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DĂRI DE SEAMĂ ALE ȘEDINȚELOR

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DES SÉANCES

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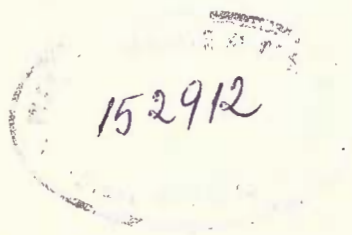
1. MINERALOGIE—PETROLOGIE



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DES

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1. MINERALOGIE—PETROLOGIE

MINERALOGIE

OBSERVAȚII MINERALOGICE ȘI GEMOLOGICE ASUPRA UNOR OCURENȚE DE OPAL, CALCEDONIE ȘI JASP DIN MUNȚII OAȘ—VĂRATEC [1]

de

OSCAR EDELSTEIN [2], KLARA EDELSTEIN [2], ANDREI GÖTZ [3]

Opal. Chalcedony. Jasper. Gems. Mineral data. Chemical composition. DTA data. Colour. East Carpathians — Neogene eruptive-Quaternary — Oaș Mountains, Gutin Mountains.

Abstract

Mineralogical and Gemmological Observations on Opal, Chalcedony and Jasper Occurrences in the Oaș-Văratec Mountains. This paper presents the main mineralogical features, established on the basis of microscopic, thermic-differential and, in some cases, diffractometric analyses of the opal, chalcedony and jasper occurrences in the zones : Dealul Dîmbanului-Bixad (Oaș), Laposa Valley (Băița), Pîrîul Nopții (Valea Roșie-Baia Mare), Vălini Valley (Firiza), Cusereada Valley (Capnic). Gemmological observations are also made.

Résumé

Observations minéralogiques et gémologiques sur quelques occurrences d'opale, calcédoine et jaspe des monts d'Oaș-Văratec. On présente les principales caractéristiques minéralogiques établies sur base des analyses microscopiques, thermique-différentielles et pour quelques cas, diffractométriques, des occurrences d'opale, calcédoine et jaspe des zones : Dealul Dîmbanului-Bixad (Oaș), vallée de Laposa (Băița), Pîrîul Nopții (Valea Roșie-Baia Mare), vallée de Vălini, (Firiza), vallée de Cusereada (Capnic). On y fait aussi des observations gémologiques.

[1] Depusă la 16 martie 1987, acceptată pentru comunicare și publicare la 1 aprilie 1987, comunicată în ședința din 24 aprilie 1987.

[2] IPEG „Maramureș". Str. Victoriei nr. 146, Baia Mare.

[3] ICITPMN. Str. V. Babeș nr. 64, Baia Mare.

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În ultimii ani în țara noastră au crescut preocupările de ordin gemologic (Papiu, Mastacan, 1973; Moțiu, Ghiurcă, 1979; Ghiurcă, 1981; Ghiurcă, Ghiurcă, Fulga, Fulga, 1984; Ghiurcă, Ghiurcă, 1982).

În cele ce urmează se prezintă cîteva date geologice, mineralogice și gemologice asupra unor ocurențe de opal și calcedonie asociate magmatitelor neogene din munții Oaș-Văratec. În intenția autorilor, nota de față ar fi prima lucrare dintr-o suită în cadrul căreia să fie expuse caracteristicile mineralogice și gemologice ale unor minerale și roci din această zonă, ce pot fi utilizate ca pietre decorative și ornamentale. Lucrarea se referă la ocurența de opal din Dealul Dîmbanului — Bixad (Oaș), la jaspul din zona valea Laposa — Băița, la calcedonia agatiformă de pe Pîrîul Nopții (Valea Roșie — Baia Mare), la ocurențele de jasp de la Firiza (valea Vălini), la calcedonia de pe valea Cuscreada (Capnic) și la calcedonia din zona Trestia. În figură este prezentată localizarea ivirilor sau a ariilor la care se referă nota. Pe măsură ce se vor acumula date, autorii doresc să prezinte și alte varietăți, precum și alte ocurențe. Întrucît determinările s-au făcut pe un număr mic de eșantioane, lucrarea are un caracter preliminar.

1. Opalul din Dealul Dîmbanului (versant stîng valea Custurii-Bixad)

Apare ca filonașe neregulate, sub 0,3 m grosime, discontinue, sau în cuiburi dispuse într-o rocă eruptivă care a fost descrisă de către Istvan et al. (1982) ca fiind hialodacit. Coloratia este variată; mai frecvente sînt culorile verde-gălbui și maronie, aceasta din urmă în două nuanțe; una mai deschisă și alta mai închisă; la unele eșantioane se observă o trecere gradată iar la altele o limită netă între culoarea maronie și cea verzuie. Mineralul este casant și prezintă spărtură concoidală. Unele sînt fin poroase (eș. 7281—3E), cu textură ușor nodulară; adesea se observă pete și reticule brune-gălbui, pe un fond verzui; mici impurități limonitice apar în benzi. Cîteva eșantioane prezintă fie cruste, fie mici cuiburi de silice albă criptocristalină.

În secțiune subțire, se disting cîmpuri fine, sub 0,005 mm cu forme neregulate, cu structuri fibroradiare vizibile doar la diafragmare puternică. Unele impurități solzoase, în agregate fin sferoidale prezintă dimensiuni sub 0,03 mm (pl. I, fig. 1). Limonitul apare sporadic (eșantion 7280, 7281—3E), ca pete sferoidale brune, cu halouri gălbui, mai rar în cuiburi neregulate. În eșantionul 7282—3E, se observă pete sferoidale de limonit, brune-opace, dispuse în jurul unor granule de sulfuri. Această probă are în secțiune subțire o culoare slab gălbuie-verzuie, cu nuanțe gălbui-brune. În proba 7284, se observă un început de cristalizare sub formă de fibre șterse și agregate fibroase (pl. I, fig. 2).

În secțiune lustruită, materialul prezintă o suprafață lucioasă, cu textura foarte slab vizibilă, nodulară (7282—3E), sau în benzi fine și cu lentile și nodule mici. În tabelul 1, sînt prezentate principalele caracteristici rezultate la analiza termogravinometrică.

Analizele difractometrice au pus în evidență în probele 7280—3E, 7281—3E și 7284—3E, pe un fond de silice amorfă, tridimit iar în ultimele două și goethit.

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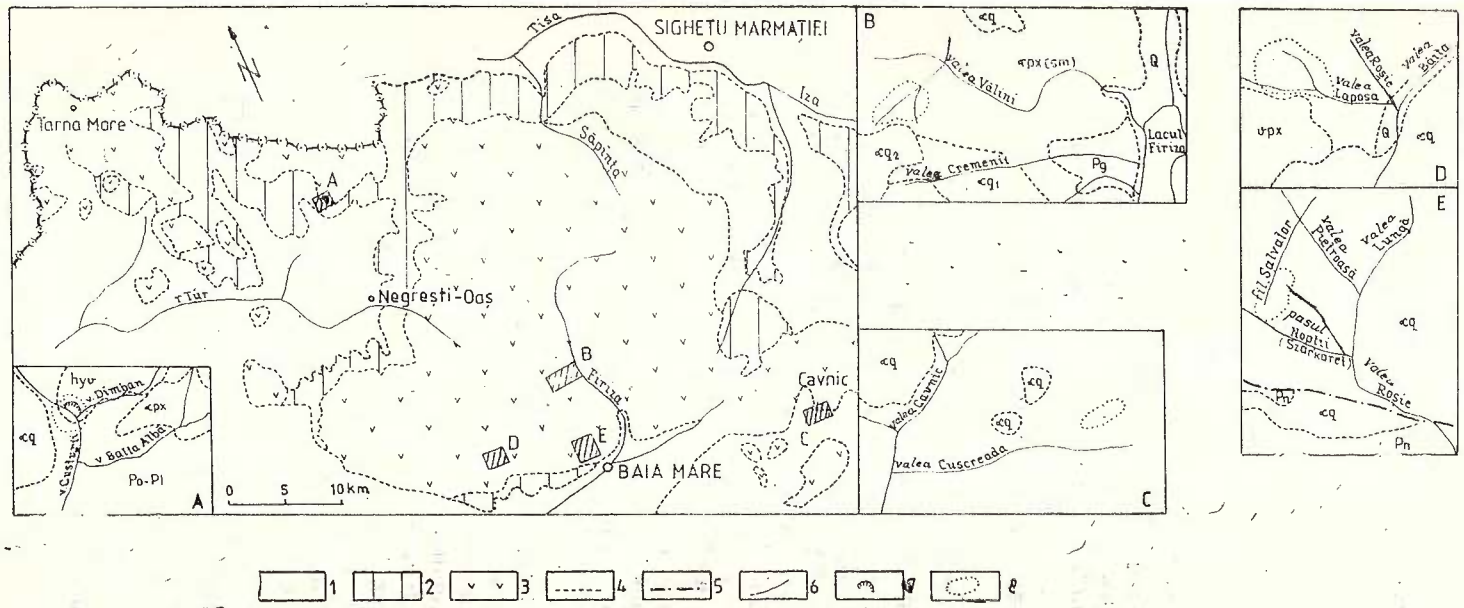

Localizarea ivirilor de opal, calcedonie și jasp.

A) Dealul Dîmbanului (Bixad), după Istvan et al. (1982); B) Valea Vălini (Firiza) după Edelstein et al. (1979 — 1980); C) Valea Cuscreada (Capnic) după Bernad et al. (1987); D) Valea Laposa (Băița) după Edelstein et al. (1979 — 1980); E) Pîriul Nopții (Valea Roșie, Baia Mare) după Edelstein et al. (1979—1980).

1, Depozite cuaternare; 2, Formațiuni sedimentare paleogene și neogene; 3, Roci eruptive; 4, Limită geologică; 5, Falie; 6, Filon hidrotermal; 7, Carieră; 8, Aria de răspîndire a fragmentelor de opal, jasp și calcedonie. Q = depozite cuaternare; Po—Pl = sedimentar pontian pliocen; Pn = = sedimentar pannonian; Pg = sedimentar paleogen; apx = andezit piroxenic; aq = andezit cuarțifer; hid = hialodacit; aq_2 = andezit cuarțifer de Tocastru; dpx = dacit de Sindileu; apx (sm) = andezit piroxenic sarmațian.

Location of the opal, chalcedony and jasper occurrences.

A) Dealul Dîmbanului Hill (Bixad), after Istvan et al. (1982); B) Vălini Valley (Firiza) after Edelstein et al. (1979—1980); C) Cuscreada Valley (Capnic) after Bernad et al. (1987); D) Laposa Valley (Băița) after Edelstein et al. (1979—1980); E) Pîriul Nopții Brook (Valea Roșie, Baia Mare) after Edelstein et al. (1979—1980).

1, Quaternary deposits; 2, Paleogene and Neogene sedimentary formations; 3, eruptive rocks; 4, geological boundary; 5, fault; 6, hydrothermal vein; 7, quarry; 8, spreading area of opal, jasper and chalcedony fragments.

Q = Quaternary deposits; Po-Pl = Pontian-Pliocene sedimentary; Pn = Pannonian sedimentary; Pg = Paleogene sedimentary; apx = pyroxene andesite; aq = quartz andesite; hid = hyalodacite; aq_2 = Tocastru quartz andesite; dpx = Sindileu dacite; apx (sm) = Sarmatian pyroxene andesite.

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TABELUL 1

	7280—3E	7281—3E	7282—3E	7284—3E
Culoarea după ardere	cărămizie	slab gălbuie	portocalie	cărămizie
Temperatura de eliminare a apei	220°	250°	250°	305°
Pierderea în greutate % (prin eliminarea apei)	5,5	6,2	5	3,8

Compoziția chimică este prezentată în tabelul 2.

Din punct de vedere gemologic se caracterizează printr-o culoare plăcută, verde, cu tentă gălbuie (asemănătoare cu cea a lizarditului), sau maronie, luciu bun, opac. Se prelucrează satisfăcător în sistemele „baroc" și „cabochon" putînd fi utilizate la confecționarea medalioanelor, cerceilor, butonilor pentru manșetă ca și a altor obiecte de podoabă.

2. Calcedonia de pe Pîrîul Nopții (Valea Roșie-Baia Mare)

Formează umplutura părții superioare a filonului Salvator din cadrul zăcămîntului Valea Roșie, pe tronsonul cuprins între pîrîul Szarkaret și Pîrîul Nopții. Filonul are grosimi pînă la 3 m și poate fi observat astăzi în cîteva șanțuri vechi și pe hălzile a două galerii. Calcedonia este asociată cu cuarț; adesea, fragmente frumos rubanate de calcedonie albă, cu lungimi de 1—10 cm și grosimi de 1—2 cm constituie elemente ale unei brecii filoniene cimentate cu cuarț și, în parte, cu limonit. Calcedonia are culoare albă-lăptoasă și este alcătuită din cruste submilimetrice.

Local, prezintă desene bizare, determinate probabil de micromorfologia peretelui pe care s-a depus sau de prezența unor impurități în timpul formării. Apar și aspecte agatiforme. O ușoară impregnare cu limonit conferă unor eșantioane o nuanță galbenă-maronie.

În secțiune subțire, se observă benzi din agregate fibroradiare, lungi pînă la 1 mm, în alternanță cu benzi mai mult granulare (pl. I, fig. 4). Majoritatea acestora au alungire pozitivă, fiind reprezentate prin cuarțină; benzile de calcedonie au alungire negativă. Se intercalează și benzi de cuarț granular, hipidiomorf-xenomorf cu granulație de 0,5 mm. Cu excepția unor pelicule limonitice, foarte fine sau a unor benzi cu impregnări limonitice, de asemenea foarte fine nu apar alte impurități vizibile. Analiza chimică pune în evidență conținuturi de Au și Ag mai ridicate decît în alte calcedonii din zona munților Oaș-Văratec (tab. 2). Conținutul de apă, așa cum rezultă din analiza termică este redus, 0,7%—cantitate ce se elimină la 305°. Se prelucrează relativ ușor atît în sistemul cabochon, cît și în cel baroc. Tăierea se face astfel încît să se surprindă suprafața rubanată; marea varietate a desenului oferă o gamă largă de modele. Prin șlefuire, capătă un luciu bun. Dacă este tăiată în plăcuțe subțiri, prezintă și o semitransparență (pl. III, fig. 3, 4). Poate fi colorată artificial, obținîndu-se pseudoagate.

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TABELUL 2

*Conținutul în Au, Ag, Pb, Zn, Cu și S al unor ocurențe de opal și calcedonie din munții Oaș—Văratec**

Proba	Localizare	Au g/t	Ag g/t	Pb %	Zn %	Cu %	S %
7274-3E	valea Laposa-Băița	0	2,0	u	0,020	0,020	0,40
7276-3E	Valea Roșie—Baia Mare	0,2	15,8	u	0,010	0,015	0,19
7277-3E	Trestia	0	3,0	u	0,010	0,015	0,26
7278-3E	valea Vălini	0	2,0	u	0,015	0,015	0,18
7279-3E	valea Cuscreada	0	2,6	0,034	0,065	0,015	0,62
7280-3E	valea Custurii-Oaș	0	6,4	u	0,010	0,010	0,19
7281-3E	,,	0	2,8	u	0,015	0,010	0,21
7282-3E	,,	0	4,0	u	0,020	0,015	0,20
7284-3E	,,	0	0	u	0,015	0,015	0,14

* Analize efectuate în laboratorul de chimie al IPEG Maramureș în anul 1983.

3. Jaspul cărămiziu din valea Laposa (Băița)

Apare ca mici filonașe discontinue și cuiburi neregulate în andezitul piroxenic. Aceeași silice roșie este relativ larg răspîndită și în zona Ilba—asociată ariei în care apar andezitele piroxenice pannoniene; de obicei, apare sub formă de fragmente. În loc a fost observată în carierele Valea Hotarului și Ilba—gară. Probele la care ne referim în cele de mai jos provin din fragmente de pe valea Laposa—Băița.

Apare ca o silice criptocristalină de culoare roșie-cărămizie intensă; frecvent, prezintă o serie de impurități reprezentate prin zone amorfe, piritizări și silice de culoare albă sau neagră.

În secțiune subțire, apare ca o masă criptocristalină de culoare roșie-sîngerie, puțin transparentă. Calcedonia este fin cristalizată (0,02 mm) și prezintă mase de impregnare de hematit fin solzos sub formă de granule sau sferule fine. Este străbătută de vinișoare, de mase xenomorfe sau fibroase de calcedonie relativ pură sau slab impregnată cu hematit (pl. II, fig. 1).

În secțiune lustruită, se obține o suprafață foarte lucioasă, roșie-cărămizie, cu o nuanță mai deschisă decît a eșantioanelor neșlefuite, uneori cu granule disperse de goethit. Analiza DTA nu pune în evidență apă; au fost observate două efecte exoterme: la 470° — transformarea polimorfă a oxizilor de fier și la 545° — oxidarea fierului.

Analiza difractometrică a pus în evidență prezența cuarțului și a hematitului.

Prin prelucrare în stilul baroc capătă o nuanță mai vie, cu luciu puternic; este deosebit de apreciat. Deoarece este casant, probabil că este mai greu de prelucrat în alte tehnici.

4. Ocurențele de jasp de pe valea Vălini (Firiza)

Nu au fost pînă în prezent cartate detaliat, dar se știe că apar în cantitate relativ mare, în orice caz mai mare decît tipurile menționate la

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punctele 1, 2 şi 3. Fragmentele au dimensiuni între 1—30 cm, mai rar 0,5 m. Paleta coloristică este relativ variată: galben (două nuanţe), roşu (o altă nuanţă decît jaspul din valea Laposa), negru intens, maronie (o altă nuanţă decît a opalului din Dealul Dîmbanului, mai puţin plăcută), galben cu pete roşii, milimetrice şi altele.

În secţiune subţire (eşantion 7278—3E), calcedonia apare în granule de cca 0,05 mm, xenomorfe sau fibroradiare; se observă o alternanţă de benzi cu granulaţie puţin diferită. Benzile sînt impregnate în mod variat cu limonit, în mase fin granulare sau sferoidale; pe benzi înguste şi în cuiburi, apar acumulări mai puternice (pl. II, fig. 2, 3).

Este străbătută de o altă generaţie de calcedonie sub formă de vinişoare, cu grosimi pînă la 0,02 mm, cu granulaţie fină, filonaşe în care direcţia fibrelor este perpendiculară pe planul filonaşului.

În secţiune lustruită, se obţine o suprafaţă lucioasă, în care uneori se evidenţiază textura rubanată şi cuiburi roşcate de goethit (pl. II, fig. 4). Prin analiza DTA, s-a pus în evidenţă un conţinut total de apă de 11,2% din care 6,1 este probabil asociată oxizilor de fier iar restul unor impurităţi de caolin.

Prelucrarea în sistemul „baroc" se face uşor obţinîndu-se suprafeţe cu luciu foarte bun şi cu o gamă variată de culori, adesea cu desene.

5. Calcedonia de pe valea Cusereada (Capnic)

Este incoloră sau are foarte slabe tente albăstrui. A fost observată doar în fragmente foarte neregulate care probabil au mulat configuraţia pereţilor fisurilor din andezitele piroxenice în care s-au depus. Fragmentele — destul de rare — au dimensiuni de ordinul centimetrilor.

Sub microscop apare sub formă de benzi fibro-radiare de 0,1 mm, sporadic lentile granulare de 0,01 mm (pl. III, fig. 1, 2). Prezintă alungirea negativă. Conţinutul de apă este de 0,8% (pus în evidenţă prin efectul endoterm DTA de la 95°).

6. Calcedonia albastră de Trestia

A fost studiată detaliat relativ recent de către Moţiu şi Ghiurcă (1979) şi noi nu aducem date noi asupra ei dar o reamintim ca fiind unul din materialele foarte apreciate pentru confecţionarea pietrelor ornamentale.

7. Modul de prelucrare

Materialul supus prelucrării a fost în prealabil spart cu ciocanul la dimensiuni convenabile (sub 4 cm lungime). Deşi prin prelucrare, muchiile ascuţite se îndulcesc, este de preferat totuşi ca înainte de şlefuire fiecare piesă să fie tăiată şi, eventual polizată, conferindu-i-se o formă ce poate fi uneori în acord cu desenul ei natural.

S-a utilizat un aparat vibrator (vibratory tumbler) marca „Graves" care prelucrează o cantitate de aproximativ 1300 grame în cursul a 204 de ore. Şlefuirea se face în două trepte, cu abraziv de tip carborund de

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1/400 mm și respectiv 1/250 mm iar lustruirea cu oxid de aluminiu sau cu alte substanțe echivalente. După prelucrare, fiecare piesă este prinsă cu ajutorul unui adeziv de tipul rășinilor epoxidice într-o capsă metalică astfel încît aceasta să corespundă formei și culorii piesei.

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MINERALOGICAL AND GEMMOLOGICAL OBSERVATIONS ON OPAL, CHALCEDONY AND JASPER OCCURRENCES IN THE OAȘ—VĂRATEC MOUNTAINS

(Summary)

Opal occurring in the Dealul Dîmbanului (Bixad—Oaș) presents very fine fibroradial structures and limonitic impurities; α — tridymite and goethite have been rendered evident by diffractograms. It displays green and brown hues. Water content varies between 3.8 and 6.2 per cent.

Jasper in the Laposa Valley (Băița) shows a specific light red-brick red colour, with a very good lustre, highly appreciated.

Chalcedony in the Pîrîul Nopții (Valea Roșie-Baia Mare) forms the filling of the Salvator vein in the Valea Roșie ore deposit. In thin section 1 mm bands of fibroradial aggregates alternate with more granular bands; some of them display a positive elongation being represented by quartzine and others a negative elongation, being represented by chalcedony; bands of hipidiomorphic-xenomorphic granular quartz are intercalated. Water content represents 0.7 per cent.

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Jasper in the Vălini Valley (Firiza) presents various colours : yellow, red, black, brownish and combinations of these colours. Water content reaches 11.2 per cent.

Chalcedony in the Cuscreada Valley (Capnic) is colourless and transparent; it consists of fibroradial bands with negative elongation. Water content represents 0.8 per cent.

All of them, by polishing, present nice colours and a lustre surface which give them gem qualities.

EXPLICAȚIA PLANȘELOR

Planșa I

Fig. 1 — Opal cu impurități fin solzoase și agregate. Dealul Dîmbanului (eș. 7282—3E). Nic. II, ×64.
Opal with fine scaly impurities and aggregates. Dealul Dîmbanului (sample 7282 — 3E). Nic. II, ×64.

Fig. 2 — Opal cu început de cristalizare sub formă de nodule sferoidale și cu impurități în benzi și sferule fibroase. Dealul Dîmbanului (eș. 7284—3E). Nic. II, ×64.
Opal with incipient crystallization as spheroidal nodules and impurities as bands and fibrous spherules. Dealul Dîmbanului (sample 7284 — 3E). Nic. II, ×64.

Fig. 3 — Opal fin nodular cu impurități înșirate vermicular : Dealul Dîmbanului (eș. 7284—3E). Nic. II, ×125.
Fine nodular opal with vermicular impurities ; Dealul Dîmbanului (sample 7284—3E). Nic. II, × 125.

Fig. 4 — Calcedonie de pe Pîriul Nopții (eș. 7276-3E); alternează benzi granulare cu benzi fibroradiare. Nic. +, × 21.
Pîriul Nopții chalcedony (sample 7276-3E); alternation of granular bands and fibroradial bands. Nic. +, × 21.

Planșa II

Fig. 1 — Silice criptocristalină impregnată cu hematit fin solzos, străbătută de vinișoare de calcedonie pură. Valea Lapoșa (eș. 7274-3E). Nic. II, × 21.
Cryptocrystalline silica impregnated with fine scaly hematite, penetrated by veinlets of pure chalcedony. Lapoșa Valley (sample 7274-3E). Nic. II, × 21.

Fig. 2 — Calcedonie fin fibroasă, în benzi fine, impregnate cu limonit, străbătută de calcedonie pură. Valea Vălini (eș. 7278—3E). Nic. II, ×21.
Fine fibrous chalcedony, as fine bands, impregnated with limonite, penetrated by pure chalcedony. Vălini Valley (sample 7278-3E). Nic. II, × 21.

Fig. 3 — Calcedonie de pe valea Vălini (eș. 7278-3E) cu benzi fin fibroase, în alternanță cu benzi impregnate cu limonit. Nic. II, × 21.
Vălini Valley chalcedony (sample 7278-3E) with fine fibrous bands alternating with bands impregnated with limonite. Nic. II, × 21.

Fig. 4 — Calcedonie cu textură rubanată și cuiburi de goethit. Valea Vălini (eș. 7278-3E). Secțiune lustruită. × 6.
Chalcedony with banded structure and goethite nests. Vălini Valley (sample 7278-3E). Polished section. Nic. +, × 6.

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Planșa III

Fig. 1 — Calcedonie de pe valea Cuscreada (eș. 7279-3E); se observă cîmpuri fibro-radiare. Nic. II, × 21.
Cuscreada Valley chalcedony (sample 7279-3E); fibroradial fields can be observed. Nic. II, × 21.

Fig. 2 — Calcedonie de pe valea Cuscreada (eș. 7279-3E); se observă structura nodulară. Secțiune lustruită. Nic. II, × 5.
Cuscreada Valley chalcedony (sample 7279-3E); nodular structure can be observed. Polished section. Nic. II, × 5.

Fig. 3 — Calcedonie rubanată de pe Pîrîul Nopții (eș. 7276-3E). Secțiune lustruită, Nic. +, × 5.
Pîrîul Nopții banded chalcedony (sample 7276-3E). Polished section. Nic. +, × 5.

Fig. 4 — Aceeași. Secțiune lustruită. Nic. II, × 5.
Idem. Polished section. Nic. II, × 5.

Planșa IV

Fig. 1, 2, 3, 4 — Opaluri și calcedonii prelucrate în sistemul baroc și montate în capse.
Opals and chalcedonies processed in the baroque style and mounted in caps.

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1. MINERALOGIE—PETROLOGIE

MINERALOGIE

15 2912

THE CHALCEDONY OF TRESTIA—MARAMUREŞ[1]

by

GHEORGHE ILINCA[2]

Silica. Chalcedony. Quartz. Opal. Recrystallization. Paramorphism. Organic materials. Colour. Chemical composition. DTA data. X-ray data. Infrared spectra. East Carpathians — Transcarpathian Flysch Zone — Ţibleş and Lăpuş Mountains. Baia Mare-Cavnic Zone.

Abstract

The chalcedony's mode of presentation and its relationship with the surrounding rocks suggest that it was formed by solidifications of silica gels precipitated from hot springs, either in subaerial conditions or in fissures and cavities of a preexisting alteration crust. From mineralogical point of view, the silica from Trestia consists of quartz, chalcedony, opal and probably relics of melanophlogite, which form a complex recrystallization sequence. The existence of some relic microstructures and the striking similarity between forms from Trestia and cubes of melanophlogite, led us to consider the cubic crystals as paramorphs after melanophlogite. There are also the results of chemical analysis, X-ray diffraction and infrared absorption spectroscopy that plead for the same idea. The blue colour of the chalcedony is due to selective absorption and diffusion of light by colloidal dispersed particles. When heated, chalcedony turns from blue to white and its strong opalescence becomes less obvious suggesting the instability of the dispersed substance to high temperatures, probably due to its fluid or organic nature.

Résumé

Calcédoine de Trestia—Maramureş. La forme de gisement de la calcédoine et ses relations avec les roches environnantes suggèrent qu'elle est le résultat de la solidification de quelques gels de silice précipités des sources chaudes, soit en des conditions subaériennes, soit en des fissures et cavités d'une écorce d'altération préexistente. Minéralogiquement, la silice de Trestia

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comporte du quarz, de la calcédoine, opale et probablement reliquants de mélanophlogite qui forment une séquence complexe de recristallisation. La présence de quelques microstructures reliques et la grande ressemblance entre les formes de Trestia et les cristaux de mélanophlogite nous déterminent de considérer les cristaux cubiques comme les résultats des paramorphoses après le mélanophlogite. Les résultats des analyses chimiques, par diffraction de rayons X et spectrophotométrie d'absorbtion en infrarouge sont aussi en faveur de cette idée. La couleur bleu de la calcédoine est le résultat de l'absorbtion et de la diffusion sélective de la lumière au niveau de quelques particules dispersées de manière coloïdale. Par échauffement, la couleur de la calcédoine passe de bleu vers blanc et l'opalescence devient moins accentuée, fait qui suggère l'instabilité de la substance dispersée à des températures hautes, fonction de son caractère fluide ou de sa nature organique.

Introduction

In a limited area, north-east of Trestia (Baia Mare—Cavnic zone), scattered in soils or along brook-beds appear numerous fragments of blue silica, rendering varied morphological aspects : sinter-like reniform masses, thin radial-fibrous or parallel banded structures, surfaces imprinted by an irregular relief or plates and druses inlaid with unwonted cubic crystals.

Among other silica occurrences which have compelled attention in the course of time, the chalcedony of Trestia won a certain renown. It is not only the colour or the unusual crystallographic forms which state the reason for this celebrity, but that for almost two centuries the attempts made to identify their origins usually remained in a stage of contradictory assumptions. Even today the genesis of the blue hexahedral crystals is not yet obvious, in spite of so many dedicated studies and maybe is for this reason that a new paper dealing with the same subject does not run the risk of being a subject matter of little interest.

Historical [3]

Three kinds of matter will be followed both by the historical references in this chapter and by our own observation : 1. The chalcedony's mode of presentation and its relationship with the country rocks, 2. The mineralogical composition and the origin of the crystallographic forms, 3. The cause of the blue colour.

