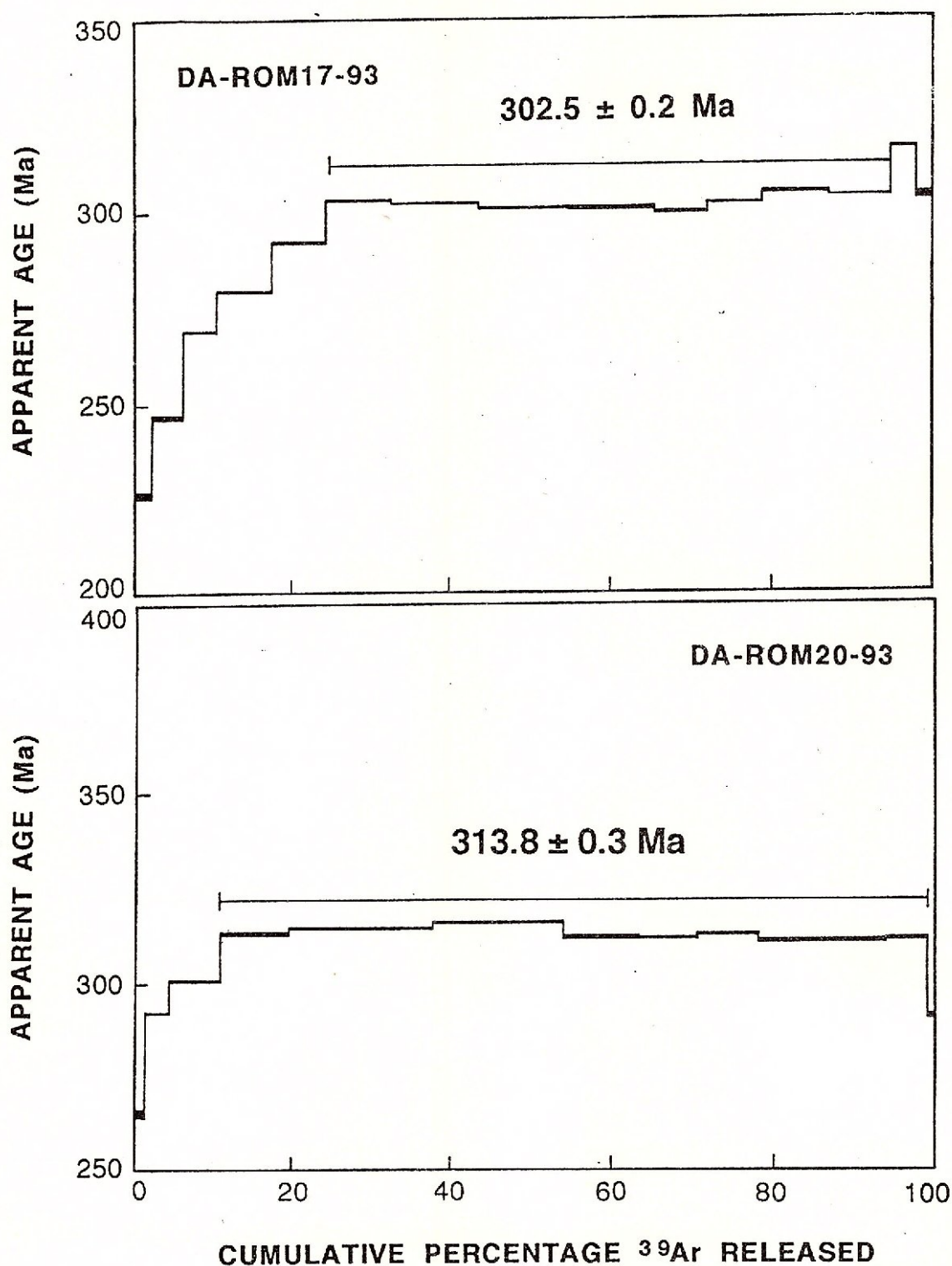


Fig. 3 - $^{40}\text{Ar}/^{39}\text{Ar}$ apparent age spectra of muscovite concentrates prepared from the Someş and Arada Series, Bihor autochthon/parautochthon. Data plotted as in Figure 2.

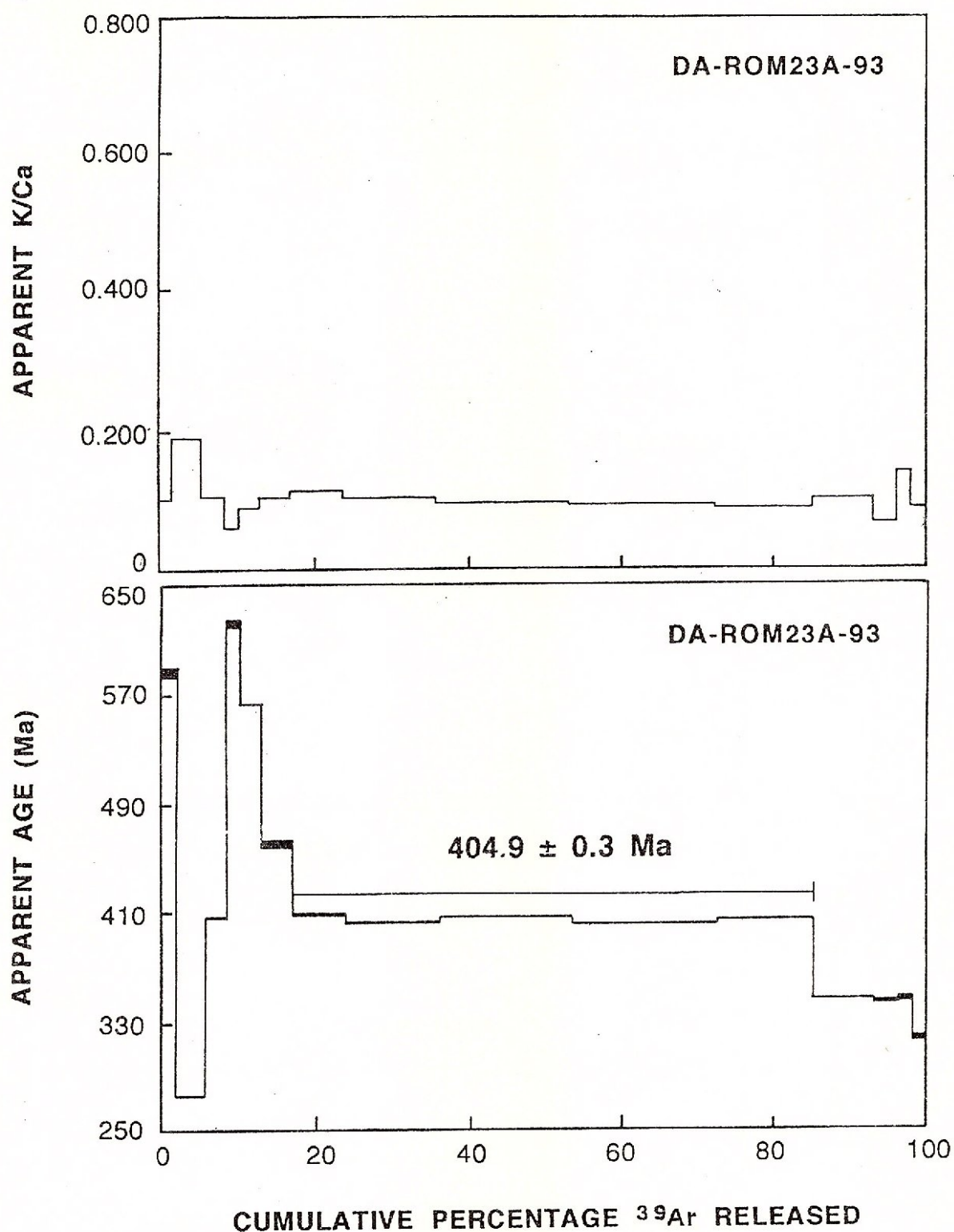


Fig. 4 -- $^{40}\text{Ar}/^{39}\text{Ar}$ apparent age and apparent K/Ca spectra of a hornblende concentrate from migmatitic amphibolite of the Codru Nappe Complex. Data plotted as in Figure 2.

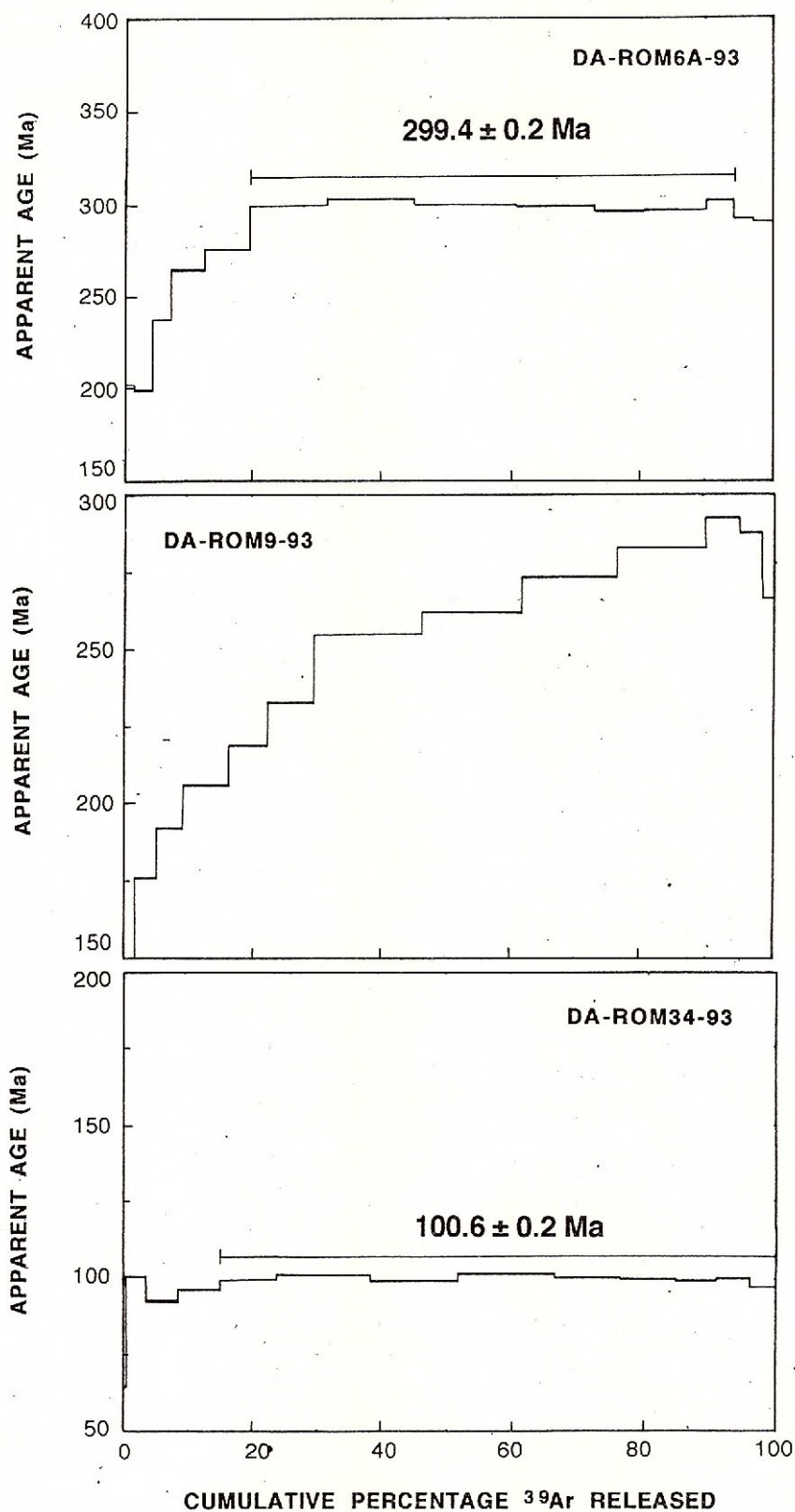


Fig. 5 - $^{40}\text{Ar}/^{39}\text{Ar}$ apparent age spectra of muscovite concentrates from various lithological elements of the Biharia Nappe Complex. Data plotted as in Figure 2.

results in other areas suggest that initial cooling through appropriate argon retention temperatures (c. $375 \pm 25^\circ\text{C}$) at c. 314 Ma was followed by slight ($\rightarrow 10\%$) Mesozoic rejuvenation of intracrystalline argon systems. A muscovite concentrate was also prepared from a sample of sericite-bearing quartzite collected within the Arada Series exposed in southeastern sectors of the autochthon/parautochthon near Rusesti (DA-ROM17-93; Fig. 1). This concentrate displays an internally discordant age spectrum (Fig. 3) with characteristics generally similar to that from sample DA-ROM20-93, and an 302.5 ± 0.2 Ma intermediate- and high-temperature plateau age is defined. However, the low-temperature increments are more discordant suggesting more extensive Mesozoic rejuvenation.

Codru Nappe Complex

A hornblende concentrate was prepared from a sample of amphibolite collected within the Codru Nappe Complex near Valea Ierii (DA-ROM23A-93, Fig. 1). The sample displays internally discordant apparent age and apparent K/Ca spectra (Fig. 4). However, the five intermediate-temperature increments are characterized by generally similar apparent K/Ca ratios and record similar apparent ages corresponding to a plateau of 404.9 ± 0.3 Ma. The plateau data yield an only slightly younger isotope correlation age of 400.8 ± 0.6 Ma. This is considered geologically significant and interpreted to date post-metamorphic cooling through temperatures appropriate for argon retention.

Biharia Nappe Complex

Muscovite concentrates have been prepared from three samples collected within the Biharia Nappe Complex (Fig. 1). These include mylonitic orthogneiss (DA-ROM34-93), quartz-clasts bearing schists (DA-ROM6A-93) and schist (DA-ROM9-93). The concentrate from metaconglomerate (collected near Salcuia de Sus) records an intermediate- and high-temperature plateau age of 299.4 ± 0.2 Ma (Fig. 5). The low-temperature gas fractions suggest slight ($\rightarrow 10\%$) Mesozoic rejuvenation of intracrystalline argon systems. The concentrate from schist (collected near Vadu Motilor) displays more extensive Mesozoic rejuvenation of intracrystalline argon systems that initially cooled through temperatures appropriate for argon retention at c. 300 Ma. The muscovite concentrate from mylonitic orthogneiss (collected near Tarnita) displays complete Mesozoic rejuvenation and records a plateau age of 100.6 ± 0.2 Ma.

Baia de Arieş Nappe Complex

Hornblende concentrates were prepared from two samples of amphibolite collected within a structural unit of the Baia de Arieş Nappe Complex near Alciua de Sus (Fig. 1: DA-ROM32A-93) and Sartes (DA-ROM33A-93). Both are characterized by similar internally discordant age spectra (Fig. 6). Low-temperature gas fractions evolved from both concentrates are characterized by anomalously old ages and variable apparent K/Ca ratios reflecting compositional controls. All intermediate- and high-temperature gas fractions display only minor intrasample variations in apparent K/Ca ratios and record similar apparent ages which define plateaus of 119.0 ± 0.1 Ma (32A) and 118.2 ± 0.3 Ma (33A). The plateau data define isotope correlation ages which are similar within analytical uncertainties. A muscovite concentrate was prepared from a sample of mylonitic gneisses collected at location 33 (DA-ROM33B-93). It records a plateau age of 110.7 ± 0.1 Ma (Fig. 7).

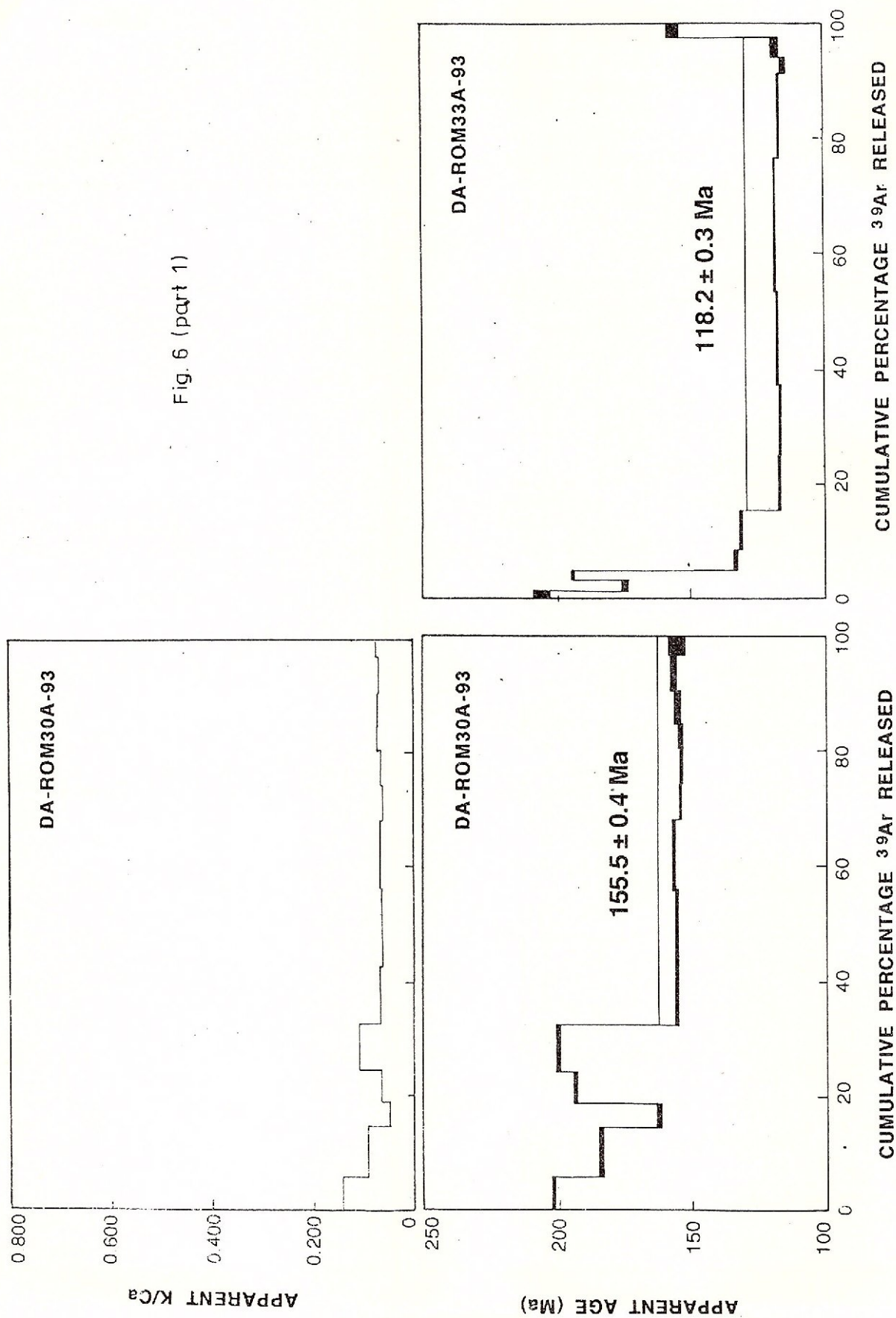
A hornblende concentrate was prepared from a sample of amphibolite collected within a different structural unit of the Baia de Arieş Nappe Complex near Surduc (Fig. 1: DA-ROM30A-93). It displays an internally discordant age spectrum (Fig. 6) with considerable low-temperature variations in both apparent age and apparent K/Ca ratios. The intermediate- and high-temperature gas fractions are characterized by similar apparent K/Ca ratios. These record similar apparent ages which define a plateau of 155.5 ± 0.4 Ma. The plateau data correspond to an isotope correlation age of 151.6 ± 0.6 Ma.

Tectonic implications

The preliminary $^{40}\text{Ar}/^{39}\text{Ar}$ results suggest that regionally penetrative high-temperature mineral assemblages and associated ductile structural elements within structural units of the Bihor autochthon/parautochthon and the Biharia Nappe Complex are all likely related to Late Paleozoic (late Variscan) tectonothermal activity. The Alpine record appears to be restricted to discrete, intermediate- or low-temperature ductile shear zones developed in the basement elements. An early Variscan history appears to be recorded within the Codru Nappe



Fig. 6 (part 1)



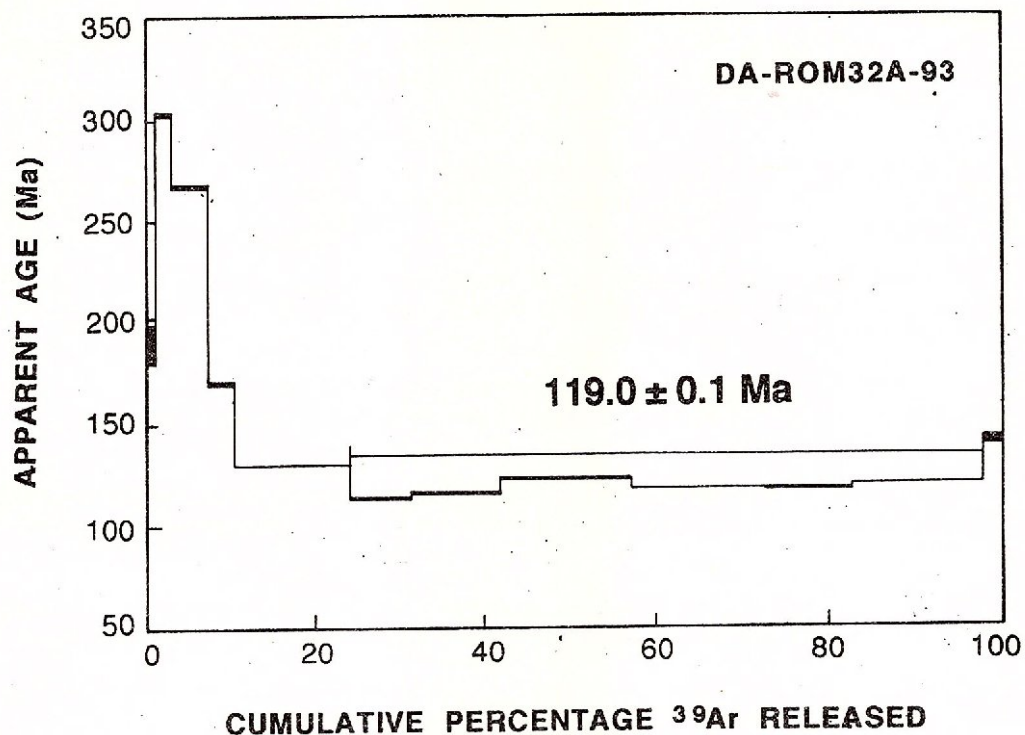


Fig. 6 (part 2)

Fig. 6 (part 1, part 2) - $^{40}\text{Ar}/^{39}\text{Ar}$ apparent age and apparent K/Ca spectra of hornblende concentrates from amphibolite of the Baia de Arieș Nappe Complex. Data plotted as in Figure 2.

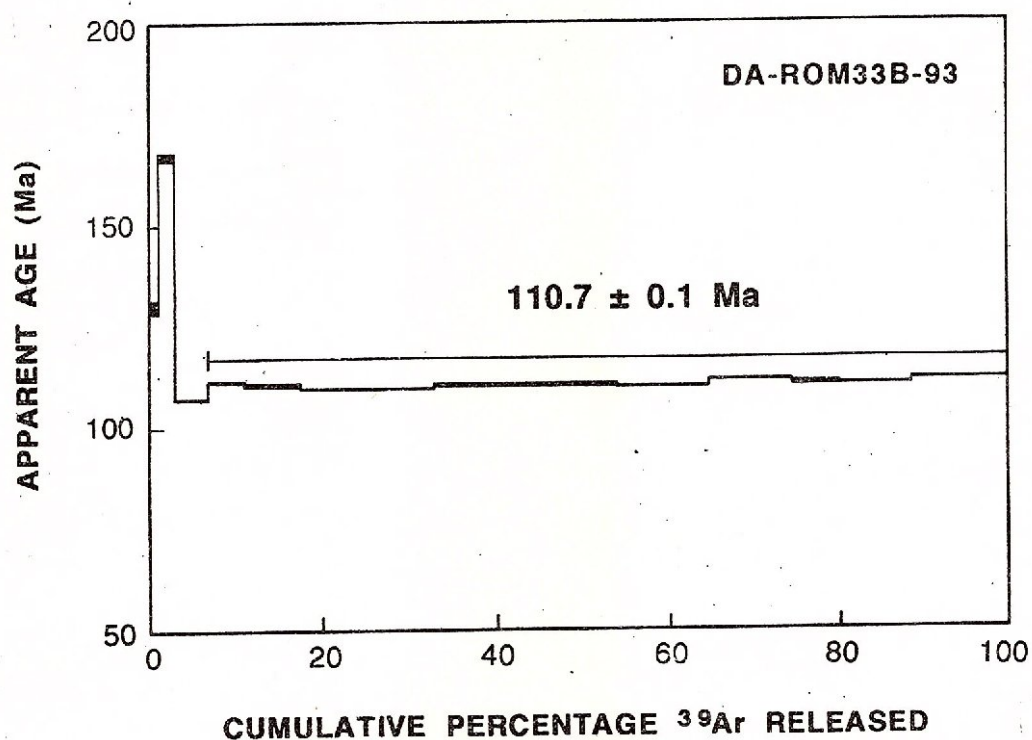


Fig. 7 - $^{40}\text{Ar}/^{39}\text{Ar}$ apparent age spectrum of a muscovite concentrate from mylonitic gneiss within the Baia de Arieș Nappe Complex. Data plotted as in Figure 2.

Complex. A more complete Alpine tectonothermal record is present in the Baia de Arieş Nappe Complex, however it appears to have been polyphase with both early and late Alpine episodes.

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$^{40}\text{Ar}/^{39}\text{Ar}$ MINERAL AGE CONTROLS FOR THE PRE-ALPINE AND ALPINE TECTONIC EVOLUTION OF NAPPE COMPLEXES IN THE SOUTHERN CARPATHIANS

by

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Introduction

A collaborative field and $^{40}\text{Ar}/^{39}\text{Ar}$ geochronological research program is currently being carried out along systematic traverses across the Southern Carpathians. Preliminary results are available from several areas to be examined during this excursion. Because they provide important constraints for distinction between various pre-Alpine and Alpine tectonothermal events the preliminary data are briefly discussed herein.

Geological setting

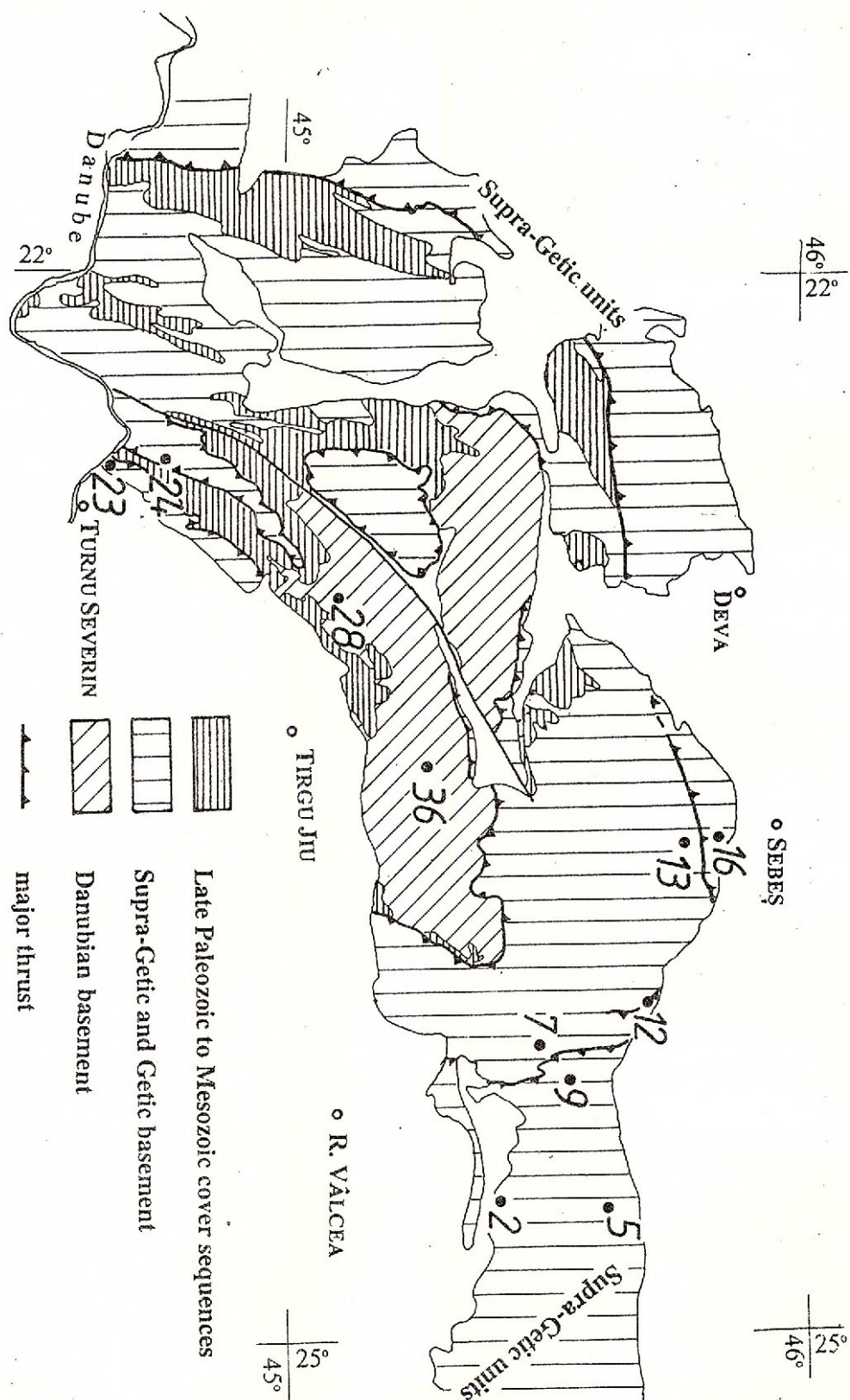
The Southern Carpathians are comprised of a number of metamorphic basement nappes which are separated by minor intercalations of Late Paleozoic to Mesozoic cover sequences (for a general overview on Southern Carpathians, see Berza, this volume; Kräutner et al., 1988; Săndulescu, 1984). Their original palinspastic setting was between the European plate with the Moesian promontory and the Vardar-Mureş-Highiş-Meliata oceanic domain, the latter considered as a western extension of Tethys. Results of recent field work and collaborative $^{40}\text{Ar}/^{39}\text{Ar}$ mineral dating have enabled especially calibration of Late Variscan tectonothermal events. These new data require significant revision of previous interpretations of the geologic evolution, and constrain large-scale Late Variscan block uplift following Variscan orogenic events.

Basement rocks exposed within the Southern Carpathians comprise structurally upward three major nappe complexes (Fig. 1): (1) The Danubian nappe complex with Cadomian granitoids and metamorphic sequences; (2) the Getic nappe complex with mainly medium-grade metamorphic sequences, too; and (3) the Supragetic nappe complex with mainly medium-grade metamorphic sequences.

Analytical Methods

The techniques used during $^{40}\text{Ar}/^{39}\text{Ar}$ analysis of mineral concentrates generally followed those described by Dallmeyer and Gil-Ibarguchi (1990). Optically pure ($\approx 99\%$) mineral concentrates and sized whole-rock powders were wrapped in aluminium foil packets, encapsulated in sealed quartz vials, and irradiated in the TRIGA Reactor at the U.S. Geological Survey in Denver. Variations in the flux of neutrons along the length of the irradiation assembly were monitored with several mineral standards, including MMhb-1 hornblende (Samson and Alexander, 1987). The samples were incrementally heated until fusion in a double-vacuum, resistance heated furnace following methods described by Dallmeyer and Gil-Ibarguchi (1990). Measured isotopic ratios were corrected for total system blanks and the effects of mass discrimination. Interfering isotopes produced during irradiation were corrected using factors reported by Dalrymple et al. (1981). Apparent $^{40}\text{Ar}/^{39}\text{Ar}$ ages were calculated from corrected isotopic ratios using the decay constants and isotopic abundance ratios listed by Steiger and Jäger (1977).





Intralaboratory uncertainties have been calculated by statistical propagation of uncertainties with measurement of each isotopic ratio (at two standard deviations of the mean) through the age equation. Interlaboratory uncertainties are ca. $\pm 1.25 - 1.5$ % of the quoted age. Total-gas ages have been computed for each sample by appropriate weighting of the age and percent ^{39}Ar released within each temperature increment. A "plateau" is considered to be defined in the ages recorded by two more contiguous gas fractions (with similar apparent K/Ca ratios) each representing $\rightarrow 4$ % of the total ^{39}Ar evolved (and together constituting $\rightarrow 50$ % of the total quantity of ^{39}Ar evolved) are mutually similar within a ± 1 % intralaboratory uncertainty. Analyses of the MMhb-1 monitor indicate that apparent K/Ca ratios may be calculated through the relationships $0.518 (\pm 0.00005) \times (^{39}\text{Ar}/^{37}\text{Ar})$ corrected for the TRIGA reactor and $0.505 (\pm 0.003) \times (^{39}\text{Ar}/^{37}\text{Ar})$ corrected for the Ford Reactor.

Plateau portions of the analyses have been plotted on $^{36}\text{Ar}/^{40}\text{Ar}$ isotope correlation diagrams. Regression techniques followed the methods of York (1969). A mean square of the weighted deviates (MWSd) has been used to evaluate isotopic correlations.

Results

Danubian Autochthon/Parautochthon

A hornblende concentrate was prepared from a sample of diorite exposed within the Tismana Valley within the Danubian autochthon/parautochthon (DA-ROM28-92; Fig. 1). It is characterized by an internally discordant age spectrum which is matched by fluctuations in apparent K/Ca ratios (Fig. 2) which suggests experimental evolution of argon occurred from compositionally distinct, relatively non-retentive phases. These could be represented by: 1) very minor, optically undetectable mineralogical contaminants in the concentrate; 2) petrographically unresolvable exsolution or compositional zonation within constituents amphibole grains; 3) minor chloritic replacements of amphibole; and/or, 4) intracrystalline inclusions. The intermediate- and high-temperature fractions are characterized by similar apparent K/Ca ratios suggesting experimental evolution of gas occurred from compositionally uniform intracrystalline sites. The intermediate-temperature gas fractions record systematically decreasing apparent ages (from c. 640 Ma down to c. 610 Ma). The 835–905°C increments comprise slightly less than 50 % of the total gas evolved from the concentrate. These record generally similar apparent ages which range between c. 596 and 591 Ma. These are considered geologically significant.

A muscovite concentrate was prepared from a basement gneiss sample collected within the Polatiștea Valley in the Danubian autochthon/parautochthon (DA-ROM36B-92; Fig. 1). The concentrate displays an internally discordant age spectrum (Fig. 3) in which apparent ages systematically increase throughout low-temperature portions of the analysis. The intermediate- and high-temperature increments define a plateau age of 296.0 ± 0.2 Ma. Apparent K/Ca ratios are very large with considerable associated uncertainties. Consequently, they are not shown in Figure 3. However, the ratios display no significant or systematic variations suggesting that experimental evolution of gas occurred from compositionally uniform populations of intracrystalline sites. The character of spectra discordance displayed by the muscovite concentrate is identical to that which has been described from partially rejuvenated muscovite in other polymetamorphic settings (e.g., Dallmeyer and Takasu, 1992). Comparison of the present discordant muscovite spectrum from results in other areas suggest that initial cooling through appropriate temperatures for argon retention (c. $375 \pm 25^\circ\text{C}$) at c. 296 Ma was followed by slight (~ 10 %) Mesozoic rejuvenation of intracrystalline argon systems.

Getic Nappe

Făgăraș Mountains

Hornblende concentrates have been prepared from three samples of amphibolite collected along a northward traverse through exposures of the Getic Nappe in the Făgăraș Mountains (Fig. 1: DA-ROM2A-93, DA-ROM7A-93, DA-ROM9A-92). The three concentrates display similar internally discordant age spectra (Fig. 4) as a result of low-temperature experimental evolution of gas from compositionally variable phases. However, the intermediate- and high-temperature gas fractions display minor intrasample variations in apparent K/Ca ratios indicating experimental evolution of gas occurred from compositionally uniform intracrystalline sites. These



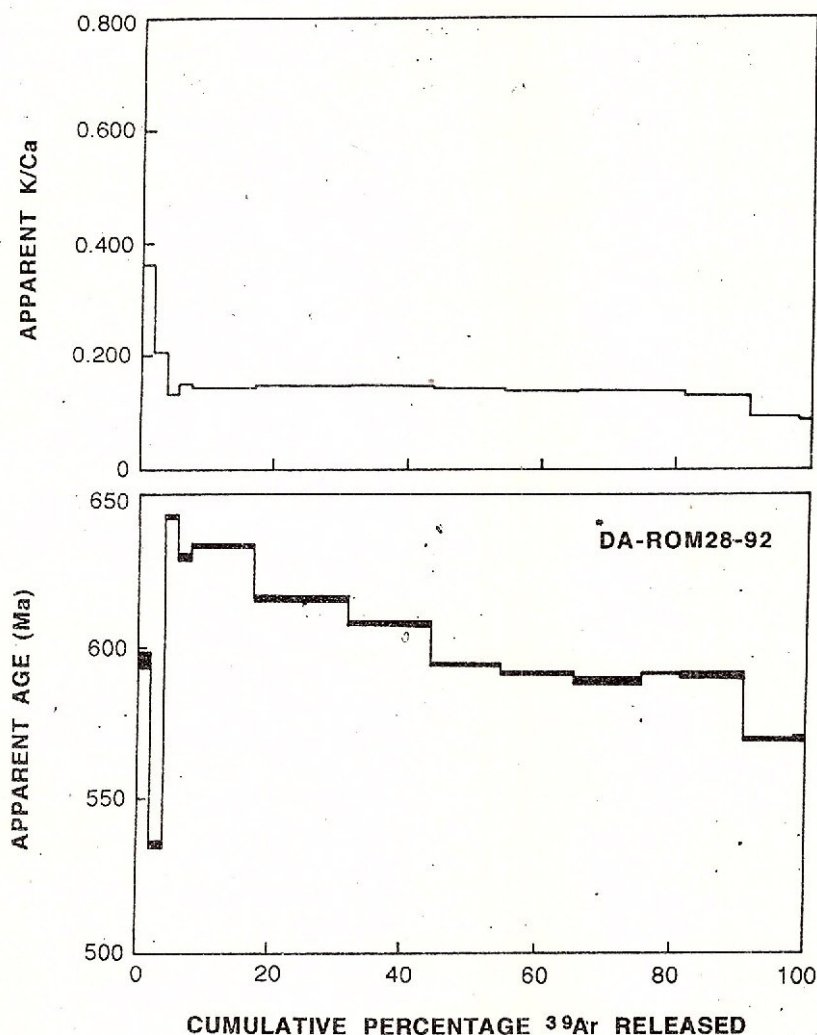


Fig. 2 - $^{40}\text{Ar}/^{39}\text{Ar}$ apparent age and apparent K/Ca spectrum of hornblende concentrate from the Tismana diorite, Danubian parautochthon. Analytical uncertainties (two sigma, interlaboratory) represented by vertical width of bars. Experimental temperatures increase from left to right.

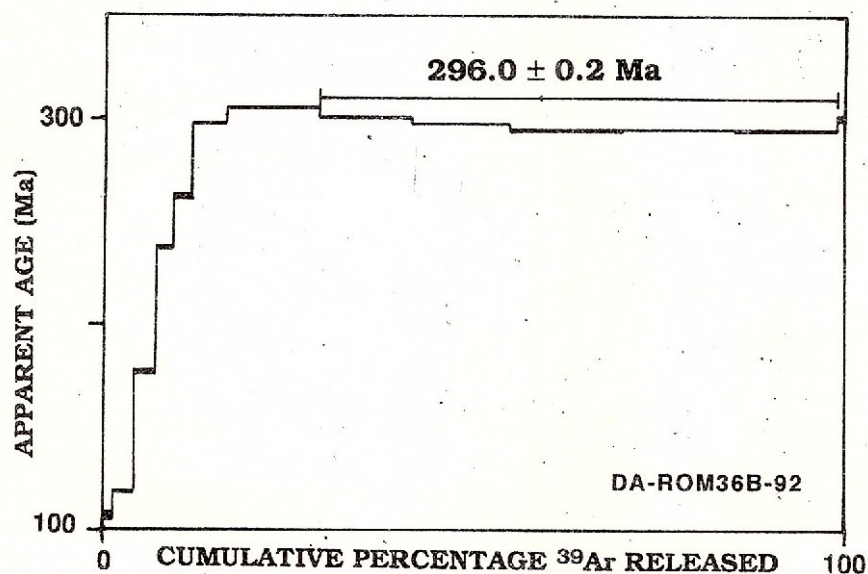


Fig. 3 - $^{40}\text{Ar}/^{39}\text{Ar}$ apparent age spectrum of a muscovite concentrates prepared from the Danubian parautochthon. Data plotted as in Figure 2.

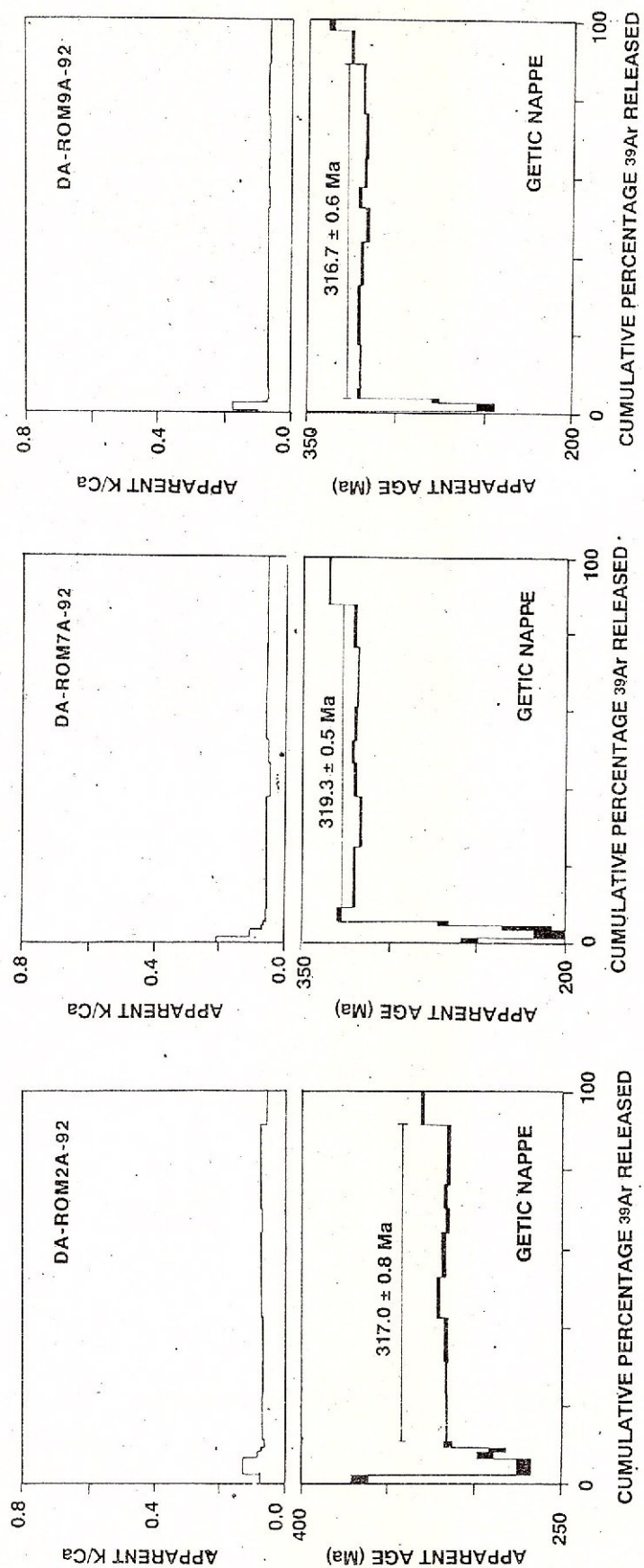


Fig. 4 - $^{40}\text{Ar}/^{39}\text{Ar}$ Ar apparent ages and apparent K/Ca ratios from hornblende concentrates from the Făgăraș Mountains. Data plotted as in Figure 2.

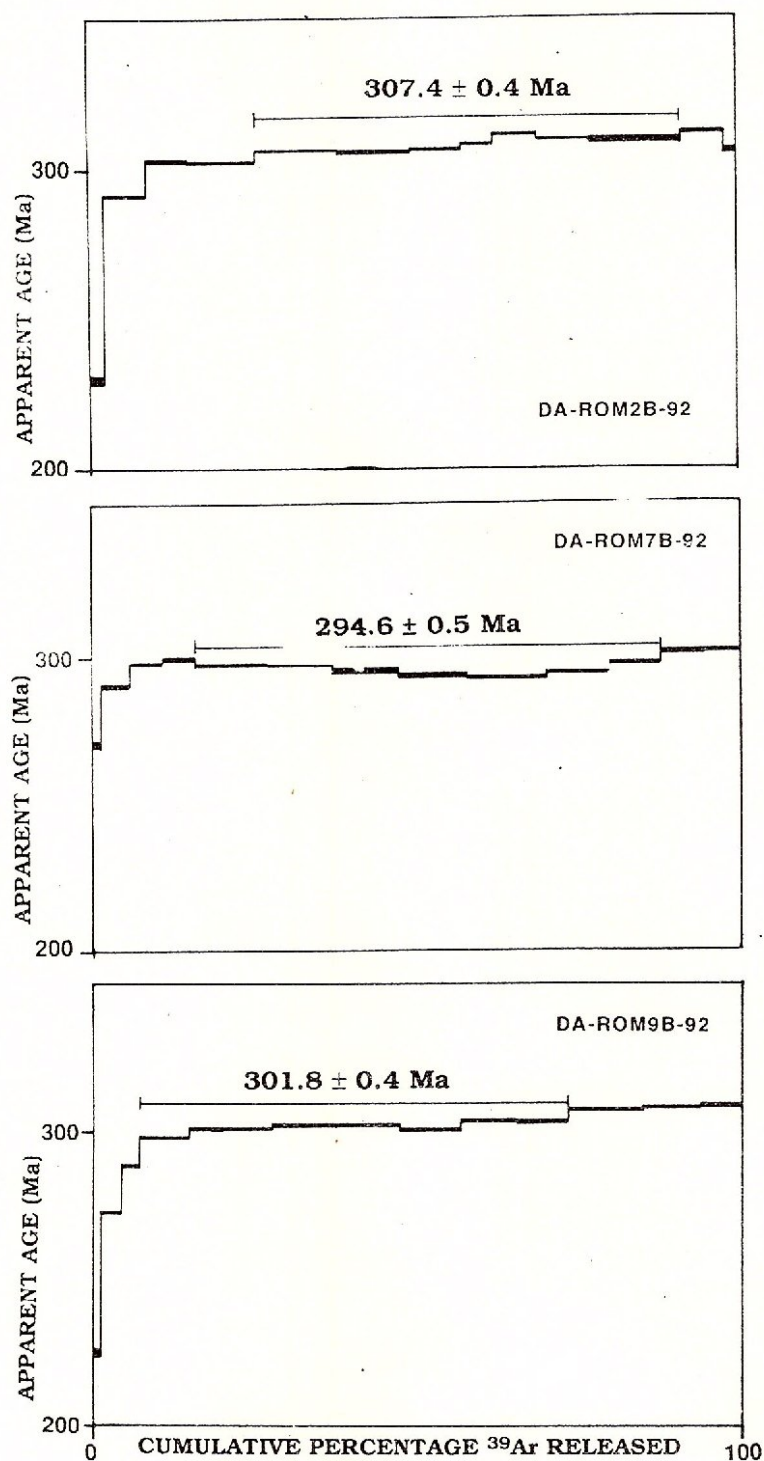


Fig. 5 - $^{40}\text{Ar}/^{39}\text{Ar}$ apparent age spectra of muscovite concentrates from gneisses of the Făgăraș Mountains. Data plotted as in Figure 1.



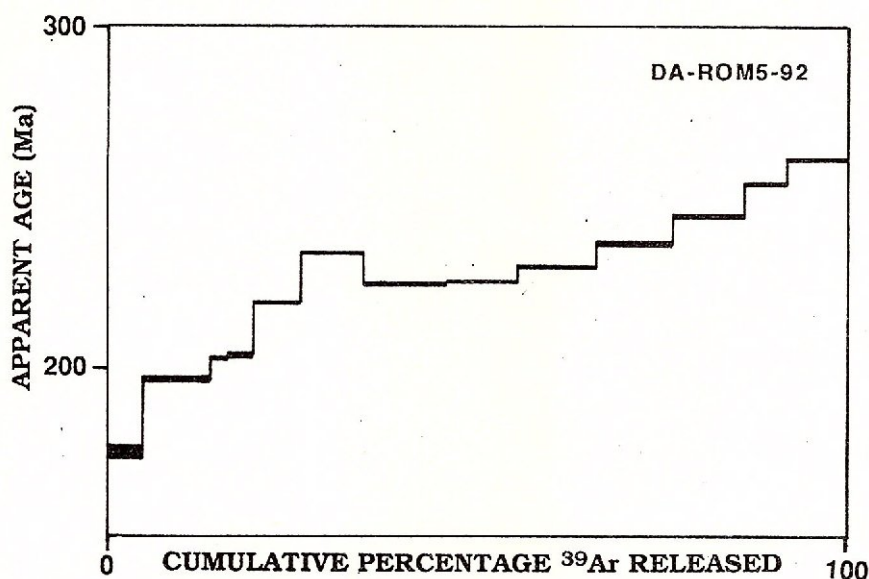


Fig. 6 - $^{40}\text{Ar}/^{39}\text{Ar}$ apparent age of a muscovite concentrate of a mylonitic quartzite from the northern Făgăraș Mountains. Data plotted as in Figure 2.

intermediate- and high-temperature gas fractions yield plateau ages of 317 ± 0.8 Ma and 318.7 ± 0.6 Ma. $^{36}\text{Ar}/^{40}\text{Ar}$ isotopic correlations of the plateau data are well-defined (MSWD ; 2.0), with inverse ordinate intercepts ($^{40}\text{Ar}/^{36}\text{Ar}$ ratios) that are only slightly larger than the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio in the present-day atmosphere. This suggests insignificant intracrystalline contamination with any extraneous argon components. Using the inverse abscissa intercepts $^{40}\text{Ar}/^{39}\text{Ar}$ ratios in the age equations yields plateau isotope correlation ages that are the same as the calculated values within analytical uncertainties. The plateau ages are considered geologically significant and are interpreted to date the last cooling through temperatures required for intracrystalline argon retention. Harrison (1981) suggested that temperatures of c. 500°C are appropriate for argon retention within amphibole at cooling rates which likely characterize most regional metamorphic settings.

Muscovite concentrates were prepared from samples of schist collected from the same three exposures along the Făgăraș Mountains traverse (Fig. 1: DA-ROM2B-92, DA-ROM7B-92, DA-ROM9B-92). These record well-defined plateau ages (Fig. 5) of 307.4 ± 0.4 Ma, 294.6 ± 0.5 Ma and 301.8 ± 0.4 Ma. These are interpreted to date the last cooling through appropriate argon retention temperatures. Slightly younger apparent ages are recorded at lowermost experimental temperatures and suggest slight ($\sim 10\%$) Mesozoic rejuvenation of intracrystalline argon systems.

A muscovite concentrate was prepared from a mylonitic quartzite collected at Bilea Cascada in more northern sectors of the Făgăraș Mountains (Fig. 1: DA-ROM5-92). The concentrate is characterized by a very discordant age spectrum (Fig. 6) which suggests extensive Mesozoic rejuvenation of intracrystalline argon systems which had initially cooled through appropriate closure temperatures sometime prior to c. 250 Ma.

Several samples were collected in the vicinity of a regional ductile shear zone exposed in northwestern sectors of the Făgăraș Mountains (Fig. 1). These included schist of the Sebeș-Lotru Series (DA-ROM13-92), retrogressed mylonitic schist proximal to the shear zone (DA-ROM16-92), and very fine-grained phyllonite within shear zone (DA-ROM12-92). A muscovite concentrate from the Sebeș-Lotru schist records a 299.4 ± 0.5 Ma plateau age (Fig. 7). Low-temperature gas increments suggest only slight ($\sim 10\%$) Mesozoic rejuvenation. A muscovite concentrate from the retrogressed schist is characterized by an internally discordant spectrum which suggests extensive Mesozoic rejuvenation of intracrystalline argon systems which had initially cooled through appropriate argon retention temperatures sometime prior to c. 200 Ma. A whole-rock sample of the phyllonite was analyzed. Low-temperature gas fractions are characterized by considerable variations in apparent age and relatively low apparent K/Ca ratios. These are ascribed to gas evolved from constituent chlorite. The intermediate- and high-temperature gas fractions are characterized by higher apparent K/Ca ratios and are interpreted to represent gas evolved from constituent, very fine-grained white-mica. The intermediate- and

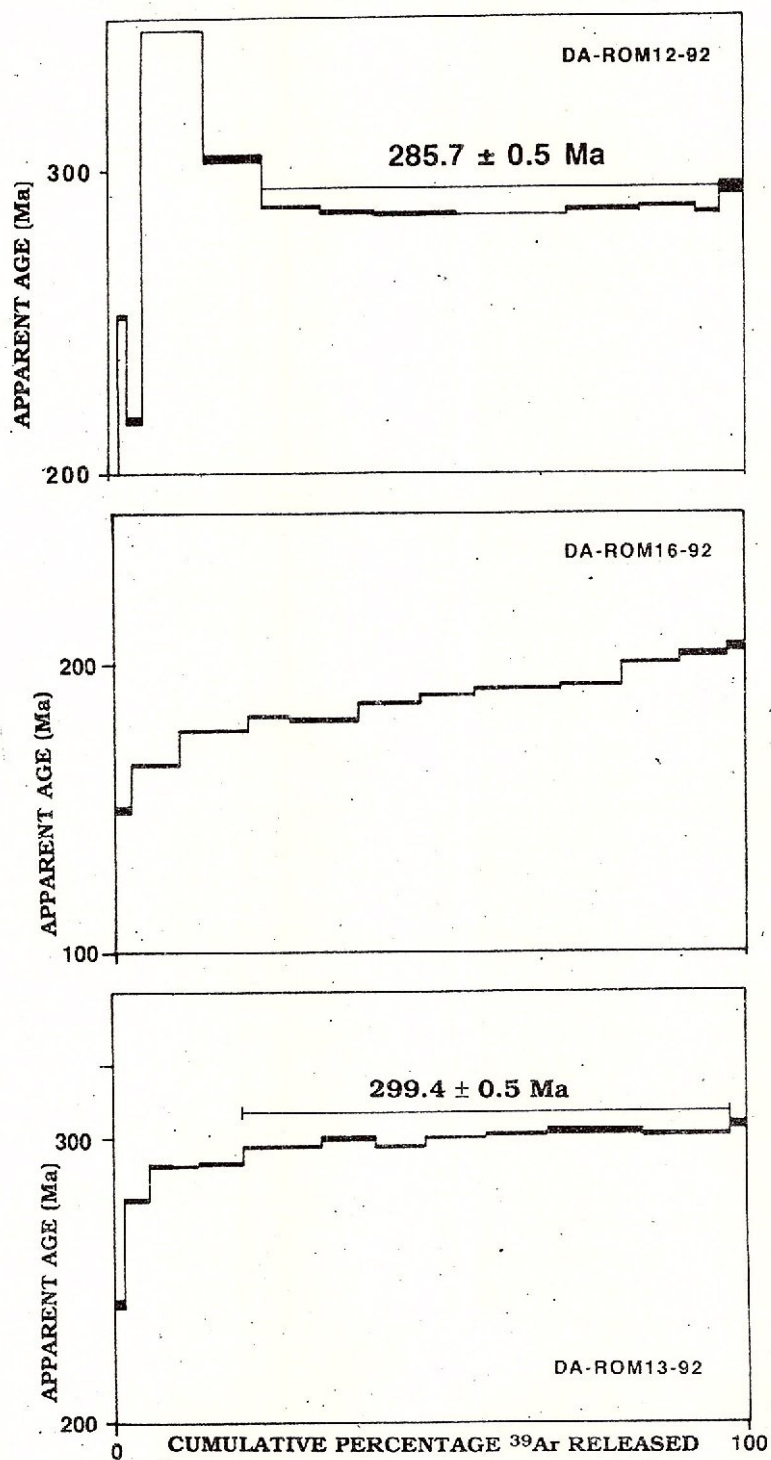


Fig. 7 - $^{40}\text{Ar}/^{39}\text{Ar}$ apparent age spectra of muscovite concentrates from variably retrogressed, mylonitic schists from northwestern Fagaras Mountains. Data plotted as in Figure 2.

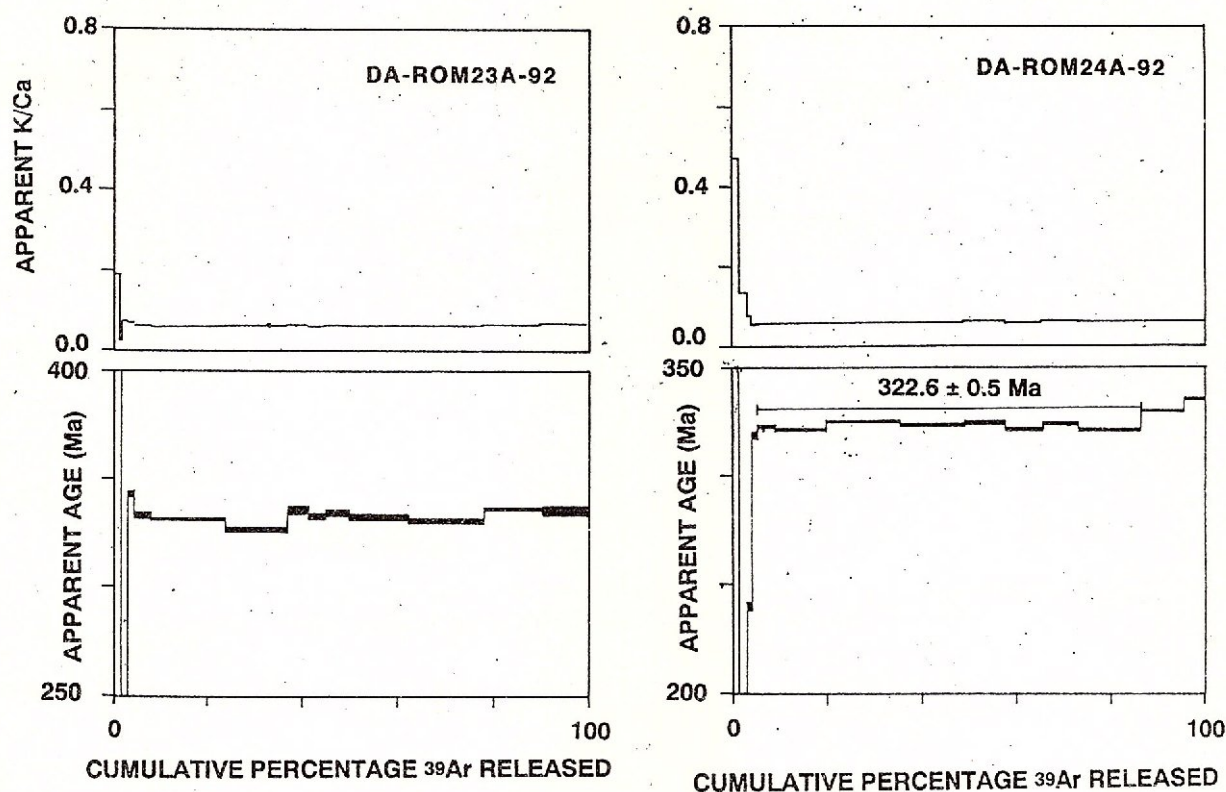


Fig. 8 – $^{40}\text{Ar}/^{39}\text{Ar}$ apparent ages and apparent K/Ca ratios from hornblende concentrates from the Vulcan Mountains. Data plotted as in Figure 2.

high-temperature gas fractions record generally similar apparent ages which define a plateau age of 285.7 ± 0.5 Ma.

Southwest Vulcan Mountains

Hornblende concentrates have been prepared from amphibolite collected within southwestern exposures of the Getic Nappe north of Turnu (Fig. 1: DA-ROM23A-92, DA-ROM24A-92). Both concentrates record well-defined intermediate- and high-temperature plateau ages of 329.4 ± 0.9 Ma and 322.6 ± 0.5 Ma (Fig. 8). A muscovite concentrate from a gneiss sample collected at location 23 (DA-ROM23B-92) is characterized by an internally concordant age spectrum (Fig. 9) which defines a plateau age of 309.5 ± 0.5 Ma.

Tectonic Implications

The preliminary $^{40}\text{Ar}/^{39}\text{Ar}$ results suggest that regionally penetrative high-temperature mineral assemblages and associated ductile structural elements within structural units of the Danubian autochthon/parautochthon and the Getic Nappe Complex are all likely related to Late Paleozoic (late Variscan) tectonothermal activity. The similar Late Variscan mineral ages constrain large-scale block uplift and exhumation of metamorphic sequences contemporaneous with localized deposition of Late Carboniferous terrestrial deposits, which are present, e.g., within the western Supragetic nappe complex. These relationships may record large-scale extension following Variscan orogenic events, especially medium-grade metamorphism. The Alpine record appears to be restricted to discrete, intermediate- or low-temperature ductile shear zones developed in basement elements. The record of any pre-Variscan tectonothermal events is apparently restricted to emplacement of Cadomian

plutons within the Danubian autochthon/parautochthon.

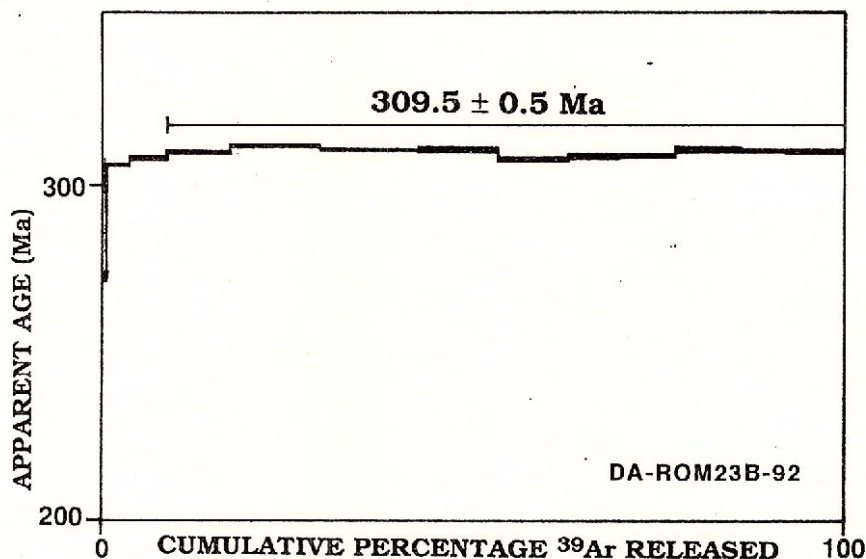


Fig. 9 - $^{40}\text{Ar}/^{39}\text{Ar}$ apparent age spectrum of a muscovite concentrate from the Vulcan Mountains (same locality as the hornblende concentrate DA-ROM23A). Data plotted as in Figure 2.

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PRE-ALPINE LITHO-TECTONIC UNITS AND RELATED SHEAR ZONES IN THE BASEMENT OF THE GETIC – SUPRAGETIC NAPPE (SOUTH CARPATHIANS)¹

by

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1. Alpine Setting

In the Alpine nappe pile of the South Carpathians, which includes Getic and Supragetic units, discontinuous sequences of Paleozoic and pre-Paleozoic age are well preserved. From the stratigraphic record and the structural data a polyphase alpine evolution is suggested. The Alpine nappe pile, from the structural top to the bottom, includes: Supragetic and Getic units, Infragetic units (Severin and Arjana nappe complexes) and Danubian units. The main tectogenetic events are interpreted as result of closure of an oceanic and zone continental collision during Upper Jurassic (Kimmerian), Mid-Cretaceous (Austrian event) and Senonian (Laramian event). The nappe stack was ultimately overthrust onto the Moesian Platform during the Miocene. The Austrian nappes were generated in the Getic realm by thrusting of the Supragetic onto the Getic units synchronously with shortening and subduction of the oceanic crust in the Severin realm (Rădulescu, Săndulescu, 1973) and underplating of the Arjana and Danubian realm as well as the western part of the Moesian Platform. An important dynamic reactivation of the pre-Maastrichtian rock-associations (pre-Alpine basement and Mesozoic cover) resulted because of the Laramian overthrust of the entire nappe pile onto the Moesian Platform (Berza et al., 1983; Iancu, 1986; Balintoni et al., 1989).

The Getic – Supragetic units show folding of the Mesozoic sedimentary sequences and local dynamic metamorphism, while Alpine polydeformations and polyphase metamorphism are well developed in the Infragetic and Danubian units.

Although polycyclic reactivation of the pre-Mesozoic rocks during the Alpine orogeny is well expressed, Paleozoic nappe structures and Pre-Paleozoic shear zones are well preserved in the basement of the Getic-Supragetic nappes.

Alpine and pre-Alpine shear zones, representing the main thrust faults are presented in Plate I(A,B).

2. Pre-Alpine basement of the Getic-Supragetic Units

The basement of the pre-Triassic-Liassic cover includes:

- Upper Paleozoic (Westphalian-Permian) coal bearing, coarse sedimentary deposits of molasse type and scarce magmatic rocks, (Năstăseanu et al., 1981);
- Paleozoic granitoids (Sichevița-Poniasca) whose intrusion (U-Pb age: 350 Ma; K-Ar age: 310 Ma, in Stan et al., 1992) post-dates the emplacement of the Paleozoic nappe pile in the Bozovici zone (Pl. IA);
- Low-grade Paleozoic sequences consisting of Cambrian – Silurian and Upper Devonian-Lower Carboniferous metasedimentary rocks and related magmatites are well exposed in the westernmost (Banat region, Pl. IA; Poiana Ruscă massif is not included) and easternmost part (Leaota mountains, Pl. IB) of the South Carpathians; the Paleozoic age of the low-grade metamorphic sequences is well documented by palynological evidence and K-Ar ages, both in good agreement;

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– Medium- to high-grade metamorphic complexes, with scarce geochronological evidence concerning protolith ages as well as metamorphic events.

Besides the clear regional and dynamic low-grade reactivation, documented by K-Ar data as Paleozoic (Variscan), in some outcrop areas of the medium- to high-grade rocks few indications exist concerning the pre-Paleozoic age of protoliths and metamorphic events: a reported U/Pb analysis on zircon crystals from a gneiss (Bota Valley) is 1000–1100 Ma (Pavelescu et al., 1983); 1600 Ma (U-Pb method), upper interception and 310 Ma, lower interception (Pavelescu et al., 1984, unpubl. report, in polymetamorphic augengneisses; 850 Ma (whole rock Rb/Sr method) (Bagdasarian, 1972), in micaschists, from the getic nappe basement.

There is a good agreement between the low-grade (greenschists facies) reactivated areas of the medium- to high-grade rock assemblages and geographic distribution of the Paleozoic low-grade formations and related shallow level shear zones.

New data concerning the protolith ages and tectono-metamorphic evolution of the pre-Alpine basement are extremely important in understanding the relationships between different types of the present juxtaposed metamorphic terrains. A conventional generalisation of the Proterozoic ages for the medium- to high-grade metamorphosed protoliths from the Getic-Supragetic basement was accepted. The polymetamorphic evolution of these rocks include pre-Paleozoic and Paleozoic events, in many areas an obvious overlapping of metamorphic parageneses or metamorphic zones and structures being described.

2.1 Paleozoic formations; low-grade metamorphism.

This paper gives only a short presentation of the undifferentiated low-grade (anchimetamorphic to greenschists facies) Paleozoic sequences, separated in the sketch (Pl. I, II). They are very diversified from the lithologic-stratigraphic point of view; discontinuous sections crop out in some Alpine and Paleozoic (Variscan) tectonic units, as shown in published materials (1:50 000 geologic maps; Maier, 1974; Iancu, 1986; Iancu, Mărunțiu, 1989). The following lithostratigraphic (and tectonostratigraphic) units crop out in the Getic-Supragetic nappes:

Caraș Group (outcrop areas in the westernmost part of the South Carpathians, Banat region, in the Bocșa alpine nappe). Scarce palynological data indicate Cambrian-Ordovician-Silurian(?) ages, (Visarion, Iancu, 1984). -This group includes the following formations with contrasting lithologies: Nădăș-Rafnic volcano-sedimentary Fm (at the base), Dognecea-Zlatița terrigenous Fm, Tâlva Mare quartzitic Fm. The characteristic feature of the Caraș Group is the development of a bimodal magmatic association (transitional basalts and high-Si alkaline rhyolites) in the volcano-sedimentary formation. Scarce bodies of mafic (metagabbros, metadolerites) and ultramafic magmatites are associated with volcanic, volcanoclastic and terrigenous (metapelito-psammitic and quartzitic) rocks.

Well-preserved pre-metamorphic magmatic parageneses and structures are conserved in magmatic rocks (rhyolites, gabbros, dolerites) while metamorphic parageneses point out to a polymetamorphic history.

A polyphase evolution is proved by the existence of two generations of folds (F_1 , F_2) and foliations (S_1 , S_2) with related metamorphic greenschists facies parageneses. The first dynamo-thermal event (M_1) was estimated to occur in greenschist facies to epidote-amphibolite facies conditions: chlorite-stilpnomelane biotite and locally, garnet zones. The second event (M_2) is retrograde in character (chlorite on biotite and garnet, actinolite on hornblende, etc).

Lithological constitution and metamorphic history suggest that Călușu Formation (Călușu-Tămășel complex cf. Dimitrescu et al., 1971; Călușu Fm. cf. Tatu, Robu, 1987), which crop out in the easternmost part of the South Carpathians, in Leaota mountains (Pl. 1B) might be an equivalent of the lower formation of the Caraș Group. This formation contains a lower member with well represented metabasic rocks and an upper metaterigenous member.

Moniom Group (outcrop area in Banat region, in the Moniom Alpine nappe). This consists of a volcano-sedimentary (Valea Satului) formation (Upper Devonian-Lower Carboniferous) and a dominantly conglomeratic (Cârșie) formation, Lower Carboniferous in age (Visarion, Iancu, 1984). The volcano-sedimentary formation includes mainly basic metatufs with interlayered metapelitic and carbonate rocks, scarce acid metatufs and small bodies of gabbros and diorites-granodiorites. A very low-grade metamorphism (anchizone to greenschists facies conditions) coexists with well-preserved sedimentary structures (bedding, pebbles, etc) and a dominantly mechanical transposition with related S_1 cleavages and foliations.

The previously described Caraș and Moniom Groups belong to the basement of the Supragetic nappes. The next described Lower Paleozoic sequences (Miniș Formation and Buceava Group crop out in the Getic nappe, in Bozovici – Sichevita zone (Pl. 1A). The only proof for their presumed lower paleozoic age is a poorly presented macrofossil (Ordovician?) found in Buceava Group rocks (Iordan, Conovici, oral communication).



Miniş Formation is a monotonous metaterrigenous sequence (mainly chlorit-biotite quartzite schists) with minor bodies of metarhyolites, suggesting a continental origin. Polydeformational features (F_1 , F_2 folds and S_1 , S_2 foliations) and chlorite-biotite-blastesis may be good evidence in separating this sequence from Buceava Group rocks of the underlying Variscan nappe and from the retrograde Lotru Group rocks of the overlying tectonic unit (Iancu, Mărunțiu, 1989).

Buccava Group includes a discontinuous, tectonic slab of dismembered basaltic rocks with associated black shales, siltites and carbonate rocks (Agris volcanogenous Fm); a continental, metapsephitic to metapsammitic sequence contains reworked basalts and other magmatic rocks (e.g. gabbros), as well as clasts from the mesometamorphic basement (Șopot Fm). Unhomogenous shear bands and prehnite-pumpellyite blastesis in the basaltic rocks, as well as chlorite-albite-calcite-sericite assemblages in volcanic and terrigenous rocks associated to S_1 cleavages and synchronous F_1 folds suggest a low-grade (Paleozoic) metamorphic event.

2.2 Proterozoic formations; medium to high-grade metamorphic evolution.

Detailed mapping in Proterozoic formations enabled the separation of various lithological divisions which were given local names; as a result, equivalent sequences from various geographic areas had different names which made their correlation difficult. The most important metamorphic series and (litho)groups specified in the geologic maps and publications were presented by Balintoni et al. (1989): Sebeș-Lotru (as undivided Precambrian basement of the Getic nappe west of the Olt river), Cumpăna Făgăraș and Leaota (east of the Olt), Bocșița-Drinț (Banat), Ghelara (Poiana Ruscă massif), etc. All these series were thought to be generally delimited by Alpine tectonic lines.

Besides geologic maps at scale 1:50 000, significant regional lithologic maps were published by: Savu (1965), Bercia (1975), Savu et al (1978), Balintoni et al (1986), Gheuca (1988), Gheuca, Dinică (1986), Hann et al (1987), Kräutner (1980), Iancu et al (1987), Stelca (1992).

All the lithologic sequences quoted above contain medium-grade mineral assemblages from staurolite to kyanite or sillimanite zones (Savu, 1965, Bercia, 1975) and from andalusite to cordierite-sillimanite zones (Bercia, 1975). They show a kyanite-staurolite zone polymetamorphic (M_1 , M_2) and polydeformational evolution (Balintoni, 1975, Hărtopanu, 1975, Iancu, Hărtopanu, 1979). However, in some areas, high-grade rocks with longer metamorphic history and different metamorphic evolution were reported (Iancu et al 1987, 1989, Săbău et al 1987). Also, individualisation of the superposed low-pressure facies parageneses in some areas (static, cf. Hărtopanu, 1975; synchronous with M_2 dynamo-thermal event, cf. Iancu et al 1987) points out on important changing in the geothermal gradients in the final stages of the collisional history. Domal high temperature areas are correlated with regional metatectic migmatization and partial melting products.

In the last years, Iancu et al (1989–1992, unpublished reports), Iancu, Mărunțiu (1994) proposed a new division of the major series (Sebeș-Lotru, Cumpăna and Leaota), using lithological criteria and taking into account the existence of pre-Paleozoic shear zones. Additional proofs are lithologic associations, internal structures of individual units, metamorphic history, geochemical data, etc. These entities, which reflect different preorogenic geotectonic environments are separated as litho-tectonic units or groups joined along pre-Paleozoic tectonic lines (Pl. I, II). The tectonic lines represent mainly deep-seated shear zones interpreted as thrust faults and associated in places with a strike-slip component and marked by discontinuous occurrences of medium- to high-grade mylonites.

The simplified lithologic content of the main litho-tectonic units are shown in plate II A,B.

Sebeș Group is a crustal-supracrustal assemblage, constituted by the following formations:

a) Micaschist Fm: metapelites-metapsaminites, scarce acid leptynitic rocks, manganiferous rocks; isolated blocks of mafic (eclogite) and ultramafic rocks are spatially related (litho-type sequence in Sebeș Mountains);

b) Gneiss Fm: ortho and paragneisses, including bimodal magmatic association (leptyno-amphibolite), metaultramafites, eclogites and manganiferous rocks (litho-type: Sebeș Mountains, Strei basin sequences); Both formations contain good lithologic markers represented by tephroite concentrations originating from nodule type, manganiferous oceanic crust metasediments (Hărtopanu, Udrescu, 1981).

c) Paragneiss Fm: metapsammitic±quartzitic±carbonatic±amphibolic rocks (litho-type: Bahna, Porțile de Fier outliers, Căpățâna Mountains sequences, plate 2 A).

All these formations include bodies, of mafic and ultramafic rocks with well-preserved eclogitic and granulitic facies parageneses, of intracontinental and/or oceanic origin, tectonically enclosed in the Sebeș Group before its first metamorphic event (M_1) (Iancu et al, 1988).

Lotru Group is defined as a supracrustal megasequence including oceanic-type metasediments and tectonic slabs of dismembered oceanic crust: metaperidotites and metagabbros (displaying relict cumulus struc-



tures and acid ortho-derivates. Geochemical data supporting the oceanic crust origin for these rocks were published by Savu et al, (1982). Some of the magmatic rocks are affected by prograde eclogitisation, followed by polyphase dynamo-thermal metamorphism (M_1 and M_2 events), suggesting a subduction-exhumation type evolution of the remnants. In the Lotru Group two lithologic units can be distinguished:

- a) a terrigenous formation: metapelito-psammites, quartzite and carbonate rocks (litho-type: upper reaches of the Lotru valley);
- b) metabasic complex including dismembered ophiolites, acid orthoderivates and metasediments (litho-type: lower reaches of the Lotru basin).

Cumpăna Group, well exposed on the southern slope of the Făgăş Mountains and on the northern slope of the Sebeş Mountains (Pl. II B), includes the following subunits:

- a) a metaterigenous formation with associated porphyritic orthogneisses (KFsp. augen gneisses) (litho-type: lower part of the "Sebeş-Lotru" series, cf. Hartopanu et al, Stelea et al, unpublished maps, on the northern slope of the Sebeş Mountains);
- b) "greenstone" complex: metagabbros-metatonalites, amphibolites and related metasediments (litho-type: Topolog + Cumpăna formations, cf. Dimitrescu et al 1985; Gheuca, 1988);
- c) orthogneiss (KFsp augen gneisses) complex and subordinate metasediments (litho-type Cozia Fm, cf. Dimitrescu et al, 1985).

Cumpăna Group is a composite terrain including gneissic domes of magmatic protholits (garnet bearing Brebu plagiogneisses, Cozia-type augen gneisses, Cumpăna layered metagabbros); metamorphic evolution started with a prograde high PT event, preserved in the eclogitic and granulitic nuclei, followed by the polyphase amphibolite facies dynamo-thermal metamorphism.

Ursu Group, a complex of metapsammitic and mafic-ultramafic rocks (Hann et al, 1987), contains nuclei of low- to medium-pressure granulites (garnets, cordierite, sillimanite or pyroxene bearing) dynamically re-equilibrated under amphibolite facies conditions (cordierite-sillimanite bearing blastomylonites, cf. Săbău et al, 1987). This unit might represent a thinned continental crust involved in the last evolution in metamorphic events (M_1/M_2) of a Proterozoic orogeny together with the Sebeş, Lotru and Cumpăna Groups. This group was thought as a continuously equilibrated gneissic mass as the result of cooling and uplift (Săbău et al, 1987).

Făgăraş Group, defined by Dimitrescu (1987) as Făgăraş Series, was restricted to the "Suru"-type lithologies (Stelea, 1992; Balintoni, Dinu, 1993) which include: metapsammities, carbonate rocks, paramphibolites and, subordinate, graphite quartzites. This group is a supracrustal, epicontinental platform-type association of terrigenous and carbonate rocks, characterised by medium-grade polymetamorphism and dynamo-thermal greenschist facies reactivation during the Variscan event (Balintoni et al, 1986; Stelea, 1992). Dynamic alpine retrogression overprints earlier medium- and low-grade metamorphic parageneses involved in the Alpine nappes (Balintoni, 1993).

2.3 Pre-Alpine shear zones.

Stratigraphic record (transgressive unmetamorphosed Westphalian deposits) and presence of Carboniferous magmatic intrusions enabled the distinction between pre-Alpine and Alpine shear zones (pl. I, A,B).

A first approximation elaborated in 1989 (Iancu et al, unpublished report) helped to separate a set of Paleozoic low-grade shear zones representing thrust planes (Iancu et al, 1988) and a group of "deep-seated shear zones" considered at this moment pre-Paleozoic.

Low- grade Paleozoic shear zones are represented by low-grade, ductile mylonites (chlorite, biotite zones) developed along Paleozoic thrust faults between low-grade Paleozoic units or between these and polymetamorphic Proterozoic units, retrogressed or not. Mineral assemblages of mylonites suggest relatively shallow level of nappe emplacement.

Pre-Paleozoic shear zones are included in several main groups based on their relationship with structural elements connected to regional metamorphic processes (S_1 , S_2 foliations and associated F_1 , F_2 folds) and on the common evolution of the lithotectonic units during the M_2 Proterozoic dynamo-thermal collisional event:

- a) M_2 related shear zones appear as structural discontinuities between the main lithotectonic units (pl. I,II); they represent thrust faults and possibly combined thrust and strike-slip tectonic lines. Medium- to high-grade mylonites showing linear and planar-linear fabrics and effects of non-coaxial progressive deformation are discontinuously exposed along these shear zones. Physical conditions of deformation are documented by syn-tectonic mineral phases of the mylonites: kyanite±sillimanite, biotite+garnet, amphibole+plagioclase



(in medium-grade mylonites) or omphacite+garnet+kyanite±amphibole (in high-grade mylonites – Leaota Mountains).

b) Possible pre- M_1 shear zones are obliterated by M_2 dynamo-thermal metamorphic overprint and may coincide with some lithologic boundaries between contrasting lithologic sequences in Sebeş and Cumpăna units (pl. I,H).

c) Pre- M_1 deep-seated cryptic shear zones, whose HP-HT shearing indicators include pre- M_1 eclogites, HP granulites and pyrope-rich garnet metaperidotite bodies from Sebeş Unit; oriented fabric of the omphacite-kyanite in eclogites and banded rocks as well as decompression structures in granulitic facies conditions (clinopyroxene-plagioclase and spinel-corundum-plagioclase symplectites) older than amphibolitic facies parageneses are good evidence for lower crust shear processes followed by up-lift and exhumation of these bodies before or simultaneously with the first regional metamorphism (M_1) of the country rocks. Regional distribution of these high-grade rocks, correlated with their lithology support the idea of a pre-Paleozoic cryptic paleosuture preserved in the Sebeş Unit (Iancu et al, 1987).