1. The chalcedony was first stated by Fichtel in 1791 (in a quotation by Ackner, 1855) who described fragments of blue silica in a brook-bed where they occurred mixed with clay. The fragments resembled common limestone stalagmites and crusts.

Later, Beudant (1822) supposed that the chalcedony fragments „ ... originate in fields with porphyritic green rocks. Maybe they are to be found in veins, but there is nothing positive in this respect as the samples were collected from brooks or found in sands".

According to Ackner (1855), chalcedony appears "on ploughed fields, in the bed of Cibin [4] river or within the alluvia all over the place".

Zepharovich (1893) quoted Koch as maintaining that fragments came from "veins and druses in rhyolitic andesites". Bombicci (1900)

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stated a similar mode of presentation but also reported chalcedony in druses within silicified wood, thus finding a good resemblance with some silica cubes from Olmo di Castellucio di Capugnano, near Poretta (Italy).

Manilici (1962) considered the chalcedony as a result of hydrothermal alterations related to Neogene extrusive rocks in the area but made no definite reference to the relations with surrounding rocks.

2. As for the origin of crystallographic forms, one of the first observations belongs to Ackner (1855), who described the hexahedrons from Trestia as "pseudocrystals", to wit chalcedony pseudomorphs after fluorite.

Zepharovich (1859, 1893) agreed with Ackner's conclusions, but also quoted some different opinions as those of Zerrener and Farber, who established that the cubes were nothing else but "independent quartz crystals". Detailed goniometric measurements made by Ferber indicated angles of 94°15' corresponding to a rhombohedral variety of quartz, previously described at the time by Mohs and Breithaupt. In a paper by Scharff (quoted by Zepharovich, 1859), the author showed that the crystal-faces were "irregularly distorted and full of pits and little ridges, as a result of some superposed minute laminae" and thus tried to explain the angle deflections noticed by Ferber. As a matter of fact, even some fluorite crystals from Zchoppau or Münster Valley frequently exhibited obtuse or acute angles, generally departing very slightly from normal values. Therefore, Scharff maintained that the cubes were pseudomorphs after fluorite but did not reject the possibility that certain smaller cubes with rounded edges and angles could belong to other minerals.

In the last tome of his lexicon, Zepharovich mentioned Mallard who identified — in association with the pseudomorphs crystals — a new species of chalcedony which he called lussatite.

Behrens (1871) and Geinitz (1876) published detailed observations regarding the microscopic structure of chalcedony and stated in their turn that the cubic crystals belong to fluorite. In a band within a chalcedony sample, Geinitz found numerous minute, colourless crystals exhibiting square or hexagonal outlines. The microcrystals displayed an irregular distribution and show obvious relief and transparency. They ranged from 0.01 to 0.03 mm and had cubic or cubic-pyramidal forms which became visible when the microscope lens was slightly moved against the thin section. Not only the author assigned the microcrystals to pseudomorphs after fluorite but also specified that this mineral "should not be regarded as something anterior to chalcedony but as a synchronous phase".

For Bombicci (1900), the cubic crystals from Trestia "do not resemble fluorite at all". Because of their face morphology, the cubes are rather similar to some "...mimetic (?) hexahedral species like boleite, hauerite, pharmakosiderite or melanophlogite. Excepting the colour, a hexahedron from Giona is apt to be confused with a cubic crystal from Trestia". Anyway, as a genetic process, the pseudomorphosis was questionable, so the author suggested instead a so-called "autocrystallization", considering that the cubic forms belong to a peculiar mineralogical species which he called *cubosilicite* and to which he gave the following definition : "... cubosilicite is a form of silica, generally quartzitic with an admix-

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ture of opal, having the appearance of a blue-coloured chalcedony; it occurs as mimetic crystals of cubic form, visible to the naked eye". The author identified organic material in the samples and assigned to it an important role in the genesis of cubosilicite: "... the mixture of quartz and hydrated silica (opal), particularly when traces of hydrocarbon are present — and this is the case of cubosilicite — represents a sufficient reason for the transition from trigonal to cubic system".

Manilici (1962) noticed some "very interesting chalcedony pseudomorphs after a cubic mineral with cubic cleavage, which seems to be formed on behalf of galenite".

X-ray studies carried out by Moțiu and Ghiurcă (1979) showed the prevailing presence of α-quartz. The authors rejected previous opinions (unspecified) according to which the chalcedony of Trestia consisted of melanophlogite.

3. The colour of the Trestia chalcedony and its origin did not focus much attention until 1979 when Moțiu and Ghiurcă dedicated to this subject a detailed study. The results of the investigations made the authors draw the conclusions that the chalcedony contains finely dispersed inclusions of sylvite (KCl) and fluorite (CaF_2). Electron probe microanalysis revealed an excessive amount of potassium in the sylvite inclusions which was inferred from the lower variation amplitude of chlorine content and this was considered the reason for the presence of "F"-type colour centers [5] within the chalcedony. Moreover, the excess of calcium in fluorite determined a so-called "aditive colour". Thus, the authors stated that the two halides played the role of ground for the colour centers which generate the chalcedony's well-known characteristic hues.

Mode of Presentation. Relationship with the Country Rocks

The field observations carried out till now did not confirm the existence of any veins or druses of chalcedony within the andesites. We have identified chalcedony just as fragments scattered in soils or alluvia and never associated or welded with elements of igneous rocks. Without definitely rejecting a possible mode of presentation in the terms described by Koch or Bombicci, we suggest that the chalcedony was formed by solidification of silica gels precipitated from hot springs, either in subaerial conditions or in fissures and cavities of a preexisting alteration crust.

A supposed deposition on an already altered crust explains fairly well why the surfaces of silica fragments are almost never stuck with extrusive or sedimentary rocks [6]. Besides, certain mineralogical aspects which will be presented further on provide evidence for very low temperature genetic processes.

Mineralogical Composition. The Origin of the Crystallographic Forms

It is obvious even from a brief examination that the blue silica fragments consist not only of chalcedony as it may be inferred from the usual denomination of the samples from Trestia.

Under the microscope, the silica fragments reveal clearly separated bands with different degrees of crystallization consisting either of quartz

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or chalcedony and opal (Pl. I, Fig. 1). The bands reflect in a typical stratified structure which is macroscopically enhanced by lighter or darker blue hues.

A significant proportion of quartz is conspicuous. The substitutions of chalcedony by quartz as a result of recrystallization are always clear and eloquent (Pl. II, Fig. 1). Besides, most of the quartz crystals show chalcedony relics, proving that they result from such processes.

The chalcedony forms radial-fibrous aggregates, sometimes of considerable dimensions which reflect a very homogeneous crystallization medium. The straight mutual impingement boundaries between the aggre-

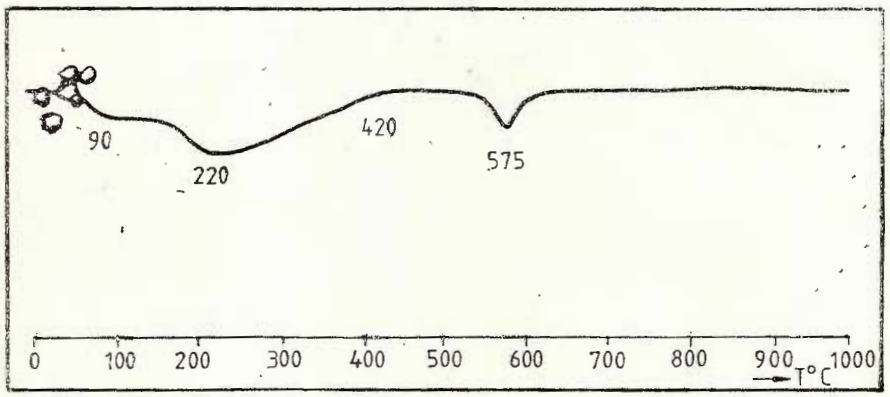

Fig. 1 — Differential thermal curve for the Trestia silica (MOM-type equipment. Analyst : C. Crăciun).

gates (Pl. II, Fig. 2) plead for the same idea. Together with the quartz resulted from recrystallization, chalcedony is part of the cubic crystals; some crystals are formed entirely of chalcedony which definitely puts of the question the assumption that the crystals represent rhombohedral forms of quartz.

Opal is quantitatively reduced and generally has a relict character biased by incipient cryptocrystalline zones. Opal is pointed out by thermal analyse too. On the differential-thermal curve (Fig. 1) between 60 and 420°C, there is a large endothermic effect, divided by several modulations at 90 and 220°C which is produced, at least partly, by the loss of absorbed water and of opal's constitution water. The later is denoted also by the absorption bands at 3500 and 1635 cm^{-1} in the infrared spectrum (Fig. 2a) which correspond respectively, to the stretching and bending vibrations of the H_2O molecule.

The infrared spectrum (Fig. 2a) points out the quartz's specific bands at 465, 510, 695, 785—801 and 1090—1175 cm^{-1}. The modulations at 557 and 670 cm^{-1} are very interesting as they are ascribable to tridymite. Such modulations, however, occur frequently in the spectra of various opals (van der Marel and Beutelspacher, 1976) and they are not necessarily explained by the presence of tridymite in the samples.

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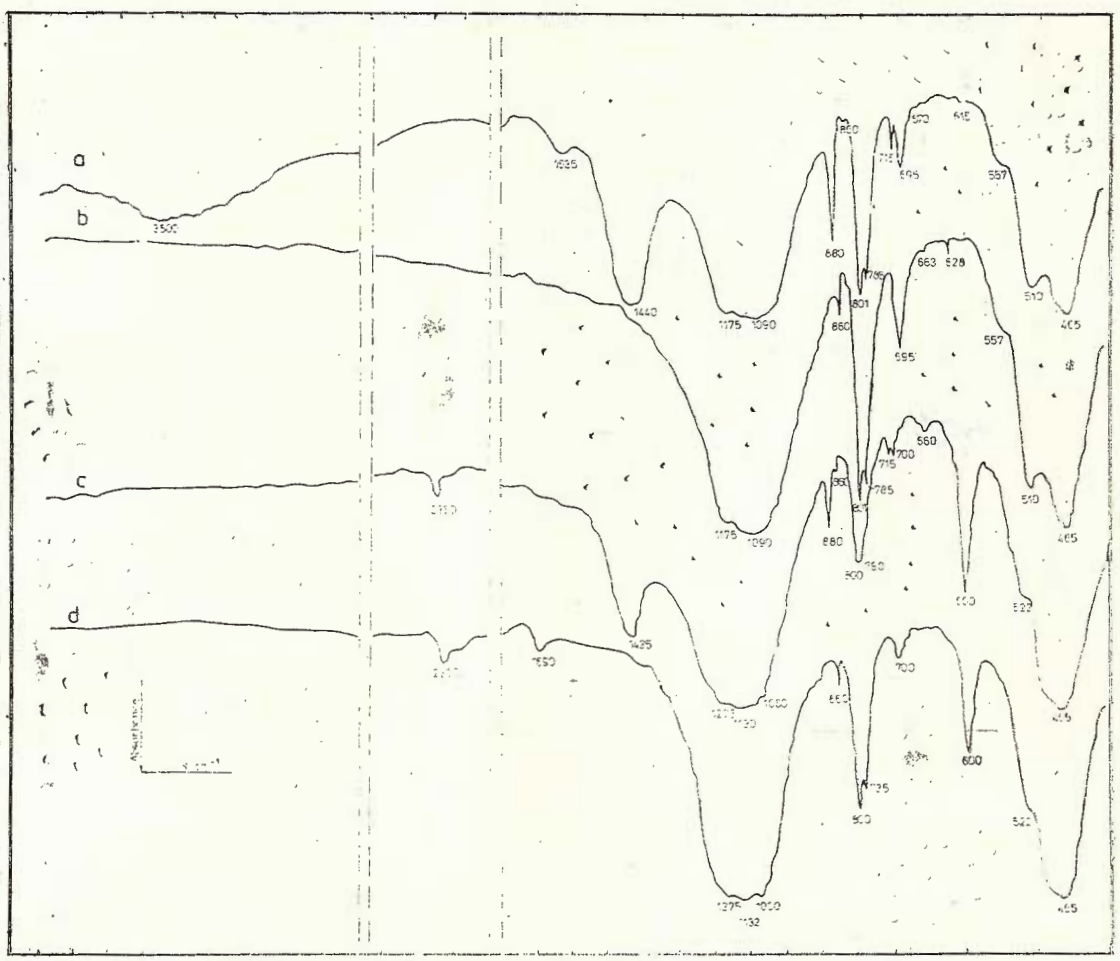

Fig. 2 — Infrared absorption spectra for chalcedony (a, b) and melanophlogite (sample 31.348— Museum of Institute of Geology and Geophysics, Bucharest), (c, d): the „b" and „d" spectra correspond to samples of chalcedony and melanophlogite respectively, calcinated for an hour at 450°C (UR-10 spectrometer. Analyst : T. Drăghici — I.G.G.).

They rather indicate the effect of some tridymite-type bonds, isolated in the opal's disordered lattice which do not form proper crystallographic planes. The absence of any indication for tridymite in repeated X-ray diffraction analysis confirm such suppositions for the Trestia silica as well.

There is evidence that within the silica aggregates from Trestia, the three identified phases : quartz, chalcedony and opal coexisted with an unstable cubic polymorph of SiO_2 : the *melanophlogite.* [7]

The above mentioned evidence and the possible genetic significance of melanophlogite will be examined further on within the discussion regarding the origin of the cubic forms.

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Certain relict microstructures localized in the cubic crystals were found to show remarkable similarity with the regular microzoning of organic films in melanophlogite crystals (Pl. III, Figs. 1—5).

In a layer of chalcedony and fine crystallized, we rediscovered the minute square and hexagonal forms described by Geinitz. Careful examination pointed out that these forms exhibit the same zoning (Fig.

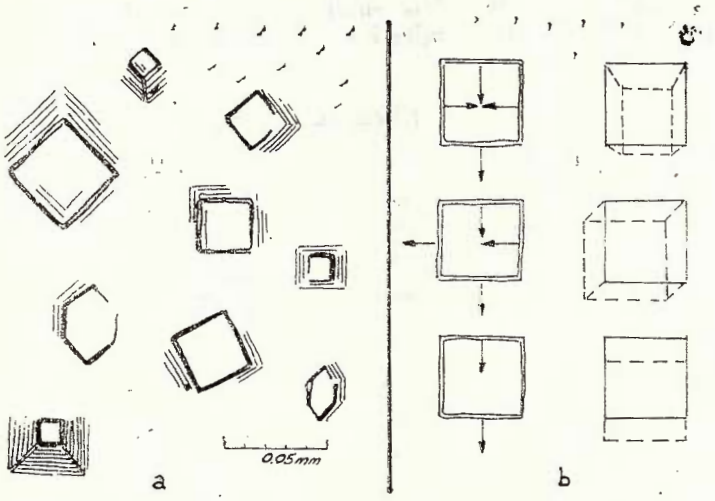

Fig. 3 — a, Microcrystals with a regular zoning resembling melanophlogite. b, Different ways in which the light band outlining microcrystals shifts at slight motions of the microscope lens from good focus, together with the inferred crystalline forms.

3a). Noteworthy that either outside or inside the outline of such "microcrystals" there is the same quartz crystal. Thus, the "obvious relief" found by Geinitz is only apparent because it does not result from the difference between the refringence of the crystal and that of the exterior medium. The narrow band of light, outlining the forms and resembling the "Becke line" is in fact the image of a very thin film of impurities which envelops the microcrystals. This could be inferred as well from the way the light band shifts at slight motions of the microscope lens from good focus (Fig. 3b).

Together with the described zoning there is the striking resemblance between the forms from Trestia and melanophlogite cubes of equal size (Pl. IV, Figs. 1—4) that led us to consider the cubic crystals as paramorphs after melanophlogite.

As an indirect evidence, it could be mentioned that chemical analyses of the chalcedony reveal 0.707 per cent organic carbon (analyst: E. Opriş, IPGG, Bucharest), suggesting that the silica was formed in a medium rich in organic material, perfectly consistent with the occurrence of melanophlogite. The considerable amount of organic material reveals itself in the opacity of the heated silica which is obvious even in thin sections (Pl. IV, Fig. 5).

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As regards the organic material, there are some very significant comparisons to be made between the infrared spectra of melanophlogite and that of the chalcedony from Trestia. The "a" and "c" spectra in Figure 2 belong to Trestia chalcedony and melanophlogite respectively, the "b" and "d" spectra belong to the same samples but after they were heated at 450° C. Few characteristic bands in melanophlogite spectrum, such as the 660, 715, 880 and 1435 cm^{-1} are missing in the heated sample's spectrum. The same bands, excepting the one at 660 cm^{-1}, are visible in the chalcedony's spectrum, too.

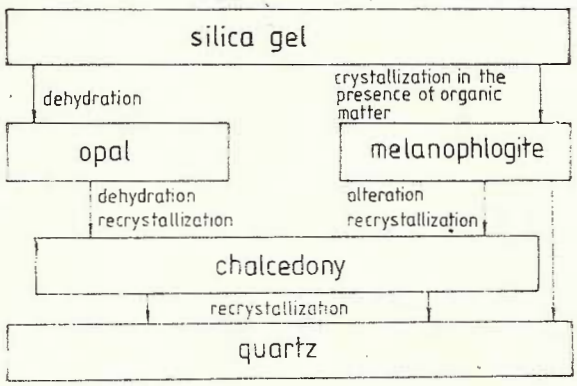

Fig. 4 — Probable phase transitions scheme that occurred in Trestia silica.

In a similar manner, the bands are missing in the spectrum corresponding to the heated sample. This fact suggests the existence of a common component, sensitive to high temperatures, in the two minerals.

X-ray powder diffraction analysis rendered, in addition to the quartz's specific reflexes, some faint peaks corresponding to d/n: 5.960—6.004—6.077; 4.700; 3.870; 3.690—3.709—3.710; 3.579—3.588 Å, which could indicate melanophlogite relics. There were eight determinations on different samples and with diffractometers of various types (Philips P.W., DRON—3, DRON—20, TUR M—61). The repeated appearance of such peaks in diverse testing conditions, increases the confidence that can be reposed in these results.

The melanophlogite entails by its occurrence low formation temperatures, consistent with the presumed surface deposition, described in the previous chapter. Moreover, it completes the transformation sequence which was identified in the silica aggregates. The supposed scheme of this sequence is rendered in Figure 4.

The Cause of the Blue Colour

The involvement of halides such as sylvite and fluorite in the generation of the chalcedony's blue colour (Moțiu and Ghiurcă, 1979) sets quite a few problems regarding both the colour saturation and the hues possible to result from such processes.

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The existence of the "F"—type colour centers in alkaline halides of various occurrence is more or less accidental; cases of large extension of such phenomena are very seldom. Hence, it is presumable that a coloured alkaline halide, moreover,finely dispersed in a colourless solid would provide nothing else but a faint chromatic effect or rather no effect at all.

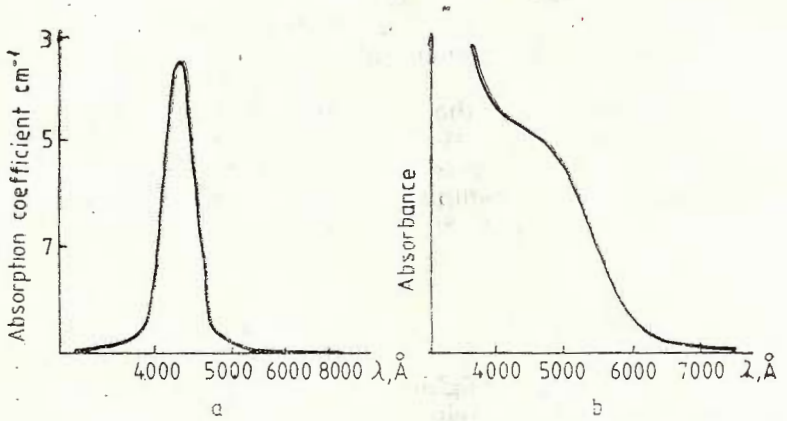

Fig. 5 — Light absorption spectra of "F"-centers in sylvite.
a) according to Jacobs and Tompkins (1955, fide Murgulescu and Segal, 1979) and b) according to Platonov (1976).

It is noteworthy that the "F" centers in different alkaline halides have specific absorption ranges within the visible spectrum. Thus, "F" centers in sylvite have an absorption maximum between 5000 and 5500 Å (green) (Jacobs and Tompkins, 1955, fide Murgulescu and Segal, 1979), therefore inducing a purple colour (Fig. 5a). Platonov (1976) indicated for the same sylvite "F" centers a strong absorption in the entire zone comprised between 4000 and 6000 Å (Fig. 5b) that makes the halide to appear pink-coloured. For this reason, it is an extremely difficult task to find out what could be mixed with purple-pink sylvite to obtain the well known blue hues of Trestia chalcedony.

There is a remarkable optical property — the *opalescence* — that considerably diminishes the range of possible causes of the blue colour. As defined by Rutzkov (1960), opalescence refers both to the dim aspect of the substance and to the fact that its colour in transmitted light as compared to the one in reflected light is always displaced towards the red zone of the visible spectrum. Such state is obvious in the case of Trestia chalcedony, too, as thin laminae always show a marked orange or reddish hue when observed in transmitted light.

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The opalescence is a defining property for colloidal systems with high dispersion degree, that makes us definitely reject any possible atomic or molecular ground for the colour of the Trestia chalcedony.

The causes of the optical phenomena which describes the opalescence of solid sols are well-known and therefore we shall not insist upon them; we shall emphasize just the fact that selective absorption of light in colloidal system does not depend too much upon the chemical composition of the dispersed substance. In this respect, a greater influence is exerted by the dimensions of the dispersed particles and by the difference between the refraction indexes of dispersed substance and dispersive medium.

Few indications exist about the nature of the dispersed phase in Trestia chalcedony. Thus, the fact that after heating fragments of chalcedony the blue colour turns to white and the opalescence becomes less obvious, suggests the instability of the dispersed substance to high temperatures, probably due to its fluid or organic nature.

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[3] Some quoted papers like those of Gautieri (1800), Blum (1825) or Hein (1903) were omitted from this chapter as they were not available.

[4] At present, an unknown toponymy.

[5] The „F"-type colour centers occur in non-stoicheiometric crystals of alkaline halides and consists of single, free electrons in lattice vacancies. Visible radiation may cause transitions of electrons' energy level in the conductivity band of the lattice and thus, determine selective absorption of light. Spectral bands that arise from "F"-centers electron transitions are characteristic of each halide.

[6] Very recent reports by O. Edelstein indicate a veinlet of blue chalcedony in a limestone fissure. It is a singular case of chalcedony on a substratus and it could be explained as resulting from silica gels trickled in the fissure *from the surface*. The welding was favoured by the fact that limestones usually exhibit clean, unaltered surfaces.

[7] Melanophlogite is a low temperature ($< 112°C$) cubic polymorph of SiO_2. The index of refraction is 1.425 and the calculated density 1.99 g/cm^3. Chemical analysis reveals C, H and S content of at least 6%. These elements are present in organic films trapped on the cube growth surfaces. According to Friedel (1890) and others, fresh crystals of melanophlogite are rare while quartz paramorphs after melanophlogite are very common (Appleman and Skinner, 1963).

[8] Bands such as 465, 522, 700 and 790—800 cm^{-1} in the melanophlogite infrared spectrum are due to quartz formed while grinding the sample of melanophlogite. Appleman and Skinner (1963) mentioned the fact that melanophlogite is unstable in a stress field and therefore very sensitive to such treatments as grinding. The authors showed that within half an hour of grinding 100 mg of melanophlogite in an ordinary mechanical driven laboratory mortar and pestle, more than 90% of the sample inverted to quartz and within an hour total inversion had occurred.

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CALCEDONIA DE LA TRESTIA — MARAMUREŞ

(Rezumat)

Între alte ocurenţe de calcedonie, care în decursul timpului au atras atenţia cercetării mineralogice, cea de la Trestia — Maramureş şi-a cîştigat o anumită celebritate. Nu numai culoarea albastră şi neobişnuitele sale forme cristalografice o justifică, dar şi faptul că de aproape două secole, încercările de stabilire a originii acestora au rămas cel mai adesea în stadiul unor ipoteze contradictorii.

Referirile de ordin istoric şi observaţiile proprii cuprinse în această lucrare au fost sistematizate urmărind trei categorii de probleme : 1. forma de zăcămînt a calcedoniei şi relaţiile ei cu rocile înconjurătoare, 2. compoziţia mineralogică şi originea formelor cubice, 3. cauzele coloraţiei albastre.

1. Cu privire la forma de zăcămînt a calcedoniei există numeroase menţiuni, mai ales în literatura veche. Acestea se referă la : fragmente în soluri şi aluviuni (Fichtel, 1971 ; Ackner, 1855), filoane şi geode în

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roci eruptive (Beudant, 1822; Koch, 1893; Bombicci, 1900), geode în trunchiuri de lemn silicifiat (Bombicci, 1900).

Observațiile proprii nu confirmă existența unor filoane sau geode de silice în rocile eruptive. Am identificat calcedonie numai ca fragmente în soluri și aluviuni și niciodată în asociație cu elemente ale unui substrat de rocă eruptivă. Fără a respinge definitiv ideea unei forme de zăcămînt de tipul celei menționate de Koch sau Bombicci, sugerăm posibilitatea formării silicei prin consolidarea unor geluri precipitate din izvoare fierbinți fie în condiții subaeriene, fie în fisuri și cavități ale unei scoarțe de alterare preexistente. O eventuală depunere pe un substrat deja alterat, ar oferi o bună explicație pentru faptul că suprafețele fragmentelor silicioase găsite nu sînt aproape niciodată cimentate de fragmente de rocă eruptivă sau sedimentară (vezi nota 6).

2. Silicea albastră de la Trestia este alcătuită din calcedonie alături de care participă în proporții însemnate cuarțul și opalul (pl.I, fig.1). Între aceste trei faze se stabilesc raporturi de înlocuire prin recristalizare (pl. II, fig. 1).

De-a lungul timpului cristalele hexaedrice de la Trestia au fost fie atribuite unor pseudomorfoze după fluorină (Ackner, 1855; Zepharovich, 1859—1893; Behrens, 1871; Geinitz, 1876; Scharff, 1893; Mallard, 1893) sau după galenă (Manilici, 1962), fie considerate forme romboedrice ale cuarțului (Zerrener, Ferber, 1873) sau forme cristalografice ale unei specii mineralogice aparte: cubosilicitul (Bombicci, 1900).

Anumite microstructuri relicte localizate în cristalele cubice atrag atenția datorită unor remarcabile similitudini cu microzonalitățile peliculelor de material organic din cristalele de melanoflogit (pl. III, fig. 1—5). Aceste microzonalități, ca și asemănările frapante care ies în evidență cînd se compară, în cadrul aceluiași ordin de mărime, formele de la Trestia cu cristale de melanoflogit (pl. IV, fig. 1—4) ne determină să considerăm cristalele cubice drept rezultatul unor paramorfoze după melanoflogit. Pentru această idee pledează și rezultatele analizelor chimice, care scot în evidență un conținut ridicat de substanță organică.

3. Cauzele colorației albastre a calcedoniei de la Trestia nu au stat în mod deosebit în atenția cercetătorilor pînă în 1979 cînd Moțiu și Ghiurcă au dedicat acestei probleme un studiu amănunțit. Determinările făcute i-au condus pe autori la concluzia că în calcedonie se găsesc fin și uniform dispersate silvină (KCl) și fluorină (CaF_2). Un anumit exces de metal în silvină — dedus din amplitudinea mai scăzută de variație a curbei de conținut a clorului în raport cu cea a potasiului, la analiza cu microsonda electronică — este considerat responsabil pentru apariția unor centri de culoare de tip „F" (vezi nota 5). Totodată, excesul de Ca din fluorină determină apariția unei așa-numite „colorații aditive". Celor două halogenuri din calcedonie li se atribuie rolul de „suport al centrilor de culoare care îi conferă acesteia nuanțele caracteristice, bine cunoscute".

Implicarea halogenurilor alcaline și alcalino-pămîntoase (silvină și respectiv fluorină) în procesele care generează culoarea calcedoniei de Trestia (Moțiu și Ghiurcă, 1979) ridică o serie de probleme ce privesc

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atît saturaţia culorii cît şi nuanţele care s-ar realiza printr-un atare mecanism.

Apariţia unor centri „F" în halogenurile alcaline din ocurenţe naturale are un caracter accidental, cazuri de largă extindere a unor astfel de coloraţii fiind rareori întîlnite. Este de presupus oricum, că prin dispersarea fină a unei halogenuri alcaline colorate într-un mediu solid incolor, efectul cromatic ar fi puternic estompat dacă nu chiar anulat.

Important este şi faptul că centrii „F" din diferitele halogenuri alcaline au *domenii proprii de absorbţie în spectrul vizibil.* Silvina cu centri „F" prezintă, după Jacobs şi Tomkins (1955—fide Murgulescu şi Segal, 1979), un maxim de absorbţie în zona 5000—5500 Å (verde), aşa încît apare colorată în purpuriu (fig. 5a). Platonov (1976) indică pentru acelaşi tip de silvină o absorbţie puternică în toată zona dintre 4000 şi 6000 Å (fig. 5b), ceea ce explică de ce silvina cu centri „F" apare colorată în roz. Din acest motiv, este dificil de imaginat cu ce s-ar putea combina silvina colorată astfel pentru a conferi calcedoniei de Trestia culoarea albastră atît de caracteristică.