Because timing of shear zones formation is a difficult task of geochronological evidence, only qualitative and descriptive data were presented on this subject in unpublished reports (Iancu et al, 1989–1993).

Complex structural and kinematic analysis of shear zones are in progress performed by Romanian geologists from GIR and French specialists from BRGM.

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Plate I A, B – Variscan and pre-Variscan shear zones (tectonic boundaries) in Getic – Supragetic Realm
(South Carpathians).

Pl. IA, western part; Pl. IB, central and eastern part of the chain.

1, fault; 2, reverse fault; 3, Alpine overthrust plane; 4, Variscan thrust (low grade shear zones); 5–6, Proterozoic medium – to high-grade shear zones; 5, M2 related overthrust plane; 6, unprecised tectonic boundary; 7, Upper Cretaceous – Paleocene magmatites; 8, a, Mesozoic and b, Upper Paleozoic sedimentary cover; 9, Paleozoic granites; 10, paleozoic, low-grade rocks associations; 11, Proterozoic sequences; 12, eclogitic rocks; 13, granulitic rocks; 14, metaultramafic rocks; 15, HP granulitic slabs (a) and LP granulitic terrains (b); 16, reverse antiform.

Abbreviations for the main litho-tectonic units: CP – Cumpăna Group; Lo – Lotru Gr.; Sb – Sebeș Gr.; Fg. – Făgăraș Gr.; Ur – Ursu Gr.

Other cartographic materials used: for Alpine tectonic lines: 1:50 000 maps of Romania; Balintoni et al. (1989); Hann (1993); Năstăseanu et al. (1981). Local pre-Alpine shear zones: Iancu et al (1987); Iancu et al (1988); Stelea (1992).

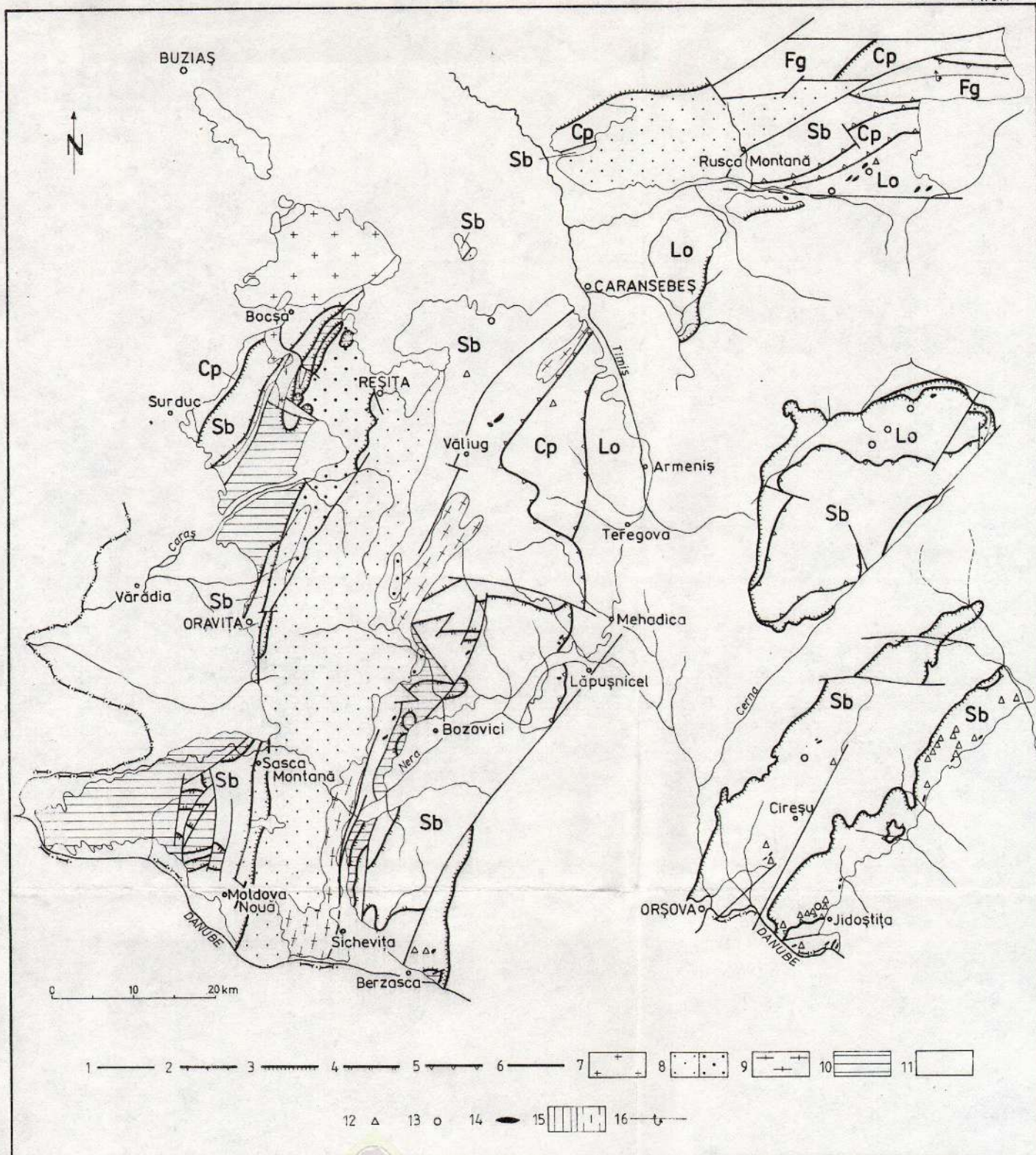
Plate II A, B – Litho-tectonic units in the pre-Alpine basement of the Getic – Supragetic Nappes.

Pl. IA, western part; Pl. IB, central and eastern part of the chain.

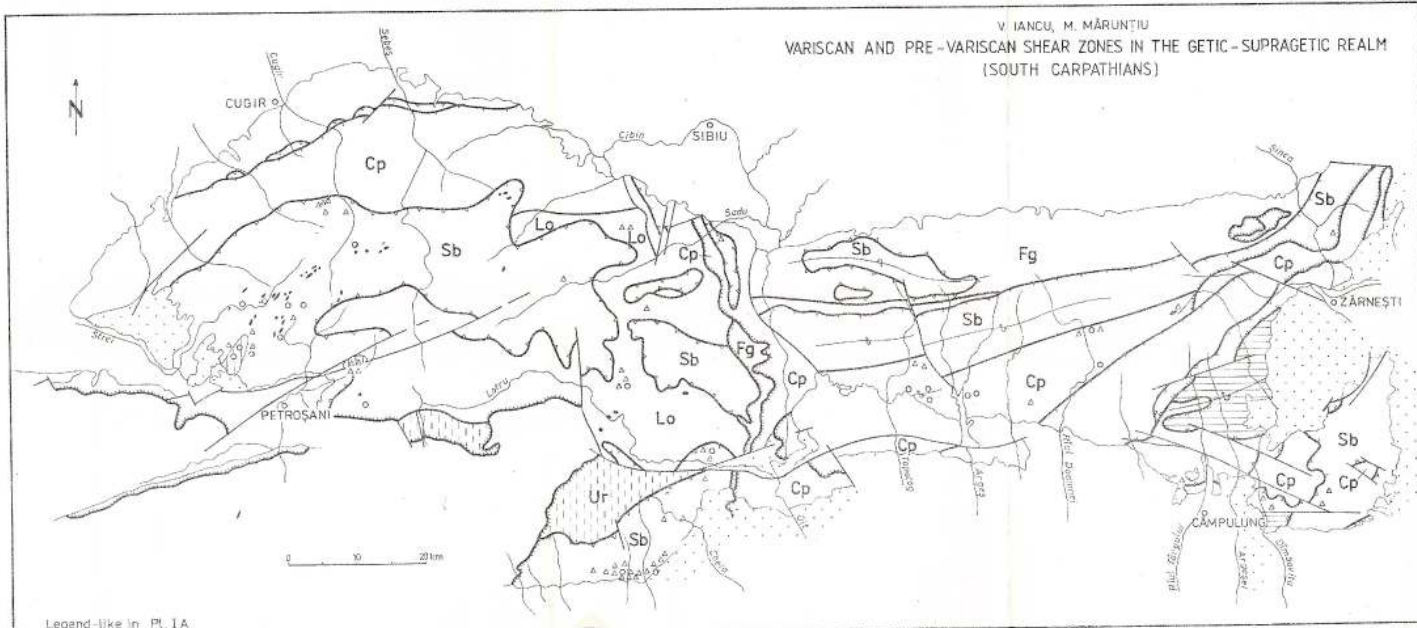
1, Upper Cretaceous – Paleocene magmatites; 2, undifferentiated cover: Upper Carboniferous – Permian + Mesozoic-cenozoic; 3, Carboniferous (Sichevița – Ponișca) Granitoids; 4, Paleozoic formations (low grade metamorphism); PRECAMBRIAN FORMATION: (Precambrian and/or Paleozoic medium- to high-grade metamorphism) Sebeș Group: 5, Micaschist formation (metapelito-psamitic + manganiferous rocks); 6, Gneiss formation (ortho + paragneisses including bimodal magmatic association, metaultramafites, manganiferous rocks); 7, Paragneiss formation (metapsamitic ± quartzitic ± carbonatic ± amphibolitic rocks); Lotru Group: 8, Terrigenous formation (metapelito-psamitic+quartzitic±carbonatic rocks); 9, Metabasic complex including dismembered ophiolites, felsic orthoderivates, metasediments; Cumpăna Group: 10, Terrigenous formation with associated orthogneisses (Kfsp. augengneisses); 11, "Greenstone" complex (gabbros-tonalites and related metasediments); 12, Orthogneiss (Kfsp augengneisses) complex + subordinate metasediments; Ursu Group: 13, Metapsamitic + mafic-ultramafic rocks; Făgăraș Group: 14, Metapsamitic and carbonatic rocks, graphite quartzites, paraamphibolites; Luchîța Group: 15, Terrigenous, quartzite formation; 16, geologic limit; 17, fault; 18, reverse fault; 19, Alpine overthrust plane; 20, Paleozoic overthrust plane (low-grade mylonites); 21, Ante-Paleozoic tectonic boundary (± medium to high-grade mylonites; M2 related shear zones); 22, lithologic discontinuities (pre-M2 shear zones); 23, medium to high-grade mylonites.

Cartographic materials used: 1:200000 map of Romania; 1:50 000 geological maps of Romania; personal data; Apostolescu et al (1976); Arion et al (1986, 1987); Bercia (1975); Balintoni et al (1986); Dimitrescu et al (1971); Gheuca (1988); Gheuca, Dinică (1986); Hann, Balintoni (1988); Hann et al (1987); Hărtopan et al (1984); Iancu, Mărmăruțu (1987); Iancu et al (1987); Savu (1970); Savu et al (1978); Stelea (1992); Tatu, Robu (1987).



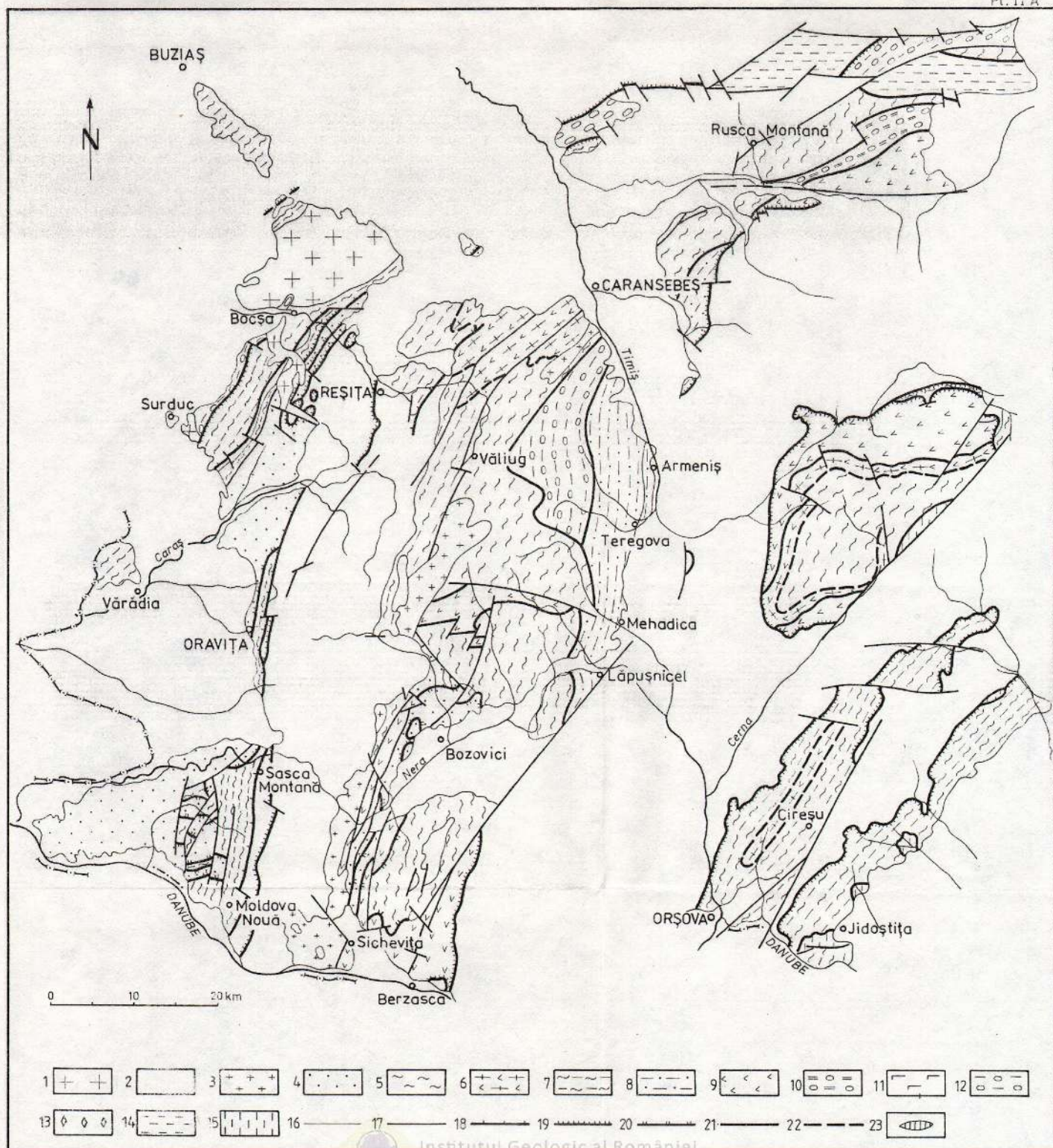


V. IANCU, M. MĂRUNȚIU
 VARISCAN AND PRE-VARISCAN SHEAR ZONES IN THE GETIC-SUPRAGETIC REALM
 (SOUTH CARPATHIANS)

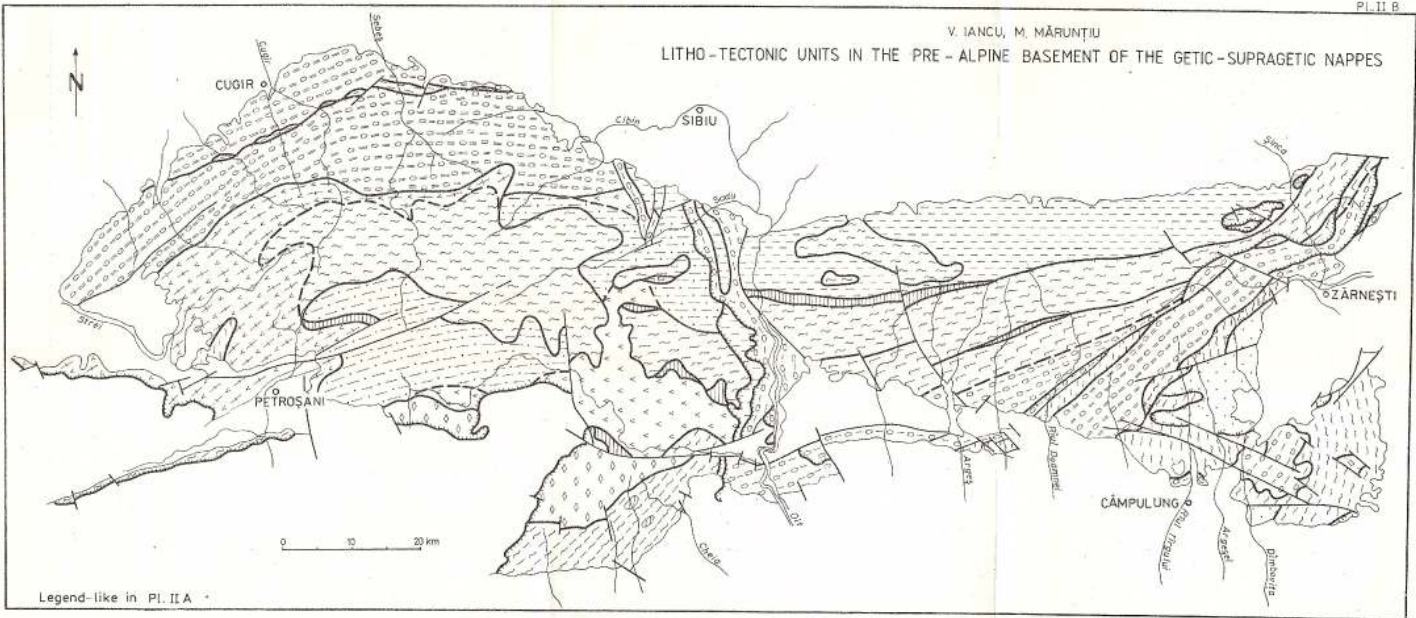


Legend-like in Pl. I.A

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V. IANCU, M. MĂRUNȚIU
LITHO-TECTONIC UNITS IN THE PRE-ALPINE BASEMENT OF THE GETIC-SUPRAGETIC NAPPES



Legend-like in Pl. II A *



Institutul Geologic al României

VARISCAN EVENTS IN THE BASEMENT OF THE DANUBIAN NAPPES (SOUTH CARPATHIANS)

by

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Geological Institute of Romania

1. Introduction

Any considerations about the effects of Variscan sedimentary, igneous, metamorphic or tectonic events in the Danubian Nappes of the South Carpathians depend on the age ascribed to the formations involved, having in mind that important rock volumes ascribed to Paleozoic have been recently reconsidered as Mesozoic and/or Precambrian (Berza et al., 1988 a, 1988 b). For a better understanding of the geological background we will present here the main structural and lithostratigraphic units composing the Danubian Alpine nappe stack.

The roof thrust above the Danubian is generally located at the sole of the Getic and/or Severin Nappes (Codarcea, 1940). The floor thrust may be the plane along which the Carpathian fold and thrust belt is thrust onto the Moesian Platform (Stefănescu et al., 1988), but considered as active in Laramian tectonics. Between these two tectonic boundaries, two sets of thrusts were defined as Upper and Lower Danubian Nappes by Berza et al. (1983) and are now described as Upper and Lower Danubian Duplexes by Seghedi, Berza (1994). Because the contents and names of the Danubian Nappes are highly controversial (see Kräutner et al., 1988, Băliptoni et al., 1989, Inacu et al., 1990; Năstăseanu, 1994, Stănoiu et al., 1992) we shall present them as follows: (figs. 1,2).

The Upper Danubian Nappes can be described, from top to bottom, as:

- Arjana Nappe, only Mesozoic cover;
- Urdele-Măru-Svinecea Nappe, with pre-Alpine basement and Mesozoic cover;
- Poiana Mărului-Cornereva Nappe, with pre-Alpine basement and Mesozoic cover;
- Godenele-Scorila Nappe, with pre-Alpine basement and Mesozoic cover.

A description of these units is presented by Berza et al (1994, this volume).

The pre-Alpine basement of Poiana Mărului-Cornereva (Presacina) Nappe involves both Precambrian medium-grade metamorphic formations intruded by granitoids and Paleozoic very low-grade formations, while the other two basement nappes are constituted only from Precambrian metamorphics.

The Lower Danubian Nappes are, again from top to bottom:

- Lainici Nappe, with pre-Alpine basement and Mesozoic cover;
- Schela Nappe, with pre-Alpine basement and Mesozoic cover.

The pre-Alpine basement of Lainici Nappe contains Precambrian medium-grade metamorphics intruded by granitoids and very low-grade Paleozoic formations; the basement of the Schela Nappe includes only



Precambrian medium-grade metamorphics intruded by granitoids.

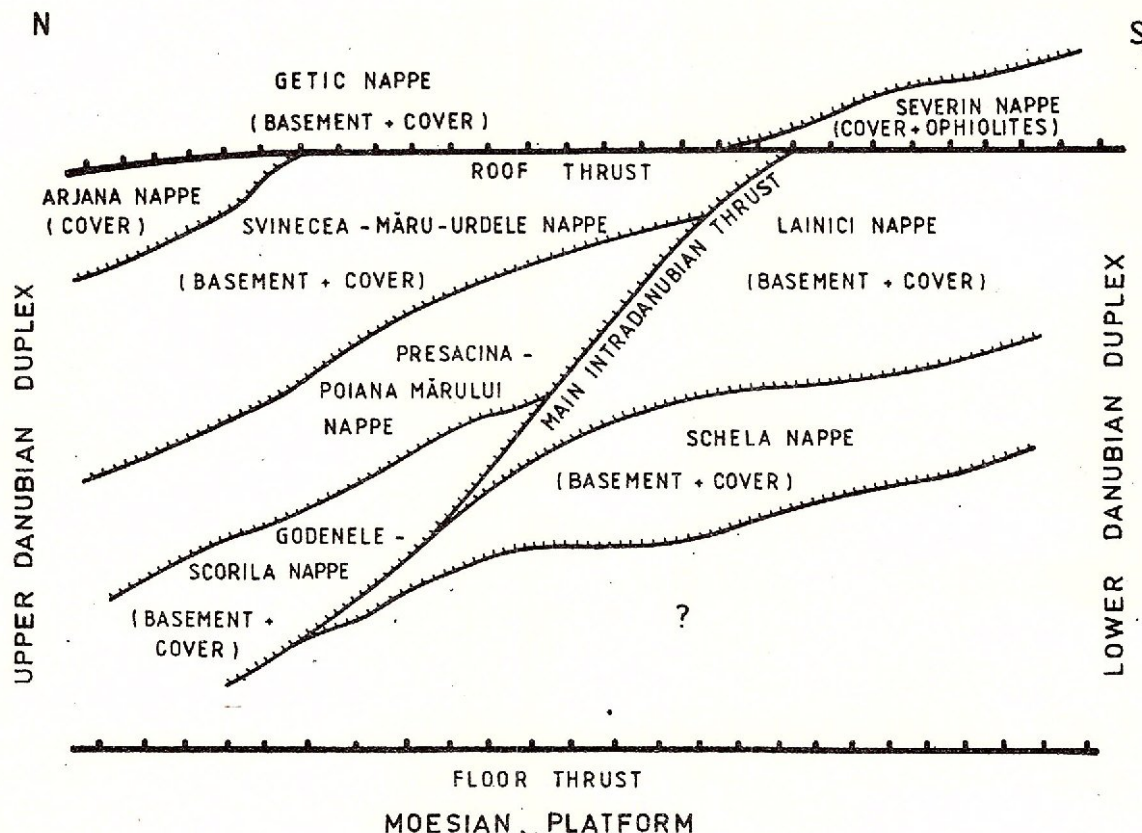


Fig. 1

2. Stratigraphy

In the last decades, metamorphic formations from the pre-Alpine basements of the Danubian Nappes have been described under many local names; until the sixties they were ascribed to the Paleozoic or Precambrian, after first K/Ar ages of over 550 Ma. appeared. However, recent synthesis (Berza, Seghedi, 1983; Iancu, 1983; Kräutner et al., 1988, Balintoni et al., 1989) tried to simplify the picture, extending the names of Drăgșan Group for all the prevailing leptyno-amphibolitic sequences and that of Lainici-Păiuș Group for most metaterrigenous sequences. Poiana Mraconia Formation and Rof Formation represent crystalline schists difficult to ascribe to one of the main groups. All these sequences are polymetamorphic, but still preserve mineralogic or structural evidence for an early medium-grade metamorphism.

Lainici-Păiuș Group has a lower Carbonate-Graphitic Formation (crystalline limestones and dolomites paraamphibolites, garnet bearing micagneisses \pm graphite \pm sillimanite \pm cordierite \pm andalusite (Savu, 1972; Berza, 1978) and an upper Quartzitic Formation (quartzites, biotite gneisses amphibolites). Distinct features include pervasive and regional migmatization by leucogranite injections and the general distribution of high-T low-P associations. Large granitoid bodies intruded the Lainici-Păiuș Group around 600 m.y. age (Grünenfelder et al., 1983), demonstrating the Precambrian age of this group.

Drăgșan Group includes a lower Augengneiss Formation, an Amphibolitic Formation (amphibolites, leptynites, serpentinites) and an upper Micagneiss Formation (muscovite, biotite, garnet gneisses \pm kyanite \pm staurolite) (Iancu, 1974; Berza, Seghedi, 1983). The most obvious mineral associations are: an early kyanite-staurolite-almandine assemblage and a locally subsequent chlorite-epidote-tremolite-albite one; in places coronitic garnets occur in amphibolites suggesting retrogressed eclogites (Berza, 1978). Artheritic migmatization accompanied by sillimanite and cordierite blastesis is restricted to areas surrounding granitoid batholiths intruding the Drăgșan Group rocks.



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STRUCTURAL SKETCH OF THE DANUBIAN NAPPE

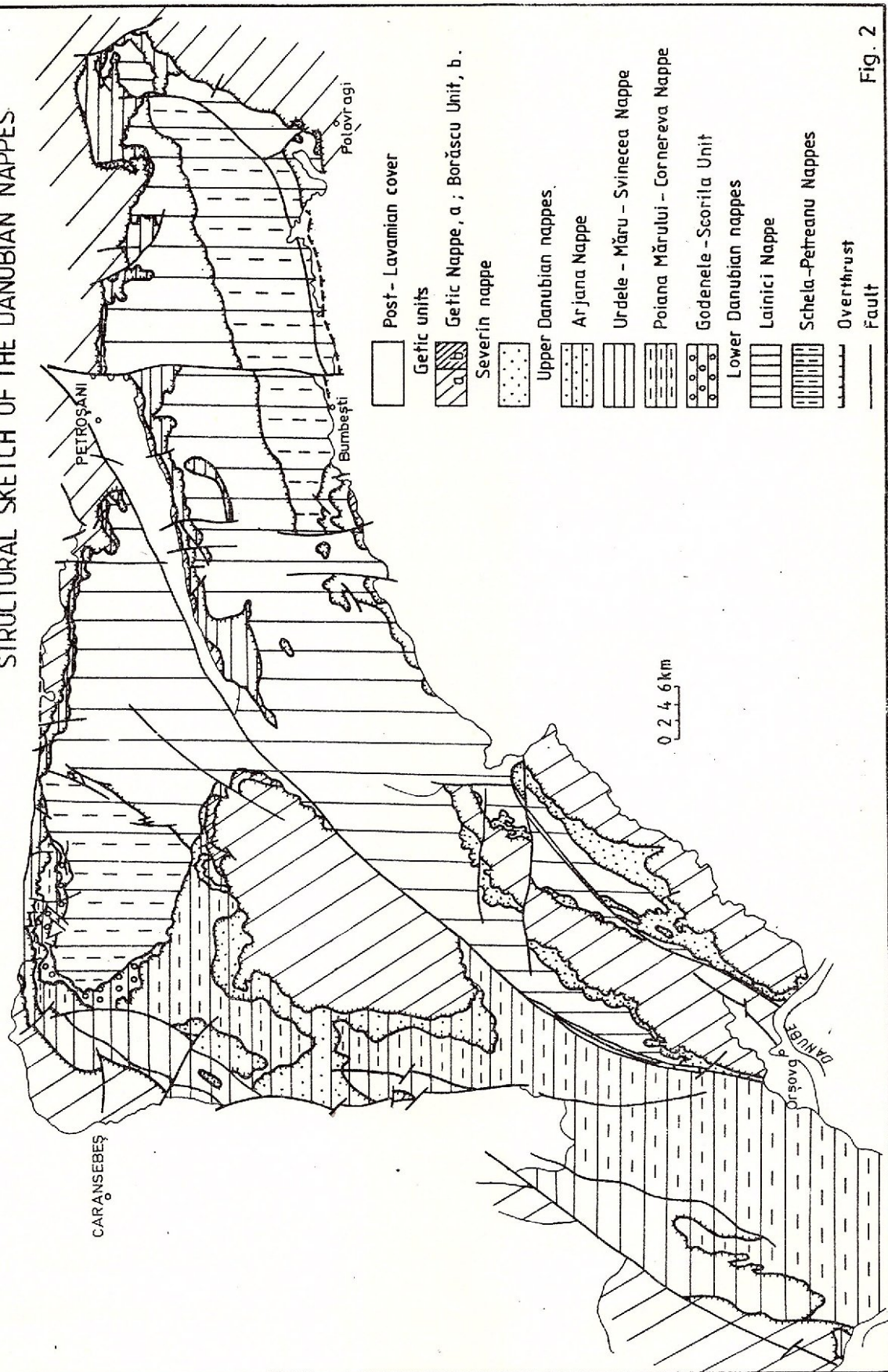


Fig. 2

K-Ar or Ar-Ar ages older than 300–400 m.y. were not recorded, but Rb-Sr and Sm-Ne ratios suggest ages of 600 m.y. for the Drăgșan protoliths (J.P. Liégeois, personal communication, 1994).

Poiana Mraconia Formation is constituted of amphibole gneisses, amphibolites, micagneisses \pm kyanite, quartzites and migmatites (Bercia, Bercia, 1980). The lithologic association and the reduced retrogression, suggest for this sequence resemblance with metamorphics from the basement of the Getic Nappe.

Rof Formation consists of leptyno-amphibolites and micagneisses with centimetric garnets and a pin-nitised aluminosilicate, but a retrogression in biotite zone is evident (Dimitrescu, 1986).

For both formations a Precambrian age is not yet proved, but is very likely, regarding regional correlations (Kräutner et al., 1988). Unconformably overlying the polymetamorphic (Precambrian) basement, very low-grade Paleozoic formations are recorded in the pre-Alpine basements of the Cornereva-Poiana Mărului (Upper Danubian) Nappe and of the Lainici (Lower Danubian) Nappe (Plate). The identification of the Paleozoic formations is not easy, as they can resemble the underlying retrogressed Precambrian formations and the very low-grade metamorphic Mesozoic formations of the Alpine cover. The best example in this sense is the "Tulișa Series", separated by Pavelescu (1953), who included here a Precambrian mylonitized Lainici-Păiuș type basement, a real low-grade Paleozoic sequence (Valea de Brazi Formation) and Mesozoic very low-grade formations (Oslea Limestone, Schela Formation). Even a special paper devoted to this topic (Kräutner et al., 1981) is still carrying such confusions, induced by the strong Alpine dynamic overprint related to nappe emplacement. However, fossil fauna and flora identified in several formations indicate Upper Ordovician to Lower Carboniferous ages (Codarcea et al., 1961, Stănoiu, 1972, 1976, 1982, Iancu, Visarion, 1988).

In the Upper Danubian Nappes Paleozoic sequences are: Râul Alb, Brustur, Drencova, Râul Rece, Ideg, Sevastru formations.

Râul Alb Formation (Năstăseanu, 1975) was defined in western Țarcu Mts as a mainly volcano-sedimentary sequence (phyllites, conglomerates, sandstones, basic tuffs) with palynological associations pointing to an Ordovician-Silurian age (Visarion, in Kräutner et al., 1981). Nijudimu Formation (Iancu, Visarion, 1988) is an equivalent sequence from south-eastern Țarcu Mts.

Brustur Formation (Iancu et al., 1990) represents probably a synchronous sequence from northern Țarcu Mountains, but with large development of thick conglomerate facies (known as "Baicu conglomerates" of Gherasi, 1937). Conglomerates grade upwards to or are interbedded with hornblende rich sandstones and black slates; acid concordant layers (flows or sills) of rhyolites recorded. Conglomerate clast petrography and sandstone mineralogy indicate that the main source area for this formation is the Iuți ophiolite complex from the South Banat (considered Caledonian by Mărunțiu, 1987) and Precambrian Drăgșan rocks (Morariu, 1976, Iancu et al., 1990). The age of the Brustur Formation is still controversial, being ascribed to the Ordovician-Silurian, based on achritarchs and chitinozoans content (A. Visarion 1974, unpublished data).

Drencova Formation from Danube Valley includes a lower conglomerate member and an upper pelitic member with basic tuffs interlayers topped by limestones; palynological association and *Plectogyra* macrofauna point to a Devonian to Lower Carboniferous age (Kräutner et al., 1981).

Râul Rece Formation from Ideg (=Râul Rece) Valley in southern Țarcu Mountains is a volcano-sedimentary sequence of sandstones, conglomerates and slates with interlayered basic tuffs and flows and with subordinate cryonidal limestones; Devonian macroflora is preserved in slates (Năstăseanu, 1979).

The Râul Rece Formation is overlain by the Ideg Limestones known for a rich Tournaisian macrofauna (Codarcea et al., 1961).

Sevastru Formation overlying the Ideg Limestones consists of slates interlayered with sandstones and basic volcanics ascribed to Visca (Kräutner et al., 1981). Trahyandesitic dykes and sills, as well as porphyritic sienites are recorded in Râul Rece and Sevastru formations (Russo-Săndulescu, 1985, unpublished data).

Structures and mineralogical associations in the described Paleozoic formations indicate a very low-grade metamorphism, coexisting in places with well preserved sedimentary and diagenetic features. Structural features are dominated by meso to macrofolds (generally vertical, open folds) with associated S_1 cleavages and foliations, pre-Permian in age. The amplitude and style of folds are variable; S_1 cleavages and foliations are partly penetrative (e.g. Brustur, Ideg formations) or well expressed (e.g. Nijudimu Formation) (Iancu et al., 1990; Iancu, Visarion, 1988). Metamorphic minerals partly grow along cleavages and foliations (axial-planar or fanning in fold hinges) or are associated with preexisting sedimentary minerals. Metamorphic blastesis is in the range of anchizone to the low greenschist facies represented by: carbonates, chlorite, sericite, stilpnomelane albite, epidote-clinozoisite.



However, the frequent preservation of sedimentary structures, (cross bedding, graded bedding) and textures features of igneous rocks, as well as of macroflora and macrofauna, are relevant for the uneven but mostly incipient character of the metamorphism. Even if this metamorphism was traditionally ascribed to Variscan events locally overprinted by Alpine metamorphism (Codarcea et al., 1981; Kräutner et al., 1981; Iancu et al., 1990), the structural and mineralogical resemblance to the Upper Paleozoic (Upper Carboniferous coal bearing terrigenous deposits and Permian volcano-sedimentary sequences) and Mesozoic formations determined some authors (Berza, Seghedi, 1983) to have doubts about the existence of widespread Variscan metamorphism.

In the Lower Danubian Nappes, Paleozoic sequences include: Valea Izvorului, Coarnele, Poiana Mică and Valea de Brazi formations.

Valea Izvorului Formation (Stănoiu, 1972) consists of a lower quartzitic member (white quartzitic sandstones) and an upper pelitic member, rich in Upper Ordovician-Lower Silurian macrofauna (coelenterates, bryozoans, brachiopods, crinoids and trilobites, cf. Stănoiu, 1972) and palynomorphs. It outcrops in Mehedinți Mts, overlying Lainici-Păiuș Group rocks and underlying Jurassic-Cretaceous formations.

Coarnele Formation (Stănoiu, 1976) was correlated to the Valea Izvorului Formation because of its quartzitic mineralogy and palynological content. (Kräutner et al., 1981). It crops out in northern Vâlcan Mts, overlying Drăgășan Group rocks and underlying Jurassic formations.

Poiana Mică Formation (Devonian or Permian) underlying a Liassic sequence in Culmea Cernei Mountains, is a thin conglomeratic to psammitic sequence resting unconformably over medium-grade rocks of Lainici-Păiuș Group. Very low-grade blastesis (chlorite, sericite, albite) may be induced by Alpine events, having in view similar physical conditions suggested by the directly overlying pyrophyllite-bearing Liassic phyllites.

Valea de Brazi Formation is exposed on the southern slope of the Eastern Retezat Mts, overlying Lainici-Păiuș Group rocks and (tectonically) underlying Drăgășan Group rocks. Stănoiu (1982) described initially this sequence under the name "Tusu Formation" and quoted an Upper Devonian macroflora; later (unpublished paper with A. Lejal Nicol) he introduced the name "Valea de Brazi Formation" for this sequence and reconsidered the age as Westphalian. As Tusu Valley is located in the Vâlcan Mts and "Tusu Formation" is now considered an Alpine mylonite zone on Lainici-Păiuș Group protoliths (Berza et al., 1988 b), we retain the name Valea de Brazi from the Retezat Mts and favour the Devonian age. The sequence consists of conglomerates, sandstones and shales, with important concordant layers (flows or sills) of rhyolites.

Just as for Paleozoic formations outcropping in Upper Danubian Nappes, the structural and mineralogical characters of the Paleozoic formations from Lainici (Lower Danubian) Nappe are consistent with shallow level of deformations and metamorphism of these sequences.

If stratigraphic unconformities between pre-Permian formations and overlying Permian or Liassic formations are well exposed, a distinction between a possible Variscan low-grade metamorphism and Alpine metamorphism is really difficult in Lainici Nappe.

Some observations can be underlined:

- The great difficulty to separate the greenschist facies rocks of the Lower Paleozoic, especially Coarnele Formation, in respect with retrogressed rocks of Drăgășan Group because of their imbrication by folding and dynamic reactivation in Variscan tectogenetic events (shear zones);
- the prevalence of dynamic Paleozoic metamorphism implying Paleozoic protoliths (sedimentary or magmatic rock assemblages) and Precambrian basement; common foliations (S_1 in Paleozoic rocks and S_{2-3} in basement) connect with low-grade blastesis: chlorite, albite, sericite, actinolite, epidote-chinozoit, tourmaline and recrystallisation of quartz and carbonates;
- contrasting mineralogy of Paleozoic terrigenous sequences in respect with some terrigenous Mesozoic (especially Liassic) rocks, induced by different conditions of sedimentation and whole rock compositions; as a result, some metamorphic phases (e.g. pyrophyllite, chloritoid, meta-antracite, pre-graphite) in Liassic rocks, are good markers for low-grade Alpine metamorphism (Iancu et al, 1984), whose history is essentially different in respect with final Variscan tectonic evolution.

In conclusion, the existence of Paleozoic (Ordovician-Lower Carboniferous) formations in the basement of the Danubian Alpine Nappes from the South Carpathians is well documented by paleontological evidence. In our opinion their area of development is small compared to the underlying Precambrian polymetamorphic formations and with the overlying Upper Carboniferous-Permian-Mesozoic formations (see Pl. I). The distinction between Lower-Middle Paleozoic (Ordovician-Lower Carboniferous) formations and Upper Paleozoic (Upper Carboniferous-Permian) formations was introduced by Codarcea et al. (1961) and Kräutner



et al. (1981) as separating two independent sedimentary and tectonic cycles, the first representing Variscan metamorphosed sequences and the second the post-metamorphic Variscan molasse. This distinction might be correlated with orogenic Variscan cycle evolution followed by collapse.

But even if there is no obvious metamorphic discordance between formations from Ordovician to Mesozoic (for example in Mehedinți Mountains, from Ordovician Valea Izvorului to Jurassic and Cretaceous formations) a tectonic phase was active in Carboniferous. This is best demonstrated in Eastern Retezat Mts, where the overthrust of Precambrian Drăgșan Group onto Upper Devonian Valea de Brazi Formation and its Precambrian Lainici-Păiuș Group basement is sealed by Permian and Mesozoic deposits. In this way the mentioned distinction between Lower-Middle Paleozoic formations and Upper Paleozoic (molassic) Formations is valid.

3. Igneous activity

As it was already mentioned on several occasions, volcanic products are found in several Paleozoic formations from basements of both Upper and Lower Danubian Nappes.