Observaţii simple arată că faţă de culoarea calcedoniei în lumină reflectată, cea văzută în lumină transmisă este deplasată spre zona roşie a spectrului. Acest fenomen, împreună cu aspectul tulbure al agregatelor de silice, defineşte *opalescenţa* calcedoniei de Trestia, proprietate definitorie pentru un sistem coloidal cu un grad înalt de dispersie şi care exclude în acelaşi timp, eventuale cauze de ordin atomic sau molecular ale coloraţiei.

Aşadar, culoarea albastră a calcedoniei de la Trestia se datorează absorbţiei şi difuziei selective a luminii la nivelul unor particule dispersate coloidal.

Asupra fazei dispersate coloidal în silicea de Trestia există puţine indicaţii : comportamentul termic particular al culorii (prin încălzire culoarea în lumină reflectată trece de la albastru la alb, iar opalescenţa mai puţin accentuată) sugerează instabilitatea substanţei dispersate la temperaturi înalte, probabil în legătură fie cu caracterul său fluid fie cu natura sa organică.

EXPLANATION OF PLATES

Plate I

Fig. 1 — Layers with different crystallization degrees within a silica aggregate from Trestia, resulting in a parallel banded structure. Q-quartz; C-chalcedony; O-opal; RC-relics of chalcedony; RO-relics of opal; RM-relics of melanophlogite microstructures; RC-recrystallizations. N+, 15 ×.

Plate II

Fig. 1 — Chalcedony (c) replaced by quartz (q) in a process of recrystallization. In the upper part of the image one can perceive incipient cryptocrystalline zones developed on opal. N+, 90 ×.

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Fig. 2 — Radial-fibrous aggregates of chalcedony with straight compromise boundaries. In the cubic crystal from below, one can distinguish relics of chalcedony in quartz. N+, ×45.

Plate III

Fig. 1 — Cubic crystal of melanophlogite within which organic films cause a characteristic regular zoning (after Appleman and Skinner — 1963). N ∥, 150 ×.

Fig. 2—5 — Relict microstructures identified in the cubic crystals from Trestia; a remarkable similarity with the melanophlogite zoning is obvious. N ∥, 150 ×.

Plate IV

Fig. 1 — Melanophlogite crystals from Sicily ... (sample 31348 — Museum of the Institute of Geology and Geophysics Bucharest), 10 ×.

Fig. 2—4 ... and cubic crystals of similar size order from Trestia, 10 ×.

Fig. 5 — Calcinated silica showing obvious opacity even in thin sections, N//, 70 ×.

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D. S. Inst. Geol. Geofiz. vol. 74/1 (1987), 1989, pag. 35—49
1. MINERALOGIE—PETROLOGIE

CONTRIBUTIONS TO THE KNOWLEDGE OF SULFATES IN ROMANIA [1]

by

MIHAI TATU [2]

Sulfates. Rozenite. Szmikite. Halotrichite. Pickeringite. Copiapite. Sulfides. Oxidation zone. Chemical composition. X-ray data. Infrared spectra. East Carpathians — Neogene eruptive-Quaternary — Gutin Mountains. Apuseni Mountains — Lower Cretaceous-Paleogene magmatites — Băiţa.

Abstract

In the oxidation zone of some base-metal deposits rosenite, szmikite, halotrichite, pickeringite (Borzaş-East Carpathians-Gutîi Neogene volcanic zone) and zincocopiapite (Băişoara—Apuseni Mountains—Gilău massif) were identified, the last three new minerals in Romania.

Résumé

Contributions à la connaissance des sulfates de la Roumanie. Dans la zone d'oxidation de certaines minéralisations de sulfures métalliques on a identifié la rosénite, la szmikite, l'halotrichite, la pickeringite (Borzaş—Carpathes Orientales, zone volcanique néogène de Gutîi) et zinccopiapite (Băişoara-monts Apuseni, le massif de Gilău); les derniers trois minéraux sont nouveaux pour la Roumanie.

In the Borzaş auriferous stock, hosted in Pannonian andesitic structures, rosenite, szmikite, halotrichite, and pickeringite occur in the oxidation zone, covering the walls of a temporarily abandoned gallery. They appear

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as efflorescences, crusts, concretionary or stalactitic formations. Zincocopiapite was also identified in the oxidation zone of a quartz and sulfate vein (Băișoara), affiliated to banatites affecting the Precambrian metamorphites of the Baia de Arieș Series. It is of note that halotrichite, pickeringite and zincocopiapite are new minerals in Romania. The complex (X-ray, IR, chemical) study made possible the diagnosis and emphasizing of the characteristic features of the mentioned minerals.

Morphological Aspects

Rosenite occurs as white-yellowish individuals with a long-prismatic and acicular habitus (Pl. I, Fig. 1) disposed as a rosette within stalactitic and crustal formations. Under the microscope (3000 x) the development of the prism face (010) is observed (Pl. I, Figs. 2, 3), as in case of Iacobeni (Bălan, 1976).

Szmikite was first described in the world at Baia Sprie by Schroeckinger in 1877. At Borzaș, szmikite crystals show a short-prismatic habitus; they are of a white-pinkish colour and form mamelon-like, saccharoidal concretions (Pl. II, Figs. 1, 2, 3).

Halotrichite is found as fibrous and acicular crystals (Pl. III, Figs. 1, 2, 3), of a white-yellowish up to yellowish-reddish colour, forming concretions, locally with textile or radial aspect.

Pickeringite occurs as white-slightly yellowish acicular individuals, constituting compact aggregates (Pl. IV, Figs. 1, 2, 3).

Zincocopiapite appears as yellow-slightly greenish, friable concretions made up of individuals flattened after (010) (Pl. V, Figs. 1, 2, 3).

X-ray Data

X-ray diffraction emphasized for rosenite, szmikite, halotrichite, pickeringite and zincocopiapite reticular distances (dÅ) and intensities (I) comparable with those found in the relevant literature (Tab. 1, 2, 3, 4, 5).

IR Spectra

The IR[3] absorption spectra analysis allowed the identification of the bands characteristic of each of the study mineral.

Rosenite (Fig. 1) shows two strong absorption bands at 1110 cm^{-1} molecule SO_4), 3440 cm^{-1} (crystallization water) and one strong absorp- (tion band at 1620 cm^{-1} (crystallization water).

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TABLE 1

X-ray data of rosenite

dÅ	hkl	I/I_1	dÅ	I	dÅ	I
	ASTM (1)		Friedland (2)		Borzaș (3)	
6.85	011,020	50	6.80	20	6.85	4.4
5.46	110	90	5.45	55	5.44	7.9
5.17	021	5			5.19	1.5
4.73	101	10			4.74	2.5
4.47	111,120	100	4.50	100	4.46	10.0
3.97	002	70	3.99	60	3.98	8.9
3.61	130	10			3.61	1.7
3.40	040	60	3.39	37	3.40	4.8
3.27	131	10			3.28	2.2
3.22	$\bar{1}12$	50	3.24	42	3.22	4.0
2.985	032	40	2.99	25		
2.953	140	50	2.95	28	2.96	4.5
2.906	210	5			2.91	1.1
2.770	$\bar{1}41$	10			2.77	1.6
2.722	220,211	10	2.72	5	2.72	1.4
2.673	$\bar{1}32$	5			2.68	0.9
2.569	051,221	40	2.58	25	2.58	2.9
2.470	150,023	5			2.475	1.0
2.430	$\bar{1}30$	30			2.44	2.4
2.371	$\bar{2}31,\bar{1}42+$	20	2.36	5	2.38	2.0
2.360	142,151+	20				
2.333	212	5			2.285	2.0
2.286	033,$\bar{1}23$	5				
2.266	123,060	30	2.26	25	2.245	2.4
2.236	052,240+	10			2.190	0.7
2.179	061	5				
2.142	241,$\bar{1}33$	5			2.120	0.9
2.112	160,232	5				
2.049	$\bar{1}61,161$	5	2.07	20		
1.969	$\bar{1}43,062+$	20	1.97	10	1.972	1.5
1.944	213,$\bar{2}51+$	10			1.945	1.0
1.890	053,223	20			1.890	1.2
1.871	$\bar{1}62,162$	10			1.875	0.9
1.819	$\bar{2}33$	5				
1.798	153,260+	20			1.800	1.2
1.755	$\bar{2}61,261$	10			1.762	0.7
1.725	063	20			1.727	1.5

(1) ASTM
(2) Störr et al. (1983)
(3) Tatu (1987)

Diffractometer TUR M—61, Rad. Co—Kα, filter Fe, 30Kv, 20 mA.
Analyst: I. Vanghelie, I.G.G.

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TABLE 2

X-ray data of szmikite

dÅ	hkl ASTM (1)	I/I_1	Å Borzaș (2)	I
4.90	011	35	4.90	3.1
4.83	$1\bar{1}1$	10		
3.80	020	6	3 82	0.5
3.499	200, 111	100	3.52	10.0
3.433	$21\bar{1}$	25	3.44	2.6
3.343	120	20	3.36	2.4
3.212	002	4		
3.135	$20\bar{2}$	40	3.145	3.4
2.602	$12\bar{2}$	25		
2.573	220	35	2.585	3.2
2.358	031	8	2.367	1.1
2.246	$11\bar{3}$	12	2.240	1.6
2.230	$21\bar{3}$	4		
2.135	131	8	2.140	1.0
2.122	$23\bar{1}$	10		
2.105	122	18	2.103	1.7
2.067	$32\bar{2}$	4	2.067	0.4
2.017	$31\bar{3}$	12	2.010	1.5
1.984	320	8		
1.974	202	6	1.965	0.7
1.875	311	10	1.873	1.0
1 852	$41\bar{1}$	4		
1.835	140	6		
1.786	113	4		
1.782	$20\bar{4}$	4	1.781	0.5
1.754	222	8	1.751	0.8
1.746	400	6		
1.740	$41\bar{3}$	8		
1.719	$42\bar{2}$	16	1.719	1.6
1.679	$14\bar{2}$	6	1.678	0.7
1.672	240	8		
1.638	042, 033	8	1.642	1.2
1.614	$22\bar{4}$, $33\bar{3}$	10	1.616	1.0
1.607	004	6	1.600	0.8
1.583	$12\bar{4}$	6		
1.571	$40\bar{4}$	4		
1.563	$32\bar{4}$	6	1.566	0.6
1.541	331	10	1.542	0.9
1.525	431	6		

(1) ASTM

(2) Tatu (1987)

Diffractometer DRON-3, Rad. Cu—Kα, filter Ni, 30 Kv, 20 mA.

Analyst: I. Vanghelie, I.G.G.

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TABLE 3

X-ray data of halotrichite

dÅ	hkl	I/I_1	dÅ	I	dÅ	I
	ASTM (1)		California (2)		Borzaş (3)	
10.0	200,120	15	10.40	17		
9.5	210	15	9.50	13	9.50	4.4
7.82	220	10	7.82	11	8.05	3.5
6.02	040,320	30	6.02	28	6.03	4.6
5.24	240,121	15	5.24	16	5.31	4.2
4.77	150,420	100	4.77	100	4.80	10.0
4.62	$23\bar{1},32\bar{1}$	30	4.62	31	4.58	4.2
4.29	311,401	55	4.29	55	4.29	9.5
4.09	510,321	45	4.09	45	4.09	7.0
3.95	160,350	35	3.95	33	3.97	3.1
3.75	401,260	40	3.75	40	3.77	4.8
3.48	$\bar{3}51,360+$	100	3.48	100	3.49	10.0
3.30	620,261	20	3.30	18	3.27	4.9
3.16		15	3.16	17	3.17	2.2
3.05		15	3.05	17	3.05	2.9
2.96		20	2.96	18	2.98	3.7
2.86		30	2.86	27	2.885	4.5
2.76		20	2.76	20	2.78	2.5
2.67		25	2.67	23	2.675	4.0
2.61		20	2.61	18	2.602	2.0
2.55		20	2.55	18	2.565	1.8
2.447		10	2.447	10	2.460	1.1
2.390		5	2.390	6	2.405	0.8
2.279		10	2.279	12	2.280	1.7
2.231		5	2.231	6	2.232	0.8
2.009		15	2.009	14	2.015	1.5
1.947		5	1.947	7		
1.868		20	1.868	20	1.873	1.7
1.776		5	1.776	6		
1.663		5	1.663	6		

(1) ASTM
(2) Baur, Sand (1957)
(3) Tatu (1987)

Diffractometer DRON-3, Rad. Cu-K, filter Ni, 30 Kv, 20 mA.
Analyst: I. Vanghelie, I.G.G.

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TABLE 4

X-ray data of pickeringite

dÅ	hkl	I/I_1	dÅ	I
	ASTM (1)		Borzaș (2)	
10.6	200,100	6	10.53	1.9
9.7	210	18	9.70	2.4
8.0	220	12	8.05	2.3
7.12	Ni	4		
6.73	310	4	6.67	1.3
6.08	040,320+	20	6.025	3.1
5.87	140	10	5.86	1.2
5.31	240	14	5.28	1.5
4.97	211	20		
4.82	$\bar{3}11$,420	100	4.80	10.0
4.69	131,221	10		
4.62	$\bar{2}31$	14	4.60	1.6
4.40	250	14	4.40	2.1
4.32	$\bar{1}41$,311+	35	4.33	2.5
4.18	141,$\bar{4}01$	20	4.16	1.8
4.122	$\bar{2}41$,321	30		
3.97	160,350+	20B	3.95	2.9
3.791	$\bar{1}51$,$\bar{3}41$	30	3.77	3.3
3.67	$\bar{2}51$	6	3.62	0,9
3.510	251	90	3.493	9.9
3.459	$\bar{4}41$	12		
3.344	$\bar{1}61$,620	12	3.34	1.5
3.285	161,270+	12	3.3	1.1
3.182	521,630	10	3.185	0.9
3.05		8B	3.04	
3.03		6B		
2,978		10	2.98	0.9
2.898		20	2.893	1.5
2.836		14	2.840	1.2
2.788		10	2.780	1.4
2.766		10		
2.72		8B		
2.687		20	2.682	1.8
2.618		8B	2.615	0.9
2.56		10B	2.565	1.1
2.52		6B		
2.47		8B	2.460	0.8
2.40		8B	2.40	1.6
2.304		10		
2.284		14	2.284	0.9

(1) ASTM
(2) Tatu (1987)

Diffractometer DRON-3, Rad. Cu-Kα, filter Ni, 30 Kv, 20 mA.
Analyst : I. Vanghelie, I.G.G.

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TABLE 5

X-ray data of aluminoçopiapite (1, 2) and zincoçopiapite (3, 4)

dÅ	hkl	I/I_1	dÅ	I	dÅ	I	dÅ	I	dÅ	I
ASTM(1)			Alaska(2)		Tsudam(3)		Frieden(4)		Băișoara(5)	
18.1	010	80	17.891	8			17.8	8	17.9	10
9.20	020	100	8.945	10	9.25	10	8.86	10	9.1	9.7
7.13	001	5	7.063							
6.82	$0\bar{1}1$	10	6.812				6.83	2		
6.51	Ni	5								
6.17	110	70	6.194	7	6.12	8	5.99	7	6.2	4.9
5.58	$\bar{1}01$	80	5.456		5.66	5	5.51	7	5.46	1.2
5.32	$\bar{1}\bar{1}1$	30	5.317				5.35	3		
4.68	$\bar{1}21$	30	4.681	3			4.68	3	4.756	1.0
4.48	040	20	4.473	2						
4.31	$1\bar{2}1$	20	4.223				4.37	2	4.22	0.9
4.20	$\bar{1}31,111$	30	4.196	3			4.15	3	4.18	1.1
4.01	$\bar{1}3\bar{1}$	30	4.011	3			4.03	3	4.00	0.9
3.88	131	10	3.770				3.85	1	3.675	0.1
3.58	$\bar{1}41,050$	50	3.578	5	3.56	6	3.55	7	3.57	0.9
3.50	$200,\bar{1}02$	50	3.484				3.52	8	3.48	1.3
3.45	$\bar{2}10,\bar{2}20$	10	3.432				3.44	3	3.43	0.9
3.36	$\bar{2}21,210$	5	3.399						3.39	1.3
3.29	$\bar{1}\bar{2}2$	5	3.272	<1						
3.22	$\bar{1}22$	10	3.222	1			3.21	2		
3.14	$\bar{2}31$	5	3.140	<1			3.12	4		
3.04	$1\bar{5}1,1\bar{3}2$	10	3.057				3.08	4	3.05	0.9
2.92	$0\bar{4}2\cdot1\bar{6}0$	10	2.921	1			2.94	3		
2.83	230,202	5	2.830	<1						
2.77	$\bar{2}11,\bar{2}12$	5	2.770	<1					2.765	0.7
2.73	$\bar{2}\bar{2}2,\bar{2}\bar{5}0$	5	2.728	<1						
2.67	$\bar{2}\bar{2}2,0\bar{5}2$	10	2.687							
2.62	$221,1\bar{4}2$	5	2.619	<1						
2.508		5	2.508	<1					2.51	0.5
2.485		5	2.486	<1						
2.360		10	2.360	<1						
2.289		5	2.289	<1						
2.165		5	2.165	<1						
2.101		5	2.101	<1						
2.063		20	2.063	2					2.08	0.5
2.009		5	2.009	<1						
1.989		5	1.989	<1						
1.825		5	1.825	<1						
1.727		5	1.727	<1						
1.668		5	1.668	<1						

(1) ASTM
(2) Jolly, Foster (1967)
(3) Tu Kwang-chin et al. (1964)
(4) Störr et al. (1983)
(5) Tatu (1987)

Diffractometer Tur M—61, Rad. Co—Kα, filter Fe, 30 Kv, 20 mA.
Analyst: I. Vanghelie, I.G.G.

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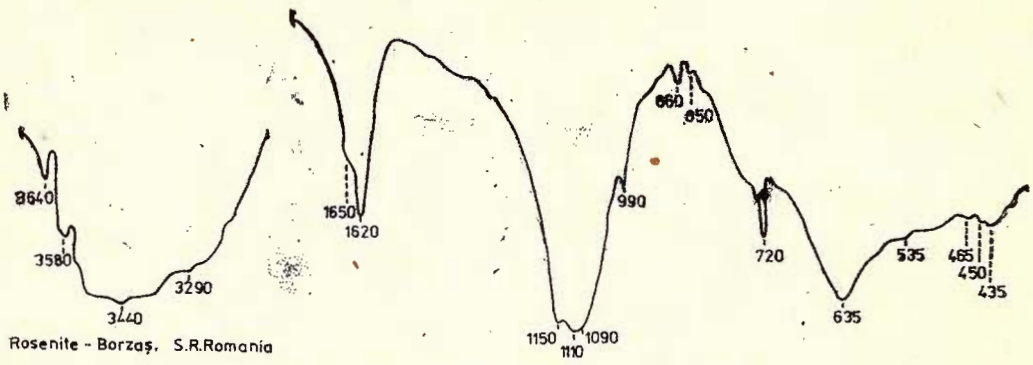

Fig. 1—IR absorption spectra of rosenite.

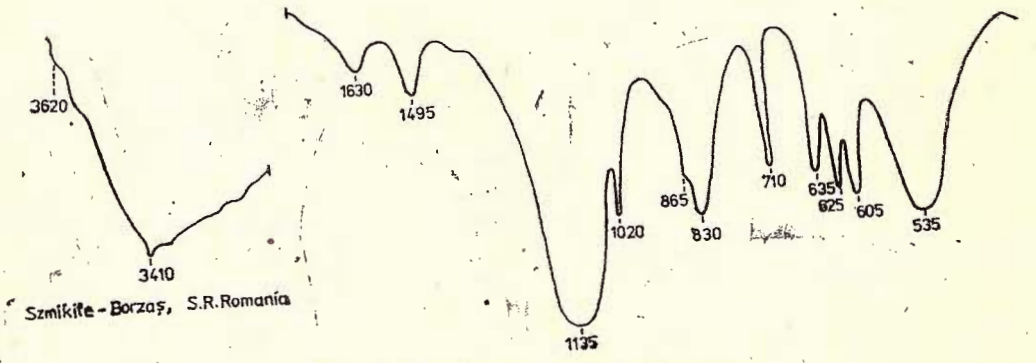

Fig. 2—IR absorption spectrum of szmikite.

Szmikite (Fig. 2) presents two very strong absorption bands at 1135 cm^{-1} (molecule SO_4), 3410 cm^{-1} (crystallization water) and one weak absorption band at 1630 cm^{-1} (crystallization water).

Halotrichite (Fig. 3) occurring at Borzaş, as compared with that at Almeda (California; Marel, Beutelspacher, 1976), shows the band for crystallization water shifted from 3380 cm^{-1} to 3425 cm^{-1} and much more intense, the inflexion being maintained at 3575 cm^{-1}. Another band typical of the crystallization water, with a high intensity, occurs at 1640 cm^{-1}. As to molecule SO_4 it has a very strong absorption band at 1110 cm^{-1} without the complications from 1075 cm^{-1} and 1048 cm^{-1}, but with the inflexions from 1153 cm^{-1} and 950 cm^{-1}.

Pickeringite (Fig. 4) found at Borzaş displays a strong absorption band at 3385 cm^{-1} for the crystallization water, slightly shifted as compared with that of pickeringite found at Tucumari (New Mexico; Marel, Beutelspacher, 1976); the inflexion at 3580 cm^{-1} is preserved. The existence of an intense band at 1640 cm^{-1} for the crystallization water is also observed. As regards molecule SO_4 it shows a very strong absorption band at 1115 cm^{-1} with inflexion at 1074 cm^{-1}.

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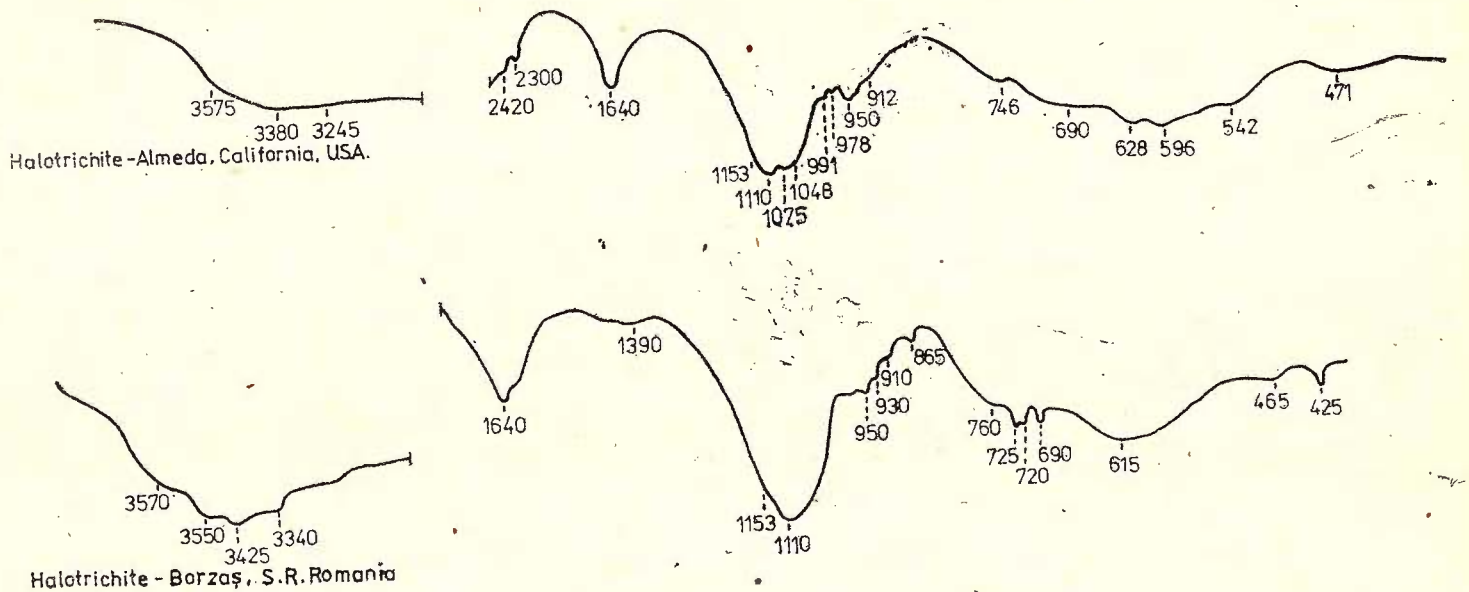

Fig. 3—IR absorption spectrum of halotrichite.

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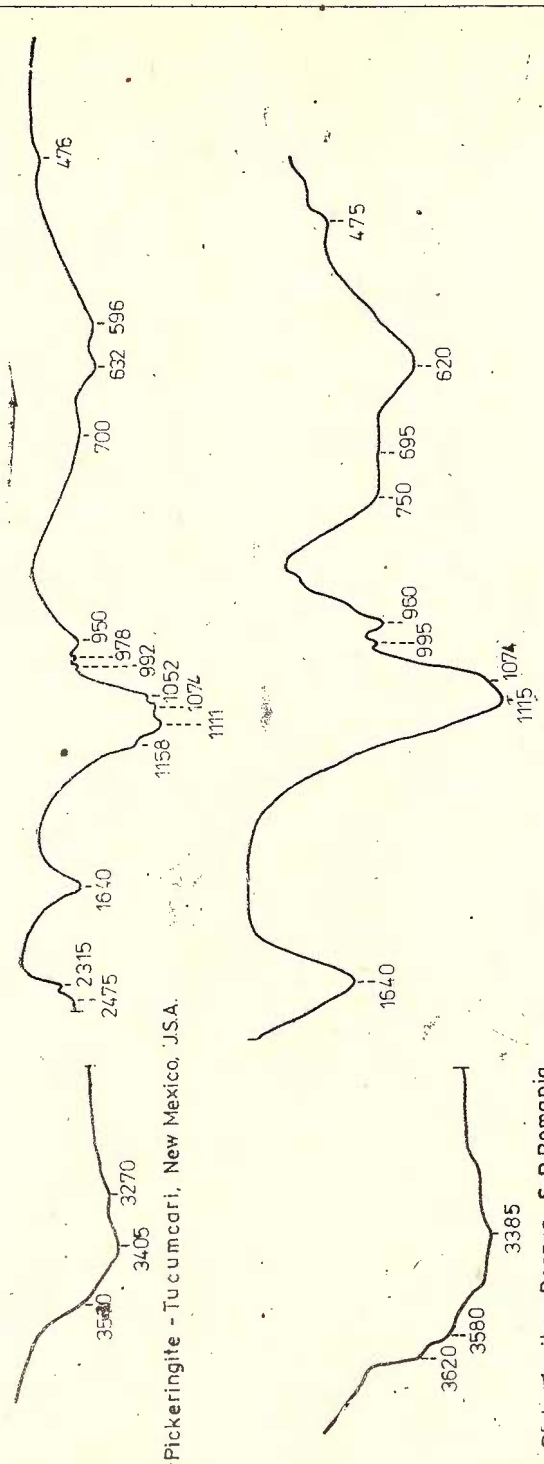

Fig. 4 — IR absorption spectrum of pickeringite.

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Zincocopiapite (Fig. 5) found at Băişoara presents a very strong absorption band at 3430 cm^{-1} (with inflexions at 3490 cm^{-1}, 3230 cm^{-1} and 3140 cm^{-1}) and an intense band at 1630 cm^{-1} for the crystallization water; the intensities differ from those of the copiapite found at Borate (California; Marel, Beutelspacher, 1976). As to molecule SO_4 a very strong absorption band occurs at 1115 cm^{-1} with the characteristic doublet at 1020 cm^{-1} and 1000 cm^{-1}.