In Poiana Mărului-Cornereva (Upper Danubian Nappe) basic volcanics are found in the Silurian Râul Alb and Nijudimu formations, in the Devonian Drencova and Râul Rece formations and in the Lower Carboniferous Sevastru formation (Năstăseanu, 1979); acid volcanics are recorded in Brustur (Silurian) Formation as concordant lenses of rhyolites (Iancu et al., 1990). Alkaline rocks (trahandesites, porphyric sienites) were described by Russo-Săndulescu (1985, unpublished data) in the Râul Rece and Sevastru Formations. The molassic Permian sequence of the Poiana Mărului-Cornereva Nappe includes important bodies (necks, dykes, sills, lava flows) of rhyolites (Stănoiu, Stan, 1986) and basaltic pyroclastics (Kräutner et al, 1981).

The Permian deposits of the Lainici Nappe are devoid of volcanics.

A special issue concerns the large mafic-ultramafic complex from the Danube Valley (Iuți and Plavișevîța gabbros and Țișovîța peridotites (Codarcea et al., 1961). Recent field, petrological and geochemical studies of Mărunțiu (1987) documented pseudostratified character of the ophiolitic complex from Țișovîța-Iuți zone and its allochthonous character. Savu, *** Strusievicz (198) underlain the existence of remnants of Lower Paleozoic oceanic crust in the Danubian basement, reworked in Upper Carboniferous and Permian conglomerates.

The presence of gabbros and serpentinites as pebbles in Baicu conglomerates of Silurian (?) Brustur Formation (Iancu et al., 1990), indicates even a pre-Silurian age.

In the absence of geochronological evidence, these ophiolites might be either Lower Paleozoic or Precambrian. Metamorphosed (in amphibolite facies conditions) mafic-ultramafic bodies up to 1 km wide frequently occur in the Precambrian Drăgșan Group sequences in Almaj, Țarcu, Vâlcău and Parâng Mountains.

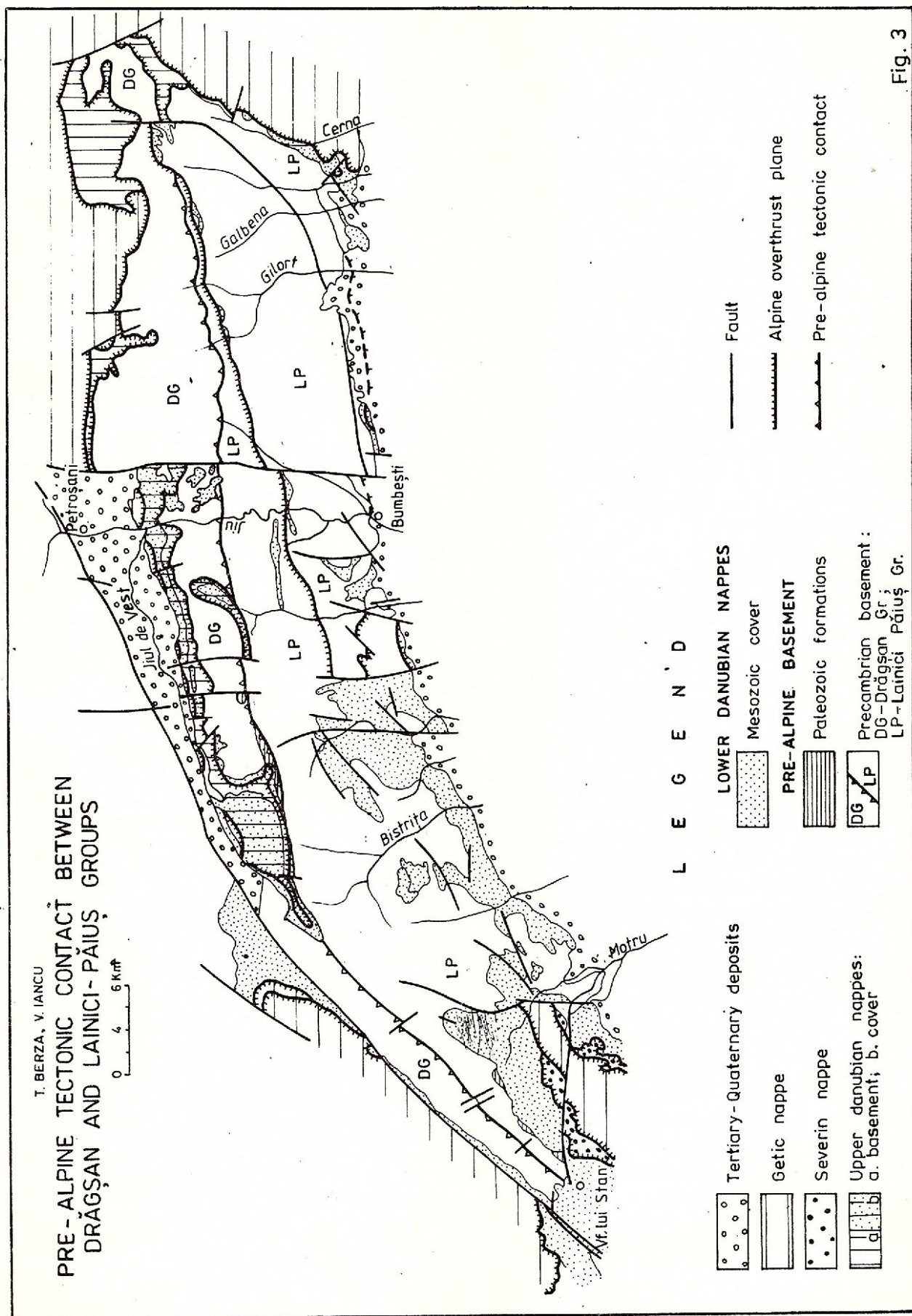
Plavișevîța-Iuți ophiolitic complex is exposed, between tectonic contacts, along 30 km in N-S direction and its counterpart across Cerna-Jiu fault system in NE Serbia is cropping out on 25 km at Deli Iovan.

The Sfârdin, Cherbelezu and Ogradena granitoid plutons from the Almăj Mountains were considered as Paleozoic (Variscan), based on mineral Rb/Sr and K/Ar ages (393 Ma. – Cherbelezu; 372 Ma. – Sfârdin (Stan et al., 1985). Additional evidence is the assumption that they probably intrude the ophiolitic mafic-ultramafic complex of (supposed) Lower Paleozoic age and crosscut Corbu blastomylonitic zone (Mărunțiu, Seghedi, 1983). Not rejecting this possibility, it is however not yet radiometrically proved that they do not belong to the Upper Precambrian family of granitoid plutons exposed in other Danubian areas. In this case, a revision of interrelationships between Iuți ophiolitic complex, Precambrian medium-grade rocks of the continental blocks and granitoid intrusions will be necessary.

4. Tectonics

There are some structural elements in the Danubian basements which can be assigned to Variscan events. Because of the uncertainties concerning the formations involved, some of these elements might be older, but for well dated sequences the Middle Carboniferous interval (late Variscan events) seems to be critical for important compressive tectonics including folding and shearing of Precambrian basement and Paleozoic sequences (Plate, Fig. 3).





A Caledonian spreading represented by the Iuți Ophiolite complex was claimed by Mărunțiu (1987), Savu and Strusievicz (1987) and Iancu et al. (1990). Pebbles from this body are common in Upper Carboniferous and Permian Conglomerates (Streckeisen, 1934; Codarcea, 1937) and in Baicu Conglomerate, of possible Silurian age (Iancu et al., 1990). In the absence of radiometric ages, older (Precambrian) timing can also be envisaged.

A blastomylonitic shear zone, up to 1 km wide, borders eastwards the Iuți complex. Separated as "Corbu phyllitic zone" by Codarcea (1940), this belt was reconsidered as retrogressed medium-grade rocks by Gunnesch, Gunnesch (1978) and as mylonitic belt by Mărunțiu, Seghedi (1983). The Corbu "Zone" includes polymetamorphic (Precambrian ?) rocks, mafic-ultramafic rocks and carbonatic and silicic rocks seemingly low-grade metamorphosed (Lower Paleozoic ?). A parallel belt ("Vodna phyllites" – Codarcea, 1940), up to 500 m wide, occurs 3 km eastwards, in a Lainici-Păiuș type (locally named "Neamțu Series") basement and was interpreted by Dinică (1987) as a fault bounded klippen. Surely pre-Alpine, the biotite-garnet zone mylonitisation (Seghedi, in Berza et al., 1984) is again loosely dated at present.

Several regional thrusts in the basements of Upper and Lower Danubian Nappes are assigned to Variscan compressions.

In the Urdele-Măru-Svinecea Nappe, a pre-Alpine tectonic contact between Drăgșan and Lainici-Păiuș rock sequences was supposed in Parâng (Berza et al., 1986) and Tarcu Mts (Săndulescu, 1984). The latter was considered Alpine by Berza et al. (1983) and Kräutner et al. (1988), but no Mesozoic formations are involved and the thrust might be Variscan. In Parâng Mts, a Mesozoic cover seals the boundary and the Variscan age is clear. In both areas, chlorite zone mylonitization point to low-temperature thrusting.

In the Poiana Mărului-Cornereva Nappe, the pre-Permian thrust of the Poiana Mraconia Formation metamorphic rocks on to the Iuți basic-ultrabasic complex was established by Streckeisen (1934) and Codarcea (1940). The contact is sealed by Upper Carboniferous coal-bearing deposits and Permian red-beds, but there are no geological constraints for the oldest possible age of this thrust. Cataclastic to ductile deformations with insignificant recrystallisation point to a shallow level of tectonics.

In the basement of the Lainici Nappe from the Retezat Mts, a clear Variscan thrust emplaces Drăgșan Group rocks onto Lainici-Păiuș Group rocks and its (Upper Devonian ?) cover, the Valea de Brazi Formation (Krautner et al., 1988). This thrust is sealed by Permian and Mesozoic rocks (Berza et al., 1988 a). A similar geometry is followed in the Vâlcan-Parâng Mts (Manolescu, 1937; Berza, 1978; Iancu et al., 1984) along 150 km (fig. 2). Here a stack of imbricates involving Drăgșan Group amphibolites and quartzites of the (Ordovician ?) Coarnele Formation is thrust directly onto Lainici-Păiuș Group rocks. This thrust is sealed in several places by Lower Jurassic to Cretaceous sequences belonging to the Mesozoic cover of the Lainici Nappe (Vârful lui Stan, Futeciu, Pietrele Albe, Latorița (Iancu, 1984; Berza et al., 1986, 1989).

In both outcrop areas (Retezat and Parâng-Vâlcan Mts) the structural and mineralogic overprint show areal development (tens to hundreds of meters in the hanging wall and hundreds to thousands meters in the footwall). A mylonitic foliation parallel with the thrust and outlined by neoformation of chlorite, white mica or quartz ribbons curves around K-feldspar, hornblende or plagioclase porphyroclasts. Stilpnomelane frequently occurs as needles in K-feldspars and albite-epidote aggregate replaces plagioclase feldspars; the newly formed associations suggest lower greenschist facies (chlorite Zone) conditions for this thrust.

In the Schela-Petreanu Nappe, a special horse-shoe structure (bounded eastwards by an Alpine thrust) was interpreted by Codarcea and Gherasi (1945), Berza et al., (1983) and Kräutner et al. (1988) as a pre-Alpine nappe-stack, while Dimitrescu (1986) considered it a recumbent fold. The geometry of the structure involves, from top to bottom: Petreanu Augengneiss, a thin band of biotite-garnet schists, Furcătura Gneiss and Rof Formation. All sequences show widespread newly formed biotite paralleling lithologic contacts and euhedral garnet is typical of schists interlayered between the gneisses. It is possible that Variscan mylonitisation in garnet-biotite zones is connected to the thrusts at the sole of Petreanu and Furcătura Gneisses, respectively, but the Paleozoic or Precambrian age of this shearing is difficult to establish in the absence of geochronological evidence.

5. Conclusions

The pre-Alpine basements of the Danubian Nappes show several characteristics concerning the sedimentary and tectonic environments for the Ordovician-Lower Carboniferous interval, respectively the traditional Variscan orogenic cycle.



The Iuți basic-ultrabasic complex, of probable Lower Paleozoic age, represents fragments of an old oceanic plate, testifying for an early period of oceanic spreading (Mărunțiu, 1987, Savu, Strusievicz, 1987). If the eastwards bordering Corbu Zone preserves associated and contemporary carbonatic and silicic deposits, the Iuți-Corbu sequences suggest the stratigraphy of oceanic crust.

The sedimentary or volcano-sedimentary formations dated as Ordovician to Lower Carboniferous present the following constitution:

- Ordovician-Silurian: detrital deposits and basic-intermediate-acid volcanics in Upper Danubian (Râul Rece, Nijudimu and Brüstur formations); detrital deposits in Lower Danubian (Valea Izvorului and Coarnele formations)

- Devonian: detrital deposits and basic volcanics in Upper Danubian (Drencova and Râul Rece formations); detrital deposits and acid volcanics in Lower Danubian (Valea de Brazi Formation);

- Lower Carboniferous: carbonatic and pelitic deposits in Upper Danubian (Ideg and Sevastru formations), not known in Lower Danubian.

The tectonic environment for these formations can be alternatively considered as active or passive continental margins, depending on the temporal or spatial position of the considered sequence.

All Ordovician to Lower Carboniferous formations preserve paleontological and/or sedimentological records, despite pre-Upper Carboniferous folding and very low-grade metamorphism. This makes a contrast with the underlying medium grade, polymetamorphic Precambrian basement, but differences with the overlying Upper Paleozoic-Mesozoic formations are less evident in metamorphic grade at a first approximation.

Upper Carboniferous coal-bearing clastites and Permian red-beds with acid intrusions in Upper Danubian Nappes show molassic affinities and are post-tectonic regarding Variscan folding and thrusting. In Lower Danubian Nappes Permian red-beds are post-tectonic in respect with the thrust emplacing Dragsan Group onto Lainici Group with Valea de Brazi Formation. In all Danubian Nappes deformation and (very low-grade) metamorphism of Upper Carboniferous-Permian sequences is similar with that of overlying Mesozoic formations, being traditionally linked by geologists with the Mesozoic cover.

Compared to the upper (Getic-Supragetic) nappes from the South Carpathians, the Danubian basement shows Ordovician-Lower Carboniferous formations with better preserved primary features in ophiolitic and sedimentary or volcano-sedimentary sequences, different deformational features and lower-grade metamorphism. This points to a different position of the Danubian versus Getic-Supragetic basement in respect with the Variscan front.

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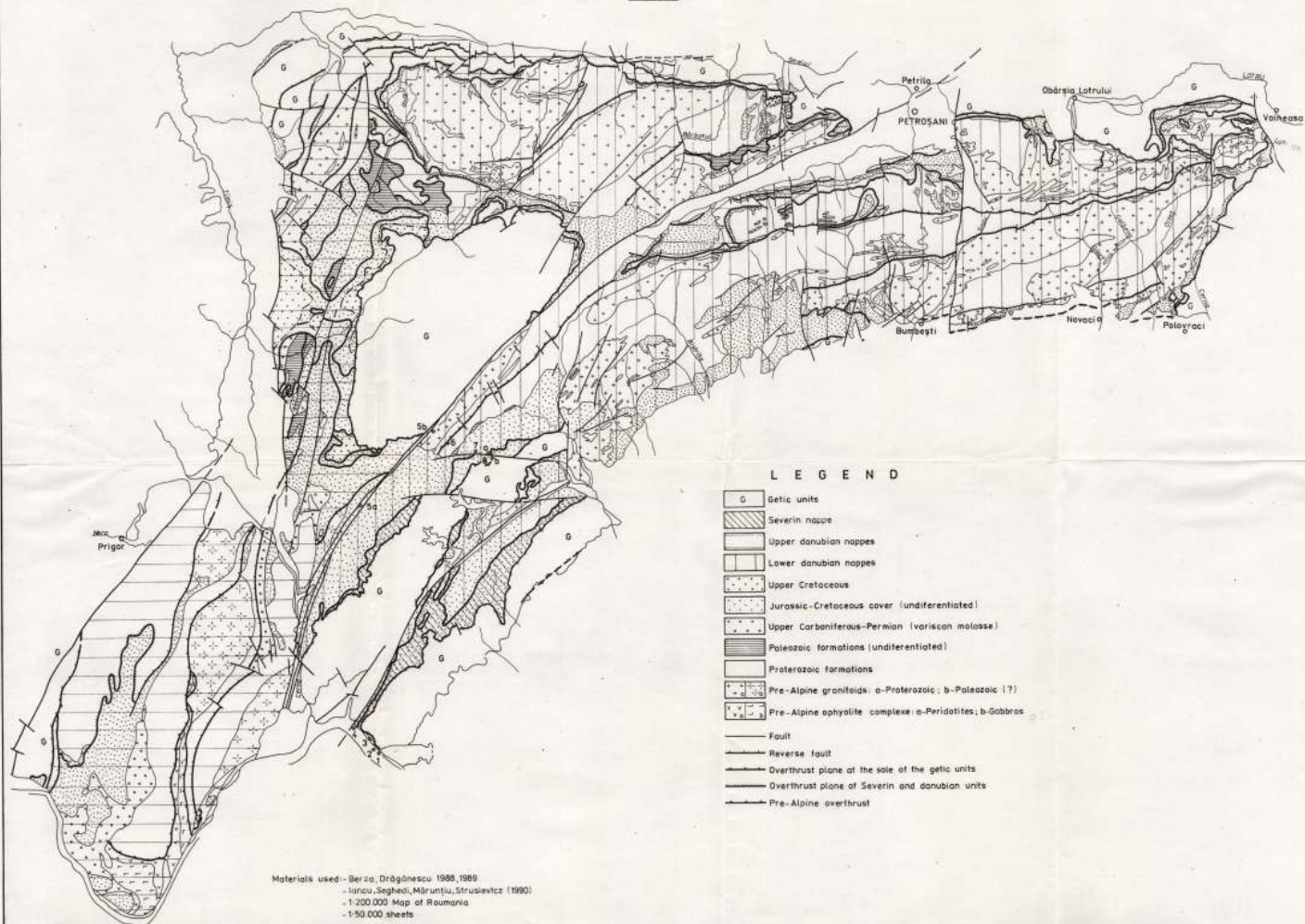
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T. BERZA, V. IANCU, A. SEGHEDI, A. DRĂGĂNESCU
STRUCTURAL MAP OF THE DANUBIAN WINDOW

0 1 2 km



EXCURSION TO SOUTH CARPATHIANS, APUSENI MOUNTAINS AND TRANSYLVANIAN BASIN : DESCRIPTION OF STOPS

T. Berza, V. Iancu, A. Seghedi, I. Nicolae, I. Balintoni, D. Ciulavu, G. Bertotti

First day

Introduction

The route follows left bank of the Danube, between Gura Văii and Orșova: continues upstream the Cerna river until the junction with the Arșasca creek; upstream the Arșasca creek-Obârșia Cloșani; downstream the Brebina river.

Between Gura Văii and the Bahna river, the road along the Danube exposes, from east to west:

- Tertiary sedimentary deposits of the Dacic Basin;
- Porțile de Fier and Bahna outliers of the Getic Nappe;
- The Severin Nappe;
- The (Lower Danubian) Lainici Nappe (Pl. I)

Cropping out in post-nappe synforms, the Severin and Getic Nappes have been overthrust onto the (Lower Danubian) Lainici Nappe during the Laramian tectogenesis (intra-Senonian). The youngest members of the Danubian cover are ascribed to the Campanian-Maastrichtian (Stănoiu, *vide* Iancu et al., 1986, unpublished 1:50.000 sheet).

The superposed Alpine structural units show contrasting lithologies and internal structures both within the pre-Alpine basement and the Mesozoic cover.

Mesozoic covers of the three super posed units reflect distinct depositional environments and show different deformational-metamorphic history.

The sedimentary cover of the Getic Nappe (Triassic; Lower Jurassic-Aptian and Albion-Upper Cretaceous) reflects the effects of mid-Cretaceous and Laramian events. These are documented by changes in sedimentary régime and by stratigraphic gaps, post-tectonic covers unconformably overlying the units successively stacked (post-Austrian cover).

Sedimentary formations preserved in the Severin Nappe include Upper Jurassic-Neocomian distal turbidites and strongly dismembered oceanic crust relics (ophiolites and pelagic deposits). The root zone of this nappe is partly preserved under the Cozla thrust sheet (Banat) of the Getic Nappe. The entire rock assemblage shows polyphase deformation and nongeneralised very low-grade metamorphism, followed by cataclastic-ductile shearing.

The Mesozoic cover of the Danubian units, showing facies changes in various units, was deposited during the Lower Jurassic-Upper Cretaceous time-span. Both Alpine deformation and very low to low-grade metamorphism, as well as pre-Alpine basement reactivation, are generalised phenomena, with higher intensities compared to the upper tectonic units.

The effects of the Laramian shearing at the sole of the upper unit (Getic Nappe) suggest shallow structural level, within the brittle-ductile transition, not accompanied by significant recrystallisation (or with slight recrystallisation of quartz and carbonates). The lower (Severin and Danubian) nappes show low-temperature mylonitisation of covers (sediments and ophiolites) at base of their pre-Alpine basement, with significant recrystallisation and neomineralisation.

A set of regional post-nappe strike-slip faults affected the Laramian nappe stack, showing a consistent NE-SW trend, parallel to the orogenic belt; the most important of them belong to the Cerna-Jiu and Bahna systems (Berza, Drăgănescu, 1988). These are followed by/or are conjugated with E-W dextral strike-slip faults (e.g. Obârșia Cloșani fault).

During the Miocene (pre and intra-Sarmatian) the last compressional events thrust the entire Laramian nappe stack onto the Moesian Platform along north-westwards dipping reverse faults; effects of these movements are found along the border of the Porțile de Fier (Iron Gates) outlier (Iancu et al., 1986) and in boreholes



(Ștefănescu et al., 1988).

Stop no. 1 (short stop): The post-tectonic cover of the Laramian nappes.

V. Iancu

Location: Orșova - Turnu Severin highway, 1 km west of Gura Văii

From Turnu Severin until Jidoștița valley and Gura Văii, Tertiary deposits of the Dacic Basin, unconformably overlying Getic Nappe basement rocks (the Iron Gates outlier) are crossed (Pl. II).

In the Schela Cladovei quarry (4 km East of Gura Văii), Neogene (Pontian) continental deposits consist of cross-bedded gravels and sandstones, overlain by horizontal Quaternary terrace deposits. Pontian deposits transgressively overlay either Middle-Miocene deposits, or directly the Mesozoic cover or the metamorphic basement of the Getic Nappe (Marinescu, in Iancu et al., 1986).

At Gura Văii, parallel bedded gravels and sands (Kossovian), showing a slight (10^0 – 15^0) eastward dip are exposed in a 50 m high escarpment. Similar deposits cropping out on the right bank of the Jidoștița Valley, overlay tuffaceous marly clays, sandy clays and fossiliferous sands and are overlain by unconformable Sarmatian deposits.

On the left bank of the Jidoștița Valley, the flatlying Miocene deposits overlay patches of the Mesozoic cover of the Getic Nappe, consisting of red ammonitic limestones (Oxfordian), white brecciated limestones (Tithonian-Kimmeridgian) and Upper Cretaceous conglomerates and sandstones, all devoid of cleavage or foliation (Codarcea, 1940).

Stop no. 2: Pre-Alpine polymetamorphic basement of the Getic Nappe.

V. Iancu

Location: 3,5 west of Gura Văii, km 12, between Scarpia and Ungureanu viaducts.

From Jidoștița to Slătiniț, the road shows 5,5 Km of continuous outcrops in the Sebeș-Lotru "Series", belonging to the pre-Alpine basement of the Getic Nappe. This "series" shows a very complex lithological constitution at regional scale. Within the Iron Gates outlier, medium and high grade polymetamorphic rocks crop out (Sebeș Group), in tectonic relationships with the Jidoștița Formation (Pl. II).

Both lithotectonic units, ascribed to the Precambrian, show polymetamorphism (M_1 , M_2) in amphibolite facies conditions: their tectonic relationship is marked by medium grade mylonites.

In the Iron Gates outlier, the Sebeș Group consists of a dominantly metaterigenous sequence (paragneisses, micashists), associated to amphibolites, quartz-feldspar gneisses and scarce marble lenses). Within this sequence tectonic lenses with a high-grade, pre- M_1 metamorphic history, occur: eclogites, banded gneisses, metaperidotites. Metatectic migmatization and in situ anatexis (on quartz-feldspar gneisses) are common, materialised by structurally controlled granitic or pegmatitic leucosome (mainly D_2 - related, but also pre- D_2); advanced anatexis produces granitoid bodies and diatexitic veins.

The Jidoștița Formation, cropping out to the north within several tectonic "windows", consists of biotitic or quartzitic paragneisses, largely with mylonitic fabrics and recrystallisation of sillimanite + andalusite.

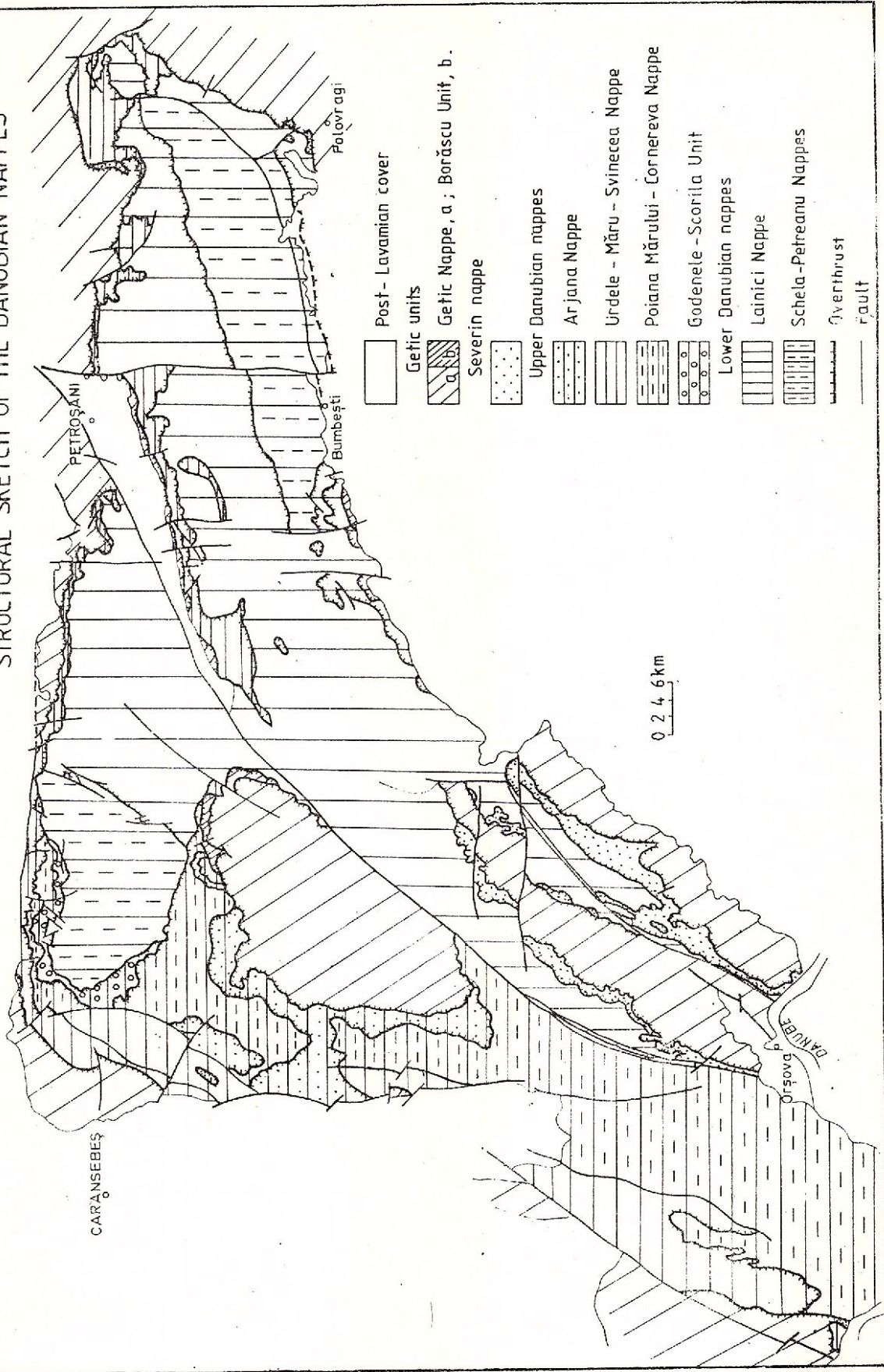
The outcrops in the Sebeș Group (300 m) show structural style and morphology of distinct fold generations (Fig. 1); S_2 foliations and related L_2 mineral lineations coexist with relict structural elements: sheared S_1 layering and B_1 folds, older mineral and stretching lineations. Stereographic representations of the regional pervasive S_2 foliations and banding, related mesoscopic folds (B_2) and lineations show a dominant NE-SW orientation and the effects of refolding (Figs. 2, 3). Zones of high shearing, with microblastic mylonitic structures, accompanied by amphibolite facies blastesis, are visible in paragneisses at Ungureanu viaduct.

Rocks show newly formed, fine grained, biotite (associated to sillimanite-muscovite, or to garnets) and planar-linear fabric: they could be related to the tectonic contact with the Jidoștița Formation. Relict mineral assemblages consist of kyanite, staurolite (in micashists west of Ungureanu viaduct) and biotite, garnet, suggesting that the first metamorphism (M_1) took place in staurolite-kyanite zone conditions: M_1 assemblage is overprinted by a sillimanite + biotite + muscovite association. On a regional scale, structural elements of



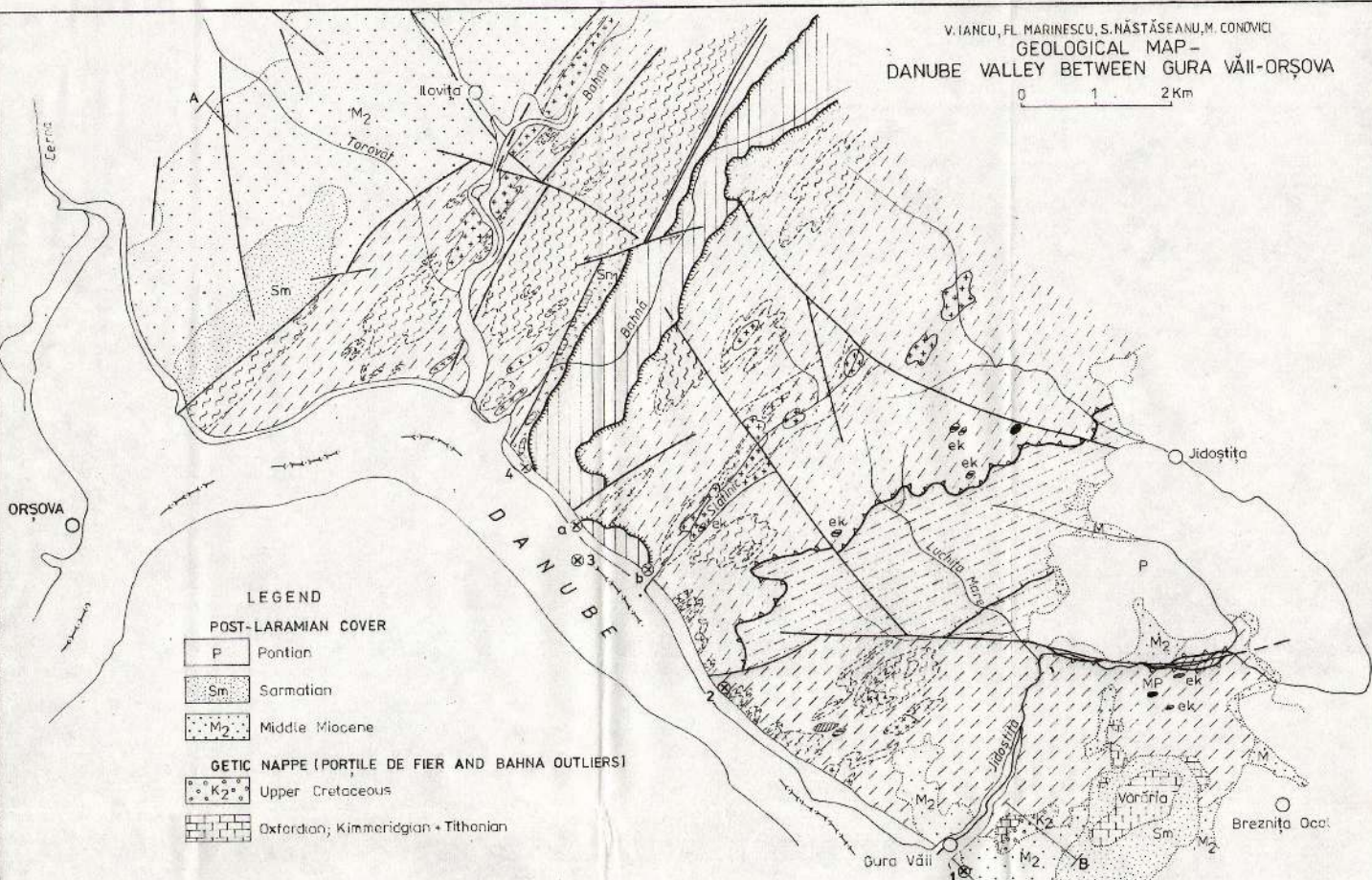
T. BERZA, V. IANCU, A. SEGHEDI, A. DRĂGĂNESCU

STRUCTURAL SKETCH OF THE DANUBIAN NAPPE



V. IANCU, FL. MARINESCU, S. NĂSTĂSEANU, M. CONOVICI
GEOLOGICAL MAP -
DANUBE VALLEY BETWEEN GURA VĂII-ORȘOVA

0 1 2 Km



LEGEND

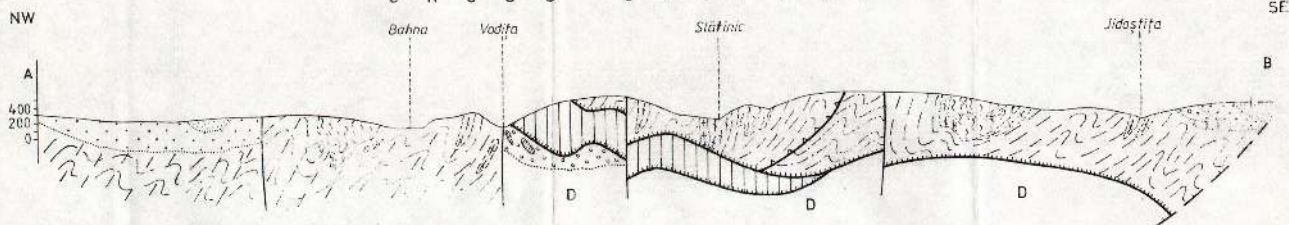
POST-LARAMIAN COVER

- P Pontian
- Sm Sarmatian
- M₂ Middle Miocene

GETIC NAPPE (PORTILE DE FIER AND BAHNA OUTLIERS)

- K₂ Upper Cretaceous
- Oxfordian, Kimmeridgian + Tithonian

C R O S S S E C T I O N



PRE-ALPINE BASEMENT
SEBEȘ-LOTRU SERIES (PRECAMBRIAN)

- Granitoids (Diatexites)
- Quartzo-feldspathic gneisses
- Micaschists
- Paragneisses
- Amphibolites
- High grade rocks: eclogites (ek); metaperidotites (MP)

JIDOȘTIȚA FORMATION (PRECAMBRIAN)

- Quartzitic, biotite bearing paragneisses

SEVERIN NAPPE

- Sinaiá beds (Tithonian - Neocomian)

LOWER DANUBIAN UNITS

- Lower Senonian; melanges (a); limestone otistoliths (J₃ ap); (b).

CONVENTIONAL SIGNS

- Unconformity
- Geological limit
- Lithologic limit
- Fault
- Reverse fault
- Alpine overthrust plane
- Pre-Alpine litho-tectonic discontinuity
- Cross section
- Stop
- Danubian units: Precambrian basement + Mesozoic (pre-Senonian) cover

D_2 are accompanied by blastesis in sillimanite-cordierite - K feldspar and sillimanite-muscovite or andalusite zones (low-pressure facies series, Iancu et al., 1987).

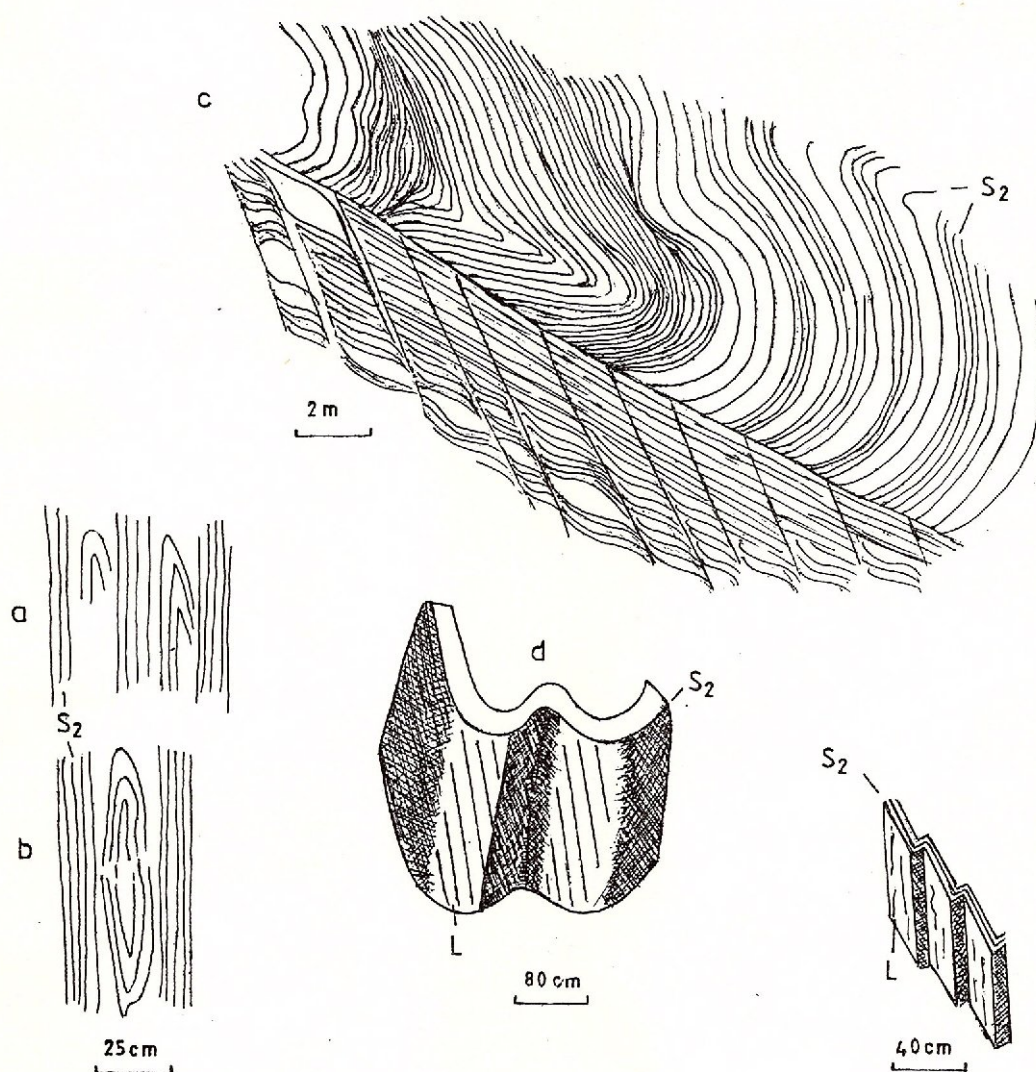


Fig. 1 – Structural style of distinct fold generations in the Sebeș-Lotru series (Sebeș Group).

a. Tight, rootless intrafolial folds, Ungureanu viaduct; b. sheath folds in gneisses and amphibolites, Padina Mică viaduct; c. Overturned B_3 fold with shallow southward plunge refolds the S_2 foliation of gneisses at Padina Scorpiei viaduct. The overturned limb is strongly sheared, showing top to the N; d. B_3 normal fold with vertical axis refolds S_2 foliation and early stretching lineation, Ungureanu viaduct; e. Asymmetrical kink-folds deform the S_2 foliation on limbs of B_n folds.