Chemistry Data

Rosenite ($FeSO_4 \cdot 4H_2O$) and szmikite ($MnSO_4 \cdot H_2O$) belong to normally hydrated sulfates with only one cation in their composition. Unfortunately up till now there are no conclusive data on the chemistry of the Borzaş szmikite or data from relevant literature for comparison in case of rosenite. The chemical analysis of rosenite points out the following composition:

$FeO = 30.06$
$Fe_2O_3 = 5.85$
$SO_3 = 43.18$
$H_2O^+ = 20.90$

Total = 99.99

Halotrichite and pickeringite also belong to normally hydrated sulfides but they mostly contain two cations in their composition. The chemical analysis of pickeringite ($MgAl_2(SO_4)_4 \cdot 22H_2O$) revcals a high MgO content in comparison with data from the relevant literature (Palache et al., 1960):

$MgO = 6.29$
$FeO = 0.18$
$Al_2O_3 = 13.19$
$SO_3 = 36.51$
$H_2O^+ = 43.82$

Total = 99.99

The chemical analysis of the Borzaş halotrichite ($Fe^{2+}Al_2(SO_4)_4 \cdot 22H_2O$) is comparable with that of Chuquicamata halotrichite (Chile; Van Tassel, 1944, fide Palache et al., 1960):

Borzaş — Romania		Chuquicamata — Chile	
MgO	0.17	MgO	
FeO	6.51	FeO	7.28
MnO	0.07	MnO	
Al_2O_3	16.02	Al_2O_3	10.26
Fe_2O_3	3.40	Fe_2O_3	0.65
SO_3	33.19	SO_3	37.28
H_2O^+	40.64	H_2O^+	43.33
		Rem.	1.59(CuO 0.66; Insol. 0.93)
Total	100.00	Total	100.39

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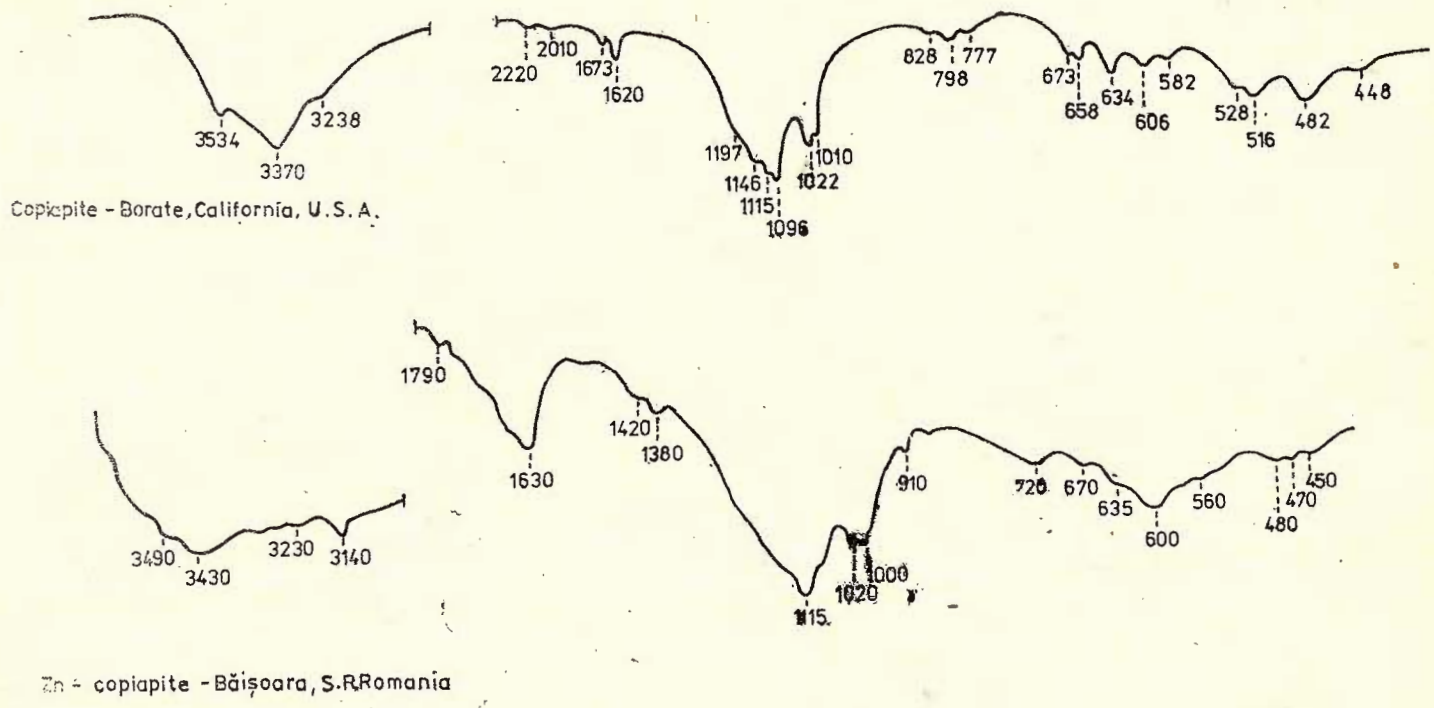

Fig. 5 — IR absorption spectrum of zincocopiapite.

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Zincocopiapite ($ZnFe^{3+}_4(SO_4)_6(OH)_2 \cdot 20H_2O$) belongs to the hydrated sulfates with two cations and hydroxide molecules in its composition. From the composition point of view the Băișoara zincocopiapite is similar with Rammelsberg zincocopiapite (Scharizer, 1919, fide Palache et al., 1960):

Băișoara — Romania		Rammelsberg — West Germany	
MgO	= 0.83	MgO	= 0.85
FeO	= 1.53	FeO	= 0.45
CuO	= 0.23	CuO	
ZnO	= 2.19	ZnO	= 2.47
Fe_2O_3	= 28.76	Fe_2O_3	= 26.19
Al_2O_3	= 2.64	Al_2O_3	
SO_3	= 39.25	SO_3	= 38.22
H_2O^+	= 24.56	H_2O^+	= 31.33
		Rem.	= 0.40 (insol.)
Total	= 99.99	Total	= 99.91

The data presented and the position at which the mentioned minerals occur within the mineralized structures suggest the oxidant character of the formation medium. The average values (slightly negative) of Eh, pH ranging between 6 and 8 and $\delta^{34}S$ strongly positive are specific to such an environment (Rye, Ohmoto, 1974). The presence of the crystallization water in the mineral composition indicates a high H_2O activity under conditions of an intense per descensum circulation.

[3] Absorption spectrometer in IR, UR 10, Karl-Zeiss Jena, KBr tablets. Analyst: T. Drăghici IGG.

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CONTRIBUȚII LA CUNOAȘTEREA SULFAȚILOR DIN ROMÂNIA

(Rezumat)

În zona de oxidație a zăcămîntului Borzaș (Carpații Orientali) au fost identificați rozenitul $FeSO_4 \cdot 4H_2O$, szmikitul $MnSO_4 \cdot H_2O$, halotrichitul $Fe^{2+}Al_2(SO_4)_4 \cdot 22H_2O$, și pickeringitul $MgAl_2(SO_4)_4 \cdot 22H_2O$. Tot în zona de oxidație a unei mineralizații de sulfuri, la Băișoara (Munții Apuseni), a fost întîlnit Zn-copiapitul $ZnFe_4^{3+}(SO_4)_6 \cdot (OH)_2 \cdot 20H_2O$. Aceștia apar ca eflorescențe, cruste, concrețiuni mamelonare și stalactitice, sau ca agregate ușor friabile (Zn-copiapitul). Datele privind chimismul și structura în acord cu cele din literatură permit reconstituirea mediului în care aceste minerale s-au format.

EXPLANATION OF PLATES

Plate I

Fig. 1 — Electronic microscope microphotograph of rosenite, × 300.

Fig. 2 — See Fig. 1, × 1000.

Fig. 3 — See Fig. 1, × 3000.

Plate II

Fig. 1 — Electronic microscope microphotograph of szmikite, × 1000.

Fig. 2 — See Fig. 1, + 3000.

Fig. 3 — See Fig. 1, + 10000.

Plate III

Fig. 1 — Electronic microscope microphotograph of halotrichite, × 1000.

Fig. 2 — See Fig, 1, × 3000.

Fig. 3 — See Fig. 1, × 10000.

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Plate IV

Fig. 1 — Electronic microscope microphotograph of pickeringite, × 300.
Fig. 2 — See Fig. 1, × 1000.
Fig. 3 — See Fig. 1, × 3000.

Plate V

Fig. 1 — Electronic microscope microphotograph of zincocopiapite, × 1000.
Fig. 2 — See Fig. 1, × 3000.
Fig. 3 — See Fig. 1, × 10000.

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PETROLOGIA ROCILOR MAGMATICE

AVERAGE CHEMICAL COMPOSITION OF THE VOLCANICS IN THE CĂLIMANI—GURGHIU—HARGHITA MOUNTAINS (EAST CARPATHIANS) AND PETROTYPES DISTRIBUTION ACCORDING TO CHEMICAL CLASSIFICATIONS[1]

by

PETRE ANDĂR[2], SERGIU PELTZ[2]

Volcanic rocks. Neogene. Classification. Nomenclature. Data processing. Statistical distribution. Major elements. East Carpathians — Neogene eruptive-Quaternary — Călimani Mountains, Gurghiu Mountains, Harghita Mountains.

Abstract

The 434 chemical analyses of all types of volcanic rocks in the whole Călimani—Gurghiu—Harghita Mts have been processed using statistical methods in order to establish the average chemical composition and the petrotypes distribution according to four systems of chemical classification (SiO_2—$Na_2O + K_2O$, TAS—IUGS and the USSR Petrographic Committee; R_1—R_2; Q′—ANOR). The average chemical composition of the Călimani—Harghita andesites (n = 399) defined on the basis of the mineralogical criterion is similar with the average composition defined by the TAS chemical classification. The systems of chemical classification evidence, in agreement with the mineralogical classification, the predominance of the andesites over the volcanic rocks in the study area (46—58%) and the almost equal percentage of the basaltic andesites (5—19%) and of the dacites (8—27%). About 40—60% of the quartz andesites plot in the dacite field. It is for the first time that chemical analyses with H_2O smaller than 2% and CO_2 smaller than 0.5%, recalculated to 100% on an H_2O and CO_2 free basis, have been used when examining the chemical classification of the neovolcanics in Romania.

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Résumé

Composition chimique moyenne des vulcanites des monts Călimani—Gurghiu—Harghita (Carpathes Orientales) et la distribution des pétrotypes en rapport avec les classifications chimiques. Les 434 analyses chimiques complètes de tous les types de roches volcaniques de toutes les régions des monts Călimani—Gurghiu—Harghita ont été interprétées automatiquement par des méthodes statistiques en vue d'établir la composition chimique moyenne et la distribution des pétrotypes conformément à quatre systèmes de classification chimique (SiO_2—Na_2O + + K_2O, TAS—IUGS et le Comitet pétrographique de l'USSR; R_1—R_2; Q'—ANOR). La composition chimique moyenne des andésites de Călimani—Harghita (n = 399), déterminée selon la composition minéralogique est semblable à la composition moyenne déterminée par la classification chimique TAS. Les systèmes de classification chimique relèvent, en accord avec la classification minéralogique, la prédominance des andésites parmi les roches volcaniques de la région (46—58%) et la participation presque égale des andésites basaltiques (5—19%) et des dacites (8—27%). 40 à 60% des andésites à quartz se projettent dans le champ des dacites. C'est pour la première fois qu'on utilise, en exclusivité, dans l'examen de la classification chimique des néovulcanites de la Roumanie, les analyses chimiques à $H_2O^+ < 2\%$ et $CO_2 < 0{,}5\%$, récalculés 100%, sans ces deux oxides.

Introduction

The geological researches, recently comprising the whole Călimani—Gurghiu—Harghita volcanic area, led to the accumulation of a significant bulk of data. These data include also complete chemical analyses of fresh volcanic rocks; they represent different petrotypes according to their participation in the formation of the volcanic units in the area of study.

The studies elaborated during this period and whose results are only partly published were based on petrochemical data especially with a view to clearing up petrological problems. Up till now no exhaustive evaluation of the bulk of chemical analyses has been achieved. In this respect one took into account that the chemical analyses were used to a less extent in the correct denomination of the petrotypes in the Călimani—Harghita area, formed about 80 per cent of volcanic rocks with a widely developed matrix.

The volcanic rocks were usually denominated on the basis of the mineralogical composition, the phenocrysts assemblage, respectively, without any comparison with the denomination resulting from the chemical classification and particularly without changing the petrographic denomination in case of a disagreement between the two diagnoses (e.g. andesite according to the mineralogical composition, dacite according to the chemical composition).

During the last 15 years one of the authors was constantly concerned with the improvement of the nomenclature of the neovolcanic rocks in agreement with the petrochemical data, some of the results being already published (Peltz et al., 1972; Peltz et al., 1973; Peltz, Peltz, 1983; Peltz et al., 1985; Peltz in Embey-Isztin et al., 1985).

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Recently, Szakács and Seghedi (1986) showed similar concerns regarding the volcanic rocks in the southeasternmost part of the Harghita Mts.

During the last two decades the above-mentioned problem was studied by the IUGS Subcommission on the Systematics of Igneous Rocks. Recently the activity of the Subcommission has been completed and recommendations on the chemical classification and nomenclature of the volcanic rocks have been given (Le Bas et al., 1986). In connection with the activity of the IUGS Subcommission numerous diagrams and chemical classifications of the volcanic rocks have been proposed. They have been tested in case of the andesitic and basaltic rocks in Romania and some of the results have already been published (see the above-mentioned references).

In recent years, the number of the chemical analyses of the fresh volcanic rocks in the Călimani—Gurghiu—Harghita Mts increased considerably and the recommendations of the IUGS Subcommission on the Systematics of Igneous Rocks for the chemical classification of the volcanic rocks were concluded. All this made us reexamine the problem of the chemical classification of the volcanics in the study area (Peltz, Peltz, 1983) using data processing by simple statistical methods facilitating the study of the petrotypes distribution established on the basis of the modal mineralogical classification according to their plotting on representative diagrams for the chemical classifications.

Thus we aimed at obtaining average chemical compositions representative for all petrotypes as well as at achieving a better agreement between the mineralogical diagnosis and the chemical one. On this basis contributions are brought to a correct estimation of the products resulting during different stages of the parental magma evolution directly reflected in the rock composition.

Sample Selection and Data Processing

The bulk of analytical data is based on fresh volcanic rocks in the Călimani, Gurghiu and Harghita volcanic massifs of Pannonian—Pliocene age, as well as in the eruptive bodies of Lower Pleistocene age occurring at Bicsad and Malnaș (southernmost part of the Harghita Mts). The volcanic rocks analysed represent, in the order of the significance of the spatial distribution, the following petrographic types: pyroxene andesite, hornblende-pyroxene andesite and/or pyroxene—hornblende andesite, hornblende andesite, basaltic andesite, hornblende, biotite ± quartz andesite, dacite, quartz andesite, pyroxene ± olivine ± hornblende ± biotite ± quartz andesite.

The selection and classification of the samples were based on the above—mentioned petrotypes as reference terms as the primary classification of the rocks is the more accessible modal mineralogical diagnosis. For the computation of the average chemical composition the primary analytical data have also been grouped according to the geographic distribution: a) samples from the south of the Călimani Mts and south—west of the Harghita Mts represent occurrences from the lower compartment; b) samples from the northern parts of the Călimani,

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Gurghiu and Harghita Mts represent especially occurrences from the upper compartment; c) samples from the Tuşnad and Bicsad —Malnaş areas represent the youngest volcanic units in the East Carpathians within which the volcanological and petrological evolution displayed particular forms.

434 complete chemical analyses have been selected; all of them represent fresh rocks containing H_2O^+ smaller than 2% and CO_2 smaller than 0.5%. According to the recommendations of the IUGS Subcommission (Le Bas et al., 1986), all the analyses were recalculated to 100% on an H_2O and CO_2 free basis.

For the data processing effectuated with a FELIX C —256 computer several FORTRAN IV programmes have been used for the computation of the average values and of the petrochemical parameters.

Average Chemical Composition

The average chemical composition has been computed separately for the petrotypes defined by the mineralogical classification (CM criterion) as well as for those defined by the TAS chemical classification (CC criterion). TAS chemical classification has been chosen by us because it is based on a simple diagram with SiO_2, K_2O and Na_2O; this classification is recommended by the IUGS Subcommission and is nowadays widely used with good results.

Table 1 shows, for comparison, the average chemical compositions of the Quaternary basalts in the Perşani Mts. Also for comparison, table 2 presents the average chemical compositions of some types of volcanic rocks referring mostly to the neovolcanic areas of the Earth.

The data presented in table 1 and 2 evidence certain characteristic features of the petrochemistry of the study rocks. Thus, the average chemical composition of the andesites in the Călimani — Gurghiu —Harghita area defined on the basis of the CM and CC criteria is similar. As regards the basalts in the Perşani Mts the differences are very important for most of the oxides.

Harghita andesites are more acid and richer in potassium; there follows those in the Călimani Mts. Gurghiu andesites are richer in iron and calcium. SiO_2 and $K_2O + Na_2O$ increase from the basaltic andesites to biotite ± quartz hornblende andesites. Except for all the petrotypes in the south of the Harghita as well as the hornblende andesites in the whole Călimani —Harghita area there are no differences within the same petrotype as regards the appurtenance to different volcanic regions taken into account. A special position occupies the pyroxene andesite ± olivine ± hornblende ± biotite ± quartz at Bicsad — Malnaş which points to the highest contents of K_2O (3.45%) and P_2O_5 (0.47%) as well as to high contents of Na_2O and MgO.

In case of the basaltic andesites there are differences between CM and CC in all oxides because, according to TAS classification, these rocks occupy a wide field of SiO_2 (52 —57%).

Referring to the acid rocks, the chemical classification emphasized the presence of a large number of cases in comparison with the petro-

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TABLE 1

Major elements representative averages for volcanic rocks from Călimani - Harghita and Perşani Mountains

I PETROTYPE, ACC TO MINERALOGICAL CLASSIFICATION

	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	K_2O	Na_2O	P_2O_5	
BASALTS PERŞANI MTS n=11	47.90	1.59	16.37	3.99	5.22	0.17	9.40	9.59	1.66	3.61	0.50	
ANDESITES CĂLIMANI+GURGHIU+HARGHITA MTS n=399	59.74	0.75	18.14	3.22	2.84	0.13	3.13	6.70	1.77	3.49	0.18	
ANDESITES CĂLIMANI MTS n=122	59.83	0.78	17.94	3.39	3.11	0.15	3.03	6.68	1.62	3.46	0.19	
ANDESITES GURGHIU MTS n=90	58.87	0.77	18.28	4.13	2.61	0.15	3.16	6.89	1.41	3.55	0.17	
ANDESITES HARGHITA MTS n=187	60.10	0.72	18.20	2.67	2.78	0.11	3.18	6.62	2.05	3.48	0.18	
BASALTIC ANDESITES												
A	51.56	1.03	20.04	2.98	5.67	0.15	5.33	8.99	1.26	2.75	0.22	(n=5)
B	53.17	1.13	18.20	3.97	4.93	0.19	4.99	8.65	1.59	2.90	0.27	(n=19)
PYROXENE ANDESITES												
A	57.99	0.87	18.15	2.86	3.92	0.16	3.48	6.93	1.92	3.52	0.20	(n=24)
B	57.17	0.84	18.42	3.94	3.33	0.15	3.73	7.38	1.55	3.20	0.19	(n=32)
D	55.81	0.97	18.71	4.55	3.29	0.19	3.83	7.81	1.18	3.46	0.19	(n=22)
E	58.87	0.81	18.14	2.76	3.41	0.12	3.52	6.99	1.71	3.47	0.19	(n=30)
F	59.25	0.73	18.77	2.42	3.19	0.11	3.33	7.10	1.64	3.33	0.14	(n=30)
G	56.31	0.77	19.70	2.27	3.70	0.12	4.26	7.73	1.72	3.26	0.15	(n=9)

II PETROTYPE ACC. TO CHEMICAL CLASSIFICATION

BASALTS, PERŞANI MTS n=11

SiO_2	TiO_2	Al_2O_3	Fe_2O_3	Fe O	MnO	MgO	CaO	K_2O	Na_2O	P_2O_5
50.99	1.16	18.52	4.33	5.47	0.21	5.92	8.69	1.54	2.93	0.23

BASALTIC ANDESITES, CĂLIMANI+GURGHIU+HARGHITA MTS n=99

55.27	0.92	18.80	3.88	3.84	0.16	4.10	8.11	1.44	3.33	0.20

ANDESITES, CĂLIMANI+GURGHIU+HARGHITA MTS n=249

59.80	0.75	18.19	3.16	2.83	0.13	3.10	6.69	1.75	3.48	0.19

DACITES, CĂLIMANI+GURGHIU+HARGHITA MTS. n=62

64.74	0.57	17.15	2.54	1.79	0.10	2.14	5.10	2.26	3.65	0.16

RHYODACITES, CĂLIMANI+HARGHITA MTS n=22

69.90	0.28	16.82	1.97	0.46	0.04	0.52	3.19	3.08	3.61	0.12

RHYOLITES, CĂLIMANI MTS n=2

77.92	0.15	13.00	1.22	0.62	0.03	0.28	2.43	1.50	2.77	0.09

A, Călimani-north; B, Călimani-south; C, Gurghiu-north; D, Gurghiu-south; E, Harghita-north; F, Harghita-south; G, Harghita SW; H, Tuşnad area; I, Bicsad-Malnaş area.

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TABLE 2

The averages for some volcanic rocks

	SiO	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	K_2O	Na_2O	P_2O_5		
	ANDESITES												
Daly	59.59	0.77	17.31	3.33	3.13	0.18	2.75	5.80	2.04	3.58	0.26	(n=87)	
Chayes	58.17	0.80	17.26	3.07	4.17		3.23	6.93	1.61	3.21	0.20	(n=1775)	1969
Nockolds	54.20	1.31	17.17	3.48	5.49	0.15	4.36	7.92	1.11	3.67	0.28	(n=49)	
Taylor	59.50	0.70	17.20	2.72	3.65	0.15	3.42	7.03	1.60	3.68			1969
Le Maître	57.94	0.87	17.02	3.27	4.04	0.14	3.33	6.79	1.62	3.48	0.21	(n=2203)	1979

Averages acc. to Le Maître 1984

BASALT											
49.45	1.80	15.35	3.28	7.95	0.15	7.71	11.01	0.52	2.57	0.21	(n=9794)
BASALTIC ANDESITE											
54.21	1.29	16.37	3.38	6.58	0.18	4.99	8.80	1.04	2.95	0.21	(n=2331)
ANDESITE											
59.94	0.82	17.03	3.18	3.78	0.13	3.21	6.56	1.69	3.46	0.19	(n=2108)
DACITE											
66.17	0.60	15.93	2.47	2.47	0.10	1.68	4.19	2.39	3.82	0.18	(n=1139)
RHYOLITE											
74.26	0.27	13.58	1.32	1.00	0.05	0.35	1.14	4.26	3.71	0.06	(n=2260)

PYROXENE HORNBLENDE ANDESITES

	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	K_2O	Na_2O	P_2O_5	
A	58.94	0.72	18.70	2.99	4.08	0.13	2.39	6.59	1.84	3.45	0.18	(n=6)
B	58.62	0.72	17.86	4.53	2.30	0.16	3.54	7.03	1.60	3.44	0.20	(n=8)
C	58.34	0.59	18.04	5.88	1.76	0.09	3.44	7.01	1.23	3.45	0.17	(n=6)
D	58.30	0.78	18.88	3.91	2.76	0.14	3.06	7.28	1.29	3.43	0.17	(n=20)
E	60.39	0.75	17.91	2.77	2.97	0.13	3.14	6.64	1.85	3.28	0.17	(n=24)
F	59.87	0.70	18.58	2.14	3.19	0.11	3.20	6.91	1.80	3.35	0.15	(n=24)
G	58.33	0.64	19.62	2.15	3.33	0.11	3.35	7.06	1.77	3.51	0.14	(n=11)

HORNBLENDE ANDESITES

A	59.96	0.80	18.63	2.29	3.60	0.14	2.67	6.49	1.83	3.39	0.20	(n=2)
B	62.30	0.72	17.47	2.93	2.60	0.12	2.47	6.18	1.50	3.51	0.19	(n=45)
C	61.37	0.64	17.78	3.63	2.18	0.13	2.53	6.03	1.68	3.57	0.15	(n=15)
D	58.26	0.80	18.37	4.74	2.43	0.16	3.06	6.93	1.42	3.65	0.18	(n=17)
E	60.55	0.82	18.43	4.67	1.74	0.21	2.11	5.70	1.84	3.68	0.25	(n=5)
F	61.00	0.63	18.41	2.60	1.91	0.10	2.83	6.97	1.89	3.52	0.14	(n=7)

HORNBLENDE BIOTITE ± QUARZ ANDESITES

G	60.74	0.78	18.97	3.95	0.76	0.15	2.08	6.12	2.39	3.72	0.13	(n=6)
H	63.85	0.55	17.00	2.79	1.25	0.07	2.40	5.14	2.97	3.83	0.16	(n=26)

PYROXENE ± OLIVINE ± HORNBLENDE ± BIOTITE ± QUARZ ANDESITES

	58.78	0.91	16.28	2.90	2.10	0.09	4.35	6.78	3.45	3.80	0.47	(n=10)

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types; consequently the average values were computed which, according to the number of the cases, can be regarded as significant especially for dacites (n = 62).

The comparison of the chemical composition according to CM criterion and to the appurtenance to the nine geographic units taken into account (Tab. 1) does not reveal significant differences. There results that in the whole Călimani-Harghita area each petrotype is characterized by a relatively homogeneous chemical composition.

The chemical composition of the andesites in the Călimani—Gurghiu—Harghita Mts ($n_{CM} = 399$ and $n_{Cc} = 249$, Tables 2 and 3) is similar with the composition of the andesites on the Earth's scale according to Daly, Taylor, Le Maître (Tab. 2). The similarity of the chemical composition of the andesites after Le Maître (1984) is remarkable. Unlike andesites, significant differences are observed in case of basalts, basaltic andesites, dacites and rhyolites.

The Petrotypes Distribution According to Chemical Classifications

For the evaluation of the concordance between the petrographic-mineralogical diagnosis and the chemical one the distribution of the absolute and relative frequency of the petrotypes in the fields of some diagrams used in the chemical classification was established. The statistical distribution of the volcanic rocks analysed to the four methods of chemical classification makes possible quantitative estimations on unitary basis both in the CM—CC ratio and between methods. The accumulated experience (Peltz, Peltz, 1983; Peltz et al., 1985; Peltz in Isztin et al., 1985) made possible the attaining of the object proposed, namely the use of the diagrams: SiO_2-$Na_2O + K_2O$ (Le Bas et al., 1986), $SiO_2 - -Na_2O+K_2O$(Bogatikov et al., 1981), $R_1 = 4Si - 11(Na + K) - - 2(Fe + Ti)$ and $R_2 = 6Ca + 2Mg + Al$ (De la Roche et al., 1980), $Q' = Q/(Q + Or + Ab + An)$ and $ANOR = An/Or + An$ (Streckeisen, Le Maître, 1979).

Due to the reduced number of samples for some petrotypes the absolute frequency distribution was considered within each volcanic massif whereas for the whole volcanic chain the relative frequency was used.

SiO_2—$Na_2O + K_2O$ Diagram (TAS—IUGS, 1986) (Tab. 3)

In the Călimani Mts most of the basaltic andesites, pyroxene andesites, pyroxene + hornblende andesites and hornblende andesites plot in the andesite field, and dacites occupy the dacite field. Very often pyroxene andesites fall into the basaltic andesites field; one third of the dacite samples occupy the rhyolite field. On the whole, quartz andesites and biotite ± quartz andesites fall into the dacite field. A similar situation is observed in case of the rocks in the Gurghiu and Harghita Mts, several particular aspects being revealed: a) in the Gurghiu Mts most of the pyroxene andesites plot in the basaltic andesites field; b) the Bicsad—Malnaș andesites fall into the trachyandesites (82%) and dacites (18%) fields.

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TABLE 3

Distribution of petrotypes based on the TAS Diagram, 1986

		B	AB	TAB	A	TA	D	TD	R
		a. absolute frequency							
	AB	5	11	4					
	Ap		20	1	27	6			
	Aph		3		11				
I	Ah		1		35		9	1	1
	Ahbq						1		
	Aq						4		
	D						14		7
	Ap		20		2				
	Aph		9		17		1		
II	Ah		3	1	22	1	5		
	Ahbq				2		6	1	
	Aq				1				
	Ap		14		48	1	3		
	Aph		4		51		4		
III	Ah		1		9		3		
	Ahbq			1	5	3	15	9	
	Apohbq					9	2		
I+II+III(n=434)		5	86	7	230	20	67	11	8
		b. relative frequency							
	AB	25	55	20					
	Ap		38	1	54	5	2		
	Aph		16		79		5		
	Ah		5	1	72	1	19	1	1
IV	Ahbq			2	16	7	52	23	
	Apohbq					82	18		
	Aq				20		80		
	D						67		33
	Total	1	19	2	53	5	15	2	2

Mineralogical diagnosis: AB, basaltic andesite; Ap, pyroxene andesite; Aph, hornblende-pyroxene andesite; Ah, hornblende andesite; Ahbq, hornblende, biotite ± quartz andesite; Apohbq, pyroxene ± olivine ± hornblende ± biotite ± quartz andesite; Aq, quartz andesite; D, dacite

Chemical diagnosis: BO, olivine basalt; BSA, subalkali basalt; B, basalt; T, tholeiite TB, trachy-basalt; AB, basaltic andesite; TAB, trachybasaltic andesite; LB, lati-basalt; L, latite; LA, lati-andesite; A, andesite; TA, trachy-andesite; D, dacite; TD, trachy-dacite; RD, rhyodacite; TRD, trachy-rhyodacite; R, rhyolite

I, Călimani Mts; II, Gurghiu Mts; III, Harghita Mts; IV, Călimani—Harghita area.

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The distribution of the relative frequency within the TAS classification system on the scale of the whole territory emphasizes a good correspondence between CM and CC in case of the basaltic andesites and andesites (except for the acid ones), as well as dacites. Moreover, it reveals the alkaline character (trachyandesites) of most of the Bicsad—Malnaș andesites and the rhyolitic character of one third of the dacites.

SiO_2—Na_2O+K_2O Diagram (Petrogr. Com. of the USSR, 1981) (Tab. 4)

Most of the pyroxene andesites in the Călimani Mts as well as of the pyroxene-hornblende andesites, hornblende andesites fall into the andesite field; the majority of the dacites plot in the rhyodacites field. In about one third of the cases pyroxene andesites fall into the basaltic andesites field. Basaltic andesites are more various, occupying almost an equal percentage in the fields of basaltic andesites, basalts and alkaline basalts. Most of the pyroxene andesites in the Gurghiu Mts plot in the basaltic andesites field, and the majority of the hornblende-pyroxene andesites and of the hornblende andesites fall in the andesite field. In the Harghita Mts there is a situation similar to that in the Călimani Mts as regards the andesites. The wide variation range of the SiO_2 contents and of alkalis in the andesites in the Tușnad area determines their plotting in the field of dacites and rhyodacites. The alkaline character (trachyandesites) of most of the Bicsad—Malnaș andesites is also revealed.

The classification system we are referring to indicates a good correspondence between CM and CC for the pyroxene andesites, hornblende-pyroxene andesites and hornblende andesites in the Călimani-Harghita area; 65 per cent of the basaltic andesites occupy the basalts field; 85 per cent of the dacites occupy the rhyodacite field. Moreover, this classification reveals that the eruptive rocks constituing the Bicsad and Malnaș bodies represent 73 per cent trachyandesites, 18 per cent dacites and 9 per cent latites.

A comparison of the two systems of classification based on the relationship between the same oxides points out similarities as regards the appurtenance of most of the petrotypes to the petrochemical classes (Tab. 3, 4). Thus, in the whole area andesites represent 53—58 per cent, basaltic andesites 17—19 per cent and dacites 8—15 per cent.

R_1—R_2 Diagram (De la Roche et al., 1980) (Tab. 5)

This is a more complex diagram as the synthetic parameters R_1 and R_2 include most of the cations of an eruptive rock. Consequently the appurtenance of the study rocks is more representative as regards the chemical nomenclature in comparison with the diagrams SiO_2—Na_2O+K_2O.

The basaltic andesites in the Călimani Mts plot mostly in the basaltic andesites field. The pyroxene andesites in the Călimani Mts and Harghita Mts fall in the andesites field, as well as in the basaltic andesites field, whereas most of the pyroxene andesites in the Gurghiu Mts plot in the basaltic andesites field. In case of the hornblende-pyroxene andesites

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TABLE 4

Distribution of petrotypes based on the SiO_2 - Na_2O+K_2O Diagram (Petrografic Committee of USSR, 1981)

		BSA	B	AB	A	L	TA	D	TD	RD	TRD	R
					a) absolute frequency							
	AB	7	6	5		2						
	Ap		1	19	29	3	2					
	Aph			3	11							
I	Ah			1	39			5	1	1		
	Ahbq				1							
	Aq							3		1		
	D							1		18		2
	Ap			18	2	2						
	Aph		1	8	17			1				
II	Ah			3	24	2		3				
	Ahbq				4			4	1			
	Aq				1							
	Ap			14	51	1						
	Aph			4	51			4				
III	Ah			1	12							
	Ahbq				11	2	4	11	3		2	
	Apohbq					1	8	2				
I+II+III(n=434)		7	8	76	253	13	14	34	5	20	2	2
					b) relative frequency							
	AB	35	30	25		10						
	Ap		1	36	58	4	1					
	Aph		1	15	79			5				
	Ah			5	82	2		9	1	1		
IV	Ahbq				37	5	9	35	9		5	
	Apohbq					9	73	18				
	Aq				20			60		20		
	D							5		85		10
	Total	2	2	17	58	3	3	8	1	5	0.5	0.5

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TABLE 5

Distribution of petrotypes based on the R_1-R_2 Diagram (De la Roche et al.,1980)

		BO	T	TB	LB	AB	L	LA	A	D	RD	R
						a) absolute frequency						
	AB	4	1	1		13			1			
	Ap	1			2	21		1	27	2		
	Aph					3			10	1		
	Ah					2			22	22	1	
	Ahbq									1		
	Aq									3	1	
	D										20	1
	Ap	2				17			2	1		
	Aph	1				9			14	2	1	
II	Ah					4	2		18	7	1	
	Ahbq								2	7		
	Aq								1			
	Ap					22		1	38	5		
	Aph					8			44	7		
III	Ah					2			9	2		
	Ahbq				1	4	1		8	18	1	
	Apohbq				3	4		1	1	1	1	
I+II+III(n=434)		8	1	1	6	109	3	3	197	79	26	1
						b) relative frequency						
	AB	20	5	5		65			5			
	Ap	2			1	42		1	47	5		
	Aph	1				20			68	10	1	
	Ah					9	2		53	34	2	
IV	Ahbq				2	9	2		24	61	2	
	Apohbq				27	37		9	9	9	9	
	Aq								20	60	20	
	D										95	5
	Total	2	0.2	0.2	1	25	0.7	0.7	46	18	6	0.2

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TABLE 6

Distribution of petrotypes based on the Q'-ANOR Diagram (Streckeisen and Le Maître, 1979)

		B	TB	AB	L	A	D	RD	R
		a) absolute frequency							
	AB	7	1	2		9	1		
	Ap	2		2		43	6	1	
	Aph					10	4		
I	Ah					12	32	2	1
	Ahbq						1		
	Aq						4		
	D						4	16	1
	Ap					20	2		
	Aph					18	7		
II	Ah			2		16	13	1	
	Ahbq			1			8		
	Aq						1		
	Ap			2		56	8		
	Aph			3		43	13		
III	Ah					9	4		
	Ahbq			7	6	6	8	4	2
	Apohbq			4	6			1	
I+II+III(n=434)		9	1	23	12	244	116	25	4
		b) relative frequency							
	AB	35	5	10		45	5		
	Ap	1		3		84	11	1	
	Aph			3		73	24		
	Ah			2		40	53	4	1
IV	Ahbq			19	14	14	40	8	5
	Apohbq			36	55			9	
	Aq						100		
	D						19	76	5
	Total	2	0.2	5	3	56	27	6	0.8

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in all the three massifs and of the hornblende andesites in the Gurghiu and Harghita Mts, most of the samples fall in the andesites field. System R_1-R_2 leads to the plotting of the Călimani hornblende andesites in equal percentages in the andesites and dacites field and also the plotting of all dacites in the rhyodacites field.

Considering the whole volcanic chain, a comparison with the SiO_2—alkalis systems points out the reduction of the andesites percentage (from 53—58% to 46%) and the increase of the basaltic andesites percentage (from 17 to 19% to 25%). Likewise, the Bicsad—Malnaș andesites mostly represent, according to this classification, basaltic andesites (37%) and latibasalts (27%).

Q' — ANOR Diagram (*Streckeisen and Le Maître, 1979*) (*Tab. 6*)

The synthetical value *Q'* and ANOR were computed from the CIPW normative composition and plotted in the fields of the diagrams proposed by Streckeisen and Le Maître. It is useful to establish by comparison with the studied classifications which is the domain of agreement and disagreement between CM and CC on the basis of the relationship between the main salic normative minerals.

In case of the majority of the andesites an agreement is observed between the two criteria; an exception is the hornblende andesites in the Călimani Mts mostly plotting in the dacite field. The Călimani basaltic andesites plot also in the basaltic fields and most of the dacites fall in the rhyodacites field. The andesites in the Bicsad—Malnaș bodies plot in different fields with close values of the absolute frequency.

According to the mentioned system of classification andesites represent 56 per cent of the total of the study rocks and at the same time the dacites percentage increases considerably.

On the whole, a good correlation CM—CC is observed only in case of the pyroxene andesites and of the hornblende-pyroxene andesites; 45 per cent of the basaltic andesites plot in the andesites field, 53 per cent of the hornblende andesites plot in the dacite field, 40 per cent of the biotite ± quartz andesites plot in the dacite field, 76 per cent of dacites fall in the rhyodacite field. The Bicsad and Malnaș andesites are represented by latites (55 per cent), basaltic andesites (36 per cent) and rhyodacites (9 per cent).

Conclusions

On the basis of 255 chemical analyses representing all types of volcanic rocks in the Călimani—Harghita area, Peltz and Peltz (1983) carried out a first examination of the correspondence between the modal mineralogical and chemical classification using the diagrams QAP (Streckeisen, 1967), Q'—ANOR (Streckeisen and Le Maître, 1979), SiO_2-K_2O (Peccerillo and Taylor, 1976), normative plagioclase-normative colour index (Irvine and Baragar, 1971). A good correspondence was inferred in case of the basaltic andesites, pyroxene andesites, hornblende-pyroxene andesites and dacites.

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Using the bulk of data constituted of 434 chemical analyses recalculated to 100% on an H_2O and CO_2 free basis and turning into account the advantages of the data processing and of the use of statistical methods, the results published in 1983 have been confirmed and completed, considering the classifications SiO_2-alkalis (TAS, USSR Petrographical Committee), R_1-R_2 and Q′—ANOR.

According to the mineralogical classification the four systems of chemical classification evidence the predominance of andesites over the volcanic rocks (46—58 per cent) and the participation in an almost equal percentage of basaltic andesites (5—19 per cent) and of dacites (8—27 per cent). Other types of volcanic rocks defined by the chemical classifications — e.g. olivine-bearing basalts, basalts, trachybasalts, latibasalts, latites, trachyandesites, trachydacites, rhyodacites, rhyolites— represent 6—0.2 per cent of the total cases examined.

According to the classifications SiO_2-alkalis and R_1-R_2, 35—42 per cent of the pyroxene andesites plot in the basaltic andesites field.

More than 40 per cent of the hornblende-biotite-quartz andesites and 60 per cent of the quartz andesites plot in the dacites field (dacites, trachydacites).

Between 76 and 95 per cent of the dacite samples in the east of the Călimani Mts represent rhyodacites.

The four systems of chemical classification emphasize the complex constitution of the Bicsad and Malnaș eruptive bodies (basaltic andesites, trachyandesites, latites and dacites); this fact has not been pointed out in the petrographic diagnosis.

The average chemical composition of the andesites in the Călimani—Gurghiu—Harghita Mts (n = 399), defined on the basis of the mineralogical composition and of the chemical classification is quite similar. A striking similarity is observed between this composition and that of the andesites according to Le Maître (1984) (n = 2108).

The chemical analyses used in this paper come from published papers (Peltz and Peltz et al., 1969—1985), reports from the archives of the IGG (Peltz et al., 1975—1986) and other analytical data not included in the mentioned papers.

Most of the analyses were carried out by E. Călinescu, E. Ianc and C. Vasiliu.

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COMPOZIȚIA CHIMICĂ MEDIE A VULCANITELOR DIN MUNȚII CĂLIMANI—GURGHIU—HARGHITA (CARPAȚII ORIENTALI) ȘI DISTRIBUȚIA PETROTIPURILOR ÎN ACORD CU CLASIFICĂRILE CHIMICE

(Rezumat)

În ultimii ani a sporit considerabil numărul analizelor chimice ale rocilor vulcanice proaspete din munții Călimani—Gurghiu—Harghita și totodată s-au finalizat recomandările Subcomisiei IUGS pentru Sistematica rocilor ignee cu privire la clasificarea chimică a rocilor vulcanice. De aceea am considerat oportună reexaminarea problemei clasificării chimice a vulcanitelor din acest teritoriu (Peltz, Peltz, 1983), utilizîndu-se prelucrarea automată a datelor prin metode statistice simple, care să faciliteze examinarea distribuției petrotipurilor stabilite pe baza clasificării mineralogice modale în acord cu proiecția lor în diagrame reprezentative pentru clasificarea chimică.

S-a urmărit să se obțină compoziții chimice medii reprezentative pentru toate petrotipurile, precum și realizarea unei mai bune concordanțe

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între diagnoza mineralogică și cea chimică. Pe această bază se pot aduce contribuții la o evaluare corectă a produselor ce au rezultat în diferitele stadii evolutive ale magmei parentale și care se reflectă direct în compoziția rocilor.

Au fost selectate 434 analize chimice complete reprezentînd în totalitate roci proaspete în compoziția cărora $H_2O < 2\%$ și $CO_2 < 0{,}5\%$. Totodată, conform recomandărilor Subcomisiei IUGS (Le Bas et al., 1986), toate analizele au fost recalculate la 100% prin eliminarea valorilor H_2O și CO_2.

Rocile vulcanice analizate reprezintă, în ordinea importanței răspîndirii spațiale, următoarele tipuri petrografice: andezit piroxenic, andezit cu piroxeni și hornblendă și/sau andezit cu hornblendă și piroxeni, andezit cu hornblendă, andezit bazaltic, andezit cu hornblendă, biotit ± cuarț, dacit, andezit cuarțifer, andezit cu piroxeni ± olivină ± hornblendă ± biotit ± cuarț.

Pentru prelucrarea automată a datelor, efectuată la un calculator FELIX C-256, au fost utilizate o serie de programe scrise în limbaj FORTRAN IV pentru calcularea valorilor medii și a parametrilor petrochimici.

În acord cu clasificarea mineralogică cele patru sisteme de clasificare chimică utilizate ($SiO_2-Na_2O+K_2O$, TAS—IUGS și Comitetul petrografic URSS; R_1-R_2; Q′—ANOR) evidențiază predominarea andezitelor printre rocile vulcanice din teritoriu (46—58%) și participarea cu pondere aproximativ egală a andezitelor bazaltice (5—19%) și a dacitelor (8—27%). Alte tipuri de roci vulcanice definite de clasificările chimice precum: bazalte cu olivină, bazalte, trahibazalte, latibazalte, latite, trahiandezite, trahidacite, riodacite, riolite reprezintă între 6—0,2% din totalul cazurilor examinate (tab. 3, 4, 5, 6).

Potrivit clasificărilor SiO_2-alcalii și R_1-R_2, 35—42% dintre andezitele piroxenice se situează în cîmpul andezitelor bazaltice. Peste 40% dintre andezitele cu hornblendă, biotit și cuarț și peste 60% dintre andezitele cuarțifere se proiectează în cîmpul rocilor dacitice (dacite, trahidacite). Între 76—95% din probele de dacite din estul munților Călimani reprezintă riodacite.

Toate cele patru sisteme de clasificare chimică scot în evidență alcătuirea complexă a corpurilor eruptive de la Bicsad și Malnaș (andezite bazaltice, trahiandezite, latite, dacite), ceea ce nu a fost sesizat în diagnoza petrografică.

Compoziția chimică medie a andezitelor din Călimani—Gurghiu—Harghita (n = 399) definite pe criteriile compoziției mineralogice și clasificării chimice este foarte apropiată. Se observă o asemănare remarcabilă între această compoziție și cea a andezitelor la scară planetară după Le Maître, 1984 (n = 2108) (tab. 1,2).

Compararea compoziției chimice pe criteriul diagnozei mineralogice (petrotip) și al apartenenței la principalele unități geografice din catena Călimani—Gurghiu—Harghita (tab. 1) nu relevă deosebiri semnificative. Rezultă că pe întreg teritoriul, independent de structură/unitate vulcanică și vîrstă, fiecare petrotip se caracterizează printr-o compoziție chimică relativ omogenă.

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1. MINERALOGIE—PETROLOGIE

PETROLOGIA ROCILOR MAGMATICE



Проект 3

PETROLOGIA ASOCIAȚIEI GRANITELOR ALCALINE DIN MASIVELE IACOBDEAL ȘI PIATRA ROȘIE (DOBROGEA DE NORD)[1]

de

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Alkali granites. Magmatic associations. Plate collision. Upper Triassic. Acidic magmas. Alkalic composition. Migration. Magmatic differentiation. Absolute age. Rb/Sr. Major elements. Minor elements. Dobrogea — North Dobrogea — Măcin Zone.

Abstract

Petrology of the Alkaline Granite Assemblage in the Iacobdeal and Piatra Roșie Massifs (North Dobrogea). The alkaline eruptive rocks belong to the North-Dobrogean Orogen, constituting small intrusions with plutonic and subvolcanic character, disposed along two NW—SE trending alignments, almost parallel to the folded zone. The plutonic bodies consist of various types of alkaline granites with riebeckite and/or egyrine and alkaline quartz syenites; the subvolcanic bodies are formed of porphyritic granites and alkaline rhyolites ± riebeckite. The petrochemical data point out the mostly alkaline character, supersaturated in SiO_2, with close contents of Na_2O and K_2O. Among the trace elements Zr, Y, Yb and La display higher contents whereas Co, Ni, Cr, V and Sr show reduced concentrations. Geochronological determinations by Rb/Sr method point out an age of 196 MA for the alkaline magmatites; it indicates their emplacement at the end of the Triassic, that corresponds to the Early Kimmerian Phase. According to the proposed petrogenetic model the alkaline granite assemblage derived from

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acid alkaline magmas formed as a result of the partial melting of the lithosphere in a consumption zone, during the post-collision stage between the Euro-Asian Plate subducted towards the Moesian Plate.

Résumé

Pétrologie de l'association des granites alcalins des massifs d'Iacobdeal et Piatra Roşie (Dobrogea de Nord). Les roches éruptives alcalines appartiennent à l'orogène nord-dogrogéen : elles forment des intrusions petites à caractère plutonique ou subvolcanique, disposées sur deux alignements orientés NW—SE, approximativement parallèles à la zone plissée. Les corps plutoniques sont constitués de différents types de granites alcalins à riebeckite et/ou égyrine et syénites quartzifères alcalines ; les corps subvolcaniques sont formés de granites porphyriques et rhyolites alcalines ± riebeckite. Les données pétrochimiques indiquent un caractère principalement alcalin des magmas, suprasaturé en SiO_2 et des teneurs proches de Na_2O et K_2O. Parmi les éléments mineurs Zr, Y, Yb et La se remarquent par des teneurs plus élevées pendant que Co, Ni, Cr, V et Sr présentent des concentrations plus réduites. Les déterminations géochronologiques par la méthode Rb/Sr indiquent l'âge de 196 m.a. des magmatites alcalines ; résulte donc que la mise en place a eu lieu approximativement à la fin du Trias, période correspondant à la phase cimmérienne ancienne. Selon le modèle pétrogénétique proposé on considère que l'association des granites alcalins est engendrée par les magmas alcalins à caractère acide formés par la fusion partielle de la lithosphère dans une zone de consommation, pendant le stade post-collision entre la Plaque Euro-Asiatique et la Plaque Moesienne.

Introducere

Cercetările efectuate în anii 1984—1986 pentru elaborarea Hărții geologice a R. S. România, scara 1 :50.000, foaia Peceneaga, au dus la obținerea unor date noi asupra rocilor magmatice din masivele Iacobdeal și Piatra Roșie, situate în vecinătatea comunei Turcoaia. Aceste roci formează o asociație de granite alcaline cu riebeckit și/sau egirin, iar date geocronologice recente obținute prin metoda Rb/Sr arată că s-au pus în loc aproximativ la sfîrșitul Triasicului. În lucrarea de față prezentăm o nouă clasificare a rocilor care alcătuiesc această asociație, caracterele geochimice și modul lor de formare.

Dintre studiile anterioare cu caracter petrografic, structural-petrologic și mineralogic, menționăm pe cele elaborate de Mrazec (1899), Murgoci (1905), Cantuniari (1915), Barbu (1959), Ianovici et al. (1969). Alte date referitoare la geneza sau vîrsta acestor roci apar în lucrări efectuate de Streckeisen (1931), Giuşcă (1935), Atanasiu (1940), Mînzatu et al. (1975), Mureşan (1975), Săndulescu (1984), Pop et al. (1985), Întorsureanu (1987), Ştefan et al., 1985 (în Constantinescu et al., 1985).

Cadrul geologic

Magmatitele alcaline din zona Turcoaia aparțin unității de Măcin a Orogenului Nord-Dobrogean, parte integrantă a unei catene care se dezvoltă mult spre est (Crimeea, Caucazul Mare), în timp ce spre NV se

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afundă sub depozitele neogene. În evoluția geotectonică a acestei catene se disting (Săndulescu, 1984) patru cicluri orogene: cadomian, caledonian, hercinic și alpin, cărora le sînt asociate magmatite intrusive sau efuzive, cu chimism calco-alcalin, alcalin sau bazic. Caracteristic pentru Orogenul Nord-Dobrogean sînt structurile în pînze sau solzi cu vergențe NE, legate de ciclul alpin, rezultate probabil, printr-o tectonică de coliziune continentală.

La alcătuirea geologică a zonei Turcoaia participă depozite paleozoice, roci eruptive alcaline și loess (planșa) (Mihăilescu et al., 1986). Depozitele paleozoice sînt reprezentate prin formațiunea de Cerna (Silurian), formațiunea de Bujoarele (Devonian inferior) și formațiunea de Carapelit (Carbonifer inferior), care sînt cutate și au suferit un metamorfism regional de grad foarte scăzut, în timpul mișcărilor hercinice.

În Pleistocen s-a depus loessul și depozitele loessoide, care acoperă, în mare parte, aria cercetată.

Petrografia magmatitelor

Rocile eruptive alcaline formează corpuri plutonice sau subvulcanice, cu dimensiuni mici, dispuse pe două aliniamente orientate NV—SE, aproximativ paralel cu catena orogenă. Primul aliniament situat spre NE este marcat de corpuri plutonice constituite din roci granitice alcaline (masivele Iacobdeal și Piatra Roșie), iar cel de-al doilea, spre SV, este format din corpuri subvulcanice alcătuite preponderent din roci cu structuri porfirice (Dealul lui Manole, Iglicioara Mare, Iglicioara Mică, Gorgane). Se admite că aceste aliniamente se continuă pe direcția sud-est, pe o distanță de cel puțin 25—30 km, pînă la Cîrjelari—Atmagea, unde afloreză granite (masivul Secarul) și riolite alcaline asemănătoare cu cele din zona Turcoaia (Cantuniari, 1915).

Din punct de vedere petrografic asociația granitelor alcaline din zona cercetată este constituită din: sienite cuarțifere alcaline, granite cu riebeckit, granite cu egirin și riebeckit, granite granofirice alcaline, grafite alcaline, microgranite alcaline, granite porfirice, microgranite granofirice cu egirin, aplite granitice cu riebeckit, riolite alcaline și riolite cu riebeckit (planșa).

Sienitele cuarțifere alcaline apar în versantul de NE al masivului Iacobdeal, sub forma unei zone alungite pe direcția VNV—ESE. Sînt roci cenușii-negricioase cu tente verzui, au textura masivă și structura fanerocristalină, cu cristale de feldspați cenușii-verzui, care ajung pînă la dimensiunea de 1 cm. La compoziția mineralogică participă feldspați sodo-potasici (75—81%) cu frecvente microstructuri pertitice, cuarț (12—18%), katoforit (5—6%) și arfvedsonit (2—4%), descrise de Ianovici et al., 1969, albit (2—4%) și magnetit, zircon, apatit, sfen, ca minerale accesorii. Feldspații și amfibolii sînt uneori substituiți parțial de albit, crocidolit, biotit, epidot, clorit, pirită.

Granitele cu riebeckit aflorează atît în versantul NE al masivului Iacobdeal, sub forma a două benzi bine conturate, orientate NV—SE, cît și în partea de N a masivului Piatra Roșie. Macroscopic se caracterizează printr-un aspect pestriț, textură masivă și structuri cu granulație

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medie, care spre periferie devin mai fine. Sînt alcătuite din microclin-pertit (62—68%), cuarț (28—39%), riebeckit (2—8%), arfvedsonit ± egirin (1—2%), zircon, apatit, magnetit, sfen, fluorină. Sînt foarte frecvente microstructurile grafice sau myrmechitice cu aspecte variate. Riebeckitul, cînd este bine dezvoltat, se caracterizează prin conture xenomorfe, fapt menționat și de Murgoci (1905), Cantuniari (1915), Ianovici et al. (1969). Această observație pledează pentru prezența unei cristalizări agpaitice (Întorsureanu, 1987) aspect subliniat și de Bard (1980 fig. 2,19) la un microgranit cu riebeckit și egirin din Alaska. Riebeckitul apare și sub formă de cristale prismatice-aciculare, sau fibroase, varietatea crocidolit (Murgoci, 1905), format, în unele cazuri, prin substituție. În granitele cu riebeckit se întîlnesc xenolite din sienitul cuarțifer alcalin, situat la contact (cariera Iacobdeal).

Granitele cu egirin și riebeckit alcătuiesc creasta principală (vîrful Victoria) a masivului Iacobdeal, cu orientarea VNV-ESE și o zonă mai restrînsă în partea mijlocie a masivului Piatra Roșie. Culoarea acestor roci este cenușie cu tente cărămizii, textura masivă, iar structura are granulație medie. În compoziția mineralogică intră feldspații sodo-potasici (52—63%), cuarț (33—39%), egirin (1—4%), riebeckit (1—3%), magnetit, zircon, sfen, ilmenit, apatit (< 2%). Egirinul apare frecvent asociat cu riebeckitul, dar relațiile nu sînt foarte clare și constante, cel mai adesea egirinul fiind succedat de amfiboli.

Granitele granofirice alcaline formează o zonă alungită între cariera Turcoaia și Valea Oilor. Roca se caracterizează prin culoarea cenușie, cu ușor aspect pătat, datorită prezenței unor cuiburi de feldspați gălbui, textura masivă și structura faneritică cu granulație medie. Microstructura granofirică, asociată adesea cu cea grafică, este caracteristică. Mineralele principale constitutive sînt feldspații sodo-potasici (58—63%), cuarț (32—37%), amfiboli alcalini (0—3%), la care se adaugă minerale accesorii (1—2%), magnetit, zircon, apatit, sfen ± fluorină.

Microgranitele alcaline („granite aplitice", după Cantuniari, 1915) sînt larg răspîndite, alcătuind versantul de SV al masivului Iacobdeal, partea sudică a dealului Piatra Roșie și mameloanele din zona Gorgane. Sînt roci cenușii, cu tente rozii, mărunt cristalizate, cu textura masivă și sînt frecvent afectate de microfisuri de contracție, dispuse în sisteme divers orientate. Mineralogic sînt alcătuite din feldspați sodo-potasici (64—61%), cuarț (35—42%), riebeckit, arfvedsonit, egirin (0—4%), iar dintre mineralele accesorii apar magnetit, zircon, sfen, ilmenit.

Granitele alcaline apar în Dealul lui Manole, corespunzînd cu „granititele" descrise de Cantuniari (1915). Roca proaspătă are culoarea cenușie cu nuanțe gălbui, dar apar și zone slab alterate hidrotermal, cînd roca devine albicioasă și friabilă (cariera Colții Babei). Mineralele principale sînt reprezentate prin felsdpați sodo-potasici (65—69%), cuarț (22—41%), arfvedsonit, katoforit (0—6%), iar cele accesorii prin magnetit, apatit, sfen, zircon ± fluorină. De remarcat că în zona carierei vechi din Dealul lui Manole, conținutul în magnetit crește mult, încît deviază puternic acul busolei. Dintre mineralele postmagmatice apar biotit, hema-

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tit, clorit, cuarț, fluorină ± pirită, formate prin substituții sau depuneri în goluri.

Granitele porfirice alcaline alcătuiesc mameloanele din versantul de SV al Dealului lui Manole, fiind descrise anterior (Cantuniari, 1915) ca „porfire roz". Apar și sub forma unor filoane orientate NV—SE situate în partea de SV a masivului Iacobdeal. Aceste roci au în mod frecvent culoarea gălbuie-cărămizie, datorită alterațiilor supergene, textura masivă și structura porfirică, cu masa fundamentală microgranitică sau granofirică, adesea estompată de un proces de silicifiere. Compoziția mineralogică este reprezentată prin feldspați sodo-potasici (58—64%), cuarț, frecvent secundar (34—41%), magnetit, apatit (< 1%). Se remarcă larga răspîndire a silicifierilor, prezența hematitului sub formă de filonașe și a impregnațiilor de pirită fin granulară, care pot trece și în roca înconjurătoare.

Microgranitele granofirice cu egirin au fost întîlnite doar în Bujorul Bulgăresc, sub forma unui filon orientat N—S, intrus în depozitele paleozoice, pe care le cornifică ușor. Au culoarea cenușie, textura neorientată, structura fin granulară și sînt alcătuite din feldspați sodo-potasici (56—62%), cuarț (36—40%), egirin (6—7%), magnetit, sfen (< 1%).

Aplitele granitice cu riebeckit formează filoane cu grosimi mici, cantonate în granitele cu riebeckit (cariera Iacobdeal), sau în depozitele paleozioce de la contact. Sînt roci albicioase, uneori ușor gălbui, cu textura masivă și structura fin granulară, allotriomorfă. La alcătuirea mineralogică participă feldspați sodo-potasici (62—67%), cuarț (21—36%), riebeckit (1—6%), zircon, apatit, sfen (< 1%).

Riolitele alcaline, descrise în trecut ca „porfire cuarțifere" (Cantuniari, 1915), alcătuiesc mameloanele din sud-vestul zonei cercetate (Iglicioara Mare, Iglicioara Mică, Gorgova etc), apărînd sub formă de corpuri subvulcanice. Foarte rar apar ca filoane cantonate în granitele porfirice alcaline (Dosul Muchiei). Culoarea riolitelor este cenușie, în nuanțe variate, textura pe alocuri laminată tectonic (Iglicioara Mare) și structura tipic porfirică, Fenocristalele sînt reprezentate prin feldspați sodo-potasici (14—18%), cuarț (11—17%) frecvent corodat magmatic, iar masa fundamentală (64—76%) are structura micro-criptocristalină, alcătuită din cuarț, feldspați, magnetit. În stadiul postmagmatic s-au format filonașe de cuarț, asociat uneori cu hematit și feldspat roz, fluorină violacee depusă pe zone de fracturi în masa riolitelor (cariera Iglicioara Mare) și mai rar pirită și calcopirită sub formă de impregnații, care pot trece și în roca gazdă (Iglicioara Mică), unde apare și malachit.

Riolitele cu riebeckit apar subordonat, fiind întîlnite ca filoane în Bujorul Bulgăresc și într-o carieră nouă situată în partea de SE a Dealului lui Manole. Sînt roci cenușii-negricioase, cu textură fluidală sau sferulitică (Bujorul Bulgăresc), concrețiile avînd dimensiuni pînă la 4—5 mm.

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Compoziția mineralogică este asemănătoare cu cea a riolitelor alcaline, la care se adaugă riebeckitul, sub formă de prisme aciculare, orientate adesea după direcția de curgere, sau cu dispoziție radială în cazul sferulitelor, cînd se poate asocia cu arfvedsonit ± magnetit.

Filoane de roci bazice. În masa rocilor granitice alcaline s-au întîlnit (cariera Turcoaia, vîrful Victoria), filoane de roci bazice, cu grosimi decimetrice, intruse pe fracturi, adesea orientate N10°E/60°SE. Aceste roci sînt puternic alterate sau milonitizate, iar la microscop se observă uneori lamele de plagioclazi alterați sau o slabă structură porfirică, aspecte care pledează pentru prezența unor anamesite și/sau dolerite. Vîrsta acestor filoane este incertă, probabil jurasică și sub aspect genetic se leagă de magme bazice, puse în loc pe fracturi adînci, în stadiul de post-coliziune.