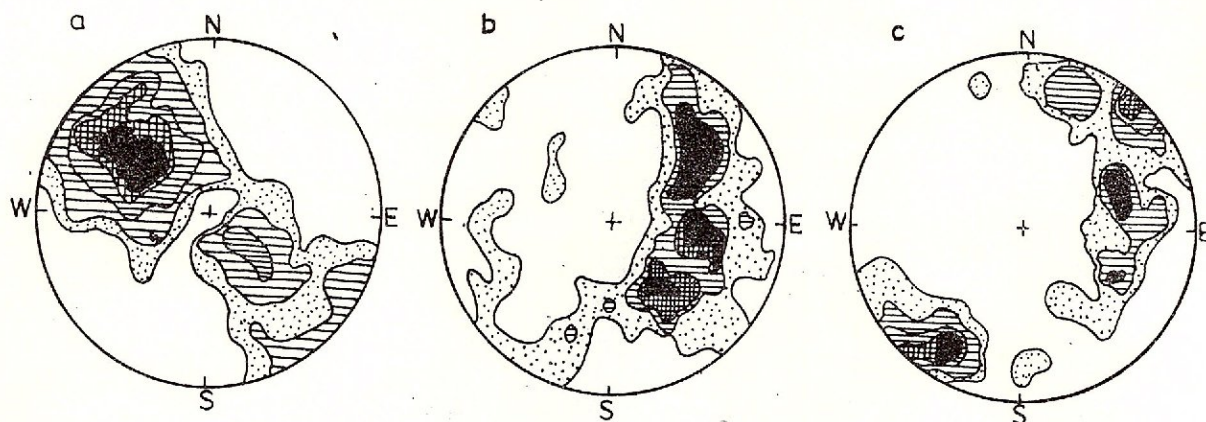


Fig. 2 - Planar and linear elements in the Sebeș-Lotru Series (Iancu, Hârtopanu, 1979). Schmidt projection, lower hemisphere. a. Stereogram of the S_2 foliations; 1258 measurements 5-4-3-2-1 %; b. Stereogram of mineral lineations; 111 measurements; 6-5-4-3-2-1 %. c. Stereogram representing hinges of the mesoscopic folds; 150 measurements; 7-6-5-4-3-2-1 %.

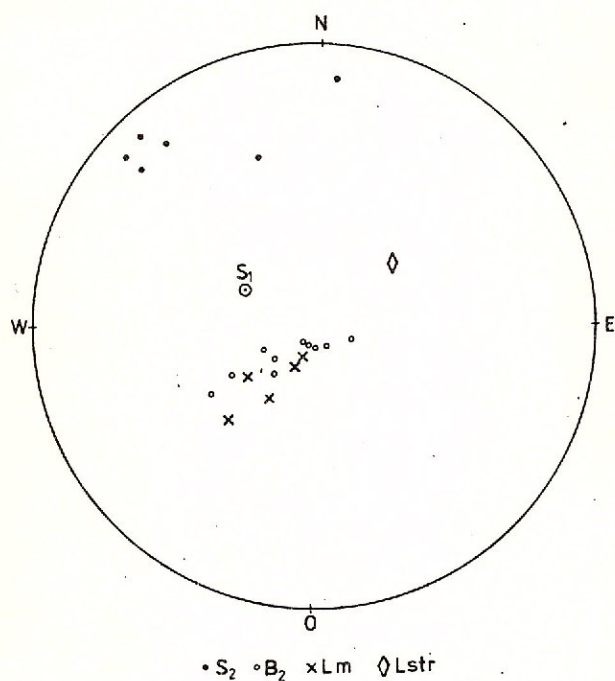


Fig. 3 - Stereogram of the planar and linear elements in the Sebeș-Lotru series; Danube river, between Padina Mică and Ungureanu viaducts. S_1, S_2 , poles; B_2 , fold hinges; L_m , mineral lineation (biotite, amphibole); L_{str} , stretching lineation.

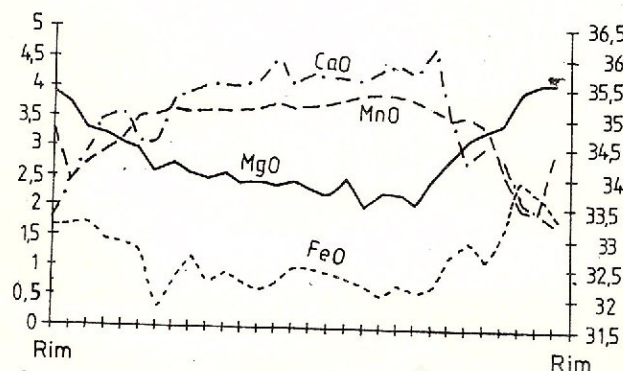


Fig. 4 - Zoning profile of garnet from kyanite-staurolite micaschist (Bahna outlier) (A. V. Bojar). Bulk chemical composition of the garnet: $(Fe_{2.12}^{2+}Mg_{0.31}Ca_{0.35}Mn_{0.25})_{3.03}(Al_{1.93}Fe_{0.06}^{3+})_2[SiO_4]_3$

Polymetamorphic evolution of these rocks are exposed by: the existence of superposed mineral parageneses related to successively metamorphic events (M_1 of barrovian type, M_2 of intermediary low pressure type; Hărtopan, 1972, Iancu et al., 1986); re-equilibration of the relict phases; overgrowth or new blastesis of some mineral phases (e.g. garnet, plagioclase). While the relict garnet from the kyanite bearing eclogites and granulites are rich in pyrop (44 core to 51 rim, in an eclogite; 25–28 to 17, in a banded granulite; Iancu et al., 1994); the zoned garnets from the micaschists are almandinic (Fig. 1 d; Bojar, unpublished data).

Post S_2 structural elements include open B_3 folds (Fig. 4), shallow crustal level kink folds (associated to crenulation cleavages and greenschist facies recrystallisation) and late, extensional boudinage.

Stop no. 3: Tectonic contact at the sole of the Getic Nappe, with lower Cretaceous turbidites of the underlying Severin Nappe: structural style of the Sinaia Beds.

A. Seghedi, V. Iancu

Location: Mouth of the Slătanic valley–Slătanicu Mare viaduct (Km 10) and Oreva viaduct (1300 m W of Slătanic Valley).

West of the junction with the Slătanic valley, the road-cut along the Danube offers continuous outcrops in the Severin Nappe, consisting here of Sinaia Beds, lying in the core of a large-scale, post-nappe, open antiformal fold.

The Sinaia Beds represent a sequence of mainly distal turbidites (facies D - Mutti, Ricci-Lucchi, 1972), dominated by Tdc and Tcde Bouma divisions; thicker sandstone beds (facies B_2), up to 40–50 cm, develop close to the Slătanicu Mare viaduct. Grey pelagic limestone interbeds also occur. In places, sandstones preserve a Chondrites ichnofauna.

The age of the sequence is ascribed to the Upper Tithonian - Lower Valanginian (Stănoiu, 1978).

Structural style of the Sinaia Beds involves tight to isoclinal recumbent folds (B_1), strongly refolded by steeply dipping normal folds (Fig. 5). Recumbent folds are highly disrupted by the normal folding, being usually preserved as isolated fold hinges. Subsequent are various types of kink and chevron folds, with axial planes steeply dipping to the East or to the West, conjugate folds being common (Fig. 6). The chevron folds exhibit hinge lines dipping gently NNE. The structures are interpreted by A. V. Bojar as result of a late compressive event (Miocene ?) after Upper Cretaceous nappe stacking. Slaty cleavages seldom occur within the pelitic divisions. Some sandstone lithofacies may show fracture cleavages (S_1) fanning in fold hinges.

The last structures are high angle normal faults. Paleostress analyse from fault planes and slickenside striations (Fig. 6 b, c) gave subvertical compressive stress, probably a later effect of loading of the crystalline on the top of the flysch deposits (A. V. Bojar, Graz, unpubl. data, 1994).

Relationships between the Getic Nappe and Sinaia Beds are visible on the right bank of the Slătanic Valley, where metamorphic rocks of the Getic Nappe are exposed, showing pervasive brittle deformation. The tectonic contact may be followed uphill, amphibolite facies Sebes Group metamorphic rocks making up the top of the hills, while the road is cutting through the sedimentary deposits from the core of the antiform. 1300 m westwards in the road-cut, (300 m east of Oreva viaduct) strongly sheared rocks of the Getic Nappe reappear in the core of a tight synformal fold; a steeply dipping shear band foliation develops both in the Sebes Group rocks and in the Sinaia Beds, several meters away from the contact (Fig. 7).



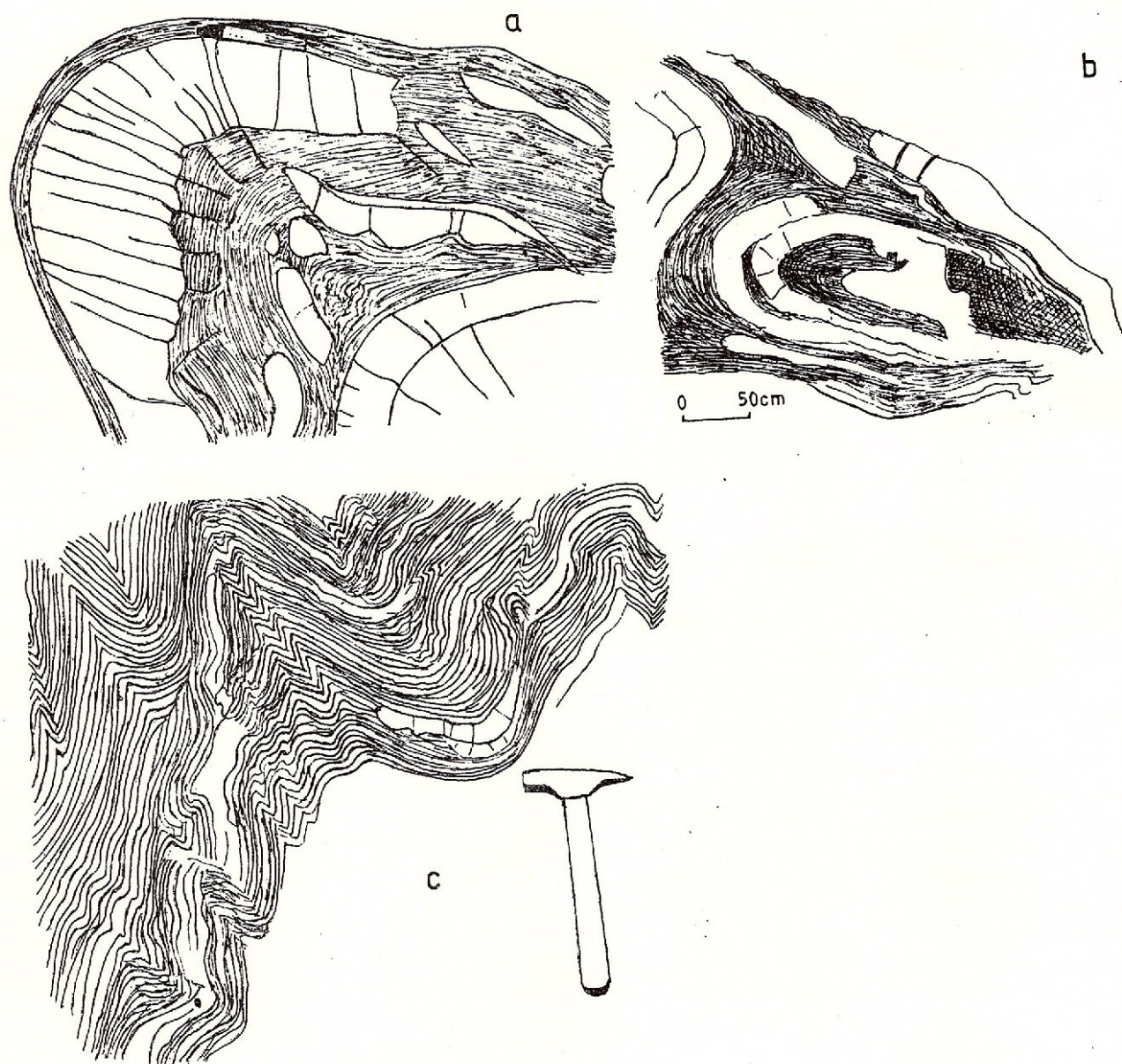


Fig. 5 - Mesoscopic folds in the Sinaia Beds.

a. Recumbent fold of sandstone beds with weakly developed fracture cleavages, Slătiniul Mare viaduct; b. Fracture cleavages fanning in sandstones from the hinge of an inclined fold, West of Slătiniul Mare viaduct; c. Assymetric kink folds with steeply dipping axial planes, West of Slătiniul Mare viaduct.

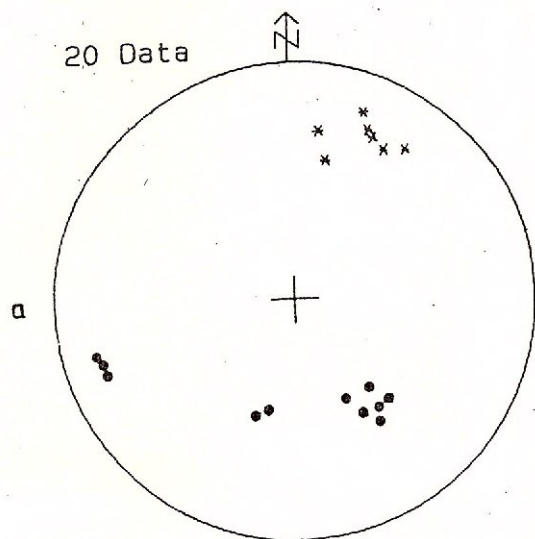


Fig.6a* Axis orientation of chevron folds

• Slaty cleavage

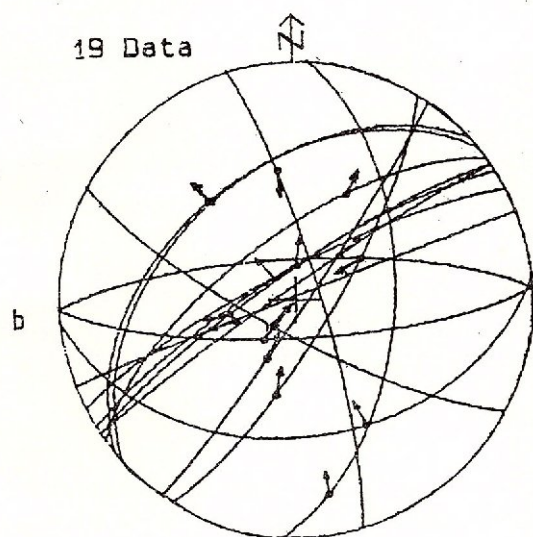


Fig.6b Fault planes, slickenside with sense of slip

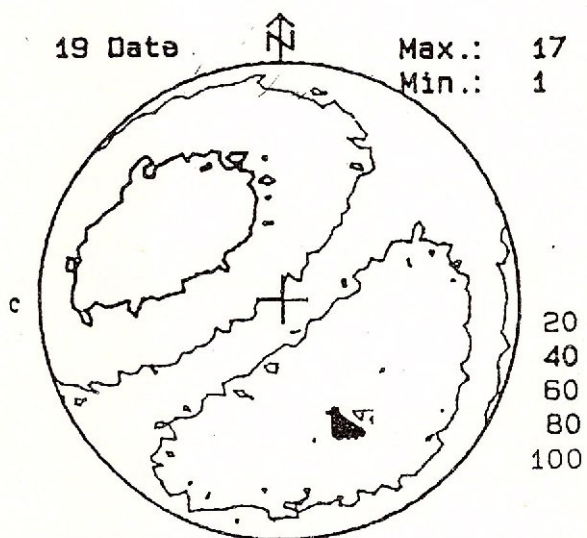


Fig.6c Paleostress directions

Fig. 6 - Structural elements (a, b) and paleostress directions (c) in Sinaia Beds (A. V. Bojar, 1994)

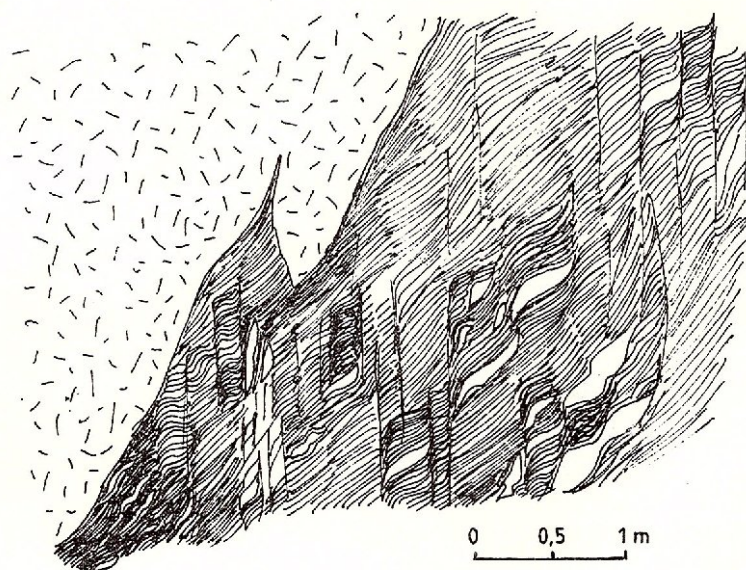


Fig. 7 – Tectonic contact between strongly brecciated polymetamorphic rocks of Getic Nappe (left) and Sinaia. Beds displaying a shear band foliation (right) East of Oreva viaduct.

Stop no. 4: Upper Cretaceous cover of the Lower Danubian Lainici Unit.

V. Iancu, A. Seghedi

Location: Junction of the Danube River with Vodița Valley, Vârciorova Viaduct. km 28.

300 m East of the Vodița Valley, Upper Cretaceous (Lower Senonian) sedimentary cover of the Lainici Nappe crops out in the core of a post-nappe antiform, beneath the Severin Nappe (in the East), affected by a fault of the Bahna system (in the West). Similar relationships are well exposed 1,5 Km upstream the Vodița Valley.

East of the Vodița Valley, highly cataclastic paragneisses and micaschists (Sebeș-Lotru "Series") of the Bahna outlier are exposed along 100 m on the road-cut. Next to this tectonic junction, situated in the prolongation of the Bahna "graben" (preserving Miocene deposits in the Mehedinți Plateau), Senonian sandstones contain "olistolites" of Upper Jurassic-Lower Cretaceous limestones or Sinaia Beds (Năstăsescu, in Hărtopanu et al., 1986; Stănoiu, in Iancu et al., 1986).

On this location, the Upper Cretaceous deposits consist of massive or graded bedded, intraclast rich, medium or coarse grained sandstones with black mudstone interbeds.

Trace fossils occur within pelitic or silty grain sizes, mainly Chondrites. Coarser grained rocks are rich in limestone lithoclasts. Plagioclase and quartz are major constituents of sandstones; detrital micas are muscovite and partly chloritised biotite; garnet, tourmaline and sillimanite bearing quartzites are minor components. Rocks have a carbonate cement, micritic to largely recrystallised.

Stop no. 5 (A, B): Cerna - Jiu Tertiary Fault System

V. Iancu, T. Berza

Location: Cerna River, upstream of Băile Herculane

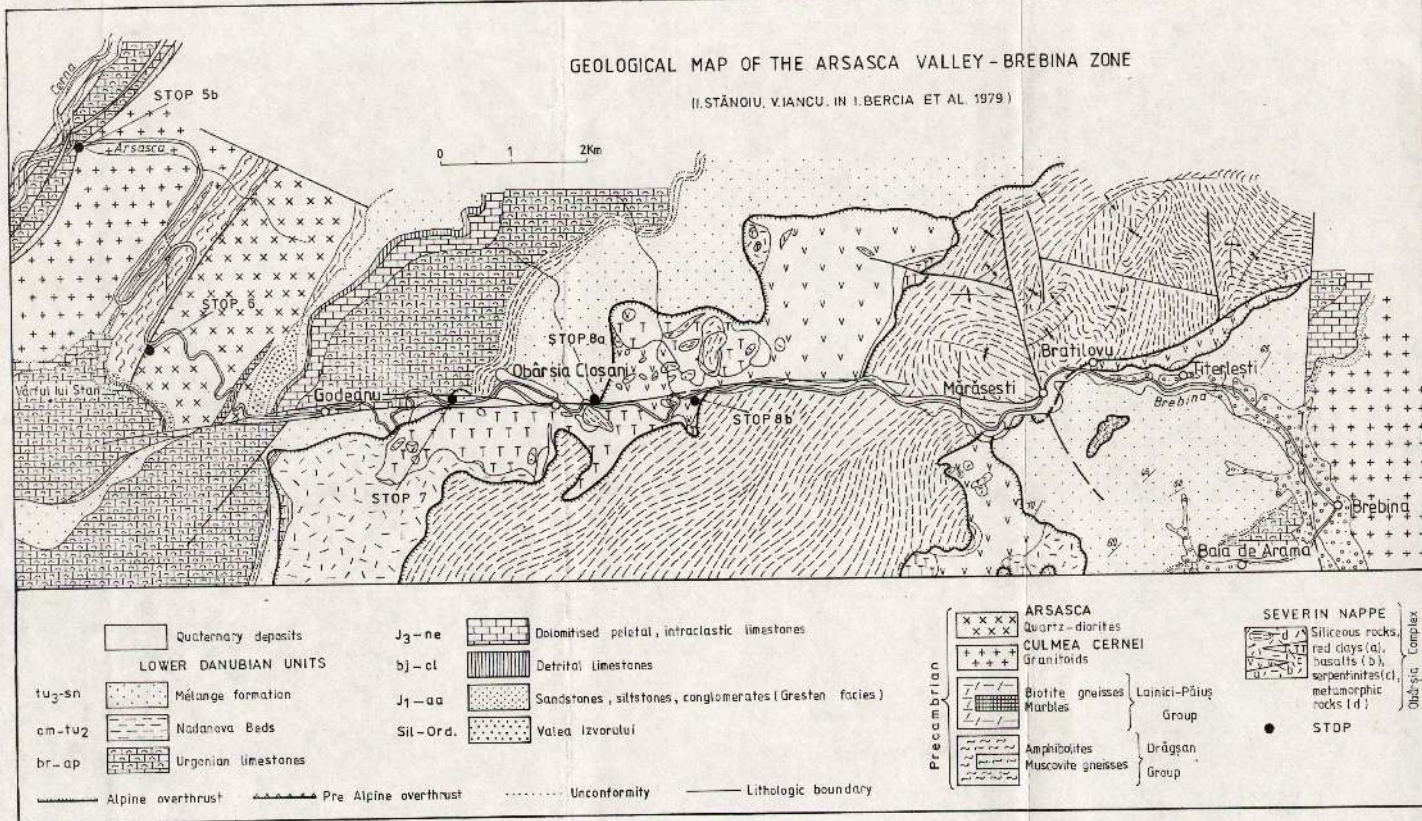
Along the Cerna River, between Orșova and the junction with Arșasca Valley, two short stops will follow.

The road along the Cerna River follows the Cerna-Jiu Tertiary strike-slip fault system. This system is a fault zone (approximately 5–600 m thick), consisting of parallel and anastomosing faults with a complex structure and history (Berza, Drăgănescu, 1988): post-Laramian (pre-Chattian) dextral strike-slip (displacement: 30–40 Km); normal extensional faults; normal or reverse compressional faults (Sarmatian) (i.e. northern border of the Petroșani Basin). Along this fault system, both Danubian terranes (basement and cover) and Getic Nappe outliers crop out.



GEOLOGICAL MAP OF THE ARSASCA VALLEY - BREBINA ZONE

(I. STĂNOIU, V. IANCU, IN I. BERCIA ET AL. 1979)



5.A. Road-cut in black marls and limestones with chert interbeds - Nadanova Beds (Cenomanian - Turonian, Năstăseanu, 1979) - showing small-scale asymmetrical folds on the normal limb of a large-scale recumbent fold (Fig. 8).

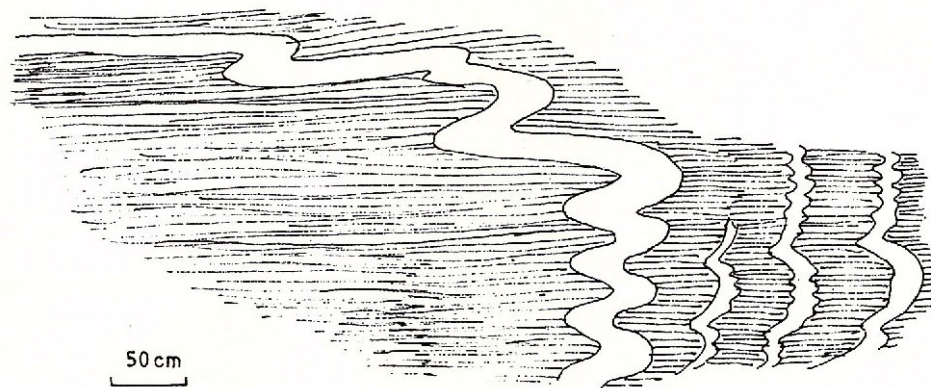


Fig. 8 - Overturned limb of a recumbent fold, with flat-lying penetrative slaty cleavage; associated microfolds are visible in siliceous interbeds. Nadanova Beds - Cerna Valley.

5.B. At the junction between Cerna and Arasca valleys, carbonatic rocks of the Mesozoic cover (Jurassic-Lower Cretaceous) show tectonic contacts with both metamorphic rocks of the Getic Nappe (to the west) and Culmea Cernei granitoids from the basement of the (Lower Danubian) Lainici Nappe (to the east) (Pl. III).

The western contact is marked by NE-SW strike-slip faults accompanied by horizontal slickensides (the right bank of Cerna River). West of the limestones, tectonic slices of Upper Cretaceous deposits (siltstones, sandstones, conglomerates or mudstones with "olistolites") crop out along the Cerna River. The eastern contact of the limestones is marked by listric normal faults (Iancu, 1977) representing extensional faults, sites of decollement of the Mesozoic cover from its metamorphic basement (Fig. 9). The carbonate rocks are Urgonian facies limestones, showing pervasive brittle deformation.

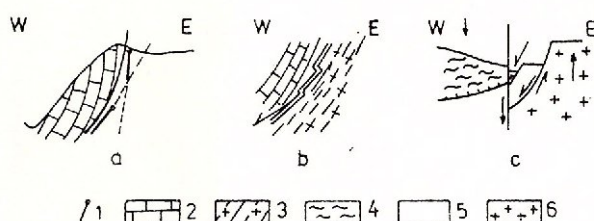


Fig. 9 - Tertiary listric faults from the Cerna-Jiu system, North of the Vârful lui Stan (Iancu, 1977).

1. fault plane; 2. Mesozoic limestones; 3. sheared Culmea Cernei granitoids; 4. Getic nappe metamorphic basement; 5. Mesozoic cover of the lower Danubian unit; 6. pre-Mesozoic basement of the Danubian

Stop no. 6: Pre-Alpine tectonic contact between two types of Danubian basement: Lainici-Păiuș and Drăgșan Groups in the basement of the (Lower Danubian) Lainici Nappe.

V. Iancu, T. Berza

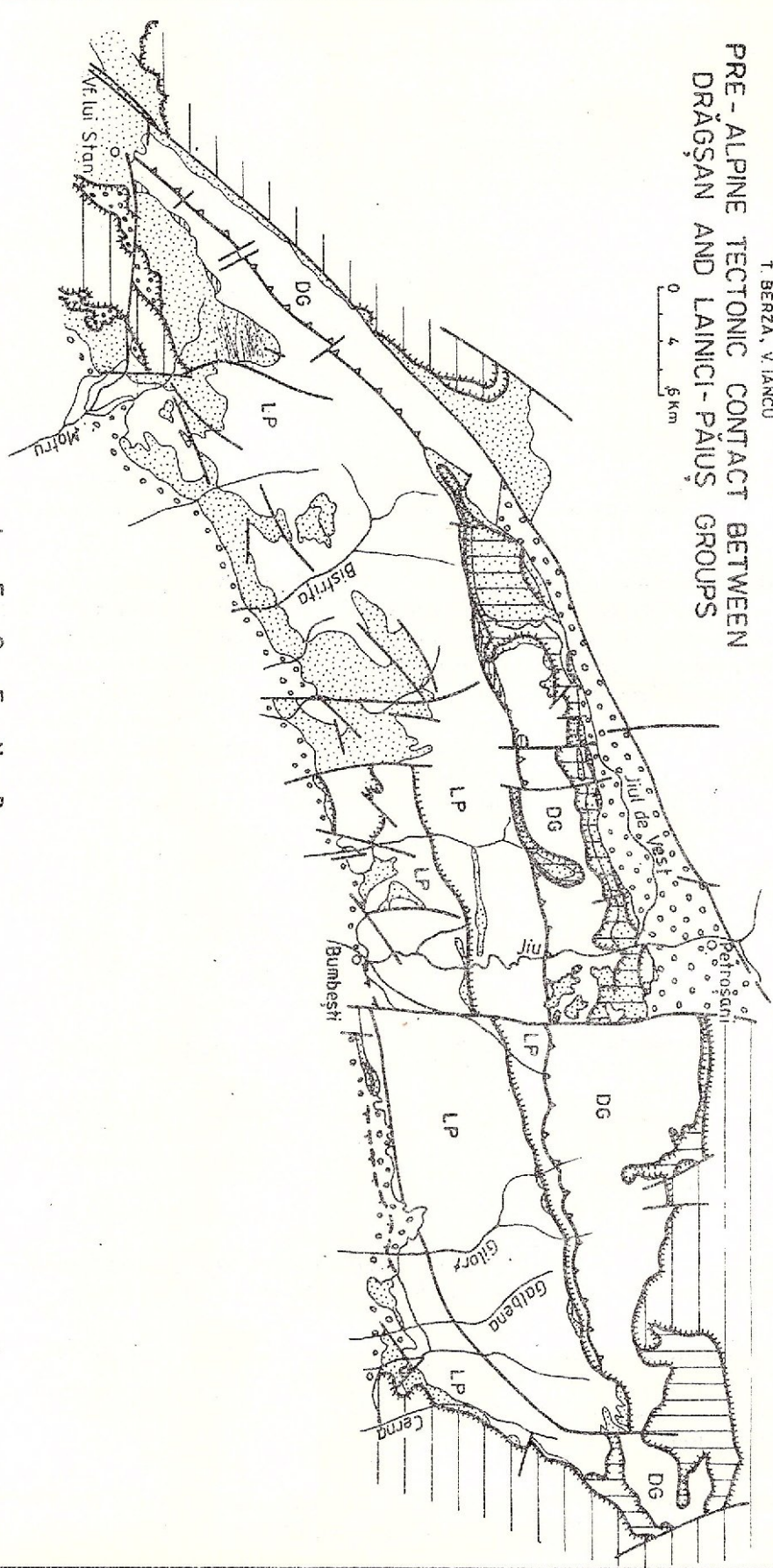
Location: Arasca valley, the road from Herculane to Baia de Arama, N of Vârful lui Stan, height 850 m (km 24 to Baia de Aramă).

From Cerna Valley to Godeanu village, the road crosses the core of an Alpine anticline, consisting of pre-Alpine basement rocks: Culmea Cernei granitoids, Drăgșan and Lainici-Păiuș Groups (Precambrian) and



T. BERZA, V. IANCU
**PRE-ALPINE TECTONIC CONTACT BETWEEN
 DRĂGȘAN AND LAINICI-PĂIUȘ GROUPS**

0 4 6 Km



- Tertiary-Quaternary deposits
- Getic nappe
- Severin nappe
- Upper danubian nappes:
a. basement; b. cover

- LOWER DANUBIAN NAPPEs
- Mesozoic cover
- PRE-ALPINE BASEMENT
- Paleozoic formations
- Precambrian basement:
DG-Drăgășan Gr.;
LP-Lainici Păiuș Gr.

- Fault
- Alpine overthrust plane
- Pre-alpine tectonic contact

Fig. 10

Fig. 10 - Simplified map of Vâlcan-Pârâng Mts.

Paleozoic sequences - Valea Izvorului (fossil-bearing, Ordovician-Silurian) and Poiana Mică (Permian ?) formations. The stop will show the pre-Alpine tectonic junction between the Drăgăsan Group (intruded by the Culmea Cernei granitoids) and the Lainici-Păiuș Group (intruded by the Arsasca diorite) (Fig. 10). The Culmea Cernei intrusive is a suite of medium K, calc-alkaline rocks, varying from biotite + hornblende granodiorites to tonalites and monzogranites (Iancu et al., 1994). The intrusion is hosted by polymetamorphic rocks of the Drăgăsan Group, consisting of amphibolites and amphibole gneisses (sometimes garnet bearing), muscovite gneisses and micaschists (with kyanite and staurolite, relics of M_1 paragenesis, within an amphibolite facies recrystallised matrix).

Granitoids are surrounded by a suite of artieritic migmatites (with granitic, aplitic and pegmatitic veins) and by weak contact metamorphic effects, seldom visible in micaschists (with sillimanite, muscovite and chloritoid).

The Lainici-Păiuș Group, showing cordierite-amphibolite facies metamorphism, consists in this area of micaceous paragneisses (sometimes graphitic), marbles and amphibolites (Carbonatic-Graphitic Formation). Subsequently to the amphibolite facies metamorphism rocks were intruded by a kilometric body of diorites (Arsasca pluton) and by younger dyke swarm of porphyritic microdiorites or microgranodiorites.

The road-cut at stop no. 6 shows low-grade mylonites on a dioritic protolith, interpreted as an effect of the dynamic metamorphism accompanying the pre-Alpine tectonic junction. Greenschist facies mylonitization yielded a retrograde association with chlorite, actinolite, epidote-clinozoizite, albite, stilpnomelane. Relict porphyroclasts - amphibole, pyroxene, plagioclase - are preserved in different stages of subgrain formation and intracrystalline deformation, due to progressive dynamic effects.

Rocks of the pre-Alpine mylonite zone show penetrative S-C or L-S fabric mylonites, with a steeply dipping mylonitic foliation deforming both metamorphic and magmatic rocks (Fig. 11).

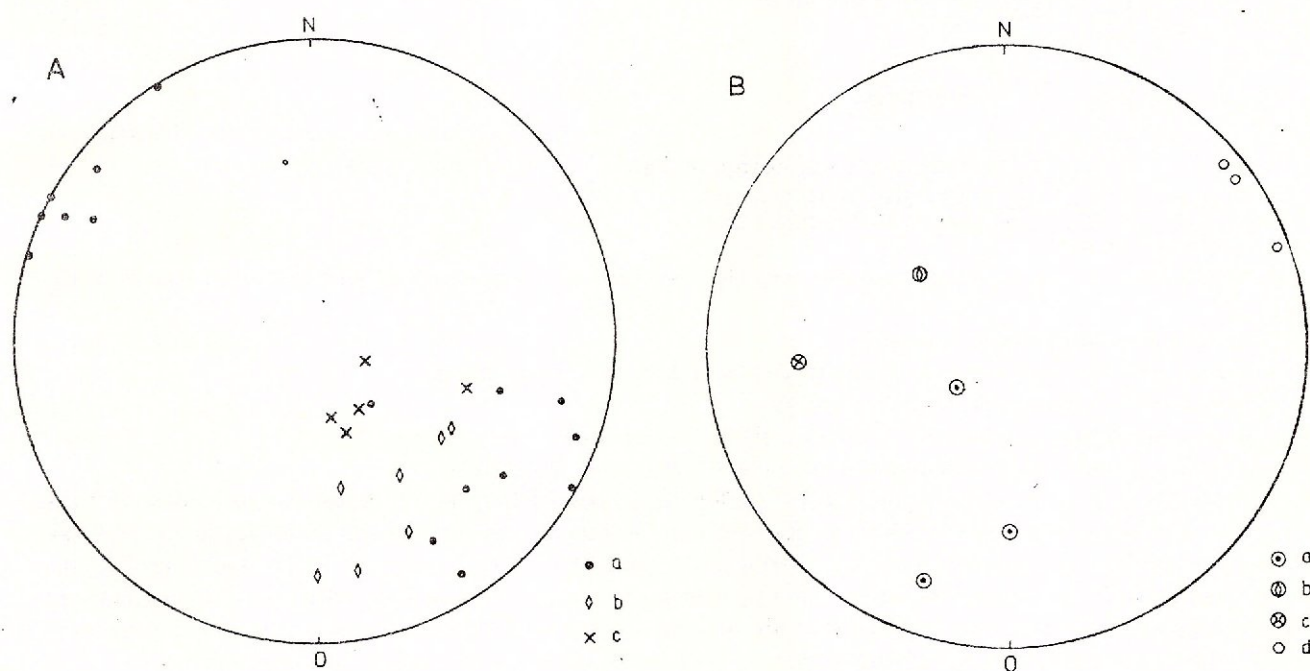


Fig. 11 - Stereogram of the planar and linear elements of the greenschist facies mylonites bordering Drăgăsan-Lainici-Păiuș pre-Alpine tectonic contact. 11A: Π poles of the planar structural elements (S // C): a, Southern part (Motru Sec basin); b, Western Jiu (Valea de Pești); c, Jiu Valley (junction with Dumitra brook). 11B: Linear elements; a, b, c, - location similar to Fig. 11A; d, quartz-chlorite fibre veins.

Stop no. 7: Mesozoic cover of the (Lower Danubian) Lainici Nappe and the Severin Nappe.

A. Seghedi, V. Iancu

Location: 2 km upstream of Obârșia Cloșani village (km 17 to Baia de Aramă).

On the southern limb of the Alpine Vârful lui Stan-Culmea Cernei anticline, the Precambrian metamorphic rocks and granitoids are overlain by a cover consisting of: the fossiliferous Ordovician-Șilurian Valea Izvorului Formation, the (probably) Permian Poiana Mică Formation and Jurassic-Upper Cretaceous formations, almost continuously outcropping between villages Godeanu and Obârșia Cloșani (Pl. III).

According to Stănoiu (in Bercia et al., 1977), the Mesozoic cover consists of several formations (Fig. 12):

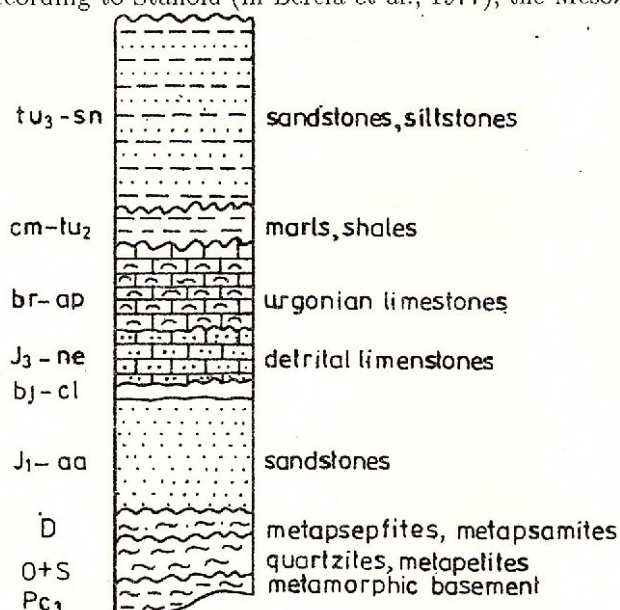


Fig. 12 - Lithologic column in the Mesozoic cover of the Lainici Nappe (Stănoiu, in Bercia et al., 1979).

- conglomerates, quartz-feldspar sandstones, siltstones and mudstones (pyrophyllite bearing) (Lower Jurassic);
- sandy limestones (Middle Jurassic);
- pelletal and intraclastic cherty limestones, (Upper Jurassic - Neocomian);
- Urganian limestones (Barremian - Aptian);
- shales, marls and sandstones - Nadanova Beds (Cenomanian - Turonian);
- siltstones, shales and conglomerates with olistolites (Lower Senonian).

Stop no. 7 shows the stratigraphic gap (disconformity) between Urganian limestones and Nadanova Beds. Urganian limestones are massive, white, seldom cleaved, Barremian-Aptian deposits (Stănoiu, in Bercia et al., 1977), overlain by black marls and shales, siltstones and sandstones of the Nadanova Beds. Beside microfauna, in the Nadanova Beds, belemnites occur, strongly deformed by a penetrative slaty cleavage. Fine laminations, representing strongly deformed bedding (S_0) are preserved between sets of spaced cleavages (S_1) in Nadanova rocks, contrasting with the underlying Urganian limestones, which show thicker bedding and no secondary cleavages.

Stop no. 8: Obârșia Complex (Severin Nappe), overlaying Upper Cretaceous mélange Formation (cover of the Lainici Nappe).

A. Seghedi

Location: Brebina valley, eastern end of Obârșia Cloșani village, Km 15 to Baia de Aramă.

The road cut along the Brebina valley exposes several tectonic units: Lainici Nappe overthrust by the Severin Nappe, overridden by the Getic Nappe (Bahna outlier).



The uppermost member of the Lower Danubian cover sequence is an Upper Cretaceous *mélange* complex (Upper Turonian-Senonian, Stănoiu, in Bercia et al., 1977). This complex is a highly broken sequence of distal or proximal turbidites, showing a chaotic block-in-a-sheared-matrix texture. Blocks include fragments of Mesozoic Danubian cover rocks, basalts, serpentinites and amphibolite facies metamorphic rocks.

Deformational history may vary for individual outcrop areas, including tectonic shearing and gravitational mass transport. Shear sense indicators suggest extensional deformation and dextral shearing, connected to displacements along the E-W trending dextral strike-slip faults of Bahna system (6-8 Km of horizontal displacement).