Metamorfismul termic

La contactul intruziunilor alcaline cu formațiunea de Carapelit sau cu depozitele devoniene și siluriene, s-au format corneene, frecvent rubanate, cu epidot, zoizit, clinozoizit, albit, tremolit, granat, cuarț, clorit, descrise de Cantuniari (1915). Această asociație arată că s-a format în condițiile faciesului corneenclor cu albit-epidot, pentru care se admit temperaturi aproximativ între 400—530°C (Giușcă, 1974). Spre exteriorul aureolei de contact se trece treptat la roci slab recristalizate cu aspect de șisturi pătate. În masa rocilor alcaline, îndeosebi în corpurile subvulcanice (Dealul lui Manole) apar xenolite, orientate NV—SE cu dimensiuni care pot ajunge pînă la ordinul zecilor de metri, provenite din fundamentul paleozoic și sînt slab cornificate. De asemenea în sienitele cuarțifere alcaline (cariera Iacobdeal), s-au întîlnit, accidental, xenolite cu dimensiuni centimetrice, de roci pelitice transformate în corneene cu biotit, avînd uneori și o slabă mineralizație dispersă de magnetit, hematit și pirită.

Geochimia

Elemente majore. Pentru studiul geochimic s-au utilizat 24 analize de silicați preluate din lucrări anterioare (Cantuniari, 1915; Întorsureanu, 1987) calculîndu-se media conținuturilor în oxizi, pe tipuri petrografice (tab. 1). Din aceste conținuturi se remarcă valoarea ridicată a SiO_2 (= = 73, 91%), în acord cu prezența cuarțului primar în toate rocile asociației. Este de reținut conținutul mic de CaO (0,62%), fapt ce a determinat, pe de o parte, absența mineralelor calcice (plagioclazi, hornblendă, piroxeni),

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cu excepţia katafo.itului şi a albit-oligoclazului care apar subordonat, iar pe de altă parte, formarea egirinului şi a amfibolilor alcalini. Alcaliile au un conţinut mediu relativ ridicat (= 8,99%), observîndu-se o valoare uşor mai mare pentru K_2O(= 4,67%), în comparaţie cu Na_2O (=4,32%).

TABELUL 1

Media conţinuturilor în elemente majore

Oxizi %	1	2	3	4	5	6	7	8	9	10	11
SiO_2	76,53	75,80	76,14	73,62	75,36	76,25	73,51	74,39	73,09	64,39	73,91
TiO_2	0,12	0,19	0,17	0,36	0,23	0,22	0,20	0,28	0,22	0,47	0,25
Al_2O_3	10,84	11,05	11,49	12,22	11,59	11,91	12,18	11,84	12,26	13,72	11,91
Fe_2O_3	2,26	1,64	1,85	2,24	2,32	1,46	2,07	1,87	2,82	3,16	2,17
FeO	0,77	1,06	0,49	1,13	0,84	0,61	1,45	1,35	1,32	3,19	1,22
MnO	0,11	0,28	0,11	0,12	0,07	0,10	0,04	0,12	0,01	0,17	0,11
MgO	0,05	0,13	0,16	0,28	0,09	0,33	0,40	0,01	0,47	0,30	0,22
CaO	0,22	0,46	0,55	0,80	0,35	0,45	0,41	0,30	0,71	1,92	0,62
K_2O	4,66	4,64	5,00	4,42	4,93	3,64	4,44	5,19	4,05	5,74	4,67
Na_2O	3,87	3,93	3,84	4,17	3,88	4,81	4,93	4,18	4,49	5,13	4,32
P_2O_5		0,04		0,07				0,02	0,15	0,22	0,05
H_2O	0,48	0,25	0,28	0,45	0,30	0,21	0,13	0,24		0,16	0,25
Total	99,91	99,47	100,08	99,88	99,96	99,99	99,76	99,79	99,59	98,57	99,70

Calculate pe baza a 24 analize extrase din Cantuniari (1915), 8 analize şi Întorsureanu (1987) 16 analize.

Tipurile de roci : 1, granite cu egirin şi riebeckit(2) ; 2, granite cu riebeckit (4) ; 3, granite granofirice (1) ; 4, granite alcaline (2) ; 5, granite porfirice alcaline (1) ; 6, microgranite alcaline (4) ; 7, aplite granitice cu riebeckit (2) ; 8, riolite cu riebeckit (2) ; 9, riolite alcaline (3) ; 10, sienite cuarţifere alcaline (3) ; 11, media asociaţiei granitelor alcaline (24). Cifrele din paranteză reprezintă numărul de analize.

Compoziţia normată C.I.P.W. a fost proiectată pe diagrama *QAP* (Streckeisen, 1967), unde toate analizele se localizează în cîmpul granitelor alcaline, respectiv al sienitelor cuarţifere alcaline (Întorsureanu, 1987), în acord cu clasificarea petrografică.

Variaţia oxizilor în raport cu SiO_2 este redată în diagrama Harker (fig. 1), remarcîndu-se o evoluţie monotonă, cu slabă tendinţă de scădere a acestora, excepţie făcînd CaO şi $FeO + Fe_2O_3$, unde descreşterea este mai accentuată.

Alcalinitatea rocilor rezultă din diagrama Wright (fig. 2) şi diagrama *al/alk* (fig. 3), întocmită cu ajutorul parametrilor Niggli. Pe prima diagramă analizele se localizează în cîmpul alcalin, în timp ce pe diagrama *al/alk*, aproximativ jumătate din probe trec în cîmpul peralcalin. De asemenea în diagrama $SiO_2/Na_2O + K_2O$ (după Miyashiro, 1978), majori-

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tatea analizelor se plasează în cîmpul alcalin, iar cîteva în domeniul subalcalin printre care și două analize reprezentînd granite cu riebeckit ± egirin (Întorsureanu, 1987, fig. 3).

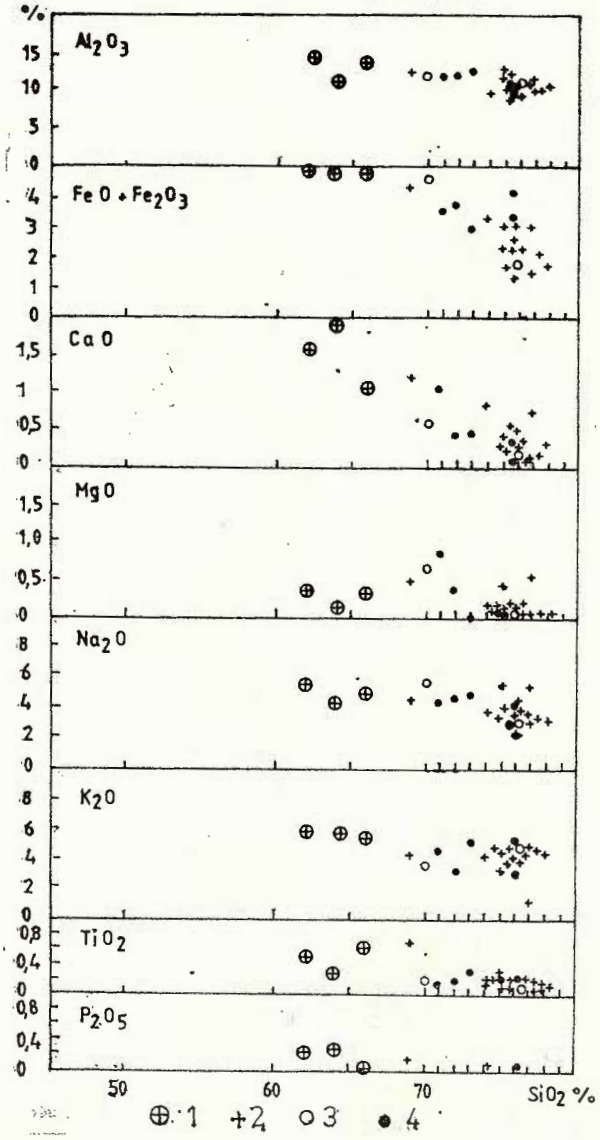

Fig. 1 — Diagrama Harker, oxizi/SiO_2.
Simboluri : 1, sienite cuarțifere alcaline ; 2, roci granitice alcaline ; 3, aplite ; 4, riolite alcaline.

Harker diagram, oxides/SiO_2.
Symbols : 1, alkaline quartziferous syenites ; 2, alkaline granitic rocks ; 3, aplites ; 4, alkaline rhyolites.

După raportul molecular $Al_2O_3 < K_2O + Na_2O$ (Shand, 1951), se constată că numai 13 probe (roci cu riebeckit ± egirin) sînt peralcaline, restul avînd o alcalinitate ușor mai scăzută. Participarea alcaliilor

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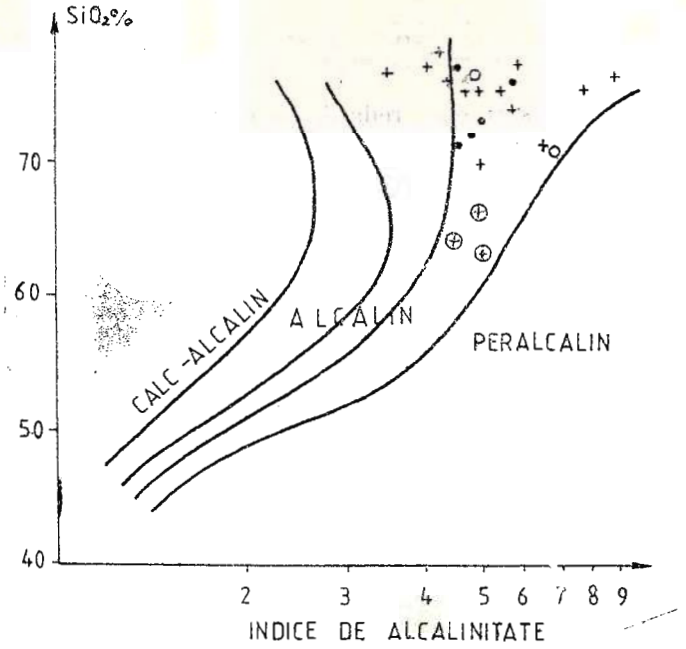

Fig. 2 — Diagram Wright. Simboluri ca în figura 1.
Wright diagram. Symbols as in Figure 1.

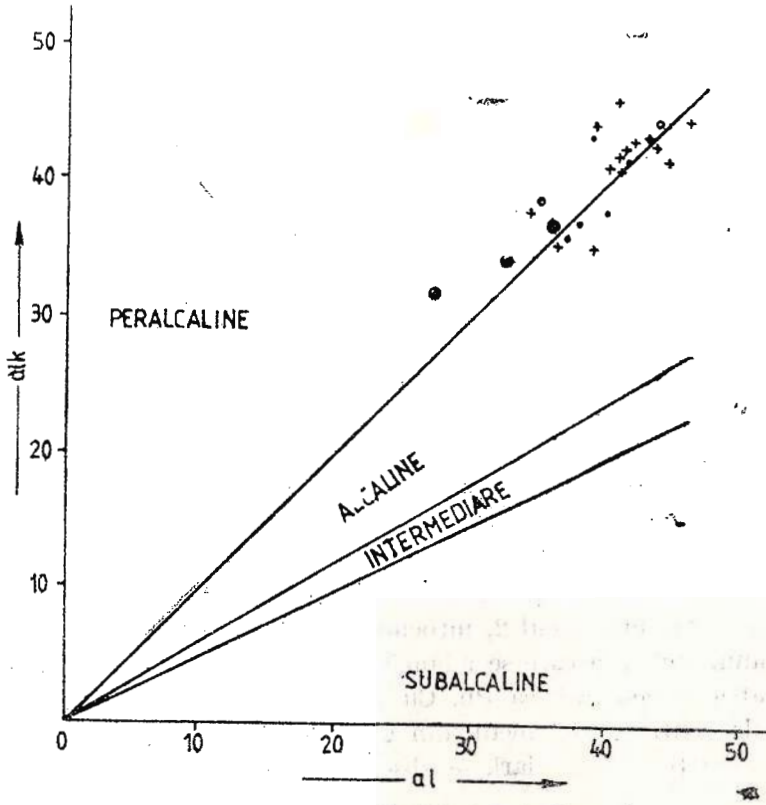

Fig. 3 — Diagrama *al/alk*. Simboluri ca în figura 1.
al/alk diagram. Symbols as in Figure 1.

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și raportul dintre aceștia este redată în diagrama K_2O/Na_2O (fig. 4), unde se remarcă o slabă tendință spre domeniul potasic.

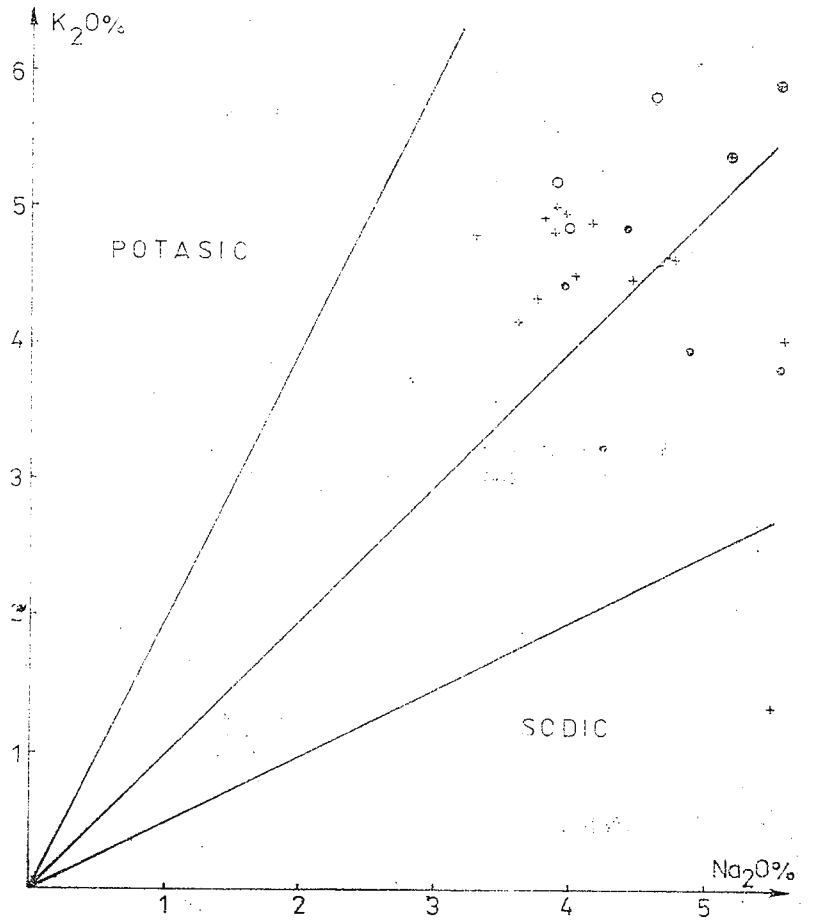

Fig. 4 — Diagrama K_2O/Na_2O. Simboluri ca în figura 1.
K_2O/Na_2O diagram. Symbols as in Figure 1.

Elemente minore. Conținutul mediu în elemente minore pe tipuri de roci este redat în tabelul 2, întocmit pe baza a 16 analize anterioare (Întorsureanu, 1987), la care se adaugă determinările recente pentru U și Th. Din tabel se constată că Pb, Cu și Zn prezintă concentrații foarte apropiate de conținuturile medii din granite (Stemprok, 1982). Concentrații mai ridicate față de clark se observă la Sn, Zr, Yb, La, în timp ce

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Ni, Co, Cr, V şi Sr au valori mici, aşa cum rezultă şi din cercetările anterioare (Berbeleac et al., 1985; Nedelcu et al., 1986). Ba oscilează în limite largi, conţinutul cel mai ridicat fiind determinat în sienitele cuarţifere alcaline.

TABELUL 2

Media conţinuturilor în elemente minore (ppm)

Roci / Elemente	1	2	3	4	5	6	7	8	9	10
a) Dozări spectrale (1) Pb	15,5	15	21	12,2	19	21	5	14,5	9	14
Cu	14	< 7	8	23,5	10	9	10	10	32	13
Zn	93	67	41	42	90	<30	30	72	<30	100
Sn	8,5	5.5	5,5	4	5	7,5	3	6	8,5	3
Ga	29	28	21	27	28	24	27	23,5	16,5	33
Mo	< 2	< 2	<2	< 2	<2	< 2	< 2	< 2	< 2	4
Ni	5,5	5	5	3,5	4,5	4	3,5	4,5	5	4
Co	< 2	< 2	< 2	<2,5	< 2	< 2	< 2	< 2	< 2	<2
Cr	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10
V	< 2	< 3	< 2	<13	<2	< 2	< 2	< 2	< 2	< 2
Sc	8,5	5,8	7	10,5	6	6,5	5,5	9	6,5	26
Y	112	71	95	77	87	87,5	37	89	100	60
Yb	13	8	10	7,5	10	11	4,1	9,5	10	5,6
Zr	1150	490	680	500	720	500	450	865	750	260
Nb	34	23	26	17	22	34,5	11	23,5	36	<10
La	66	81	70	41	110	43	<30	58,5	80	34
Be	7,5	15	4,8	3,3	5,3	6	1,6	5	< 1	2,2
B	<30	<30	<30	<30	<30	<30	<30	<30	<30	<30
Ba	41	55	85	1100	38	238	50	100	175	1300
Sr	11	<10	17	47	<10	23,5	<10	<10	19	28
Bi	< 3	< 3	< 3	< 3	< 3	< 3	<3	< 3	< 3	< 3
b) Dozări prin fluorescenţă Rx(1)										
Rb	192	164	161,2	105,4	124	162	156	170,5	79,7	88,1
c) Spectrometrie gama (2)										
U	6,5	2,7	4,9	2,3	4,4	5	2,6	3	3,9	1,4
Th	24,1	18,8	26,0	18,0	19,7	26,4	6,7	15,6	23,4	7,3

(1) Calculate pe baza a 16 analize (după Întorsureanu, 1987), (2,) Analize efectuate de I. Tiepac. Roci: 1, granit cu egirin şi riebeckit; 2, granit cu riebeckit; 3, granit granofiric; 4, granit alcalin; 5, granit porfiric alcalin; 6, microgranit alcalin; 7, aplit granitic alcalin; 8, riolit cu riebeckit; 9, riolit alcalin; 10, sienit cuarţifer alcalin.

Variaţia conţinuturilor în elemente minore faţă de SiO_2 (fig. 5) arată o dispersie accentuată pentru Rb, Zr, U, Th şi o dispunere aproximativ liniară pentru Sc, Be şi V.

Este semnificativă variaţia raportului Rb/Sr faţă de indicele Nockolds—Allen (fig. 6), care sugerează prezenţa unor diferenţieri magmatice puţin evoluate, în acord cu datele petrografice. În diagrama Rb/

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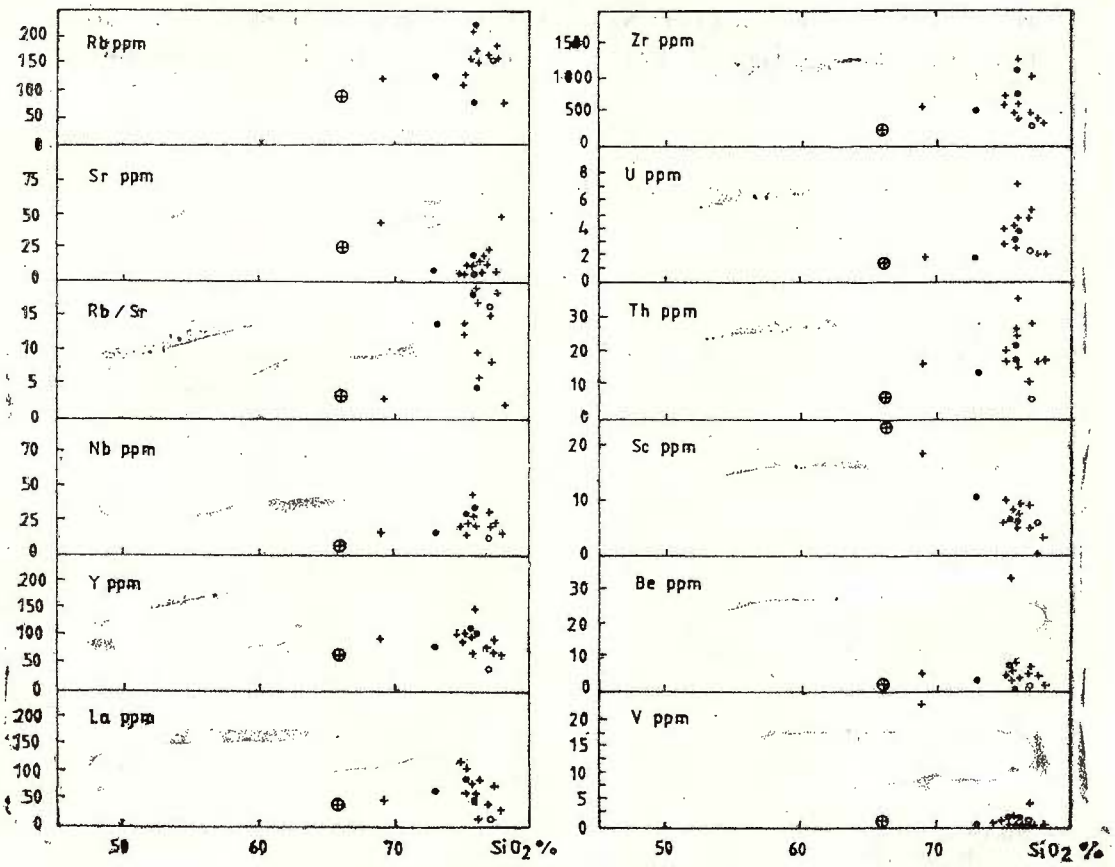

Fig. 5 — Diagrama elemente minore/SiO_2. Simboluri ca în figura 1.
Trace elements/SiO_2 diagram. Symbols as in Figure 1.

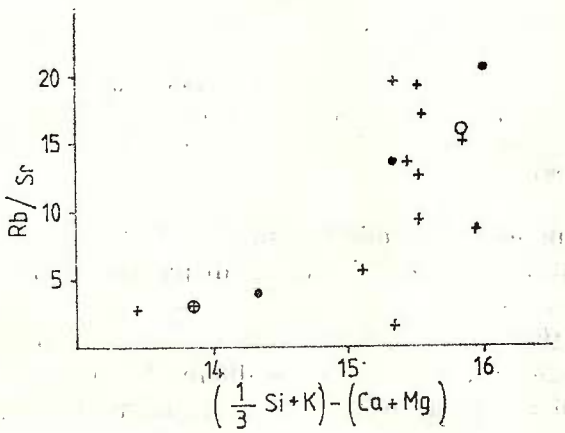

Fig. 6 — Diagrama (Rb/Sr)—indicele Nockolds-Allen. Simboluri ca în figura 1.
Nockolds-Allen index (Rb/Sr) diagram. Symbols as in Figure 1.

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/Y + Nb (fig.7), elaborată pentru interpretarea tectonică a rocilor granitice (Pearce et al., 1984), rocile se localizează în cîmpul granitelor de intraplacă. Această situație se explică prin apartenența asociației de roci alcaline la grupul granitelor post-coliziune (cum vom arăta într-un alt

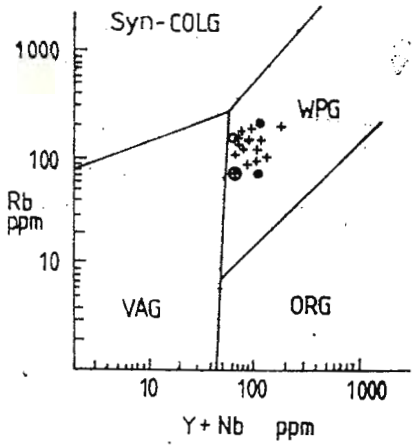

Fig. 7 — Diagrama Rb/Y + Nb. Simboluri ca în figura 1.
Rb/Y + Nb diagram. Symbols as in Figure 1.

capitol) și prin absența din diagramă a unui cîmp al acestor roci, caracterizate prin variații mai largi în componenții respectivi, comparativ cu alte tipuri tectonice de granite (Pearce et al., 1984).

Date geocronologice

În tabelul 3 sînt prezentate datele geocronologice recente obținute prin metoda Rb/Sr (Pop et al., 1985) asupra granitelor alcaline din masivele

TABELUL 3

Date geocronologice Rb/Sr pentru granitoide din Orogenul Nord-Dobrogean

Magmatism	Masivul	Roca	m.a.	Autori
Alcalin	Iacobdeal	Granite alcaline	196	Gh. Popescu (în lucrare)
		Granite alcaline	193±15	Pop et al. (1985)
	Secar	Granite alcaline (2)	197±10	Pop et al. (1985)
Calco-alcalin	Greci-Morsu	Granit (2) Granit (1)	192±10 227±10	Pop et al. (1985)
	Coșlugea	Granit (3) Granit (1)	193±10 238±10	Pop et al. (1985)
	Pricopan	Granite (3)	250±10	Pop et al. (1985)

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Iacobdeal şi Secar (Cîrjelari), la care se adaugă şi determinările efectuate în cadrul acestui studiu, ultimele indicînd vîrsta de 196 m.a. (fig. 8). După aceste date rezultă că asociaţia granitelor alcaline s-a pus în loc aproximativ la sfîrşitul Triasicului, ceea ce ar corespunde cu faza chimerică

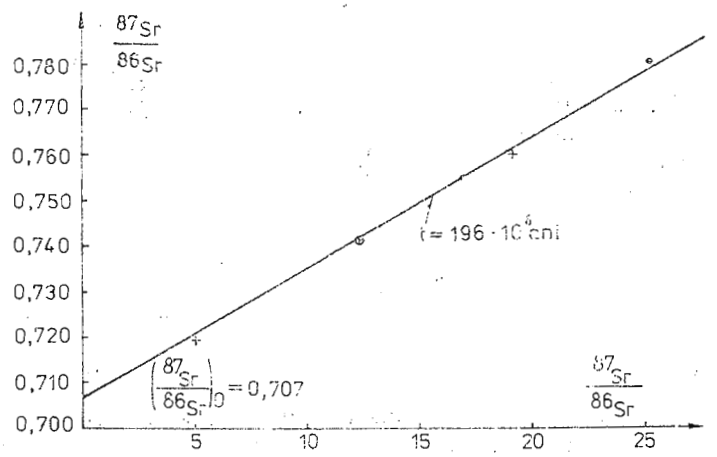

Fig. 8 — Izocrona Rb/Sr pe roci globale din zona Turcoaia. Simboluri ca în figura 1.

Rb/Sr isochrone on global rocks in Turcoaia Zone. Symbols as in Figure 1.

veche. Pentru comparaţii şi consideraţii petrogenetice menţionăm în tabel şi vîrstele unor granite calco-alcaline (Pricopan, Greci-Morsu, Coşlugea), care aparţin de asemenea Orogenului Nord-Dobrogean. Determinările K/Ar (Mînzatu et al., 1975) par discutabile întrucît arată diferenţe foarte mari (264 — 210 m.a.) pentru magmatitele din masivul Iacobdeal, iar vîrstele stabilite pentru alte masive (Greci, Pricopan) nu se corelează decît rareori cu cele din tabelul 3.

Asupra relaţiilor de vîrstă relativă dintre rocile asociaţiei dispunem doar de cîteva observaţii care permit stabilirea unei succesiuni incomplete. Astfel sienitele cuarţifere alcaline sînt anterioare granitelor cu riebeckit, întrucît acestea conţin enclave de sienit. Prezenţa xenolitelor din fundamentul paleozoic, în aliniamentul subvulcanic (Dealul lui Manole, Dosul Muchiei) pledează pentru un nivel de eroziune mai puţin avansat în comparaţie cu aliniamentul de corpuri plutonice, de unde se poate deduce că rocile porfirice s-au pus în loc mai tîrziu şi încheie activitatea magmatismului alcalin (Giuşcă, 1974). Dintre rocile subvulcanice riolitele alcaline par să fie ultimele întrucît la Dosul Muchiei apare un filon de riolite localizat în granite porfirice alcaline.

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Petrogeneza

Factorii principali care au controlat geneza asociației granitelor alcaline au fost: condițiile geotectonice în care s-au format magmele, chimismul și evoluția acestora.

Condiții geotectonice. Asociația granitelor alcaline aparține Orogenului Nord-Dobrogean situat între Placa Euro-Asiatică (NE) și Placa Moesică (SV), ale căror margini au fost active și în Mezozoic, după cum se poate deduce din încălecările alpine cu vergențe nord-estice. Acest fapt, împreună cu dispunerea corpurilor granitice pe aliniamente orientate aproximativ paralel cu structurile majore ale orogenului, ne sugerează, ținînd seama de teoria tectonicii plăcilor, că magmele s-au format într-o zonă de subducție (Întorsureanu, 1987, fig. 6). În această ipoteză fosa vechii subducții ar fi marcată de actuala Depresiune Predobrogeană, considerată ca avanfosa hercinică a Dobrogei septentrionale (Săndulescu, 1984). Planul Benioff a fost înclinat spre Placa Moesică iar subducția a început probabil în Carboniferul inferior și s-a continuat în Mezozoic printr-un stadiu de coliziune continent/continent. De remarcat că după Colman-Sadd (1982), în stadiul de coliziune continentală, se pot continua atît subducția cît și generarea de magme. Cadrul orogenic pledează în favoarea acestei interpretări, iar datele geocronologice (tab. 3), caracterele geochimice și absența complexelor inelare permit încadrarea asociației granitelor alcaline la grupul granitelor post-coliziune (Pearce et al., 1984).

Chimismul magmelor. Caracterele geochimice ale asociației granitelor alcaline sînt, foarte probabil, compatibile cu chimismul magmatismului generat în stadiul post-coliziune. Astfel conținutul ridicat de SiO_2 (= 73,91 %) se explică prin participarea materialului sialic așa cum sugerează și valoarea raportului $^{87}Sr/^{86}Sr = 0,707$ pentru magmatitele alcaline din zona Turcoaia stabilită atît anterior (Pop et al., 1985) cît și în cadrul studiului de față. Se presupune că aceste valori implică atît o participare substanțială a materialului continental la formarea magmelor în zone de subducție (James, 1981), cît și contaminarea sialică (Pop et al., 1985).

Conținutul de alcalii poate fi de asemenea realizat în stadiul de post-coliziune, cînd chimismul magmelor este controlat de gradul de participare a crustei și a mantalei la formarea topiturilor (Pearce et al., 1984). Această posibilitate rezultă în mod clar din tabelul 4 care cuprinde conținuturile medii de Na_2O și K_2O din asociația granitelor alcaline din zona Turcoaia, comparativ cu conținuturile de alcalii din unele granite și riolite, considerate că s-au format în zone de coliziune continent/continent.

Caracteristice sînt și conținuturile foarte mici de CaO (= 0,62 %), comparativ cu alte roci granitice sau riolitice calco-alcaline, deși există unele valori apropiate (tab. 4). Conținuturile scăzute de CaO se pot explica prin acțiunea mai multor factori, dar cel mai plauzibil este topirea parțială sau incipientă a litosferei, în ultimul stadiu de coliziune. De remarcat că aceste conținuturi mici de CaO nu par să confirme ipoteza asimilării unor roci carbonatice, care ar fi dus la formarea granitelor alcaline din Dobrogea (Streckeisen, 1931).

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Ceilalți componenți majori și minori prezintă concentrații în limite relativ largi, funcție de numeroase variabile, implicate în geneza magmatismului în zone de subducție (James, 1981) sau post-coliziune (Pearce et al., 1984).

TABELUL 4

Conținuturi medii de alcalii și oxid de calciu din roci granitice și riolitice din zone orogene

Roci, localizare	Nr. analize	Alcalii (%)			CaO %	Chimismul, după:	Subducția sau coliziunea, după:
		Σ	Na_2O	K_2O			
Asociația granitelor alcaline (Iacobdeal, Piatra Roșie) Orogenul Nord-Dobrogean	24	8,99	4,32	4,67	0,62	Cantuniari (1915), Întorsureanu (1987)	Întorsureanu (1987)
Granite (Coșlugea) Orogenul Nord-Dobrogean	3	8,61	4,46	4,15	1,20	Giușcă (1974)	Săndulescu (1984) Întorsureanu (1987)
Granite, aplite (Pricopan) Orogenul Nord-Dobrogean	2	8,63	2,90	5,73	0,47	Giuș.ă (1974)	Săndulescu (1984) Întorsureanu (1987)
Riolite (Camena) Orogenul Nord-Dobrogean	22	7,85	1,77	6,08	0,66	Cădere (1925)	Întorsureanu (1987)
Granite (baratite) (Bocșa, Banat) Carpații Meridionali	2	9,16	3,39	5,77	1,62	Russo-Săndulescu, Berza (1979)	Rădulescu, Săndulescu (1973)
Granite (Querigut, Pirinei)	46	8,05	3,49	4,56	1,33	Fourcade, Allegre (1981)	Pearce et al. (1984)
Granite (Manaslu, Himalaia)	23	8,56	4,06	4,50	0,50	Le Fort (1981)	Pearce et al. (1984)

Evoluția magmelor. După formarea magmelor prin reciclarea litosferei, acestea au suferit procese evolutive complexe care au condus la apariția asociației petrografice. Factorii de control au fost: migrarea intermitentă, amestecul, diferențierea și cristalizarea magmelor.

Migrarea intermitentă a magmelor este sugerată atît de relațiile de succesiune dintre tipurile de roci cît și de datele geocronologice. Acest proces, cu rol important, se explică prin: 1. condițiile de subducție ce presupun că magmele se formează într-un interval mare de timp, care ajunge pînă la ordinul zecilor de milioane de ani și la adîncimi de aproximativ 165 km (Întorsureanu, 1987, fig. 5); 2. compresiunea exercitată de mișcarea convergentă a plăcilor litosferice și care probabil a avut stadii de maxim sau de minim, determinate de ritmuri diferite de subducție (Lowrie, Hey, 1981), în anumite intervale de timp; 3. apariția unor fracturi

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majore în scoarţă, aproximativ paralele cu fosa de subducţie, facilitînd migrarea ascendentă a magmelor, în stadii de compresiune maximă, care pot coincide cu mişcări orogene. Astfel punerea în loc a magmatitelor alcaline se asociază cu mişcările chimerice vechi, în timp ce granitoidele calco-alcaline din masivele Pricopan sau Greci, generate probabil în aceeaşi zonă de subducţie (Întorsureanu, 1987, fig. 6) se leagă de faze mai vechi (saalică, pfalzică). De observat că acest model evolutiv este incompatibil cu ipoteza formării granitelor alcaline şi calco-alcaline din unitatea de Măcin, prin diferenţierea magmatică a unui „lacolit" profund (Atanasiu, 1940), întrucît acesta s-ar fi consolidat într-un interval mult mai mic de aproximativ $10^5 - 10^6$ ani (Spera, 1980, fide Gill, 1981).

Amestecul de magme, la diferite scări, este posibil în stadiul postcoliziune, cînd se pot forma topituri cu chimism mai mult sau mai puţin diferit, prin fuziunea crustei sau a mantalei superioare (Pearce et al., 1984), fără ca acestea să migreze imediat.

Diferenţierea şi cristalizarea sînt procese asociate care s-au desfăşurat în stadiul final de evoluţie. Produsele de diferenţiere magmatică par a fi reprezentate doar prin filoane de aplite granitice cu riebeckit, de unde se deduce prezenţa unei diferenţieri de slabă intensitate. De asemenea produsele postmagmatice sînt în general slab dezvoltate, aşa cum rezultă din descrierea petrografică.

După intruziunea corpurilor alcaline acestea au fost afectate de procese tectonice care au dus la apariţia unor deformări locale, reprezentate prin cuarţ cu extincţii ondulatorii, texturi orientate sau laminate, cataclazite, milonite.

Mulţumim dr. A. Ştefan pentru sugestiile şi observaţiile critice în faza de manuscris, care au contribuit la îmbunătăţirea şi definitivarea lucrării. De asemenea mulţumim colegului I. Tiepac pentru dozarea U şi Th prin spectrometrie gama.

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PETROLOGY OF THE ALKALINE GRANITE ASSEMBLAGE IN THE IACOBDEAL AND PIATRA ROȘIE MASSIFS (NORTH DOBROGEA)

(Summary)

The alkaline granite assemblage belongs to the North-Dobrogean Orogen, integral part of a chain mostly developed eastwards (Crimea, Great Caucasus). Four orogenic cycles are distinguished in the evolution of this chain (Săndulescu, 1984) — Cadomian, Caledonian, Hercynian and Alpine — to which intrusive or effusive magmatites with a calc-alkaline, alkaline or basic chemistry are associated. During the Alpine cycle occurred nappe structures or NE-trending scales, probably the result of a continental tectonic collision. The geological structure of the study area is represented by Paleozoic formations, alkaline magmatites and loess (Plate).

The alkaline eruptive rocks form small-sized plutonic or subvolcanic bodies disposed along two NW—SE trending alignments almost parallel to the orogenic chain. The first alignment situated towards NE is marked by plutonic bodies (Iacobdeal, Piatra Roșie) and the second one, towards SW, consists of subvolcanic bodies (Dealul lui Manole, Iglicioara Mare, Iglicioara Mică, Gorgane). Petrographically the alkaline granite assemblage is represented by alkaline quartziferous syenites, riebeckite-bearing granites, egyrine- and riebeckite-granites, alkaline granophyre granites, alkaline granites, alkaline microgranites, alkaline porphyritic granites, egyrine-bearing granophyric microgranites, riebeckite-bearing granitic aplites, alkaline rhyolites and riebeckite-bearing rhyolites.

At the contact of the alkaline intrusions with the Paleozoic formations hornfelses, frequently banded, formed under conditions of the albite-epidote hornfels facies.

From the chemical point of view among the major elements (Tab. 1) SiO_2 displays the highest value (73.91%) and CaO the lowest one (0.62%), whereas K_2O and Na_2O show close values (4.67 and 4.32%, respectively). On Wright diagram (Fig. 2) the analyses are plotted in the alkaline field and on the *al/alk* diagram (Fig. 3) almost half of the samples occur in the peralkaline field. The average content of trace elements (Tab. 2) points to higher concentrations of Sn, Zr, Yb, La versus clark, whereas Ni, Co, Cr, V and Sr display lower values. The Rb/Sr variation diagram versus Nockolds-Allen index (Fig. 6) suggests a poor magmatic differentiation according to the petrographic and petrochemical data. On the

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Rb/Y + Nb diagram (Fig. 7), elaborated for the tectonic interpretation of granites (Pearce, 1984), the rocks fall in the intraplate granite field. It is explained by the appurtenance of the alkaline rock assemblage to the post-collision granite group and by the absence of a field for these rocks on the diagram.

The geochronological data (Tab. 3) determined by Rb/Sr methods point to 196 MA (Fig. 8) rendering evident the emplacement of the alkaline granite assemblage at the end of the Triassic, that corresponds to the Early Kimmerian Phase.

According to the proposed petrogenetic model the acid alkaline magmas formed in a subduction zone (post-collision stage) between the Euro-Asian Plate, lying north-eastwards, and the Moesian Plate, situated in the south-west of the orogenic chain. Benioff plane dips south-westwards and the trough of the old subduction is marked by the present Predobrogean Depression with molasse deposits and is regarded as the Hercynian Foredeep of the North Dobrogea (Săndulescu, 1984). The chemistry of the generating magmas is compatible with their formation way in the consumption zones. It results from the comparison of the average contents of Na_2O, K_2O and CaO in the alkaline granite assemblage with other granites or rhyolites in different regions formed during the continent/continent collision stage (Tab. 4). The magmas formed as a result of the partial melting of the lithosphere, in the convergence zone between the two plates, characterized by stages of maximum and minimum, intermittently migrated towards the surface, as shown by geochronological data.

The tectonic movements subsequent to the emplacement of the alkaline intrusions determined the appearance of local deformations (oriented or laminated textures, cataclasites, mylonites) in their mass.

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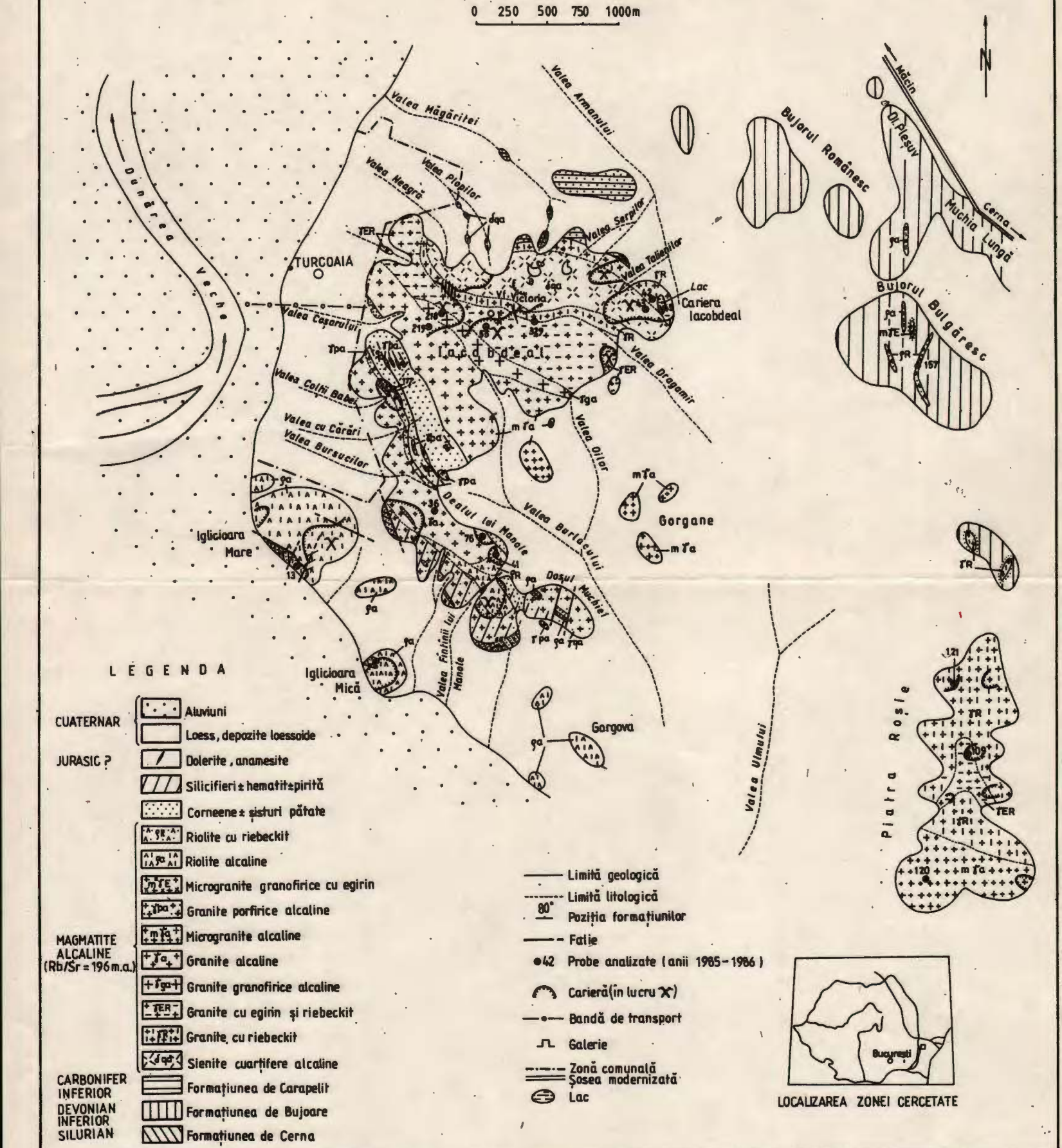
HARTA GEOLOGICĂ A ZONEI TURCOAIA (DOBROGEA DE NORD-VEST)



LEGENDA

- CUATERNAR
  - Aluviuni
  - Loess, depozite loessoide
- JURASIC ?
  - Dolerite, anamesite
  - Silicifieri ± hematit ± pirită
  - Corneene ± șisturi pătate
- MAGMATITE ALCALINE (Rb/Sr = 196 m.a.)
  - Riolite cu riebeckit
  - Riolite alcaline
  - Microgranite granofirice cu egirin
  - Granite porfirice alcaline
  - Microgranite alcaline
  - Granite alcaline
  - Granite granofirice alcaline
  - Granite cu egirin și riebeckit
  - Granite cu riebeckit
  - Sienite cuarțifere alcaline
- CARBONIFER INFERIOR
  - Formațiunea de Carapelit
- DEVONIAN INFERIOR
  - Formațiunea de Bujoare
- SILURIAN
  - Formațiunea de Cerna

- Limită geologică
- Limită litologică
- 80° Poziția formațiunilor
- Fatie
- ●42 Probe analizate (anii 1985–1986)
- Carieră (în lucru)
- Bandă de transport
- Galerie
- Zonă comunală / Șosea modernizată
- Lac

LOCALIZAREA ZONEI CERCETATE

Imprim. Atel. Inst. Geol. Geof.

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D. S. Inst. Geol. Geofiz. vol. 74/1 (1987), 1989, pag. 87—106

1. MINERALOGIE—PETROLOGIE

PETROLOGIA ROCILOR MAGMATICE

ANDEZITUL CU PLAGIOCLAZI (± HORNBLENDĂ OPACITIZATĂ) DE TIP MĂGURA (MUNȚII CĂLIMANI)[1]

de

EUGENIA NIȚOI[2]

Andesites. Neogene. Acidic magmas. Andesitic composition. Mineral composition. Major elements. East Carpathians — Neogene eruptive-Quaternary — Călimani Mountains.

Abstract

Măgura Plagioclase Andesite (± Opacitized Hornblende) (East Carpathians, Călimani Mountains). Măgura plagioclase andesite (± opacitized hornblende) is situated in the South Călimani Mts, at the boundary between the Crystalline-Mesozoic Zone and the Neogene eruptive. It is considered to represent a significant phase in the succession of the eruptions in the Călimani Mts, being situated towards the final part of those that take place in the lower compartment C_1 of the study zone. The Măgura andesite occurs as enrooted bodies and lava flows. Such andesites with an extremely simple mineralogy originate in a magma situated close to the surface, so that the constituting minerals have not had the necessary time to crystallize and remained as skeletal microlites, incompletely formed. Eight new chemical analyses are presented which, compared with analyses of some comparison terms belonging to other types of andesitic rocks in the close vicinity, prove the well-defined character of the study andesites and their special position in relation to the comparison terms.

Résumé

Andésite à plagioclases (± hornblende opacifiée) de Măgura (Carpathes Orientales, monts de Călimani). L'andésite à plagioclases (± hornblende opacifiée) de Măgura est situé dans les monts de Călimani de sud à la limite entre la zone cristallino-mésozoïque et l'éruptif néogène

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et représente une phase importante dans la succession des éruptions des monts de Călimani. Il se forme vers la fin des éruptions manifestées dans le compartiment inférieur C_1 de la zone susmentionnée.

L'andésite se présente sous forme de corps enracinés et écoulements de laves. Il y a des andésites à une minéralogie extrémement simple originaires d'un magma placé dans des conditions proches de celles de surface, quand les minéraux components n'ont pas eu le temps de cristalliser et sont restés au stade de microlites squélétiques, partialement formés. On y présente 8 analyses chimiques nouvelles qui, comparées aux analyses de certains terms de comparaison d'autres types de roches andésitiques des environs, relèvent le caractère chimique bien définie de ces andésites, la position particulière par rapport aux terms de comparaison.

1. Introducere

Andezitul cu plagioclazi (± hornblendă opacitizată) este situat în munții Călimani de sud, la limita dintre zona cristalino-mezozoică și eruptivul neogen. Primele observații asupra acestei roci la care ne referim le face Török (1944, 1948, 1950, 1951). Fără prea mari detalieri este semnalată prezența andezitului tip Măgura în raportul preliminar al foii Bilbor (Peltz et al., 1986) sub denumirea de andezit afanitic de Măgura.

Török distinge în munții Călimani mai multe cicluri de punere în loc a rocilor eruptive, pe care le descrie în ordinea succesiunii lor, grupate pe serii și faciesuri de roci, fiecare dintre acestea având caractere petrografice și mineralogice distincte. Referindu-se strict la zona de față, Török (1944) semnalează prezența ciclului de erupție I (faciesul de Drăgoiasa cu dacite și tufuri dacitice) și a ciclului IV (seria Bilbor-Sărmaș). Andezitul cu plagioclazi (± hornblendă opacitizată) de Măgura a fost situat de acest cercetător în ciclul IV, în care apar cu precădere andezitele cu biotit și hornblendă (denumire dată și acestui andezit de Măgura) sub formă de scuturi și mici neck-uri de lavă acidă și a căror venire este considerată a fi ultima din ciclul vulcanic al zonei la care ne referim. În lucrările sale, Török susține următoarea succesiune a formațiunilor geologice din perimetrul Drăgoiasa—Bilbor: fundamentul cristalin, dacitul de Drăgoiasa, formațiuni eruptive nediferențiate (corespondentul lor actual fiind formațiunea vulcano-sedimentară), curgeri de lave, mai mult sau mai puțin acide, din centrele Măgura, Văcăria și Fundoaia.

În lucrarea de față, încercăm să aducem noi date referitoare la petrografia, mineralogia și geochimia andezitului cu plagioclazi (± hornblendă opacitizată) tip Măgura, care s-a dovedit a reprezenta un tip de rocă cu caractere mineralogice și chimice bine individualizate, ce se poate alătura celorlalte tipuri de roci aparținînd aceleiași familii de andezite (în sens larg) din munții Călimani și de a cărui prezență, în teritoriul cercetat, va trebui să se țină seama în stabilirea succesiunii erupțiilor, constituind chiar o etapă importantă în cadrul acestora. Astfel, cercetările efectuate în zona Bilbor—Drăgoiasa ne-au permis stabilirea următoarei succesiuni de erupții: dacit de Drăgoiasa, formațiune vulcano-sedimentară cu alternanțe și îndințări de depozite sedimentare, tufuri dacitice, riodacite de Văcăria, andezite cu plagioclazi (± hornblendă opacitizată) tip

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Măgura. Întreaga succesiune o atribuim compartimentului inferior al munților Călimani.

Harta geologică anexată lucrării (fig. 1) reprezintă ansamblul formațiunilor geologice din perimetrul cercetat, a căror succesiune în timp și spațiu, stabilită prin cercetări de teren și laborator, are, deocamdată, o valoare locală.

Vîrsta rocilor sedimentare din imediata apropiere a zonei cercetate, pe care formațiunile eruptive le acoperă sau cu care acestea sînt îndințate, este considerată a fi cea stabilită de Bandrabur (în Peltz et al., 1987) și anume Ponțian-Romanian.

2. Cadrul geologic

Fundamentul cristalin. Este reprezentat prin roci cristaline (șisturi clorito-sericitoase, paragnaise, micașisturi, calcare și dolomite cristaline) ce aparțin din punct de vedere petrografic grupurilor de Rebra și Tulghes, iar din punct de vedere tectonic pînzei bucovinice (Gheuca în Peltz et al., 1987). A fost studiat fie în aflorimente continue, situate la cote inferioare, pe văile Fundoaia, Dobreanu, Borcut, unde este acoperit de formațiunea vulcano-sedimentară sau de curgerile de lave andezitice și dacitice, fie în cîteva petece izolate situate la cote ridicate din zonă (creasta Păltiniș, Piciorul Bilborului). Nu rareori au fost întîlnite anclave de roci cristaline în secțiunile subțiri executate pe roci eruptive. Pe baza aflorimentelor actuale, situate între cotele 1450 m și 1050 m și a răspîndirii celorlalte formațiuni, s-ar putea reconstitui cu aproximație, suprafața morfologică a fundamentului cristalin, bănuită a fi existat în timpul punerii în loc a formațiunilor eruptive. Astfel, între vîrful Răcila, vîrful Păltiniș și Măgura se pare că a existat o creastă de șisturi cristaline, care înaintea ieșirii la zi a magmei andezitice de tip Măgura, a funcționat ca o barieră între zonele depresionare de la Drăgoiasa și cele de la Bilbor. Savul (1938) prelungește această „creastă" pînă la nord de Șaru Dornei (N și NV zonei cercetate). Această creastă de cristalin, ca și restul fundamentului cristalin, a fost intens fracturată în „sloiuri dezechilibrate" (Török, 1948), fapt ce a provocat intruziunea magmelor calcalcaline și apariția la zi a fenomenelor magmatice.

Pe valea Lespezi și afluenții săi, fundamentul cristalin este afectat de o falie majoră, mai nouă, cu direcție NNV—SSE, falie marcată de intense zone de brecifiere și care poate reprezenta o reactivare a unei vechi falii (fig. 1).

Formațiunea vulcano-sedimentară (seria I). Între pîrîul Fundoaia și Dobreanu, pe versantul estic al dealului Văcăria și pe valea Lespezi apare o formațiune vulcano-sedimentară, cu dezvoltare redusă, între cotele 1080 m și 1350 m, ce stă transgresiv peste fundamentul cristalin. Dacă pe valea Lespezi și pe afluenții de dreapta ai văii Dobreanu se poate vedea clar amplasarea acestei formațiuni direct peste fundamentul cristalin, în versantul estic al dealului Văcăria este mai greu de precizat natura acestor relații, situația reală fiind mascată și complicată de existența faliei amintite, care „amputează" zona de contact, favorizînd schimbarea relațiilor spațiale dintre formațiunea vulcano-sedimentară și riodacit. Pe

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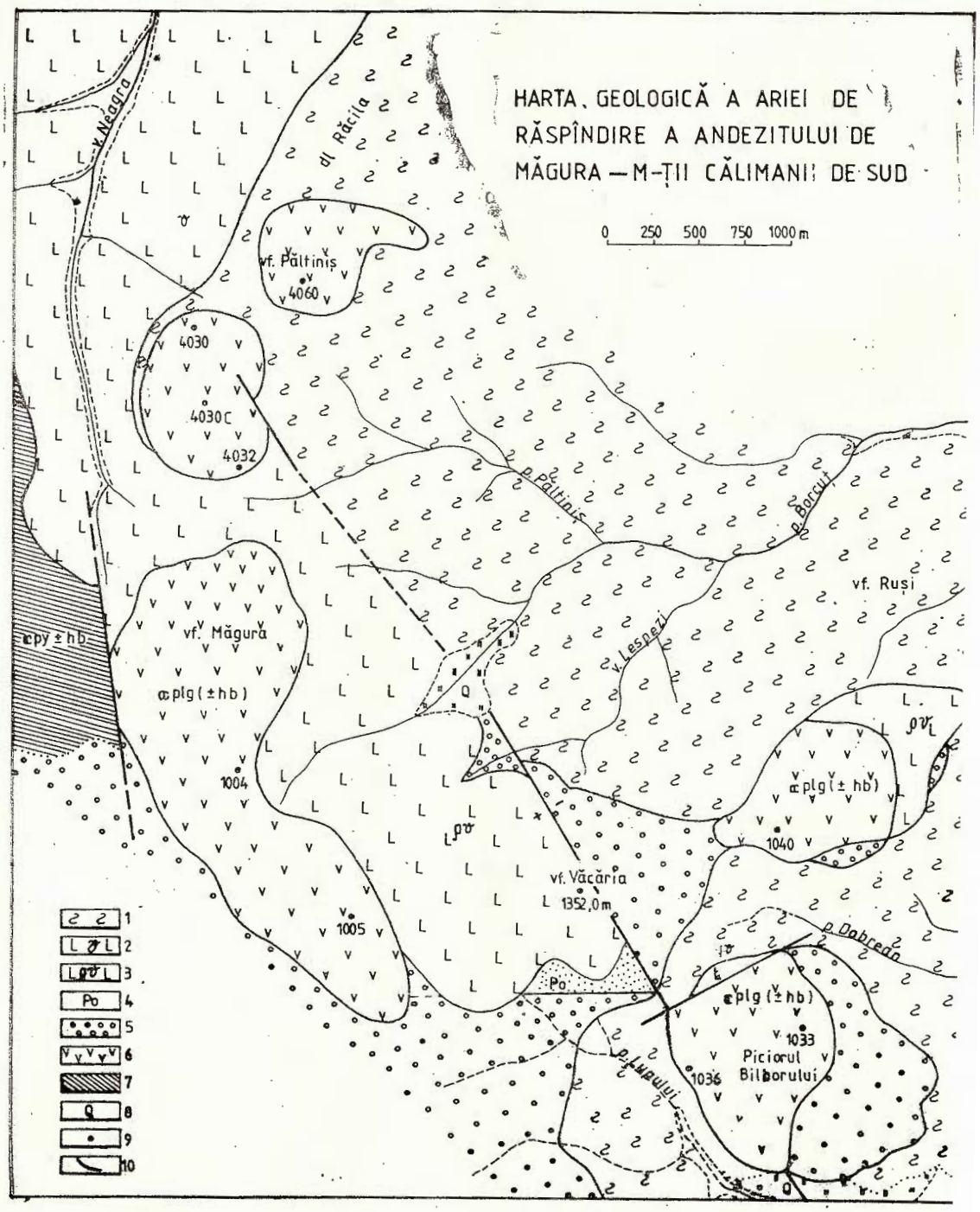

Fig. 1

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întreaga suprafață cercetată formațiunea vulcano-sedimentară este acoperită fie de riodacitul de Văcăria, fie de andezitul cu plagioclazi (± hornblendă opacitizată) de Măgura.

Dacă ținem cont de poziția formațiunii vulcano-sedimentare, de relațiile acesteia cu depozitele sedimentare (îndințare sau alternanță), de modul de depunere, de compoziția mineralogică și petrografică, putem bănui existența, în cadrul acesteia, a seriei inferioare I (serie localizată în întreg lanțul eruptiv neogen, la partea bazală a formațiunii vulcano-sedimentare). La alcătuirea petrografică a formațiunii participă : cinerite lapillice andezitice, microaglomerate cu liant tufaceu andezitic, tufuri andezitice, nisipuri andezitice cu concrețiuni de siderit. Întreaga stivă de roci nu pare a avea o orientare preferențială ci doar o ușoară sortare gravitațională. Rar apar intercalații de lave andezitice (cu piroxeni și hornblendă) cu aspect vacuolar (pe afluenții drepți ai văilor Dobreanu și Fundoaia) și mici corpuri de andezite cu piroxeni (pe valea Lespezi). Natura petrografică a fragmentelor de roci eruptive ce intră în alcătuirea acestei formațiuni vulcano-sedimentare este predominant andezitică. Rar au fost întîlnite și fragmente de dacite (Drăgoiasa ?).

Formațiuni sedimentare. În zona de izvoare a pîrîului Fundoaia au fost întîlnite argile fin-nisipoase-gălbui ce stau peste aglomeratele andezitice ale formațiunii vulcano-sedimentare și suportă un nivel de tufuri dacitice.