The Severin Nappe in this area consists of the Obârșia Cloșani Complex (Stănoiu, in Bercia et al., 1977) - a *mélange* formation with ophiolites and siliceous pelagic deposits assigned to the Upper Jurassic-Lower Cretaceous.

Ophiolitic rocks are basalts (pillow-lavas in places), harzburgitic ultramafites and minor gabbros, pervasively sheared and disrupted (Mărunțiu, 1987). Geochemical features of basalts suggest ocean floor tholeiites (Cioflica et al., 1980), considered an obducted, marginal basin crust (Mărunțiu, 1987) or a segment of transitional ridge (T type MORB) (Savu et al., 1985, 1991) (Figs. 13, 14).

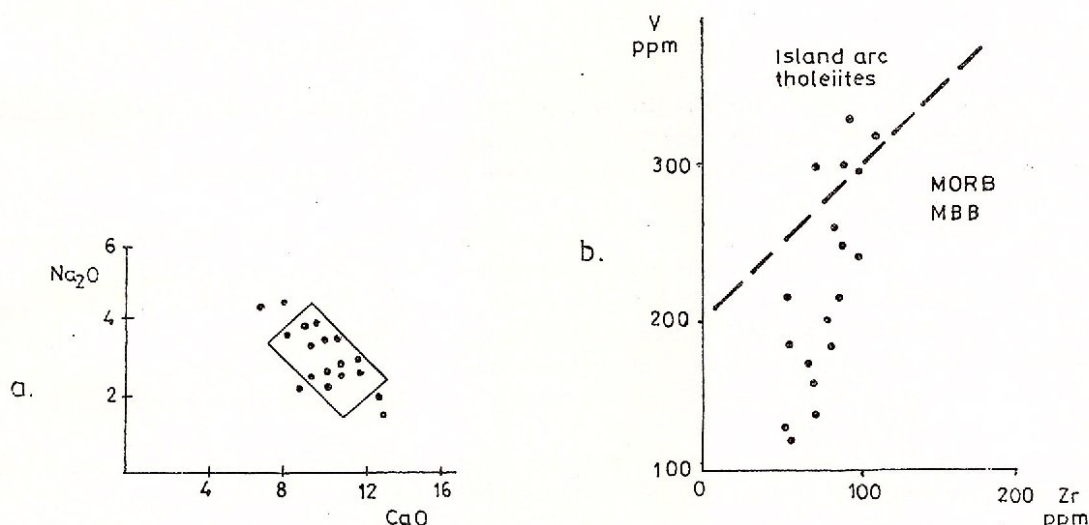


Fig. 13 Na₂O-CaO diagram (a) and V-Zr diagram for Obârșia Complex basalts (Mărunțiu, 1987).

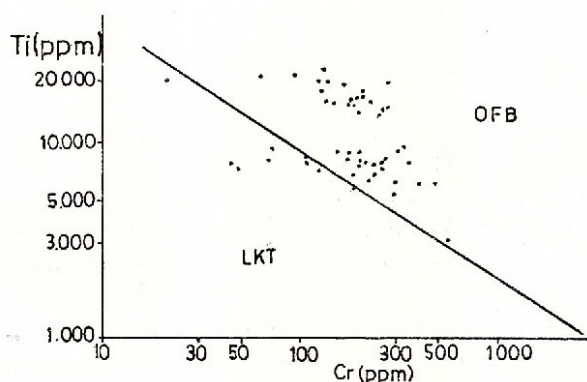


Fig. 14 Ti-Cr diagram for mesozoic ophiolites (Baia de Arama (Savu et al., 1985).

Along the Brebina valley, spotted basalts are typical occurrences, showing low-temperature metamorphism in the prehnite-pumpellyite facies.



The rocks preserve fresh pyroxenes and magmatic fabric. Metamorphic minerals replace primary minerals (chiefly feldspars). The main newly formed assemblages of these sheared green rocks involve: pumpellyite + prehnite + chlorite + epidote + titanite + calcite + albite + zeolites.

Metamorphic conditions are evaluated at 2–300°C, in the low-pressure bathozone.

Basalts included within serpentinitic rocks show metasomatic alteration products including garnet (grossular) + prehnite + epidote + clinozoisite (Mărunțiu, 1987).

Structural elements of ophiolites consist of S-C fabric, with a penetrative scaly-cleavage typical for serpentinitic rocks. Flat lying banded mylonites develop on basalt protoliths along the western border of the nappe.

Associated sedimentary deposits include red or green coloured silicious and pelitic rocks representing radiolarian cherts and muds. These rocks show a penetrative planar cleavage, axial planar to tight recumbent folds. Microstructure of cleavages reveal an anastomosed phyllosilicate foliation coexisting with a slaty cleavage, both strongly deforming radiolarian remains. Mineralogical composition of $\pm 2\mu$ fraction of these rocks indicate illite dominated mineralogy, with IC in the anchizone.

The eastern border with the Bahna outlier metamorphic rocks is marked by brittle deformation. Non-cohesive breccia and fault gouge in this area are probably related to the dextral strike-slip of the Bahna system.

Second day

General stops along three valleys - Tismana, Bistrița and Sohodol - will present the internal structure and constitution of the Lainici Nappe from the lower Danubian Units.

These streams expose sequences of the Precambrian basement and its Mesozoic cover (Lower Jurassic-Upper Cretaceous). From W to E, the sections reveal:

- Tismana stream: Precambrian granitoids of the Tismana batholit; disconformable Mesozoic cover;
- Bistrița stream: high temperature-low pressure metamorphic rocks of the Lainici-Păiuș Group, intruded by garnet bearing leucogranites, pegmatites and later (but pre-Ordovician) dyke swarms (subsequent to regional metamorphism and granite intrusion); the transgressive Mesozoic cover unconformably overlies the basement rocks; the structural and metamorphic disconformity between basement and cover rocks is obvious, but Alpine metamorphic effects are superimposed on both; The Alpine imprint on the Danubian basement is dynamic low-grade type, ductile and regional.
- Sohodol stream: the section goes through the Mesozoic cover rocks; Upper Cretaceous (Lower Senonian) show low-temperature Alpine metamorphism (prehnite-pumpellyite facies).

Stop no. 9: Tismana pluton

T. Berza

Location: Tismana stream, at junction with right tributary (Fig. 15)

The Tismana pluton consists mainly of porphyritic granites with a primary foliation; to the SW, massive granites occur. Diorites and granodiorites develop at the western border (Berza, 1978).

CIPW normative QAP diagram (Fig. 16 a) shows the low Q content of most porphyry granites and the existence of monzonitic and monzodioritic rocks. $MgFe_{tot}Alk$ diagram (Fig. 16 b) point to a calc-alkaline differentiation trend. Kuno diagram (Fig. 17 c) shows even an alkali enrichment, conspicuous against neighbouring Frumusu Massif and dykes cross-cutting both massifs, with pure calc-alkaline trend. The pluton was emplaced into low P - high T metamorphic rocks of the Lainici-Păiuș Group, subsequently to early metamorphic event. Geochronological evidence (K-Ar and U-Pb ages) suggest Upper Proterozoic age for pluton emplacement (Figs. 18, 19).



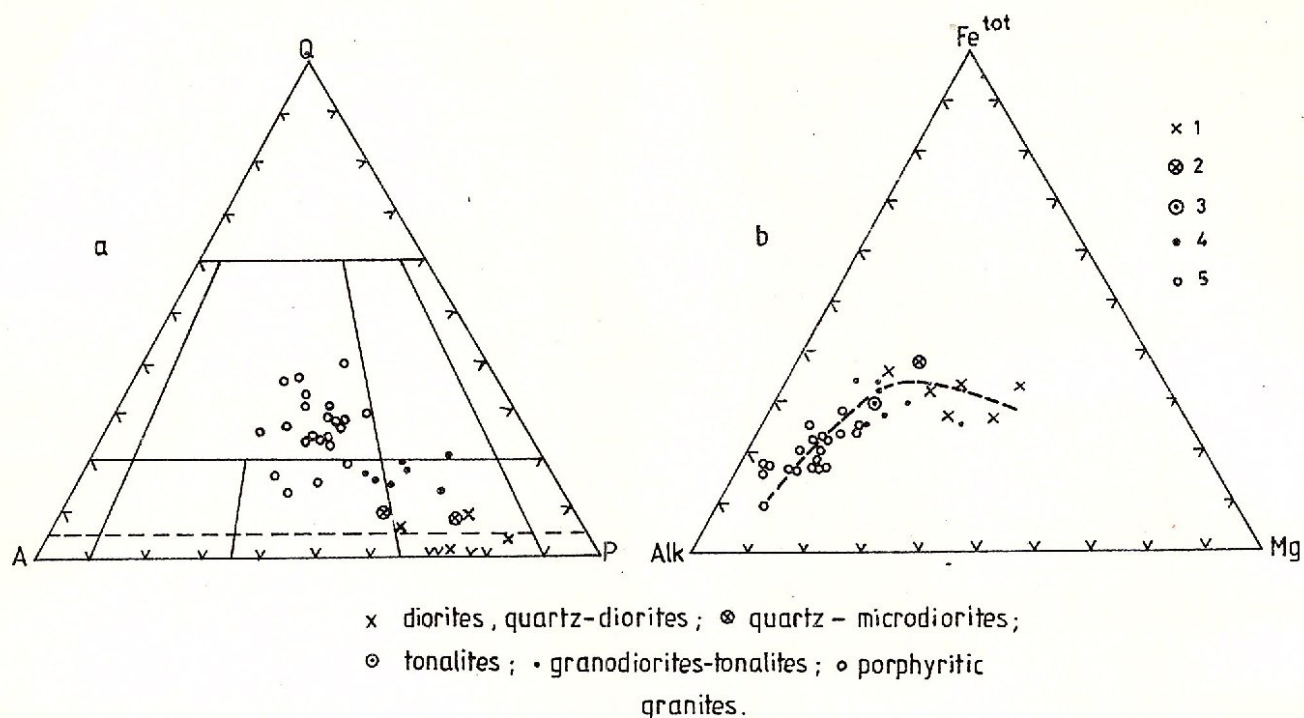


Fig. 16 Normative QAP diagram (a) and Mg-Fe-Alk diagram for the Tismăna graniteoid (b) (Berza, 1978)

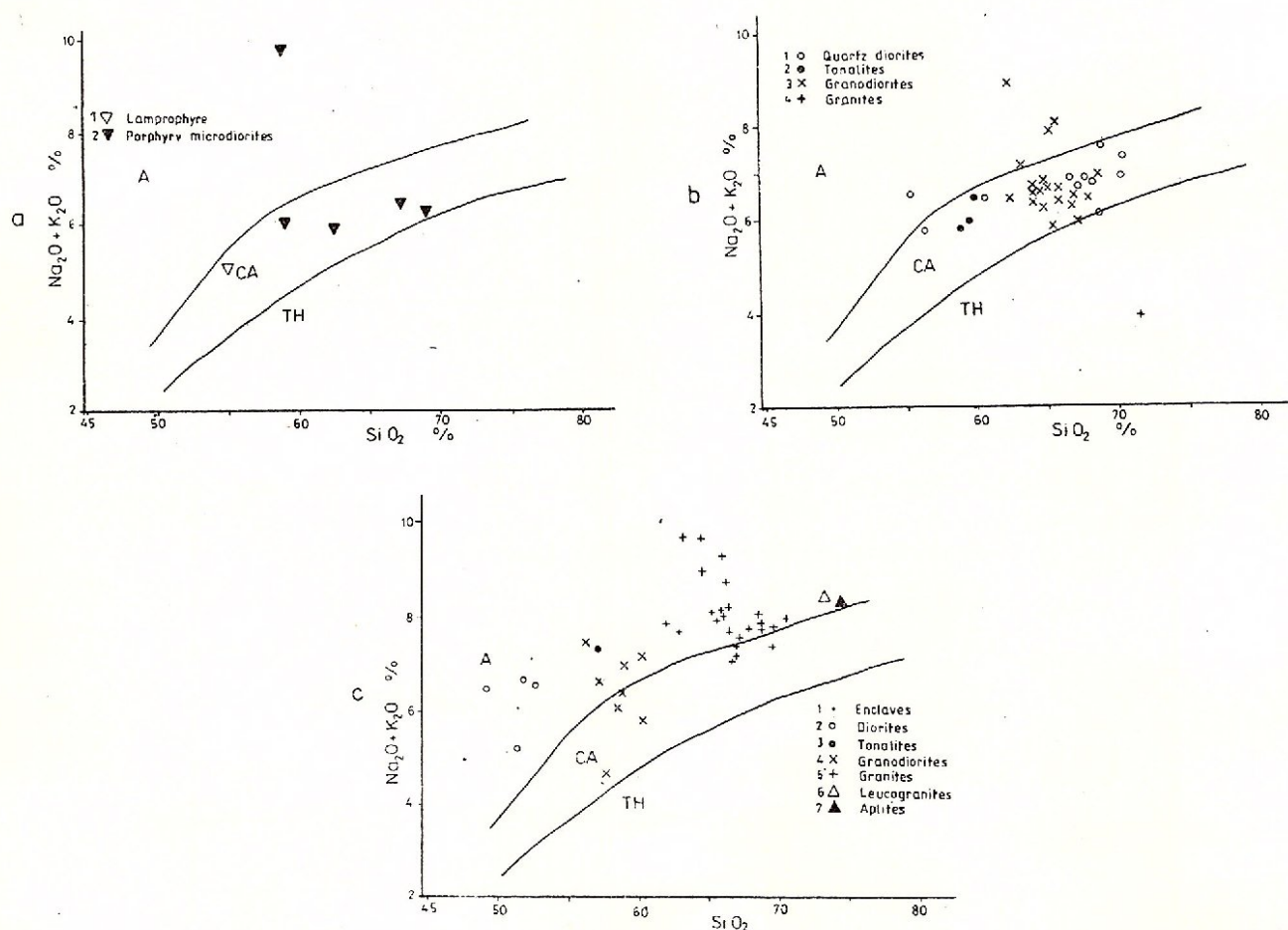


Fig. 17 - $\text{Na}_2\text{O} + \text{K}_2\text{O} - \text{SiO}_2$ diagram for: (a) dyke-swarms; (b) Frumosu Massif; (c) Tismăna Massif. (Tatu et al., 1993).



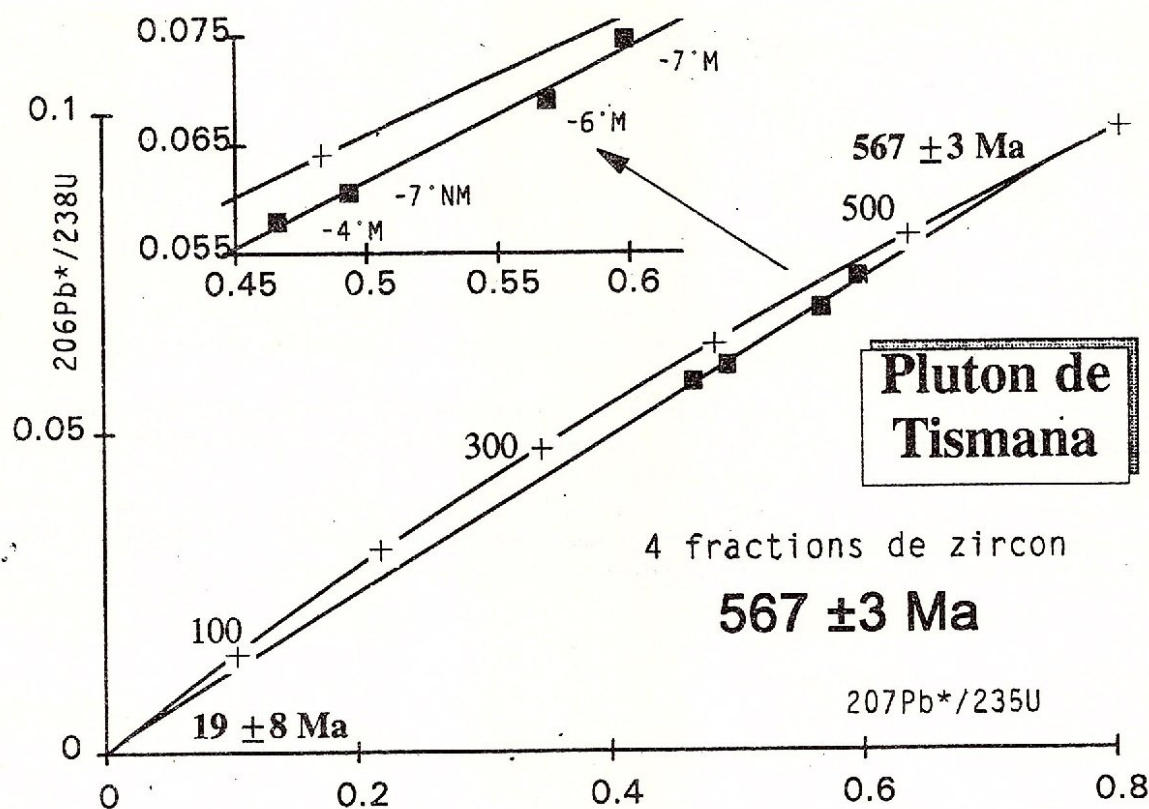


Fig. 18 Concordia diagram for a sample of Tismana granite from Poenina Valley (Zircons separated by J. C. Duchesne and analysed at Brussels by J. P. Liégeois, 1993, 1994).

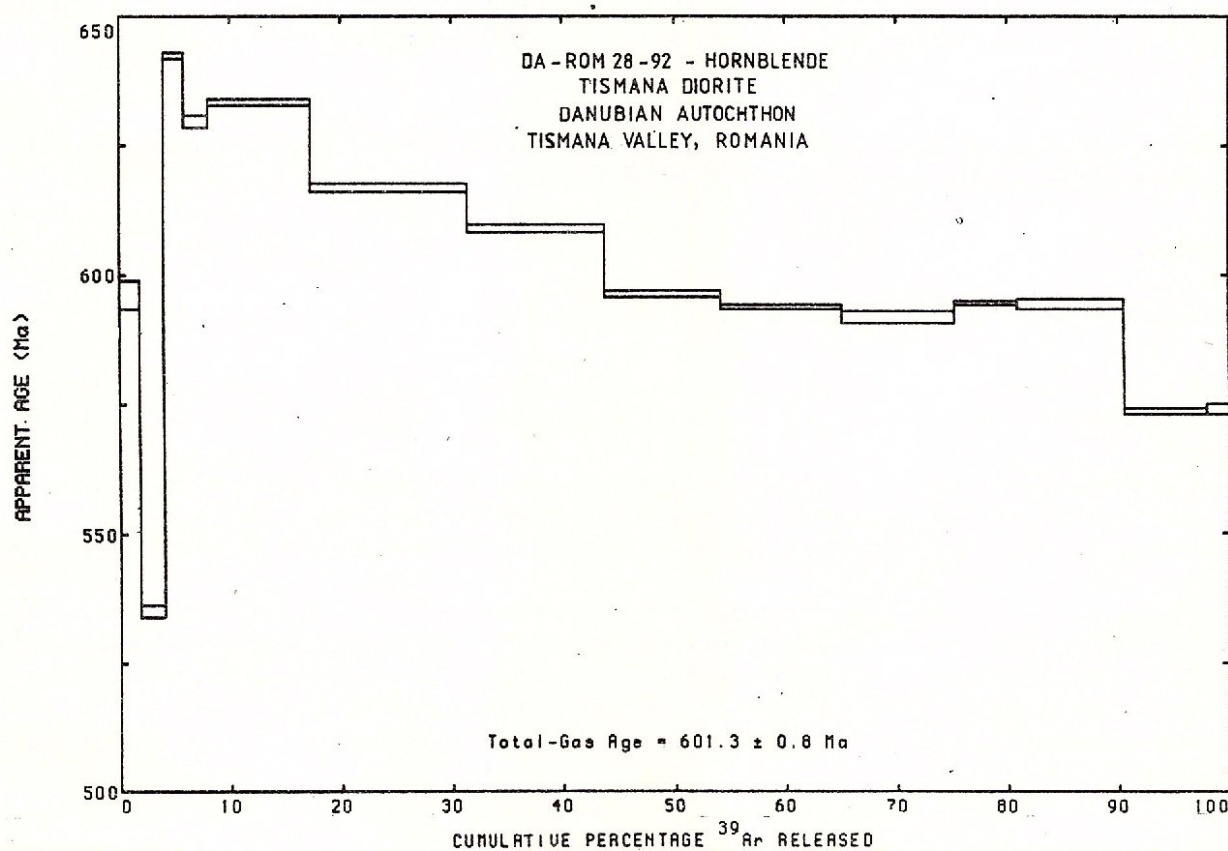


Fig. 19 - Ar-Ar diagram for amphibole fraction from a Tismana diorite, Tismana Valley (Dallmeyer, 1994).



Stop no. 10 A: Tismana granite and its Mesozoic cover.**T. Berza**

Location: right bank of the Tismana Valley, at the entrance in the in the Tismana Monastery.

Porphyritic, coarse grained Tismana granites (showing K-feldspar megacrysts), belonging to the Precambrian basement, are covered by Jurassic deposits.

The porphyritic granites consist of microcline, plagioclase (An 30), quartz, biotite and accessories. Typical Tismana granites are very coarse grained rocks, with plagioclases up to 1 cm and K feldspar megacrysts up to 10 cm.

In places, below the base of the Jurassic cover the Tismana granites preserve relics of an old weathering crust (pre-Liassic), with caolinitic alteration of feldspars and in situ bleaching of biotite, the resulting hematite films giving to the granitic rocks a red coloration (Berza, 1978).

The Mesozoic cover consists of Liassic conglomerates and arkosic sandstones, directly overlying the basement and interpreted as continental deposits (Pop, 1973). Liassic deposits are overlain by shallow water dolomites (Dogger) and limestones (Malm), representing carbonate platform deposition (Pop, 1973).

Stop no. 10 B: The Tismana granite.**T. Berza**

Location: left bank of the Tismana valley, 300 m, down stream of the Tismana Monastery.

The road-cut exposes meladiorites, related to the magmatic suite of the Tismana massif.

The rocks show massive fabric and consist of clinopyroxene, brown hornblende, plagioclase (An 40-50), biotite, interstitial quartz, apatite. Static retrograde mineral alterations include formation of chlorite-prehnite (on biotite), epidote-clinozoisite, chlorite (on pyroxenes), saussurite assemblages (on plagioclase).

Stop no. 11: Precambrian metaterrigenous sequence of the Lainici-Păiuș Group & synkinematic leucogranites; post-metamorphic dyke swarm; basement/cover relationships.

T. Berza

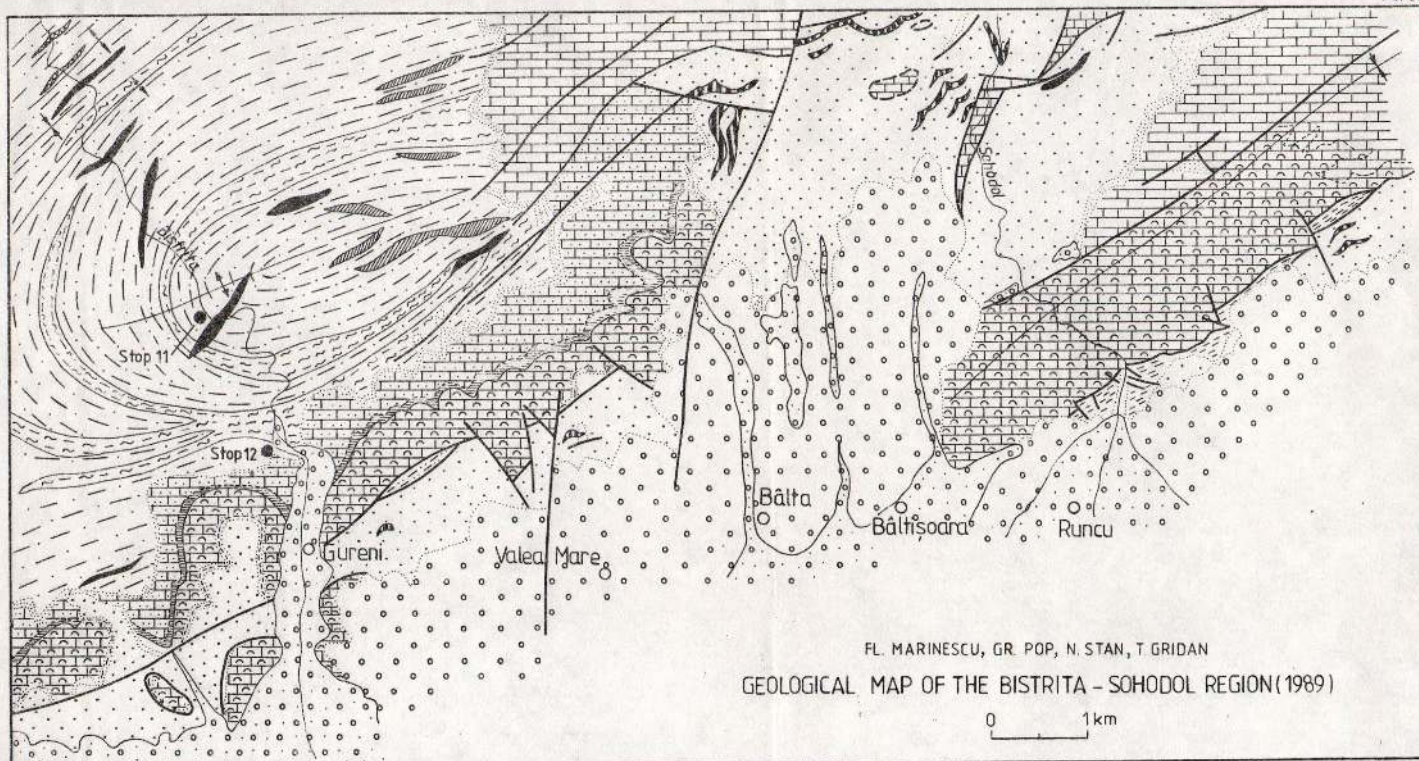
Location: Bistrița Stream, height 400 m, 500 m upstream of junction with left tributary (Pl. IV).

Along 250 m downstream, continuous outcrops in metapsamites interlayered with metapelites (Lainici-Păiuș Group) are exposed in the road-cut: biotite gneisses, amphibole-biotite gneisses, feldspar quartzites, biotite quartzites.

Garnet occurs in all rock types, usually in different stages of biotitisation. Sillimanite (highly altered to pinnite) and graphite are widespread in paragneisses. A few km eastwards, a mica gneiss layer shows a spectacular mineral association: plagioclase-quartz, biotite, centimetric sillimanite, cordierite, andalusite, almandine, muscovite, graphite.

Metasedimentary rocks are intimately penetrated by leucocratic granites and granodiorites, migmatitic leucosome with quartz-feldspar composition, pegmatitic or aplitic. The structural control of the Lainici-Păiuș Group host rocks is obvious in the development of both leucogranites and leucosome as centimetric or decimetric layers, conformable with the planar elements (S_2 metamorphic foliation) of the host (in this area, S_2 trends E-W with south-ward dips). Metamorphic foliations are cut by veins of microgranites or grey feldspar bearing pegmatites (Fig. 20 a, b, c).





FL. MARINESCU, GR. POP, N. STAN, T. GRIDAN
GEOLOGICAL MAP OF THE BISTRITA - SOHODOL REGION (1989)

0 1 km

	Post Badenian deposits		Urgonian limestones		J ₂ Detrital limestones
	Badenian gravels		J ₂ -ap Urgonian and dolomitised limestones		J ₁ Sandstones, siltstones conglomerated (Gresten facies)
	tu ₃ -sv Mélange formation		ne Micritic limestones		Pz Lower Paleozoic dyke swarms
	cm-tu ₂ Nadanova Beds		J ₂ -ne Recrystallised limestones		Lainici-Păiuș Group (Precambrian)
					PC ₃ Micashists, paragneisses
					Amphibole gneisses
					Quartzites, paragneisses

Leucogranites show pegmatitic separations and restitic garnet (included in or surrounded by biotite), muscovite, K-feldspar, plagioclase, quartz and accessories.

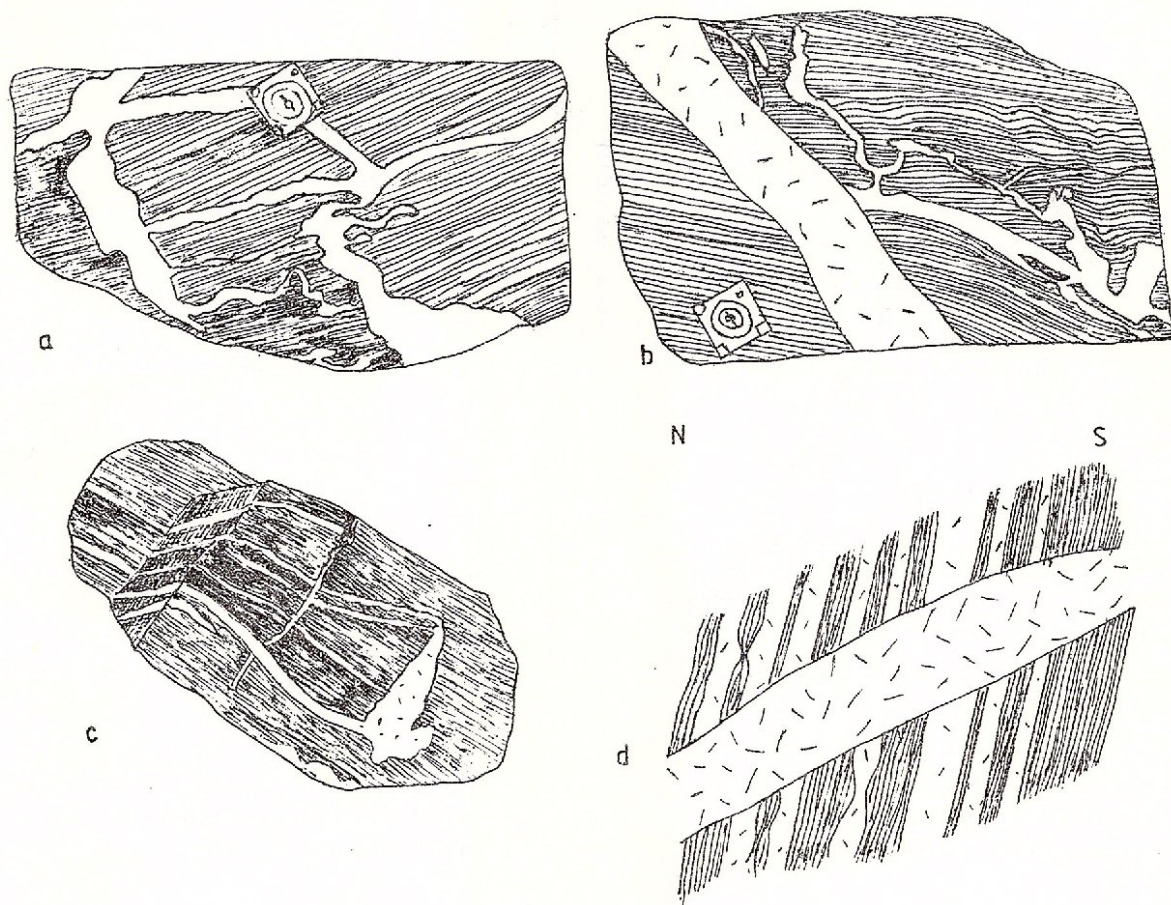


Fig. 20 - Migmatites (a, b, c) and Palaeozoic dyke (d) in a metaterrigenous sequence of Lainici-Păiuş Group (Bistrița Valley).

Further downstream, migmatized Lainici-Păiuş Group rocks are intruded by a kersantite dyke, rich in hornblende needles, which show NNE-SSW lineation, shallow plunging towards the dyke margin. (Fig. 20 d).

This is representative for countless porphyry Pre-Silurian dykes emplaced in Lainici-Păiuş rocks, migmatites or granitoid plutons, showing dioritic, granodioritic or granitic composition (Berza, Seghedi, 1975 a; Fig. 21) and a small negative Eu anomaly (Vieru et al., 1993) (Fig. 22).

All rocks described show uncomplete alterations of the primary minerals, with recrystallisation of epidote/clinozoisite, chlorite, albite.

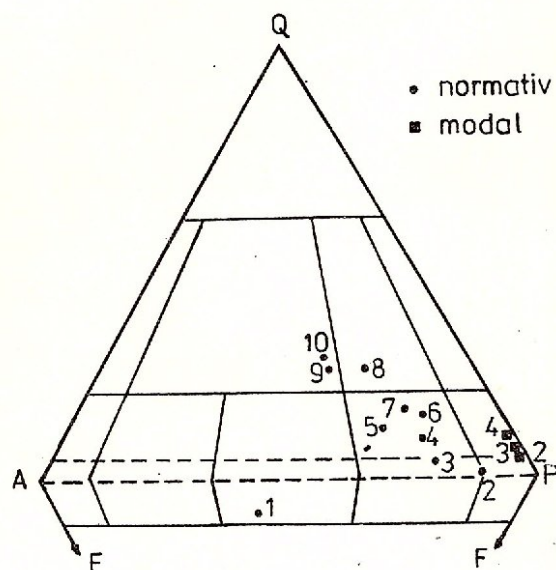


Fig. 21 - QAP normative diagram for pre-Silurian dyke rocks of the Lower Danubian (Berza, Seghedi, 1975 a).

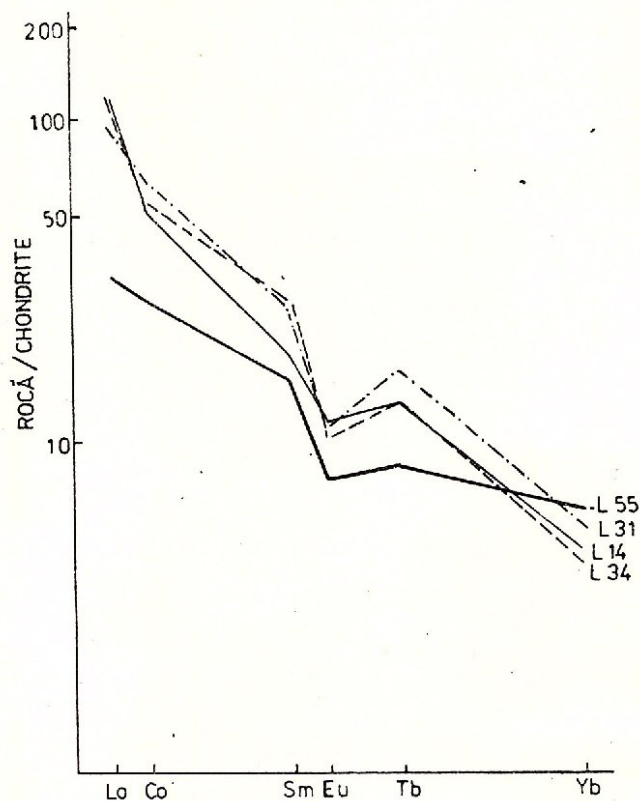


Fig. 22 - Chondrite normalized REE pattern of pre-Silurian dyke rocks (Vieru et al., 1993).

Stop no. 12: Basement/cover relationships in the Lainici Nappe.

T. Berza

Location: Right bank of Bistrița stream.

The escarpment on the right bank of the Bistrița stream exposes the structural break between the metamorphic basement and its sedimentary cover (Fig. 23). Basement rocks (migmatized Lainici-Păiuș Group quartzites and gneisses) show a planar steeply dipping, E-W trending foliation while bedding in the unconformably overlying cover rocks dips gently to the South (Fig. 24). The Mesozoic cover consists of a thin layer of Liassic sandstones (arcoses (several meters in thickness), overlain by a thick sequence of carbonate rocks (Middle Jurassic-Lower Cretaceous), Nadanova Beds (Cenomanian-Middle Turonian) and Upper Cretaceous turbidites (Pop, 1973).

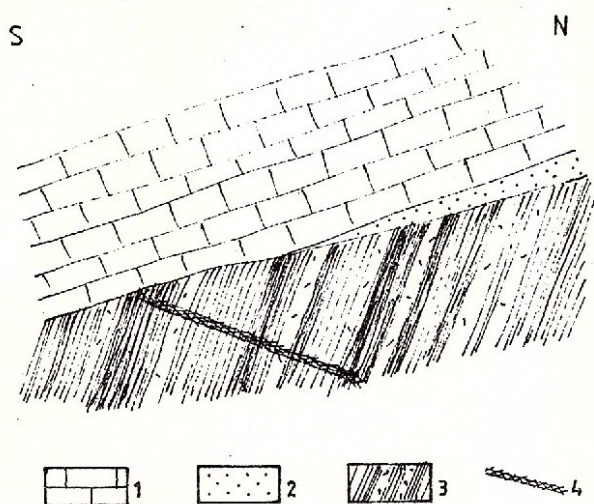


Fig. 23 - Simplified outcrop sketch showing cover-basement relationships (Bistrița valley). 1, Upper Jurassic limestone; 2, Liassic sandstone; 3, Lainici-Păiuș quartzitic paragneisses; 4, cataclases.

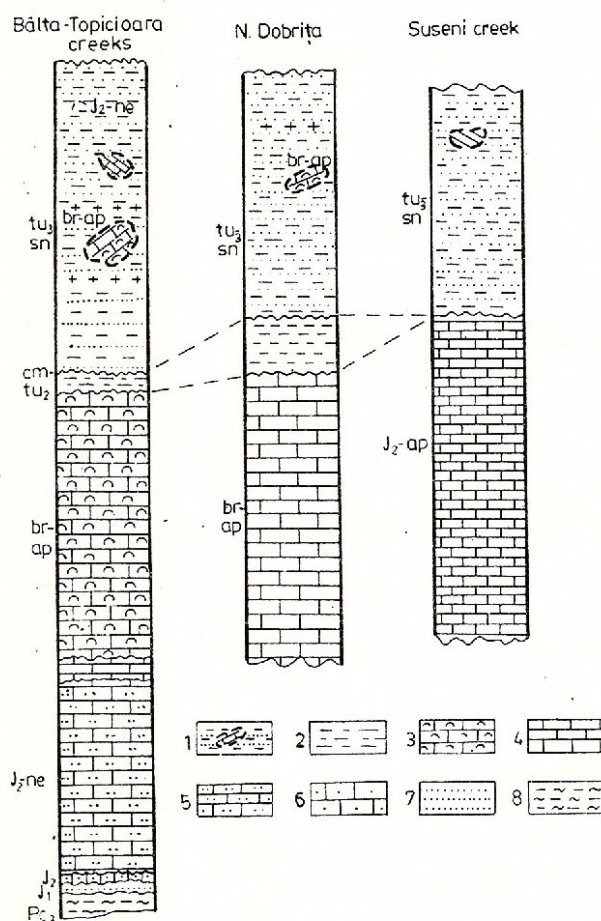


Fig. 24 - Correlation of the mesozoic cover of the Lainici Nappe (South Vâlcan Mountains) (Pop. in Marinescu et al., 1989). 1, mélangé formation, a, olistolith; 2, marls, shales; 3, Urgonian limestones; 4, recrystallised Urgonian limestones; 5, recrystallised micritic limestones; 6, detrital limestones; 7, sandstones; 8, metamorphic basement.

Stop no. 13: Upper Cretaceous cover rocks of the Lămlci Nappe, showing low temperature metamorphism in the prehnite-pumpellyite facies

A. Seghedi

Location: Sohodol valley, height 420 m, 100 m upstream of the touristic lodge.

Upper Cretaceous (Upper Turonian - Senonian) deposits exposed in the road-cuts along the Sohodol Valley include microconglomerates, sandstones, siltstones, interlayered with volcanoclastic and volcanic rocks; various blocks of carbonate rocks of different ages occur within this sequence, suggesting a mélangé origin (Pop, 1973, Pop, in Marinescu et al., 1989). In this area, Upper Cretaceous deposits are tectonically bounded, both to the North and to the East, along normal faults (Pop, in Marinescu et al., 1989), developed at the contact with Egeonian limestones.

The Sohodol Valley exposes sandstone dominated volcanoclastic turbidites, forming an upward coarsening sequence. Volcanic rocks described in this sequence include island arc basalt tuffs and basaltic andesites (Savu et al., 1987) (Fig. 25). Their geochemistry suggests derivation from a calc-alkaline magma, originating through melting of the Carpathian oceanic crust during subduction; this origin is suggested by $\text{Sr}^{87}/\text{Sr}^{86}$ values ranging between 0.704-0.708; REE abundance indicate a depleted upper mantle source (Savu et al., 1987).

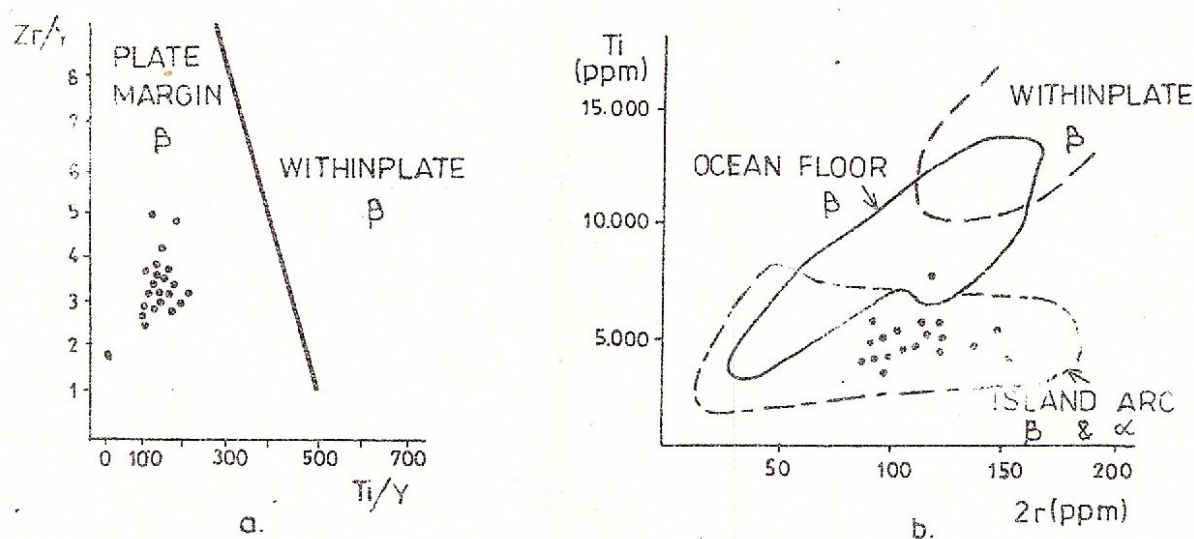


Fig. 25 - a: Zr/Y-Ti/Y discrimination diagram for the tectonic interpretation of Upper Cretaceous volcanics (Savu et al., 1987); b: Ti-Zr discrimination diagram for the tectonic interpretation of Upper Cretaceous volcanics (Savu et al., 1987)

The green sandstones cropping out in the right bank of the Sohodol Valley, 100 m upstream the touristic lodge, made up a 10 m thick sequence; individual sandstone beds attain 2-3 m in thickness and suggest channel deposits (facies B of Mutti, Ricci-Lucchi, 1972); finer grained interbeds (facies D, consisting of Tcd or Tde Bouma divisions) vary in thickness from cm to dm and are interpreted as overbank (interchannel) deposits.

Facies B sandstones are massive or graded rocks with coarse to fine grain sizes some beds are rich in mudstone intraclasts or limestone clasts dispersed at their base. Their mineralogy indicates a major volcanic source (probably a volcanic arc), supplying feldspar, pyroxene, brown hornblende, quartz, porphyritic basalt and andesite clasts, with a minor terrigenous source, providing quartzites, detrital micas, garnet and apatite. Sandstones are mainly clast supported rocks; carbonate cement seldom occurs in small amounts. The low-temperature Alpine metamorphism is in prehnite-pumpellyite facies; prehnite occurs as patches formed on plagioclase feldspar, while pumpellyite is widespread both as large, individual grains, completely or partly replacing detrital micas, or slightly replacing feldspars and amphiboles.

Stop no. 14: Normal fault at the junction between the pre-alpine basement (Lainici-Păiuș Group) and cover rocks (Urgonian limestones).

T. Berza

Location: upstream of the Sohodol Gorges.

Downstream of stop no. 13, the river crosses an Alpine anticline with Lainici-Păiuș Group rocks in its core. A NE-SW trending normal fault develops along the southern limb of the anticline, exposing the Mesozoic carbonate cover which consists of bioclastic, peletal, micritic and biolithitic limestones in Urgonian facies (Barremian-Aptian, Pop, 1989).

Upstream the Lainici-Păiuș metamorphic rocks are exposed, consisting of banded quartzites associated to micrograined granitoids (plagioclites). Feldspathic and biotitic quartzites prevail, containing muscovite, garnet, graphite. They display a lineation of metamorphic biotite, as well as a discrete lineation of biotite + muscovite lenses with annealed fabric.

Stop no. 15: Stratigraphic contact between Nadanova Beds overlaying Urgonian limestones.

A. Seghedi

Location: Downstream of the Sohodol Gorges.

Both banks of the Sohodol Valley show Urgonian limestone banks with a penetrative, steeply dipping cleavage, overlaid by a 20 m thin, flatlying, cover of Nadanova Beds. There is a stratigraphic gap between these formations, corresponding to the Albian (Fig. 24) (Pop, 1973). Nadanova Beds consist of shales, siltstones and marls, showing yellowish-grey colours due to weathering. Rocks are rich in fossil remains. Uphill towards East, Nadanova Beds are overlain by Upper Cretaceous mélange formation; to the South they are covered by Miocene gravels of the post-tectonic cover.

Third day

Introduction

This day will present a complete S-N cross section in the South Carpathians between the Getic Basin and Mureș Valley, from the lowest to the highest tectonic units. Meanwhile, a variety of metamorphic, igneous, sedimentary or very low-grade metamorphic formations will be introduced, in the Jiu, Strei and Cerna (of Hunedoara) valleys. In order to reach the last stop by day-time, participants are asked to conform to the indicated time table. So 5 structural stops of one hour each and 5 petrographic stops of a quarter of hour each can be accommodated in an 150 Km long journey by daylight.

The first seven stops are located in the Jiu Gorges, the main target of the day. They show basements and covers of three Danubian Nappes: Schela, Lainici and Urdele. In the Lainici Nappe, pre-Alpine tectonic contact of two contrasting metamorphic basements will be presented (Fig. 26). The Jiu Gorges represent a deep trench separating the Vălcău Mountains (to the West) from the Parang Mountains (to the East). At the end of 19th century a modern road was built along the river replacing the old roman road. Because of high traffic, participants to the excursion are asked to stay out of the road when looking to outcrops and watch well when they cross it. Getic basement and cover will be presented in the next two stops, and the last one will introduce Paleozoic rocks from Supragetic Units.



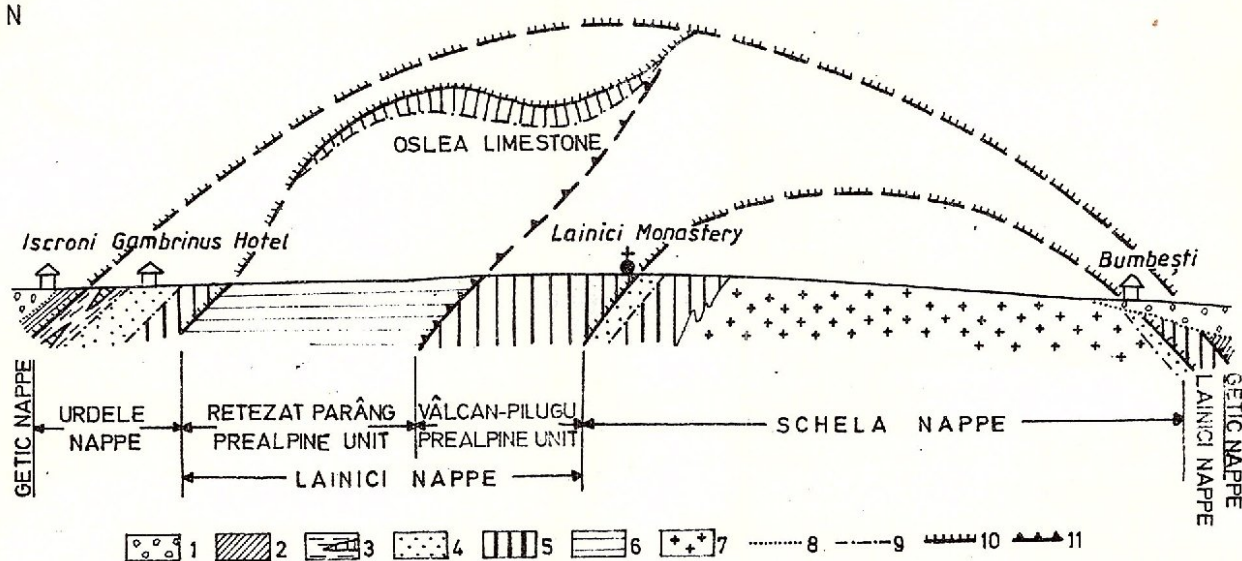


Fig. 26 Schematic cross-section along the Jiu Gorges (Berza, Drăgănescu, in Balintoni et al., 1989) 1. Tertiary deposits; 2, Sebeș-Lotru rocks; 3, Upper Cretaceous Flysch with Lupeni limestone klippen; 4, Schela Formation; 5, Lainici-Păiș rocks; 6, Drăgășan amphibolites; 7, Șușița Pluton; 8, unconformity; 9, "décollement" stratigraphic boundary; 10, Alpine overthrust; 11, pre-Alpine overthrust.

Stop no. 16 (normal stop): Schela Formation in Schela Nappe.

T. Berza

Location: Porceni quarry, 1 Km north-west of Bumbești, at the southern end of Jiu Gorges.

The first 10 kilometers going upstream in the Jiu Gorges expose within an asymmetrical antiform the lowest tectonic unit known in the South Carpathians - the Schela Nappe. Stop no. 16 is located on the southern limb of the antiform, showing, from north to south, Șușița granitoids, Schela Formation and again Șușița granitoids.

The Șușița granitoid is the largest pluton from the South Carpathians, cropping out within tectonically or sedimentary boundaries on 50 km in length, 10 km in width and on 2 km altitude difference.

The Șușița pluton consists of biotite granites and granodiorites, here massive, but intensely fractured and affected by pre-Liassic weathering (bleaching of biotite, kaolinisation of feldspars).

Schela Formation overlies the granites with a 30–50° southward dip. The sequence starts with a 2 m thick conglomerate bed, grading upwards into microconglomerate, then to black quartzitic sandstones, typical for the Schela Formation. Conglomerates are rich in centimetric, rounded, lithoclasts, suggesting a source area formed by Lainici-Păiș quartzites, granites and vein quartz. On the left bank of Porcului Valley, opposite from the quarry, pelitic interbeds in sandstones contain Lower Jurassic fossil flora, known from Manolescu (1937) and studied recently by Stănoiu et al. (1992). The upper part of the sequence is constituted by chloritoid-pyrophyllite bearing slates with anthracite boudins, evidenced by electrometric prospections (Constantinescu, in Berza et al., 1989).

To the South, the Schela Formation plunges with moderate dips below the Șușița granites; this more probably indicates reverse the limb of a north-vergent fold than a scale structure (Berza, Drăgănescu, in Berza et al., 1989).

Stop no. 17 (short stop): Șușița granitoids.

T. Berza

Location: Km 97 on national road no. 66.



This stop is located in the axial zone of the nappe antiform, presenting Șușița granitoid rocks; chemical composition of biotite granites and biotite-hornblende granodiorites and tonalites was investigated by Savu et al. (1971). K-Ar ages of Șușița granitoids range between 547 and 73 Ma (Grünenfelder et al., 1983) reflecting increasing degree of Alpine overprinting; the youngest age is Laramian, yielded by a mylonitised Șușița granite in the footwall of the Lainici Nappe thrust.

Stop no. 18 (normal stop): Lainici-Păiuș gneisses and Schela Formation in Schela Nappe, on the northern limb of the antiform.

T. Berza

Location: Rafaila-Lainici, km 104 on national road no. 66.

Between stops no. 17 and 18, Șușița granitoids become increasingly more foliated and altered, finally being transformed into quartz-albite-K feldspar-muscovite-chlorite-stilpnomelane-epidote mylonites. At km 103,700, the Șușița pluton terminates and its metamorphic horst rocks - the Lainici-Păiuș Group - are exposed on 800 m in the road-cut (Fig. 27). Quartzites \pm plagioclase \pm biotite \pm muscovite \pm garnet \pm diopside are the main rock type, but micagneisses \pm sillimanite are also frequent. Sillimanite prisms may reach 2 cm in length and show a strong horizontal E-W trending lineation. Savu (1970) considered that sillimanite formed consequently to the temperature rise in the vicinity of the Șușița Pluton; Berza (1978) revealed the regional extension of the sillimanite zone on the whole outcrop area of the Lainici-Păiuș Group. Typical leucogranitic veins invade the metamorphic rocks and augen migmatites are also found in the gneisses. Zircons from this augen gneiss yielded a Concordia diagram with an upper interception at 610 ± 30 Ma (Fig. 28) (Grünenfelder et al., 1983).

Dykes of porphyritic microdiorites, emplaced in the Lainici-Păiuș metamorphic rocks and migmatites, show a pervasive Alpine mylonitisation. They belong to the same magmatic province as the dyke swarms in Motru-Tismana area, emplaced subsequently to granitic batholiths but prior to the Silurian, as they do not intrude the Silurian Valca Izvorului Formation.

At Rafaila Cross - a monument raised in the memory of a first world war hero - the cover of the Schela Nappe is exposed in the Jiu Valley. The left bank of the river shows mostly Quaternary deposits at the mouth of a left tributary, with huge blocks of sandstones and conglomerates and smaller debris of pyrophyllite-chloritoid slates. Famous in the area, this outcrop was first described a century ago by Duparc, Mrazec (1893); since then it was subject of debate between partisans of the Lower Jurassic or the Upper Carboniferous age of the rocks. However, the sequence continues to the West at Porceni, where a Lower Jurassic age is well documented by plant remains (Manolăscu, 1973).

In the right bank of the Jiu Valley, two stripes of similar chloritoid bearing rocks outcrop repeatedly with Lainici-Păiuș rocks; they represent isoclinal cover/basement folds (Berza et al., 1989), or scales (Savu et al., 1984 b). The last strip of Schela rocks is overlain by a 100 m thick sheet of Precambrian granite rocks, covered by white arkosian sandstones (10 m) (Lower Jurassic) and 200 m of recrystallised limestones (Upper Jurassic-Lower Cretaceous). The thrust plane in the footwall of the granites continues eastwards (Berza et al., 1986 b) and westwards (Berza et al., 1989) of the Jiu Valley as the thrust fault of the Lainici Nappe on to the Schela Nappe. The line is crossing the Jiu at km 106, at Lainici Monastery, built on a terrace in the XVIII-th century.

Stop no. 19 (short stop): Lainici - Păiuș Group in Lainici Nappe.

T. Berza

Location: km 111 on national road 66.

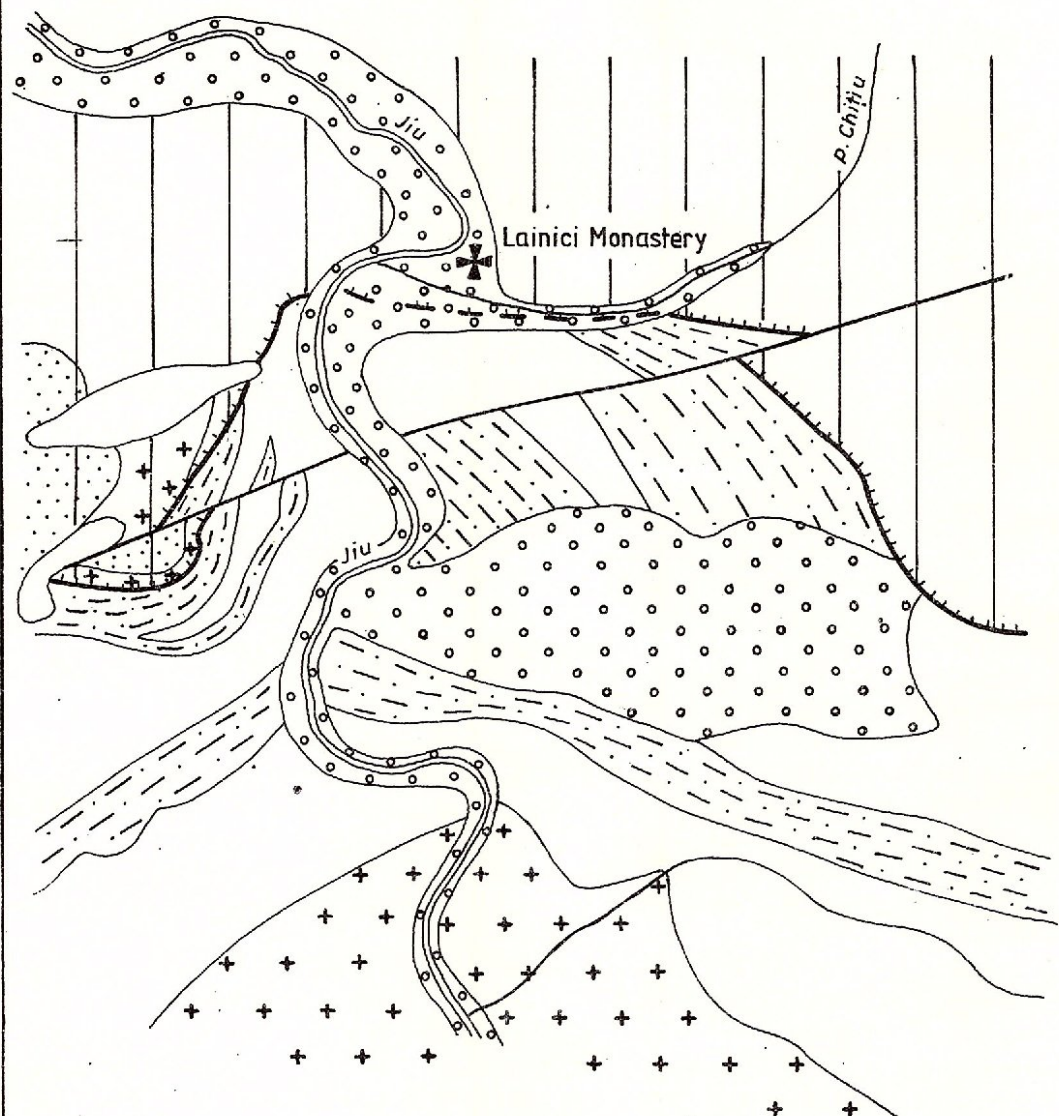
From Lainici Monastery (km 106) to Cărligul Caprei Bridge (km 115.500) the Jiu Gorges are carved in Lainici-Păiuș Group (5 km on N-S direction), exposing its main divisions: Quartzitic Formation (km 106-km 114.500) and Carbonato-Graphitic Formation (km 114.500-km 115.500). Leucogranitic injection and various migmatites are widespread in both formations (Stan, 1977), and porphyritic microdiorites cut them frequently. The general aspect of the rocks is mylonitic, with strong and pervasive mylonitic foliation and retrograde greenschist facies recrystallisation. The southern border is Alpine (the floor-thrust of the Lainici Nappe), but



GEOLOGIC SKETCH-JIU VALLEY AT LAINICI

T. BERZA, A. DRĂGĂNESCU

0 500 1km



LEGEND

- | | | |
|-------------------|----------------------|--|
| | | Quaternary deposits |
| | Lainici Nappe | |
| JURASIC | | Limestones |
| UPPER PROTEROZOIC | | a. Lainici-Păiuș Group
b. Granitoid intrusion |
| | Schela Nappe | |
| LOWER JURASIC | | Schela Formation |
| UPPER PROTEROZOIC | | a. Lainici-Păiuș Group
b. Șușița granitoid pluton |

Fig. 28



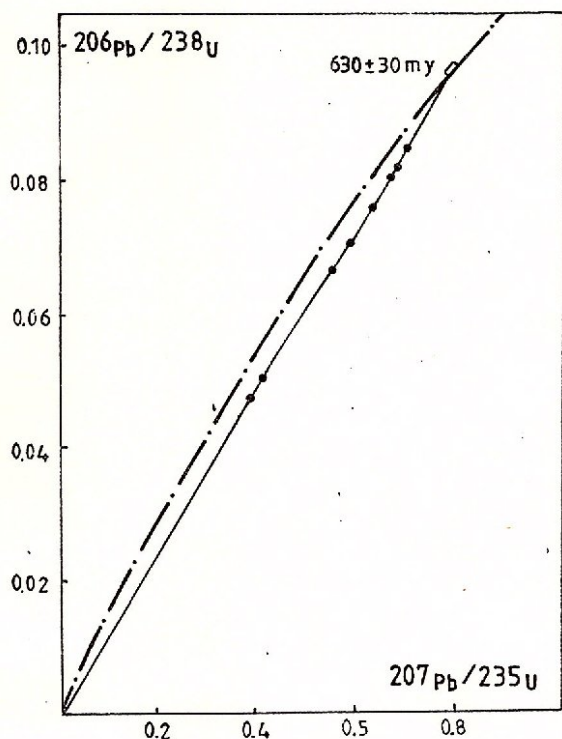


Fig. 28 – Concordia diagram for an augen migmatite from Jiu Valley, km 104 on national road 66 (Grünenfelder et al., 1983)

the northern border is pre-Alpine (stop no. 20); in between, an overprinting of Variscan mylonites by Alpine mylonitisation resulted in the "epimetamorphic" look of the rocks.

At stop no. 19, a classic outcrop of mylonitised quartzites and leucogranites with several porphyritic dykes, show the prevailing North dipping of the foliation.

Relict mineral assemblages of an early (Precambrian, high T-low P) metamorphism, typical for the Lainici-Păiuș Group rocks, are preserved in places (with andesine, garnet, hornblende, diopside, sillimanite, andalusite, cordierite) (Savu, 1970; Berza, Seghedi, 1983), strongly overprinted by later, greenschist facies, retrogression (albite, chlorite, tremolite, epidote, stilpnomelane). Higher in the left bank of the Jiu Gorges, some marble layers show a calcite + diopside + phlogopite + graphite assemblage, typical for Lainici-Păiuș Group carbonate rocks.

Stop no. 20 (normal stop): Pre-Alpine tectonic contact between two distinct basements (Lainici-Păiuș and Drăgșan Groups, respectively) of the Lainici-Nappe.

T. Berza

Location: Cărligul Caprei Bridge, km 115.500 on national road 66.

The tectonic contact exposed here was first described by Manolescu (1937) and figured as pre-Alpine, because on both banks of the Jiu River it is sealed by Mesozoic limestones.

Coming from the South, on the last hundreds of meters the road exposes amphibolites, micagneisses and silicate marbles with the ubiquitous leucogranite injections; the entire sequence is strongly sheared, with a steep foliation and E-W trending horizontal lineation.

Near the bridge, contrasting mechanical properties of carbonates and silicates and pervasive deformation produced black (amphibolite) or white (leucogranite) boudins in the marbles, resulting a (pseudo) conglomeratic aspect with carbonatic matrix.

The Carbonato-Graphitic Formation of Lainici-Păiuș Group ends 100 m north of the bridge, Drăgșan amphibolites outcropping upstream. Here the retrogression is limited to fracture zones, frequently with quartz veins, and after several hundreds of metres the South dipping banding of the amphibolites is dominant.

The first outcrops of Drăgșan Group are fine to large grained, massive garnet amphibolites (metagabbros), followed by serpentinite (metaperidotite).

Stop no. 21 (short stop): Drăgșan Group banded leptyno-amphibolites in the basement of the Lainici



Nappe.

T. Berza

Location: km 118.500 on national road 66.

This is a typical outcrop for the Amphibolitic Formation of the Drăgșan Group. Typical banded leptyn-amphibolites with milimetric to centimetric contrasting black-white layers dip 45° to SE. All proportions between 100 % hornblende and amphibole free oligoclase + quartz layers can be found. Biotite, garnet and muscovite are frequent. ICP-MS analyses of 8 samples, from dark to white, were performed by Dr. J. P. Liégeois from Brussels Free University (Table 1).

The origin of this rock association is still unclear, but metamorphic differentiation has surely operated. Savu et al. (1984 a, 1990) proposed an ocean floor origin for the basalts (now in amphibolitic facies) of the Drăgșan Group. Another question concerns the origin of the white layers, whose low K_2O content makes difficult to found either an acid volcanic or a sedimentary protolith. There are no similar formations in the Romanian Carpathians, but they may be correlated to the Leptyno-Amphibolitic Complex (LAC) described by Ilvorka et al. (1993) in the basements of the Tatrides, Veporides and Gemerides from the Slovakian Carpathians.

Scarse layers of micagneisses occur in the amphibolites. On a left tributary (Polatiște Valley), kyanite-almandine micagneisses crop out (Solomon, 1986) and on a right tributary of the Motru River kyanite-staurolite-almandine gneisses are interlayered with the amphibolites (Berza, Seghedi, 1975 b).

Contrasting with the Lainici-Păiuș Group, granitic injections are very scarce in the Drăgșan Group. However, in Retezat Mts. a kilometric body of augen gneisses with amphibolitic bands may represent a premetamorphic granitic pluton.

Stop no. 22 (normal stop): Tectonic contact between Urdele and Lainici Nappes.

T. Berza

Location: km 123 on national road 66.

Upstream of stop no. 21, banding in the Drăgșan amphibolites changes gradually to eastern dips, and then to northern dips, marking the periclinal closure of a large scale antiform refolding the early foliation.

At the northern end of Jiu Gorges, this medium-grade metamorphic banding is overprinted by a low-grade schistosity, with similar trend but with different dip; in places transposition of the banding, by gliding on closely spaced slip planes, can be observed. This is accompanied by retrogression, amphibolite facies assemblages tending to be obliterated by albite-epidote-chlorite-tremolite associations. Longtime considered a younger low-grade metamorphic formation, this zone has proven to be the result of greenschist facies retrogression of previously higher-grade rocks (Berza, 1975). Recently this retrogressive event was dated as Alpine, based on reconsideration of the cover ages (Mesozoic age for Oslea Limestone, previously considered Paleozoic) and on the reinterpretation of the so called "conglomeratic member" (in fact basement phyllonites), both overling Drăgșan retrogressed amphibolites (Berza et al., 1988 b).

This last stop in the Jiu Gorges presents the northern end outcrops of the retrogressed amphibolites; these mylonites, showing a pervasive North 45° dipping mylonitic foliation and quartz veins or boudins in a chloritic matrix, are known for long time as green "conglomerates" (Manolescu, 1937). They are overlain by grey "conglomerates", in fact mylonites with quartzitic matrix and tectonic lenses (boudins) of leucogranites, typical for the migmatites invading Lainici-Păiuș Group rocks. A few hundred meters above this outcrop, in both banks of the Jiu River, a thin layer of recrystallised limestones (Oslea Limestone) is interposed between the green and grey rocks. This geometry is followed 30 km westwards, in Vâlcan Mts., and eastwards in Parâng Mts.; in places, the Oslea Limestone has a few meters of arkosic sandstones or conglomerates at its sole.

The sequence including arkosic sandstones, Oslea Limestone, grey "conglomerate" and overlying black chloritoid rocks (Schela Formation) was considered to represent Triassic and Liassic (Manolescu, 1937) or Carboniferous (Pavelescu, Pavelescu, (1964) formations. Demonstrating that the grey "conglomerates" are in fact greenschist facies mylonites formed on Lainici-Păiuș migmatized basement, Berza et al. (1986 b, 1988 b) documented in Northern Parang and Vâlcan Mts. an Alpine nappe foreseen by Popescu-Voitești (1923) - the Urdele Nappe.



Table 1

Chemical analyses of Drăgsan amphibolites and leptynites (J. P. Liégeois, Brussels, unpubl. data, 1994)

DRAGSAN (vallée du Jiu)								
Sample	R117	R118	R119	R120A	R120B	R120C	R121	R122
SiO ₂	47.57	51.14	49.1	48.96	61.51	47.24	53.64	66.6
TiO ₂	1.75	2.65	1.8	2.73	0.89	1.5	1.52	0.31
Al ₂ O ₃	16.18	13.82	16.89	14.48	17.05	16.09	17.17	17.55
Fe ₂ O ₃	3.96	5.3	3.49	5.87	2.75	4.13	4.25	1.77
FeO	7.49	8.09	7.07	7.83	3.76	7.07	5.42	1.02
MnO	0.18	0.25	0.22	0.25	0.09	0.16	0.2	0.07
MgO	6.87	4.32	6.27	5.27	2.6	8.42	3.97	0.87
CaO	9.03	7.73	7.42	8.15	3.51	7.82	6.99	3.65
Na ₂ O	2.89	2.24	2.15	1.81	3.95	2.62	3.56	4.69
K ₂ O	0.57	0.85	1.22	1.09	1.07	1.03	0.59	1.14
P ₂ O ₅	0.26	0.37	0.37	0.47	0.15	0.26	0.33	0.14
PF	2.58	2.32	3.03	2.18	2.55	2.97	2.2	1.46
Total	99.33	99.08	99.03	99.09	99.88	99.31	99.84	99.27
Fe ₂ O ₃ t	12.28	14.28	11.34	14.56	6.92	11.98	10.26	2.91
V	349.12	398.64	281.63	428.76	140.90	273.45	213.11	45.32
Rb	14.09	20.34	29.39	22.13	22.41	20.76	12.46	22.13
Y	27.11	45.39	27.00	44.32	26.22	26.07	31.02	13.34
Zr	109.73	188.28	120.96	193.23	184.07	82.79	96.73	148.24
Nb	4.88	7.83	5.62	9.18	5.38	3.54	3.83	2.85
Ba	102.26	162.76	205.02	178.90	206.15	184.51	164.64	310.07
La	8.36	13.96	10.68	14.68	15.83	6.10	10.40	13.61
Ce	24.17	40.10	29.94	42.79	40.67	18.69	27.44	32.34
Pr	3.57	5.76	4.33	6.15	5.27	2.77	3.98	3.94
Nd	17.19	28.38	20.32	30.04	23.65	14.89	19.82	15.58
Sm	4.39	7.28	4.90	7.65	5.54	4.12	5.18	2.97
Eu	1.67	2.42	1.83	2.69	1.54	1.47	1.78	1.02
Gd	5.06	8.19	5.37	8.63	5.70	4.69	5.67	2.85
Dy	4.56	7.60	4.63	7.64	4.91	4.44	5.43	2.27
Ho	0.93	1.61	0.93	1.80	1.00	0.96	1.12	0.47
Er	2.55	4.29	2.56	4.26	2.75	2.40	3.17	1.46
Yb	2.25	4.11	2.15	3.90	2.88	2.38	3.00	1.62
Lu	0.32	0.58	0.30	0.56	0.43	0.30	0.42	0.27
Hf	2.54	4.60	2.76	4.54	4.81	2.17	2.51	3.53
Ta	0.20	0.39	0.26	0.46	0.31	0.07	0.14	0.11
W	1.11	0.62	1.64	2.18	1.32	0.98	2.13	1.09
Pb	3.04	5.40	3.65	3.82	5.91	3.41	4.72	4.99
Th	0.35	0.93	0.50	0.64	2.56	0.24	0.90	2.61
U	0.18	0.45	0.36	0.35	0.69	0.20	0.45	0.93
Sr	251.10	212.00	186.10	221.80	335.30	160.40	358.20	399.30
Sample	R117	R118	R119	R120A	R120B	R120C	R121	R122

The Lower Jurassic age of the arkosic sandstones and the Upper Jurassic-Lower Cretaceous age of the Oslea Limestone could be estimated by the perfect correlation with detrital and carbonatic deposits from the southern slope of the Vâlcân-Parâng Mts. (Berza et al., 1988 b).

East of the Jiu Gorges, the basement of the Urdele Nappe changes to green, mylonitised amphibolites, gneisses, serpentinites, cipolines, indicating that a Drăgăsan type basement is also present in this unit.

In the right bank of the Jiu Valley, the grey mylonitic pseudoconglomerates crop out only on several meters, quaternary deposits hiding their contact with the overlying cover. On the left bank, 100 m upstream from the floor thrust, the mylonitised Lainici-Păiuş migmatized quartzites are overlain by a sequence beginning with 10–20 m thick conglomerates, followed by black sandstones and slates \pm chloritoid \pm pyrophyllite, typical for the Lower Jurassic Schela Formation (Gresten type Liasic); in this location rocks bear fossil flora and fauna (Stănoiu, in Kräutner et al. 1981). This formations can reach 200 m in thickness and usually is overlain in its turn by a 10–100 m thick carbonatic formation - Lupeni Limestone, of unquestioned Upper Jurassic-Lower Cretaceous age. At the top of the cover sequence, turbidites with large hectometric slices of basement, Lower Jurassic sandstones or Upper Jurassic-Lower Cretaceous limestones (Iseroni Flysch) outcrop on 100–200 m, supporting directly Sebeş-Lotru metamorphic rocks from the basement of the Getic Nappe. In places Upper Cretaceous Gosau facies deposits cover and rework Sebeş-Lotru crystalline schists. Then follows Petroşani Tertiary basin with red Aquitanian conglomerates and coal bearing Chattian sandstones. All the planes: thrusts of Urdele and Getic Nappes, cover/basement or cover/cover limits, even coal beds in Petroşani Basin, are E-W oriented with 30°–50° northward dip, showing the recent folding of the nappe pile, with a synform at Petroşani and an antiformal between Lainici and Bumbăşti (Fig. 27).

Stop no. 23 (short stop): Sedimentary cover of the Getic Nappe.

T. Berza

Location: Livadia Bridge on national road 66.

If the Danubian covers already seen show penetrative Alpine deformation (slaty or crenulation cleavages, mylonitic foliations) and very low temperature minerals (chloritoid, pyrophyllite, meta-anthracite, illite, chlorite, prehnite, pumpellyite), the covers of the Getic-Supragetic Nappes are not at all affected by such phenomena, sedimentary and paleontologic records being well preserved. Between Băniţa and Livadia, the national road 66 crosses the Getic (Austrian) Nappe from the Lotru-Bistra (Laramian) Nappe, represented by Sebeş-Lotru Group basement and Lower Jurassic-Lower Cretaceous cover. Aalenian-Bathonian sandstones and bioclastic limestones are followed by bioclastic limestones with siliceous deposits (Lower-Middle Callovian) and Upper Jurassic-Aptian pelletal and micritic Urgonian limestones (Pop, in Berza et al., 1986 a). Upper Cretaceous deposits (Albian to Lower Maastrichtian) from Pui region have to be considered the post-Austrian cover from Lotru-Bistra (Laramian) Nappe.

Stop no. 24 (normal stop): Getic and Upper Danubian thrusts in Retezat Mts.

T. Berza

Location: Râu Bărbat Valley, 6 km South of Pui, a village located on national road 66, between Petroşani and Haţeg.

This stop is similar to stop no. 22, but here the Oslea Limestone is preserved. From South to North, Lainici, Măru (= Urdele) and Getic Nappes are crossed (Fig. 29). Retezat granitoids are epidote bearing leucogranodiorites (Berza et al., 1994) here recrystallised to quartz-albite-muscovite-chlorite-epidote greenschist facies mylonites. Mylonites are overlain by a few meters thick layer of foliated limestones - Oslea Limestone; westwards and eastwards from Râu Bărbat, sandstones, conglomerates and chloritoid-pyrophyllite bearing slates are interposed between the Retezat granite and Oslea Limestone.

The granitoids represent the basement and the sandstones, slates and limestones represent the Mesozoic cover of the Lainici Nappe.



Fig. 29 - Profile in the left bank of the Râu Bărbat valley, 1 km upstream Hobița (Berza, in Kräutner et al., 1981).

1, Sebeș-Lotru metamorphic rocks; 2, Retrogressive metamorphic rocks of the Drăgășan Group; 3-4, Mesozoic rocks: 3, black slates; 4, crystalline limestones; 5, Foliated Retezat Granite; 6, Trace of the Getic Nappe overthrust; 7, Trace of the Upper Danubian Măru Nappe overthrust.

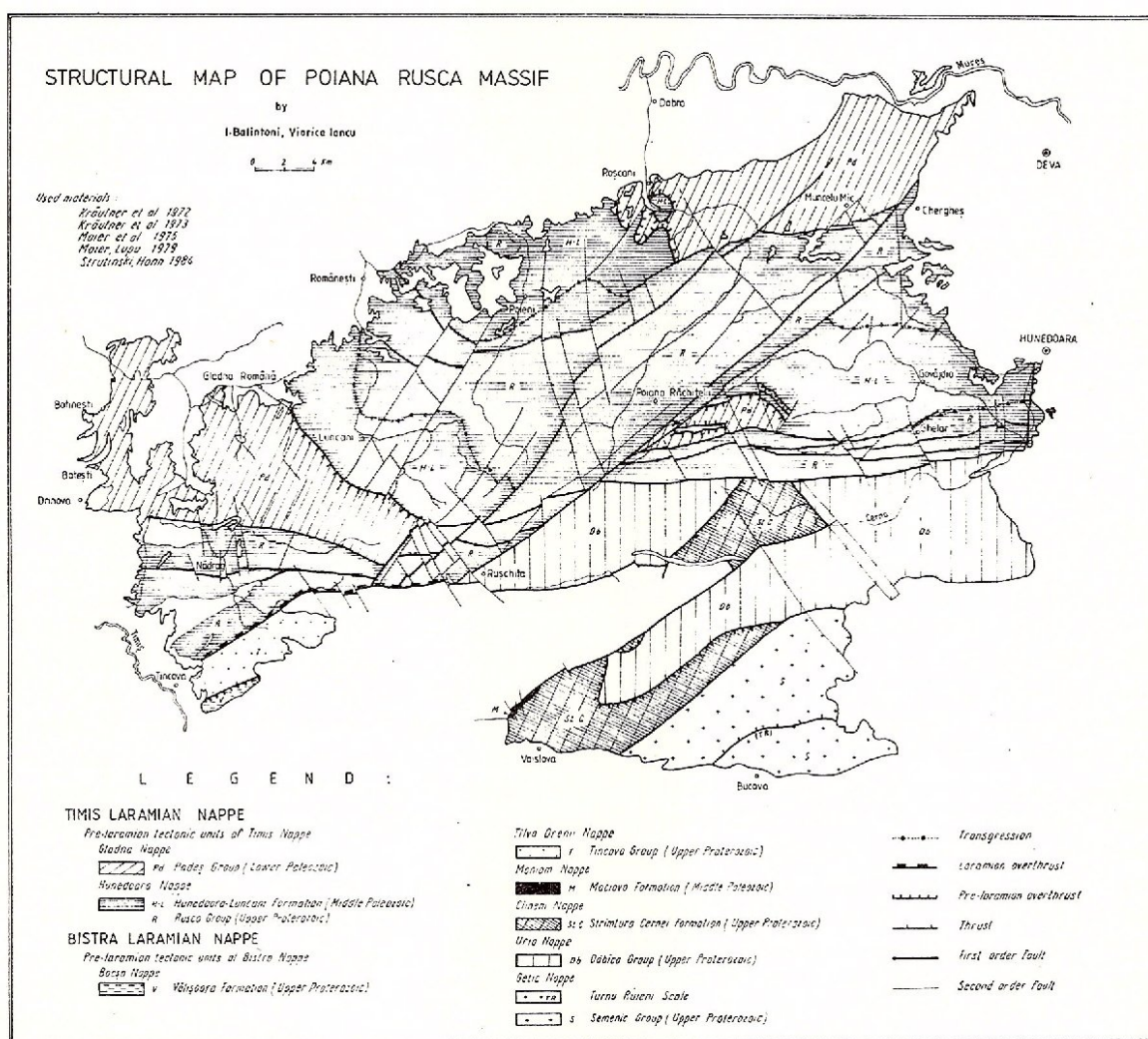
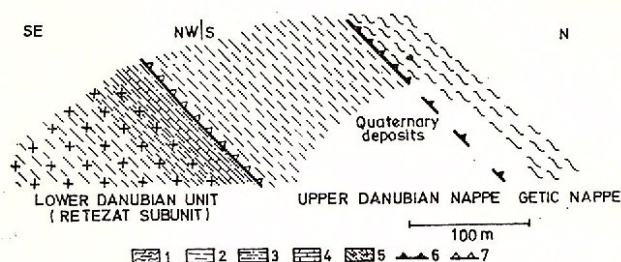


Fig. 30 – Structural map of the Poiana Ruscă Massif

They are overlaid by chlorite mylonites preserving amphibole porphyroclasts, followed by fish mylonites with garnet and staurolite porphyroclasts. These mylonites represent Drăgșan type (previously described here as Zeicani Series) basement of the Măru (=Urdele) Nappe (Berza et al., 1988 a).

Northwards, Sebeș-Lotru gneisses representing the basement of the Getic Nappe crop out on 2-4 km, covered by deposits of the Hațeg Basin. Again, all the planes dip North 30° - 40° , showing Tertiary folding of the Cretaceous nappe pile.

Stop no. 25 (short stop): Hunedoara Dolomite in the Timiș-Boia (Laramian) Nappe.

I. Balintoni

Location: Cerna Valley, below Hunedoara Castle.

Northwards from Pui, the excursion crosses the Hațeg Basin, formed on Lotru-Bistra (Laramian) Nappe. At Hunedoara, Paleozoic formations from the basement of the (Austrian) Hunedoara Nappe (a component of Timiș - Boia Laramian Nappe) are well exposed in the Cerna Valley (Fig. 30). Hunedoara Castle (XVth century) was built on reef limestones surrounded by stratified deposits. The carbonate rocks interfinger with sericite-chlorite schists, phylites and quartzites, with palynological evidence showing a Lower Carboniferous age (Kräutner et al., 1981).

This (mainly) carbonatic formation extends over a large area in the Poiana Ruscă Mts and is a typical Variscan component of the Supracrustal basement.

Fourth Day

The Ophiolitic Rocks from Mureș Valley

The ophiolitic rocks from the South Apuseni Mts. are well developed between the localities Turda, in the NE, and Lipova, in the SW. Initially, they have been assigned to three evolutionary stages (Giușcă et al., 1963). Later, the ophiolitic rocks were considered as ocean floor remnants (Rădulescu, Săndulescu, 1973; Bleahu, 1974; Herz, Savu, 1974). Subsequently (Savu, 1980, 1983), they were separated in ocean floor ophiolites (equivalent to first stage) and island arc rocks (equivalent to the second and third stages, *sensu* Giușcă et al., 1963).

Within the ophiolitic rocks from the Mureș Zone Savu (1983) distinguished a lower sheeted dyke complex (O_2) and an upper ocean floor complex (O_1). This edifice is overlain by island arc products as result of bilateral subduction: a north western branch, representing a bimodal magmatism (Savu et al., 1986), and a south-eastern one, marked by numerous volcanic structures covered by a reef barrier, which continues to the north-eastern extremity of the South Apuseni Mts. In a synthetic column within the island arc volcanics of the south-eastern branch, the mentioned author distinguishes a lower andesite-basalt complex (IAV_1), followed by an upper leucocratic volcanic complex (IAV_2).

Using the term in a wider sense, other authors admit that ophiolites may form in various tectonic settings, as island arc ophiolites (Cioflica et al., 1980; Cioflica, Nicolae, 1981) or magmatic arc ophiolites (Lupu, 1983; Lupu et al., 1993). Thus, they correlated the tholeiitic and calc-alkaline series to the first and second stages of Giușcă et al. (1963), and the marginal basin type ophiolites (spilitic complex - Cioflica, Nicolae, 1981; Lupu, 1983; Nicolae, 1983, 1985) to the third stage *sensu* Giușcă et al. (1963).

Differences in petrographic composition of ophiolitic rocks from different tectonic units have been revealed recently (Nicolae, 1992).

The field trip in the South Apuseni Mountains will present the Căpâlnaș-Techereu Nappe and the Rimetea Nappe (subunit of Bedelen Nappe).

The tholeiitic series are largely developed especially in the western part of this unit: here mainly tholeiitic, aphanitic basalt flows (with intersertal textures and often pillow lava facies), anamesites, dolerites, small gabbroic intrusions occur, and seldom very small bodies of peridotites.



















In the eastern part of the Căpâlnaș-Techereu Nappe, the rocks of the tholeiitic series cover smaller areas, usually underlying the products of calc-alkaline series, which consist of pyroclastics and porphyritic basalt

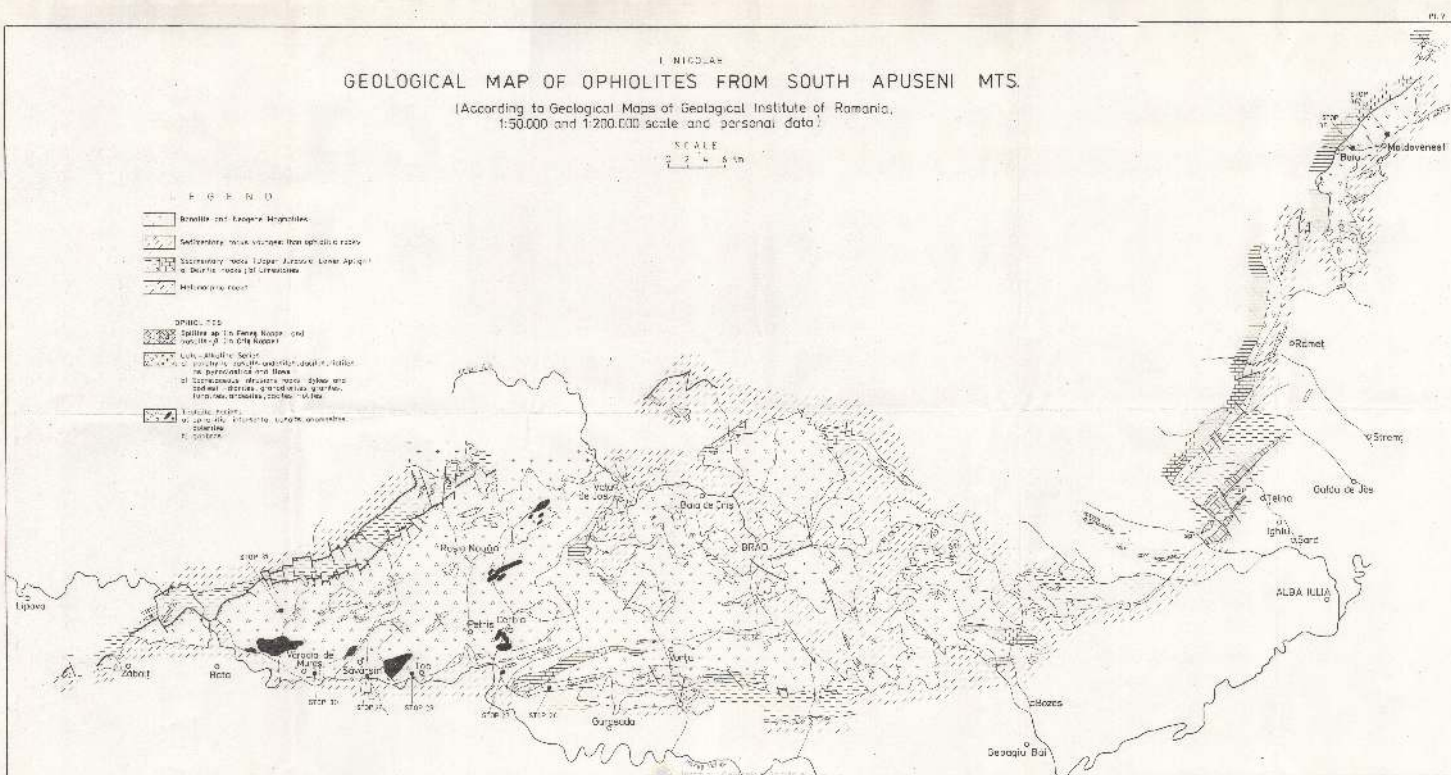


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 -  Sekundäre Metalle: Aluminium, Kupfer, Edelmetalle
 -  Sekundäre Metalle: Aluminium
 -  Metallgewinnung
- DMHC - DTS**
-  Spaltenergie: in Form Wasser und in Form Dampf
 -  Wärme, Abfall, Spezial
 -  Wärme, Abfall, Spezial, Wasser, Abfall, Spezial, Wasser, Abfall, Spezial
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flows, basaltic andesites, andesites, dacites, rhyolites, with minor trachyandesites, orthophyre dikes. Intrusive subvolcanic and plutonic bodies or apophyses which yielded Early Cretaceous K-Ar ages (Lemne et al., 1983; Savu et al., 1986 b) are widespread in this nappe and consist of andesites, dacites, rhyolites, (micro)diorites-(micro)granodiorites-tonalites (Fig. 31). Previously most of these intrusives have been considered banatites (e.g. Cerbia, Săvârşin, Căzăneşti-Pietroasele). They are typical products of the final phase of a magmatic arc. According to the K-Ar isotopic ages (Nicolae et al., 1992) and paleontological data (Antonescu in Lupu et al., 1994) the transition of the tholeiitic series to the calc-alkaline series took place in the Căpâlnaş-Techereu Nappe during the Upper Jurassic.

In the Rimetea Nappe (lying between the Turzii Gorges in the North and Poiana Aiudului in the South), the tholeiitic rocks are exposed South of the Arieş Valley on very restricted areas: the Podeni Valley, Porcului and Dracului Brooks. They consist mainly of tholeiitic basalts, locally in pillow lava facies and subordinately of dolerites and microgabbros, underlying the calc-alkaline series which is largely developed in this unit.

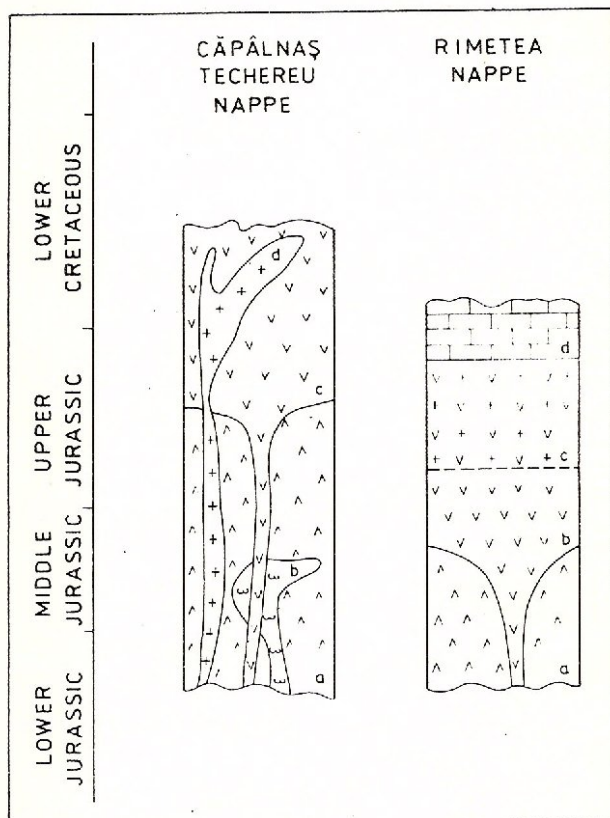
The calc-alkaline magmatites are represented by pyroclastics and porphyritic basalt flows, basaltic andesites, andesites, dacites, rhyolites; in the upper part of the sequence pyroclastics, keratophyre and rhyolite flows occur, overlain by Upper Tithonian-Berriasian limestones (the latter developed between Turzii Gorges and South of the Arieş Valley, in the Valea Albă zone). Dykes of orthophyres, oligophyres, rhyolites and rhyodacites (showing hydrothermal alterations) and porphyritic microdiorites, with a probably Early Cretaceous age, also occur (Fig. 31).

In the Rimetea Nappe, the transition of the tholeiitic series to the calc-alkaline series took place during the Middle Jurassic.

Fig. 31 – Synthetic columns of the Căpâlnaş-Techereu Nappe and Rimetea Nappe (according to Nicolae et al., 1992).

Căpâlnaş-Techereu Nappe: Tholeiitic series: basalts, andesites, dolerites (a); gabbros (b); Calc-alkaline series: pyroclastics and porphyritic basalt flows of porphyritic basalts, basaltic andesites, andesites, dacites, rhyolites, rarely trachyandesites; orthophyre and oligophyre dikes (c); Early Cretaceous intrusive magmatites: andesites, dacites, rhyolites, (micro)diorite-(micro) granodiorite-tonalite (d).

Rimetea Nappe: Tholeiitic series: basalts, in places in pillow lava facies, subordinately dolerites, microgabbros (a); Calc-alkaline series: pyroclastics and flows of porphyritic basalts, basaltic andesites, andesites, dacites, rhyolites, trachytes, latitandesites; oligophyre, orthophyre, andesite, dacite, rhyolite and rhyodacite dykes with hydrothermal alterations, porphyritic microdiorites (b); pyroclastics and flows of keratophyres and rhyolites (c); Upper Tithonian-Berriasian limestones (d).



During the Triassic-Lower Jurassic times (?), the Transylvanian rift formed, generating oceanic crust (Fig. 32 A). Subduction started in the Lower or Middle Jurassic, leading to the formation of a magmatic arc (MA); a marginal basin (MB) formed behind the arc. Initially magmas were generated through melting of the mantle above the downgoing plate, producing the tholeiitic series TH (Fig. 32 B). The process took place in some zones until the Oxfordian (inclusive) and in other zones until the Middle Jurassic. Through further melting of the subducted slab, magmas uprose directly to the surface or rested in an intermediate magmatic basin (imb), generating rocks of the calc-alkaline basin series (CA) (Fig. 32 C); in places these products occurred since the Middle Jurassic or continued to form until the Neocomian. Marginal basin ophiolites were usually produced during the Callovian and Neocomian.

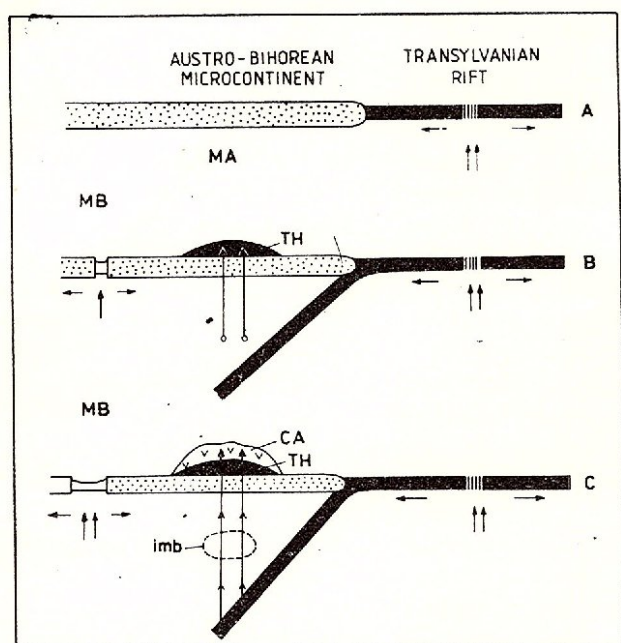


Fig. 32 – Model for ophiolite genesis in the South Apuseni Mountains (According to Nicolae et al., 1992)

Stop no. 26: Basalts in the southern part of the Capalnaș-Techereu Nappe.

I. Nicolae

Location: Deva-Arad highway, east of village Zam.

East of the village Zam, on the highway between Zam and Burjuc, basaltic rocks with aphanitic intersertal textures crop out; rocks are bounded by Early Cretaceous sedimentary rocks (Căbești Beds) along a reverse fault. The area is strongly tectonized, being folded and twisted; calcite, zeolite \pm epidote and quartz veins are widespread.

Stop no. 27: Tholeiitic basalts in pillow lava facies.

I. Nicolae

Location: Zam quarry (Almaș-Săliște Valley).

Basaltic rocks, sometimes in pillow lava facies are exposed in the Zam quarry. In places rocks are intruded by doleritic dykes.

They are cross-cut by veins and veinlets filled by secondary minerals: zeolites, calcite \pm epidote; local pyrite impregnations occur. Chemical composition of basalts is shown in table 2, samples no. 2, 5, 6, 7 and 8.

Stop no. 28: Tholeiitic basalts in pillow lava facies.

I. Nicolae

Location: Toc, Mureș Valley.

Pillow lava facies basaltic rocks are exposed between villages Toc and Cuiăș. They consist of pillow basalts and local tuffs and agglomerates with the appearance of explosion breccia.



Table 2
Chemical composition of the tholeiitic series rocks*
(Căpâlnaş-Techereu Nappe)

%/ppm	1	2	3	4	5	6	7	8
SiO ₂	41.17	45.36	46.88	48.51	49.80	51.00	52.24	52.54
TiO ₂	1.06	1.68	1.81	1.68	1.22	1.46	2.12	1.28
Al ₂ O ₃	16.67	16.50	16.83	20.74	13.50	14.45	13.84	13.10
Fe ₂ O ₃	3.88	6.56	2.12	4.27	4.62	5.98	6.90	5.99
FeO	3.69	6.59	8.26	6.64	4.78	5.26	7.09	4.04
MnO	0.12	0.18	0.23	0.11	0.14	0.19	0.22	0.12
MgO	6.20	8.20	7.60	3.90	4.90	6.40	5.26	3.62
CaO	16.10	3.12	9.38	7.56	8.04	5.53	5.30	8.29
K ₂ O	0.15	1.12	0.25	0.00	0.05	2.00	0.00	2.52
Na ₂ O	2.30	4.32	2.95	2.85	4.78	3.04	3.76	4.10
P ₂ O ₅	0.08	0.15	0.17	0.32	0.14	0.16	0.28	0.14
H ₂ O	2.03	n.d.	0.33	0.90	n.d.	n.d.	n.d.	n.d.
H ₂ ⁺	3.18	4.50	1.63	2.05	2.57	3.27	2.83	1.41
CO ₂	2.81	0.45	0.96	0.00	5.38	1.36	0.00	2.74
S	0.24	0.08	0.15	0.17	0.00	0.08	0.21	0.00
Pb	<2	3	32	<2	n.d.	12.00	7.00	n.d.
Cu	85	160	28	23	29	65	34	65
Ga	15	16	22	33	17	14	24	11.5
Ni	125	34	75	24	32	38	25	31
Co	48	42	42	37	22	40	55	38
Cr	240	35	160	13	75	75	8	55
V	240	165	360	380	240	340	600	430
Sc	40	40	44	32	38	36	43	40
Zr	50	155	145	240	95	115	260	105
Yb	2	3.6	5.4	8.5	3.4	3.9	9.5	4.1
Y	33	37	48	70	30.5	30	68	36
Sr	160	98	70	110	110	140	75	300
Ba	10	70	17	15	27	145	15	140

* According to I. Nicolae (1983, 1992)

n. d. - not determined

1, Basalt, pillow lava, Toc; 2, Basalt, external part of pillow, Zam Quarry; 3, Anamesite, Vărădia de Mureş Quarry; 4, Basalt dyke, Juliţa Quarry; 5, Dolerite dyke, Zam Quarry; 6, Basalt, internal part of pillow, Zam Quarry; 7, Basalt, Zam Quarry; 8, Basalt, Zam Quarry.

Pillows are spheroid bodies enclosed within a previously chloritized vitric matrix impregnated by zeolites, calcite and other secondary minerals. Usually the pillow lava bodies show a vitric-variolitic margin and are cross-cut by radial cracks. Veinlets of calcite, zeolites, chlorite \pm epidote cross-cut the rocks.

Chemical features are shown in table 2, sample no. 1. Whole-rock K-Ar isotopic age is 151.1 ± 4.2 Ma (Nicolae et al., 1992).

Stop no. 29: Early Cretaceous diorite-granodiorite intrusion at Săvârşin.

I. Nicolae

Location: Săvârşin, Mureş Valley.

The Săvârşin granitoid body consists of a dioritic marginal zone and an internal granite-granodiorite zone, sometimes porphyritic-microgranitic fabric (Savu et al., 1986 b).

The diorite mineralogy includes: plagioclase, few quartz, pyroxene, amphibole and in places biotite; grano-dioritic-granitic rocks consist of: plagioclase, alkali feldspar, quartz, biotite and sometimes amphibole (both



chloritized). For chemical characteristics see Table 3.

Table 3
Chemical composition of the Early Cretaceous Săvârşin granitoid body*

%/ppm	9	10	11	12
SiO ₂	75.35	71.54	66.87	52.50
TiO ₂	0.11	1.07	0.38	1.32
Al ₂ O ₃	12.54	13.79	15.61	16.26
Fe ₂ O ₃	0.93	1.95	2.42	4.99
FeO	0.00	0.73	0.94	4.20
MnO	0.03	0.04	0.07	0.17
MgO	0.24	0.82	1.17	5.12
CaO	1.03	2.60	2.95	8.93
K ₂ O	4.44	3.15	4.52	4.52
Na ₂ O	3.60	3.40	3.76	3.15
P ₂ O ₅	0.00	0.14	0.14	0.15
H ₂ O ⁺	1.08	0.57	0.62	1.45
CO ₂	0.00	0.00	0.00	0.00
S	0.18	0.23	0.19	0.19
Pb	10	2.5	4	2
Cu	19	50	19	65
Ga	20	20	20	20
Ni	5.5	6	7	48
Co	2	5	5.5	28
Cr	6	6	2.5	105
V	10	47	2.5	390
Sc	2	7	57	46
Zr	100	145	250	95
Yb	2	2	2.4	4.1
Sr	53	240	650	270
Ba	470	600	1050	95

* According to Ştefan et al., 1990, unpublished data.

9, Porphyritic microgranite, Troaş Valley; 10, Granodiorite, Troaş Valley;
11, Megaporphyritic granite, SE of Săvârşin; 12, Quartz diorite, Troaş Valley

Stop no. 30: Dolerites of Vărădia de Mureş Quarry.

I. Nicolae

Location: Quarry east of Vărădia de Mureş.

The quarry shows massive dolerites and anamesites without a sheeted dike character. Rock textures are ophitic-subophitic or intergranular. For chemical features see Table 2, sample no. 3. K-Ar isotopic age for Vărădia Basalts is 155.7 - 5.7 Ma (Nicolae et al., 1992).

Stop no. 31: Tholeiitic rocks with sheeted dike appearance.

I. Nicolae

Location: Juliţa Quarry on Juliţa.

Sheeted dyke type rocks represented by basalts, anamesites and dolerites crop out in this quarry. Dykes generally show decimetric widths. A chemical analysis for a basaltic dyke is given in Table 2, sample no. 4; three isotopic K-Ar ages exhibit 139.9 ± 6 , 147.6 ± 6 and 167.8 ± 5 Ma (Nicolae et al., 1992).



Fifth day

Cross-section of Apuseni Mts. along Arieş River

Stop no. 32: Relationships between the Biharia and Feniş tectonic units into the Huzii brook valley.

I. Balintoni

Location: Profile along the Iara valley, left tributary of the Arieş Valley.

Between villages Buru and Surduc, the steep banks of the Iara valley expose paragneisses and marbles of the Baia de Arieş Series. Then the valley broadens until the Băişoara village, because it crosses post-Mediterranean

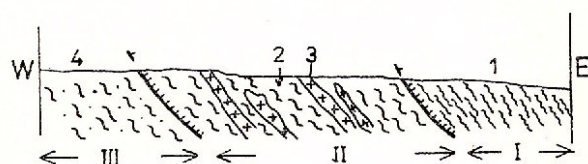


Fig. 33 – Cross section along the Huzii brook near the Săcelu Mountain (scale 1:50.000). I, Biharia Nappe: 1, mylonitized mafic rocks (ophyolitic type sequence) and granitoids of the Biharia Series; II, Feniş Nappe: 2, amphibolites of the Codru Series; 3, Codru migmatites; III, Bihor unit: 4, retrogressed paragneisses of the Someş Series.

sedimentary cover rocks. North-West of Băişoara village the road cuts again through the micaschists of the Baia de Arieş Series pierced by banatites, further crossing the Lunca Larga granitoids and metaophiolites of the Biharia Series. At the north of the Huzii brook the road cuts through the base of the Biharia Nappe. Upstream of "Muntele Fili" chalet the basal mylonites of the Biharia nappe are exposed along 750 m in the road-cut. Rocks show a penetrative mylonitic foliation parallel to the overthrust surface. Earlier structures are entirely obliterated and the rocks show a very fine grain-size. At the mouth of the Sălăşele left tributary the eastward dipping tectonic contact between the Biharia and Feniş nappes is exposed. The very hard rocks of the medium grade metamorphic Codru Series show a narrow mylonitic border (a few meters thick). A quarry upstream the Huzii brook expose Codru Series amphibolites showing syn-S₂ migmatization in the sillimanite zone. The quartz-feldspar veins invading the amphibolites, generally parallel the dominant S₂ foliation.

Stop no. 33: The Băişoara (Permian?) conglomerates, between Baia de Arieş and Biharia nappes (A) and the Biharia Nappe mylonites (B).

I. Balintoni

Location: Road along Ierţii brook, upstream Băişoara village.

Upstream the Băişoara village, the left bank of the Ierţii brook exposes clast supported conglomerates with well rounded and sorted lithoclasts and a quartzitic cement. Clast petrography indicates a source area consisting mainly of Biharia Series metamorphic rocks. Conglomerate beds are dominated by large grainsizes, but local graded bedded sandstones may occur. The sequence might represent deltaic conglomerates. A Permian age is supposed without proofs. Rocks are undoubtedly pre-Tertiary in age because they have been pierced by banatites. The outcrop is more than 200 m wide, fault bounded and shows eastward dips of the bedding planes. The conglomerates are sandwiched between the Baia de Arieş and Biharia nappes and probably represent the cover of the Biharia Series.

Six km upstream, spectacular mylonites crop out on the left bank of Ierţii brook, near the base of the Biharia Nappe. The Biharia Series metaophiolites underwent intense subgranulation, accompanied by a penetrative mylonitic foliation; the quartzitic microlithous have been subject to extensional boundinage and the rocks have a special silky aspect. In the lower part of the Biharia Nappe the mylonites can reach thicknesses around 1000 m. A deformational study of these mylonites is not accomplished yet.



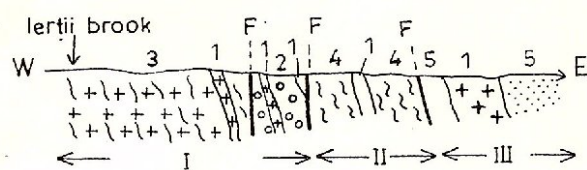


Fig. 34 - Cross section along the Iertii brook near the Băișoara village (scale 1:50.000). 1, Biharia Nappe; 3, Lunca Largă granitoids; 2, Permian (?) conglomerates; II, Baia de Arieș Nappe; 4, micaschists of Baia de Arieș Series; III, Post-Mediterranean cover: 5, Senonian deposits; 1, Upper Cretaceous, banatitic magmatites; F, fault.

Stop no. 34: Relationships between the Biharia and Baia de Arieș Nappes.

I. Balintoni

Location: Runc Gorges on the Ocoliș brook, 8 km upstream its junction with the Arieș Valley.

The Runc Gorges are dug into carbonate rocks of the Vulturese-Belioara Series. This profile presents some of the important problems of the "Biharia Nappe System". Upstream the Lunca Largă village, the Lunca Largă granitoids crop out as flât, gneissous bodies, interlayered with the metaophiolites of the Biharia Series. Moreover, the Lunca Largă granitoids contain xenoliths of metaophiolites. The contamination of the granitoids with basic material is very clear, but contact metamorphic effects are not visible between granitoids and metaophiolites, as those seen in the Highiș Massif, where the alkaline Highiș granitoids pierce the metabasic suite. On this profile the Biharia Series rocks preserve their metamorphic structures and show no mylonitization. Chemical analysis of a plagiogranite (No. 848 of Pociovaliștea Valley) and other of a metaophiolite (No. 64 of Corțești Valley) (Mârza, 1969) are shown in table 4.

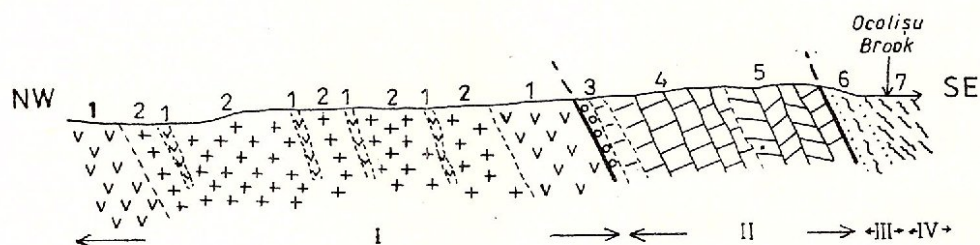


Fig. 35 - Cross section along the Ocoliș brook, between Lunca Largă and Ocoliș villages. I, Biharia Nappe: 1, metamorphosed ophiolitic rocks; 2, Lunca Largă plagiogranites; II+III, Baia de Arieș Nappe: II, Vulturese-Belioara Series (3, pseudo ?-metaconglomerates; 4, dolomites; 5, marbles); III, Baia de Arieș Series: 6, retrogressed paragneisses and micaschists; IV, Post-tectonic cover: 7, Gosau type formation.

Table 4

Chemical analysis of Biharia Series rocks*

Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MnO	MgO	CaO	K ₂ O
848	76.60	0.17	12.39	0.50	1.78	0.017	0.31	0.63	1.95
64	50.06	0.73	14.28	12.01	0.83	0.14	8.67	3.85	0.56

Na ₂ O	P ₂ O ₅	S	CO ₂	H ₂ O ⁺	H ₂ O ⁻	Total
5.21	0.10	-	-	0.82	0.12	100.597
4.33	0.10	0.20	2.11	2.10	0.38	100.35

* According to Mârza (1969)



Table 5
Chemical composition of the calc-alkaline series rocks
(Rimetea Nappe)*

%/ppm	13	14	15	16	17	18	19
SiO ₂	53.30	53.30	54.79	58.73	61.13	63.80	63.92
TiO ₂	0.36	0.32	0.29	0.74	0.42	0.66	0.58
Al ₂ O ₃	19.10	18.12	19.91	16.85	16.73	14.56	17.08
Fe ₂ O ₃	4.33	4.40	3.95	4.94	3.11	4.15	3.11
FeO	3.28	3.62	2.93	1.30	1.15	1.30	0.79
MnO	0.10	0.14	0.10	0.18	0.12	0.08	0.13
MgO	3.70	5.48	0.80	1.72	1.90	1.94	1.22
CaO	7.37	8.50	7.10	6.16	5.32	2.38	2.80
K ₂ O	0.37	0.29	0.75	1.70	1.53	0.93	1.73
Na ₂ O	3.30	2.79	4.17	3.84	3.26	8.06	6.63
P ₂ O ₅	0.08	0.08	0.07	0.10	0.09	0.11	0.12
H ₂ O ⁻	1.26	n.d.	0.46	n.d.	n.d.	n.d.	n.d.
H ₂ O ⁺	2.95	3.09	2.56	1.61	4.39	1.35	1.34
CO ₂	0.55	0.00	2.06	2.06	0.79	0.43	0.43
S	0.00	0.03	0.00	0.07	0.05	0.21	0.12
Pb	2.5	7.5	2	4	2	7	6
Cu	63	50	75	46	11	67	16
Ga	17	12	20	11.5	13	14	14.5
Ni	30	40	29	12	3	3	2
Co	22	22	20	11	5.5	5.5	3.5
Cr	28	65	29	34	4	3	2
V	250	380	210	135	65	35	24
Sc	32	52	30	26	9	13	12
Zr	83	115	83	120	95	42	100
Yb	2.4	1.7	2.2	1.3	2.2	3.2	3.1
Y	30	21	20	31	17	22	25
Sr	260	460	210	320	130	85	140
Ba	65	75	130	340	220	95	250

* According to I. Nicolae, 1994 (unpublished data)

* n. d. - not determined

13 and 14, porphyritic basalts, Arieș Valley; 15, Albitized andesite, Arieș Valley; 16, Albitized dacite, Arieș Valley; 17-19, Albitized rhyodacites, Arieș Valley.

The Biharia Series is overlain by the Vulturese-Belioara Series beginning with quartzitic mylonites derived by conglomerate protoliths. Some authors (e.g. Pană, unpublished data 1993) consider them pseudo-conglomerates resulted through pervasive shearing of the Baia de Arieș Series rocks. In our opinion it is not so easy to demonstrate such a point of view. Strongly mylonitized carbonate rocks lie above the quartzitic rocks of the Vulturese-Belioara Series. In places these mylonites are banded and folded. We have attached the Vulturese-Belioara Series to the base of the Baia de Arieș Series, but it might be a completely independent sequence. Downstream of the Runc Gorges, lying above the carbonate rocks of the Vulturese-Belioara Series, strips of retrogressed paragneisses and micaschists of the Baia de Arieș Series crop out. The metamorphics are overlain by the Gosau facies Senonian deposits of the post-tectonic cover.



Stop no. 35: Calc-alkaline series of Arieș Valley (Rimetea Nappe, subunit of Bedeleu Nappe).

I. Nicolae

Location: km 16+5 on the Turda-Abrud railroad, upstream the tunnel.

Albitized andesites

Volcanics show both massive and block lava structure; in places the andesites show amygdaloid structures. The albitized andesites show porphyritic texture with vitric or cryptocrystalline groundmass of pyroclastic or hyalopilitic character. Plagioclase phenocrysts and microlites show albitic alteration; mafic phenocrysts include pyroxene \pm hornblende, with secondary alteration, opacitisation, chloritisation and calcitisation. Chlorite is widespread through the groundmass.

Chemical characteristics are presented in Table 4, sample no. 15.

Porphyritic basalts

These are rocks with andesine-labradorite phenocrysts (An_{45-75}), showing polysynthetic twinning and zoning; albitization occurs in places and is accompanied by secondary alteration with chlorite, epidote, calcite. Mafic minerals are represented by clinopyroxene \pm orthopyroxene. In places rocks exhibit prismatic columns (e.g. km 16+1); in other cases (near Borzest Valley) spheroidal alterations appear.

For chemical features see Table 5, samples no. 13-14.

Stop no. 36: Albitized ryodacites of the calc-alkaline series (Rimetea Nappe).

I. Nicolae

Location: km 13+6 on the Turda-Abrud railroad.

These are rocks with strongly recrystallized groundmass and altered albitized, argillized plagioclase phenocrysts and chloritized biotite and/or amphibole. K_2O contents are restricted to the groundmass. For chemical characteristics, see Table 5, samples no. 17-19.

Sixth day

Stop no. 37: Badenian deposits overlying Upper Cretaceous deposits.

D. Ciulavu, G. Bertotti

Location: Geoagiu Valley (ca 20 km N of Alba Iulia).

A quite complete section from the Mesozoic to Badenian deposits can be seen along the valley. Outcrops around the village of Geoagiu de Sus show Upper Cretaceous deposits folded and faulted during the Latest Cretaceous-Paleogene, discordantly overlain by Badenian deposits. These consist of sandstones and limestones.



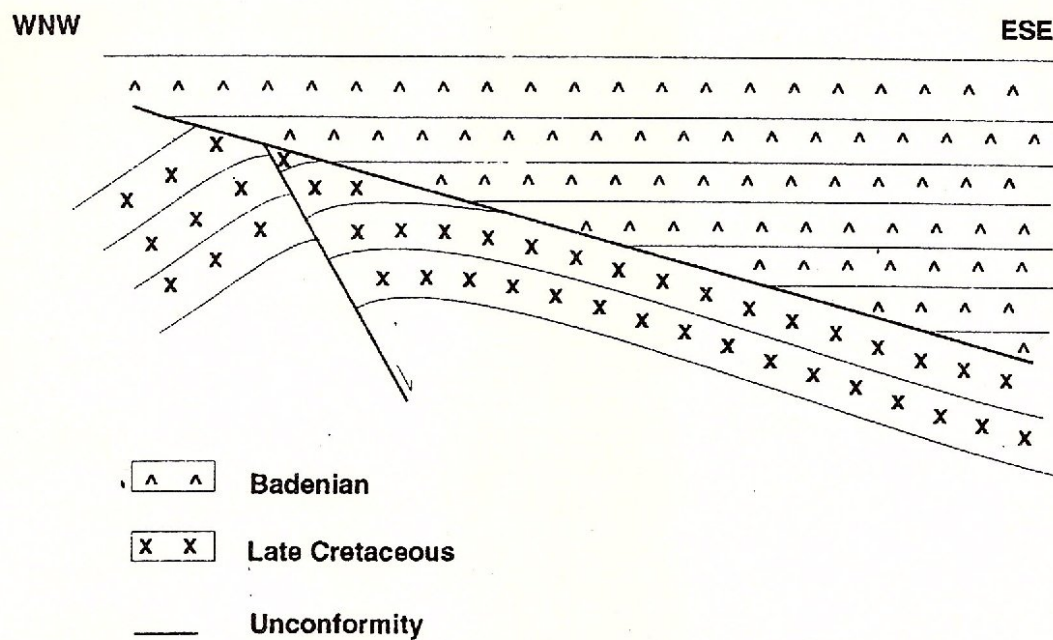


Fig. 36 - Geoagiu Valley

Stop no. 38: Post-Badenian thrust.

D. Ciulavu, G. Bertotti

Location: Ciugud quarry, Alba Iulia.

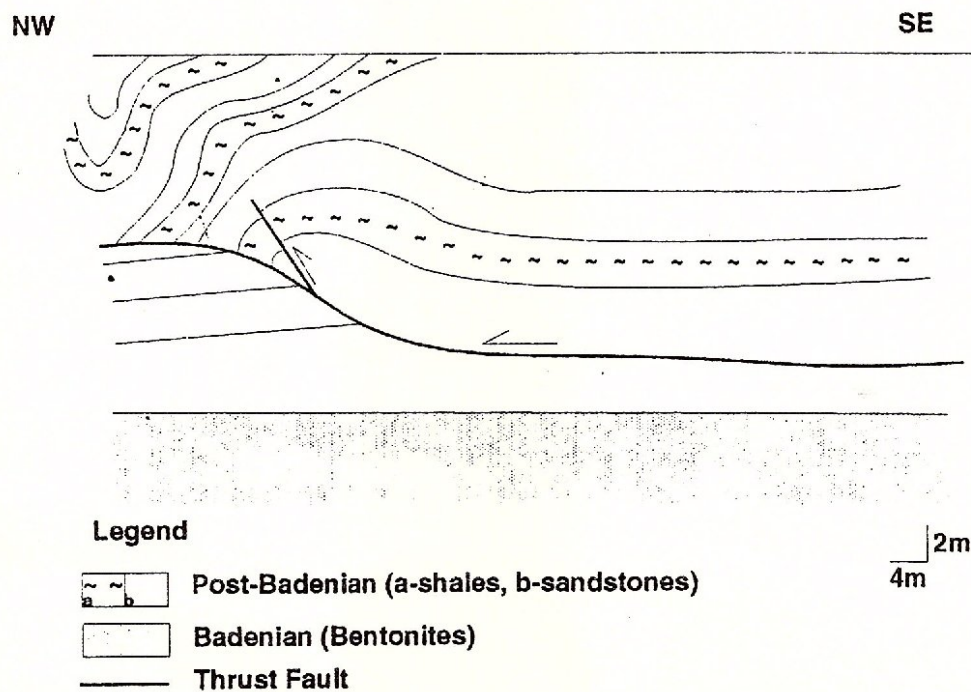


Fig. 37 - Ciugud Quarry



Badenian to post-Badenian silts and sandstones outcrop in this bentonite quarry lying few kms east of Alba Iulia. The sediments are mostly flat-lying but a nicely-developed thrust can be observed in the northern part of the quarry. The thrust is related to WNW-ESE shortening and was formed in post-Badenian times, possibly in the Pliocene.

Stop no. 39: Quaternary Racoș basalts.

D. Ciulavu, G. Bertotti

Location: Racoș quarry, 40 km NNW of Brașov.

This stop is in the SE-part of the Transylvania where, in a very limited area, Quaternary basalts are found. Here beautifully developed columnar lava flows are overlapped by lapilli tuffs. These are faulted, stressing the importance of very young tectonic activity in the Transylvanian Basin and surrounding regions.

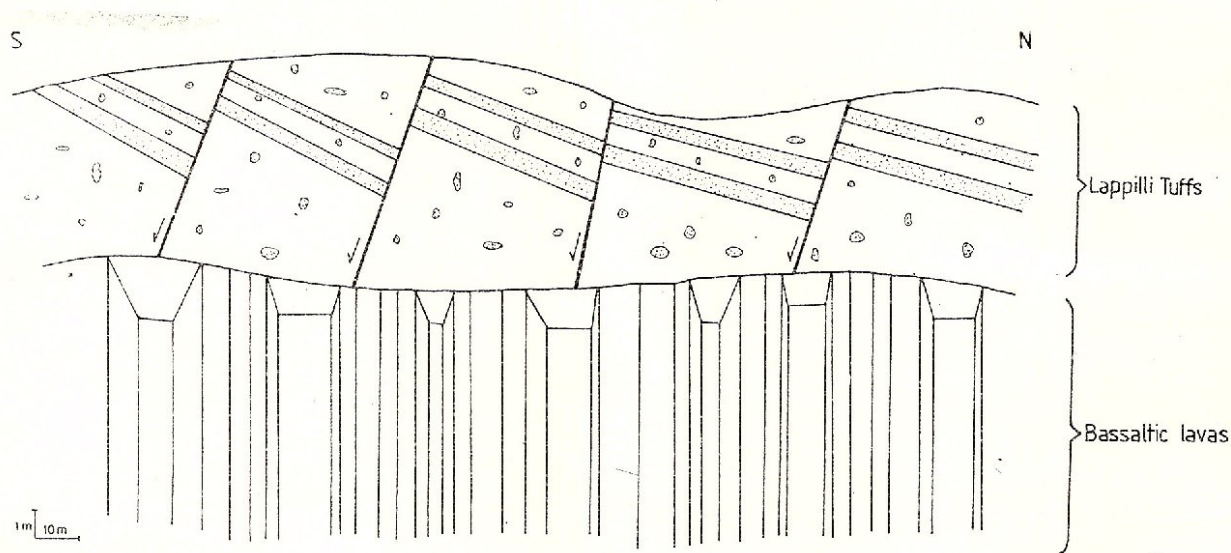


Fig. 38 Racoș Quarry

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