Tufuri dacitice. Pe clina vestică a dealului Văcăria, în apropierea cotei de 1250 m apare de sub riodacitul de Văcăria un tuf dacitic alcătuit din fragmente mărunte (0,50 cm) de dacite, roci cristaline și cristale sfărîmate sau ușor rotunjite de cuarț, piroxeni, biotit și sticlă izotropă.

Riodacite de Văcăria. Aceste roci ocupă o mare parte din suprafața cercetată fiind situate sub lavele andezitului cu plagioclazi (± hornblendă opacitizată). Modul de zăcămînt al riodacitelor are o mare uniformitate în ansamblul aflorimentelor întîlnite. Apar curgeri de lavă puse în loc de-a lungul unei zone fisurale (dealul Văcăria-valea Lespezi-valea Măgura) cu direcție NV—SE și a unor mici apofize (Piciorul Bilborului, versant drept al văii Rușii). Curgerile de lavă par a fi alcătuite din mai multe veniri succesive, cu o superpoziție regulată între ele, fiecare curgere prezentînd o separație scoriacee pe fața superioară. Aparent orizontale, curgerile

Fig. 1 — Harta geologică a ariei de răspîndire a andezitului de tip Măgura (munții Călimani de sud).

1, fundamentul cristalin ; 2, dacitul de Drăgoiasa ; 3, riodacitul de Văcăria ; 4, formațiuni sedimentare ; 5, piroclastite nediferențiate ; 6, andezite tip Măgura ; 7, andezite cu piroxeni și hornblendă ; 8, Cuaternar ; 9, analize chimice ; 10, falii.

Geological map of the distribution of the Măgura-type andesite (South Călimani Mts).

1, crystalline basement ; 2, Drăgoiasa dacite ; 3, Văcăria rhyodacite ; 4, sedimentary formations ; 5, nondifferentiated pyroclastites ; 6, Măgura-type andesites ; 7, andesites + pyroxenes + hornblende ; 8, Quaternary ; 9, chemical analyses ; 10, faults.

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sînt ușor înclinate către NE. Sînt frecvente aspectele microrubanate, mai accentuate macroscopic decît microscopic (alterare diferențială) și mai rar, aspectele brecioase.

Ținînd cont de aspectele structurale diferite ale pastei, riodacitele de Văcăria pot fi separate în : a) riodacite efuzive hialine cu texturi ușor fluidale ; b) riodacite efuzive cu structură pilotaxitică și textură fluidală; c) riodacite de corp cu structura microgranulară și textura masivă.

a) Gradul scăzut de cristalinitate al acestor roci arată o răcire rapidă a lavei și acumularea acesteia în zone apropiate de suprafață (versant drept al pîrîului Rușii). Macroscopic rocile au aspect afanitic, leucocrat și microrubanat. Microscopic, se observă predominanța masei fundamentale constituită din sticlă, microlite de feldspat plagioclaz și biotit în care plutesc rare fenocristale mărunte de plagioclaz, biotit și hornblendă.

b) Acest aspect al riodacitelor predomină în zona cercetată, observîndu-se pregnant textura fluidală cu microlitele de feldspat plagioclaz și biotit alungite după axa „c" în direcția curgerii.

c) Riodacitele cu structura microgranulară sînt roci leucocrate al căror aspect macroscopic nu prezintă caracteristici distincte față de celelalte două tipuri descrise, însă microscopic sînt net diferite. Într-o masă fundamentală formată din feldspat plagioclaz, cuarț, biotit apar fenocristale bine dezvoltate de plagioclaz, biotit și hornblendă verde. Ca minerale accesorii se întîlnesc cristale idiomorfe de apatit și zircon.

3. Andezitul cu plagioclazi (± hornblendă opacitizată) de Măgura

În literatura de specialitate despre această rocă s-a scris puțin. Astfel, la noi în țară, Rădulescu (1964) referitor la geologia părții centrale a munților Harghita, le denumește „andezite lipsite de minerale fero-magneziene" care „nu constituie un element de importanță în structura regiunii și nu par a fi un produs independent al aparatului vulcanic". Manilici (1962) denumește acest tip de roci „andezite feldspatice" și atrage atenția asupra lor, în regiunea Baia Sprie. Le consideră roci cu caractere mineralogice și chimice net diferite de alte tipuri de andezite și observă (fără a da însă explicații) „importante perturbări" în diagrama de variație a parametrilor Niggli, în care valorile acestor andezite au o poziție aparte față de restul rocilor analizate.

În munții Călimani de nord, andezitele cu plagioclazi (± hornblendă opacitizată) tip Măgura au fost întîlnite fie ca elemente în roci vulcanoclastice, pe valea Zăpodiei, (Nițoi, Constantinescu, 1984), fie ca un facies local al andezitelor cu hornblendă, hipersten și augit de pe valea Haita (Nițoi, 1987).

În zona cercetată andezitele tip Măgura apar sub formă de corpuri înrădăcinate și curgeri de lave. Morfologic, structural și textural se evidențiază în vîrful Păltiniș și Măgura centre de emisie a unor lave andezitice, destul de vîscoase ce s-au răspîndit pe o suprafață restrînsă, dar cu grosime mare (100—150 m). Grosimea aceasta a putut fi observată, mai ales, în partea estică a vîrfurilor amintite unde relieful prezintă abrupturi accentuate. În apropierea acestor centre, andezitele cu plagioclazi (± hornblendă opacitizată) tip Măgura acoperă parțial fundamentul cristalin, dacitul de Drăgoiasa, cît și formațiunea vulcano-sedimentară din peri-

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metrul Bilbor, inclusiv riodacitul de Văcăria. La limita dintre andezitele tip Măgura și riodacitul de Văcăria, pe toată suprafața cercetată a fost întîlnit un nivel de brecii de culoare roșie. În aceste brecii se pot observa, prin studiul microscopic, fragmente de riodacite prinse într-o masă fundamentală andezitică. Andezitele tip Măgura au fost întîlnite atît pe suprafețe continue cît și în puncte izolate, depărtate de centrele de emisie, constituind probabil, mici apofize ale aceleiași veniri.

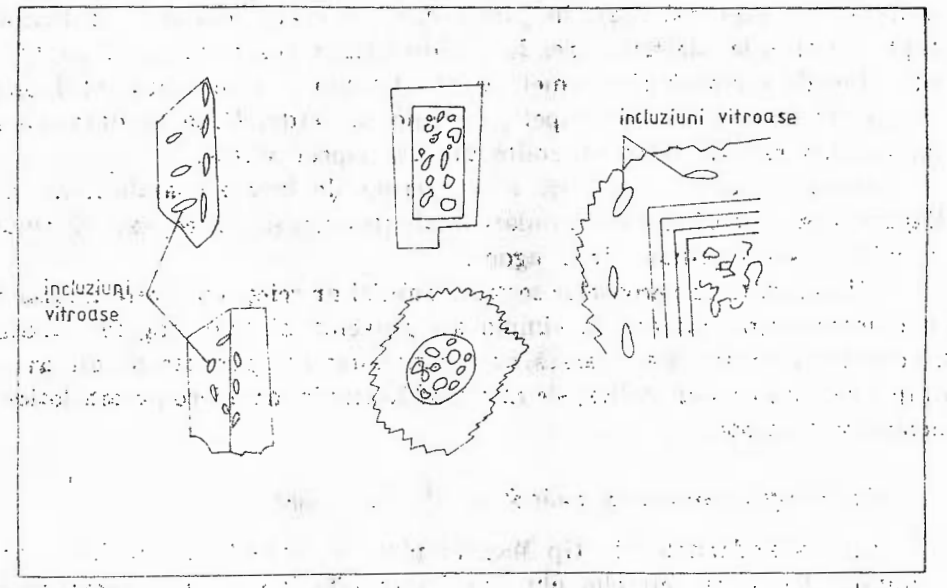

Fig. 2. — Aspecte ale dispunerii incluziunilor vitroase în fenocristalele de feldspat plagioclaz.
Aspects of the disposition of the glass inclusions in plagioclase feldspar phenocrysts.

Caracterul petrografic esențial al acestor andezite este acela că sînt foarte omogene din punct de vedere al texturii, structurii și mineralogiei (pl., fig. 1, 2). Macroscopic, roca este compactă, de culoare cenușie-negricioasă, cu ochiul liber neputînd fi observate mineralele componente. La curgerile de lavă structura este microlitică și textura fluidală, iar la corpurile înrădăcinate apare structura microgranulară și textura masivă.

Fenocristalele lipsesc în general, iar cînd apar, cu totul sporadic, sînt reprezentate prin plagioclaz și hornblendă opacitizată. Microlitele din masa fundamentală sînt de feldspat plagioclaz (99%) și rar de piroxeni (1%). Mineralele opace împînzesc roca mai ales sub forma unei pulberi fine de culoare neagră.

După procentajul fenocristalelor, andezitele cu plagioclazi (± hornblendă opacitizată) pot fi numite afirice (0—1% fenocristale) și subafirice (1—5% fenocristale). Fenocristalele de plagioclaz (1—1,5 mm) sînt proaspete, rar idiomorfe, maclate în principal din doi indivizi, cu incluziuni vitroase, izotrope, dispuse diferit în interiorul cristalului (fig. 2);

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microlitele de plagioclaz sînt clare, nealterate, cu numeroase și submilimetrice crăpături umplute cu o substanță amorfă cărămizie-negricioasă, cu macle tip Karlsbad sau chiar nemaclate (predominante). Fiind de talie foarte mică (sub 1 mm) studiul optic este îngreuiat. Totuși prin metoda clasică optică a unghiului de extincție s-a putut determina un conținut în An de 45% la microlite și 62% An la fenocristale. Biotitul apare accidental în paiete mici, submilimetrice diseminate în spații intergranulare în masa fundamentală sau în jurul unor minerale complet opacitizate (fapt ce împiedică determinarea lor). Hornblenda apare răzleț, cu conture rotunjite, de dimensiuni milimetrice (1—1,5 mm), opacitizată total sau cu mici nuclee centrale proaspete, de culoare verde. Masa fundamentală din spațiile rămase între microlite are un aspect vitros.

Pe suprafața superioară a curgerilor de lave ale andezitelor de Măgura apar uneori spații veziculare goale (texturi scoriacee) sau umplute cu mici cristale radiare de aragonit.

Din studiul numeroaselor secțiuni subțiri ar reieși că aceste andezite cu o mineralogie extrem de simplă au provenit dintr-o magmă situată în condiții apropiate de suprafață, cînd mineralele componente nu au avut timpul necesar de a cristaliza, ele rămînînd la stadiul de microlite scheletice, incomplet formate.

Compoziția chimică a andezitelor de tip Măgura

Asupra andezitelor de tip Măgura pînă în prezent nu s-a făcut un studiu detaliat al caracterelor chimice. În această lucrare sînt prezentate un număr de 8 analize chimice noi (tab. 1), care luate în comparație cu analize ale unor tipuri de andezite din compartimentul inferior și superior al munților Călimani (tab. 2)), demonstrează caracterul chimic bine definit al acestor andezite și poziția aparte pe care o au față de andezitele folosite ca termeni de comparație. În diagramele ce vor fi prezentate în continuare se poate observa plasarea punctelor reprezentative ale andezitelor de tip Măgura în cîmpuri bine marcate, individualizate, lucru ce demonstrează că andezitele cu plagioclazi (± hornblendă opacitizată) de tip Măgura nu numai mineralogic se individualizează în familia andezitelor din munții Călimani ci și din punct de vedere chimic.

Urmărind conținutul în SiO_2 al acestor roci se remarcă valori destul de ridicate cuprinse între 59,73% și 62,16%.

Din amplasarea probelor în diagramele de clasificare rezultă oscilația caracterelor chimice a andezitului de Măgura între o rocă acidă (diagrama R_1/R_2 — fig. 3 și diagrama Q′ — Anor, fig. 4) și o rocă intermediară (diagrama $SiO_2/Na_2O + K_2O$ din figura 6 și diagrama SiO_2/K_2O din figura 5).

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TABELUL 1

Analizele chimice ale andezitelor de Măgura

Nr. crt.	Nr. probei	Localizare	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	K_2O	Na_2O	P_2O_5	H_2O	CO_2	Total
1	4032	Dealul Măgura	59,73	0,64	18,77	5,92	0,83	0,17	0,78	4,88	1,78	3,01	0,18	2,93	—	99,78
2	1040	Pîriul Dobreanu	59,75	0,64	18,62	4,22	1,64	0,13	1,11	6,26	1,80	3,44	0,23	1,58	—	99,55
3	4030	Dealul Măgura	60,95	0,80	18,66	1,78	3,99	0,19	1,41	6,34	1,71	3,17	0,26	0,55	—	99,98
4	4030c	Dealul Măgura	61,03	0,88	18,31	3,39	4,40	0,16	1,31	5,93	1,82	3,17	0,25	0,87	—	99,76
5	1036	Piciorul Bilborului	61,40	0,62	17,81	4,00	1,98	0,20	1,04	5,22	1,81	3,32	0,25	2,01	—	99,77
6	1033	Piciorul Bilborului	61,60	0,56	18,21	2,85	2,84	0,19	1,19	5,75	1,75	3,48	0,24	0,86	—	99,63
7	1004	Dealul Măgura	61,65	0,84	18,14	3,46	2,06	0,18	2,28	5,84	1,82	2,92	0,33	0,43	—	100,65
8	4060	Dealul Păltiniș	62,10	0,53	17,97	3,18	2,45	0,16	1,20	5,56	1,88	0,16	0,22	1,09	—	99,68

Analizele au fost executate de chimist E. Călinescu (I.P.G.G.).

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TABELUL 2

Analize chimice ale termenilor de comparație

I. Andezite din compartimentul superior

Nr. crt.	Nr. probei	Localizare	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	H_2O^+	H_2O^-	Total
1	134	vîrful Pietrosu	54,36	0,99	19,93	3,02	6,42	0,13	4,55	8,52	2,83	1,83	0,13	0,43	0,39	100,58
2	313	creasta 12 Apostoli	56,66	0,69	18,02	4,85	2,90	0,13	4,30	6,34	3,10	1,56	0,11	0,82	0,37	99,91
3	131	vîrful Tamău	57,78	0,94	18,76	3,05	4,45	0,18	1,63	6,75	4,14	1,60	0,16	0,16	0,32	99,92
4	133	Pietrele Roșii	59,70	0,83	19,05	1,75	2,89	0,14	2,52	6,37	3,43	1,92	0,17	1,12	0,43	100,32
5	310	valea Tamău	59,85	0,46	18,30	4,16	1,79	0,07	3,50	5,46	3,12	1,87	0,07	—	0,80	99,92
II. Andezite din compartimentul inferior																
6	232	valea Haita	49,91	1,10	16,76	6,48	3,20	0,16	5,95	9,42	2,69	1,32	0,28	1,88	—	100,03
7	233	valea Haita	50,42	1,14	16,91	3,71	4,17	0,16	5,68	9,66	2,67	1,13	0,14	1,63	—	99,50
8	208	valea Haita	53,37	0,89	17,56	4,43	3,48	0,16	4,94	8,91	3,04	1,59	0,32	0,70	—	99,62
9	306	valea Tamău	54,75	0,59	17,62	3,62	4,01	0,11	5,80	7,56	2,85	1,56	0,10	0,36	0,52	99,99
10	236	valea Haita	55,96	0,88	17,56	2,36	3,64	0,13	4,57	8,28	3,16	1,57	0,16	1,21	—	99,55
11	141	valea Neagra	57,33	0,85	19,64	2,77	3,92	0,15	1,50	7,91	2,51	2,26	0,20	1,22	0,36	100,26
12	400	valea Tamău	59,73	0,66	17,41	1,29	3,21	0,16	1,96	5,98	3,36	1,88	0,19	1,17	1,95	99,40
13	206	valea Neagra	59,98	0,52	19,26	1,85	2,69	0,14	1,52	6,68	3,46	1,60	0,22	1,55	—	100,07
14	29	valea Neagra	60,33	0,58	20,47	1,37	2,83	0,04	1,87	6,65	3,58	1,54	0,18	0,68	0,44	100,57
III. Dacitul de Drăgoiasa																
15	3034	valea Stînceni	67,69	0,25	16,11	3,13	0,56	0,03	0,69	2,30	2,83	3,11	0,11	2,75	—	100,05
16	3020	valea Curmăturii	68,58	0,23	17,13	2,17	0,61	0,04	0,29	2,53	3,46	3,33	0,08	1,43	—	99,83
17	3046	valea Neagra	69,19	0,22	16,59	2,11	0,31	0,03	0,54	3,02	3,60	3,24	0,11	0,86	—	99,94
18	3016	valea Curmăturii	69,31	0,22	16,48	1,61	0,34	0,03	0,49	3,40	3,64	3,35	0,22	0,14	—	100,30
19	3005	valea Neagra	69,38	0,26	16,44	2,50	0,17	0,04	0,53	2,71	3,30	3,28	0,12	1,00	—	99,88
20	3033	valea Tomnatec	69,45	0,21	16,47	1,50	0,62	0,03	0,55	3,07	3,47	3,30	0,13	1,18	—	100,06
21	3006	valea Curmăturii	70,41	0,18	16,51	1,31	0,16	0,02	0,48	2,50	3,55	3,58	0,11	1,05	—	99,99
IV. Riodacitul de Văcăria																
22	1045	valea Ruși	67,00	0,24	17,31	2,44	0,10	0,03	0,48	1,90	3,12	2,28	0,12	4,55	—0,09	99,66
23	1057	Piciorul Bilborului	67,83	0,30	16,44	2,32	0,23	0,04	0,75	2,79	3,00	3,62	0,12	2,05	—0,11	99,60
24	1052	valea Ruși	68,47	0,24	16,14	2,17	0,14	0,03	0,51	2,30	3,19	3,34	0,11	2,68	—0,12	99,74
25	1053	valea Lespezi	69,00	0,22	15,40	1,82	0,10	0,02	0,33	2,45	3,25	3,55	0,14	2,01	—0,10	99,55
26	1055	valea Lespezi	69,53	0,20	16,24	1,95	0,12	0,02	0,25	2,49	3,25	3,60	0,11	1,83	—0,09	99,68
27	1024	valea Fundoaia	71,10	0,25	16,20	1,79	0,09	0,03	0,29	2,23	3,23	3,06	0,16	0,64	—0,05	99,00

Analiști: probele 131 și 134 după Nichita (194); celelalte probe au fost executate de E. Călinescu I.P.G.G. (din Nițoi, Constantinescu, 1984); Nițoi (1986); Nițoi (1987).

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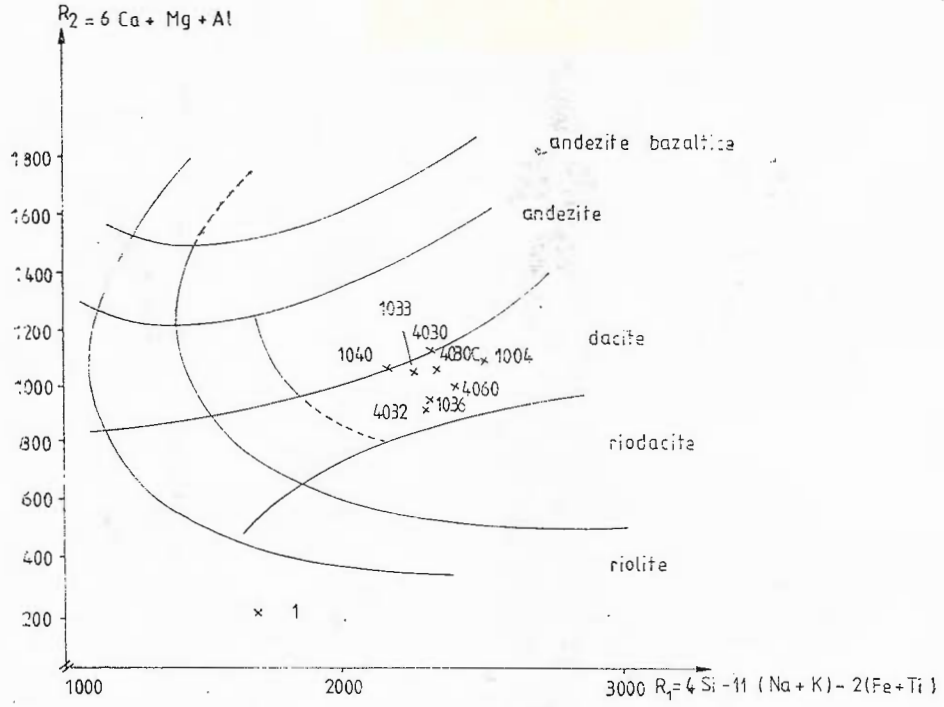

Fig. 3 — Diagrama R_1/R_2 ($R_1 = 4Si - 11\ (Na + K) - 2(Fe + Ti)$; $R_2 = 6Ca + Mg + Al$) (după H. de la Roche et al., 1980).

1, andezite cu plagioclazi (± hornblendă opacitizată) tip Măgura.

R_1/R_2 diagram ($R_1 = 4Si - 11\ (Na + K) - 2(Fe + Ti)$; $R_2 = 6Ca + Mg + Al$) (after H. de la Roche et al., 1980).

1, andesites + plagioclase (± opacitized hornblende) of Măgura type.

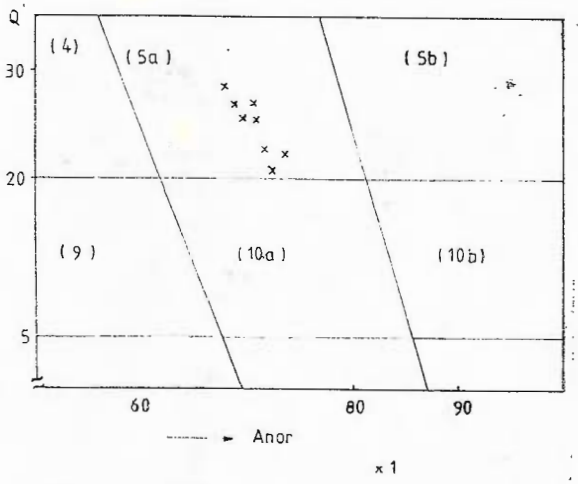

Fig. 4 — Diagrama Q'/Anor ($Q' = Q/Q + Or + Ab + An) \times 100$; $Anor = An/(An+Or) \times 100$ (Streckeisen, Le Maitre, 1979);

1, idem fig. 3; 5a, cîmpul plagiodacitelor.

Q'/Anor diagram ($Q' = Q/Q + Or + Ab + An) \times 100$; $Anor = An/(An+Or) \times 100$ (Streckeisen, Le Maître, 1979).

1, see Fig. 3; 5a, plagiodacite field.

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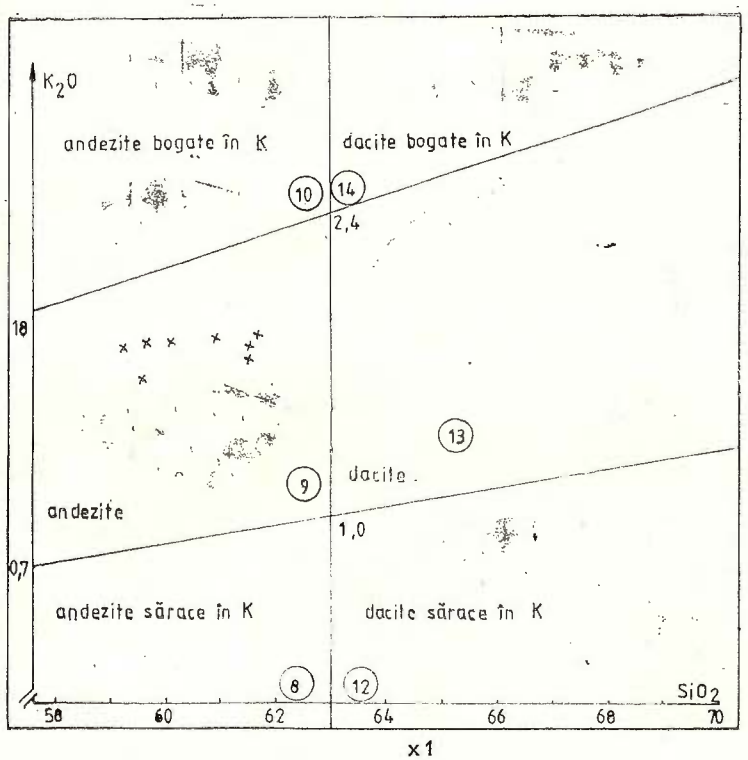

Fig. 5—Diagrama SiO_2/K_2O (Taylor, 1969 ; Pecerillo, Taylor, 1976).
1, idem fig. 3.
SiO_2/K_2O diagram (Taylor, 1969 ; Pecerillo, Taylor, 1976).
1, see Fig. 3.

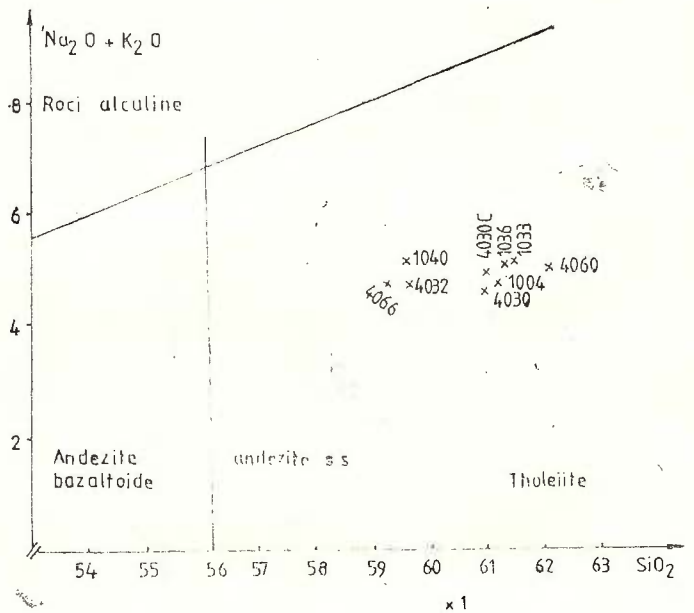

Fig. 6 — Diagrama $SiO_2/Na_2O + K_2O$ (Kuno, 1978).
1, idem fig. 3.
$SiO_2/Na_2O + K_2O$ diagram (Kuno, 1978).
1, see Fig. 3.

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În diagrama din figura 7, construită pe baza valorilor Niggli *al*, *fm*, *c*, *alk* raportate la *si* (tab. 3) și luate în comparație cu valori ale altor tipuri de andezite (andezite cu hornblendă și piroxeni, andezite cu piroxeni și hornblendă, andezite cu piroxeni etc.) aparținînd celor două compartimente structurale ale munților Călimani, și anume compartimentului supe-

TABELUL 3

Parametrii Niggli ai andezitelor de Măgura

Nr. crt.	Nr. probei	*si*	*al*	*fm*	*c*	*alk*	*k*	*mg*	*ti*	*p*	*qz*	*c/fm*	*Q*	*M*	*L*
1	4032	222,96	41,28	24,09	19,51	15,12	0,28	0,18	1,80	0,28	62,47	0,81	48,31	13,26	38,43
2	1040	210,60	38,67	21,91	23,63	15,79	0,26	0,27	1,70	0,39	46,43	1,07	45,78	9,76	44,45
3	4030	211,03	38,07	24,02	23,51	14,41	0,26	0,30	2,08	0,38	55,40	0,97	46,64	11,45	41,92
4	4030c	217,86	38,51	23,71	22,67	15,11	0,27	0,29	2,36	0,38	57,43	0,95	47,18	11,38	41,45
5	1036	229,67	39,25	23,49	20,91	16,35	0,26	0,25	1,74	0,40	64,27	0,89	48,17	11,27	40,55
6	1033	222,20	38,70	22,90	22,21	16,19	0,25	0,28	1,52	0,37	57,45	0,97	47,21	10,49	42,30
7	1004	214,23	37,14	27,26	21,73	13,86	0,29	0,45	—	0,49	59,78	0,79	47,62	13,06	39,33
8	4060	228,82	38,97	23,42	21,92	15,68	0,28	0,28	1,55	0,34	66,09	0,93	48,47	10,89	40,65

rior C_2 și compartimentului inferior C_1, se remarcă amplasarea andezitelor de tip Măgura către ultimii termeni de diferențiere progresivă a lavelor andezitelor calcoalcaline aparținînd compartimentului inferior C_1. Nu numai mineralogia simplă a andezitelor de Măgura le conferă un loc aparte ci și diagrama *mg/alk*, *k*, *ti*, *p* (fig. 8) în care se conturează clar cîmpue acestor roci, departajat de cîmpurile rocilor luate ca termeni de comparație. Se remarcă conținuturi mai scăzute în *mg* și ceva mai ridicate în *alk*, *k* și *p*

Reprezentativă este diagrama FAM din figura 9 în care au fost trasate, pe bază de analize chimice, cîmpurile rocilor, din perimetrul cercetat, care aparțin pe de o parte magmei andezitice (rocile andezitice aparținînd compartimentului C_1 și C_2 care au fost luate ca termeni de comparație) și pe de altă parte magmei dacitului de Drăgoiasa și a riodacitului de Văcăria (tab. 2). Se observă o ușoară abatere de la domeniul de amplasare a cîmpurilor amintite, andezitele de Măgura fiind situate deasupra limitei de domeniu, marcînd astfel o tendință de îmbogățire în fier. Această îmbogățire în fier face ca în diagrama $MgO/FeO + 0{,}9\ Fe_2O_3$ (fig. 10) grupul andezitelor de Măgura să aibă o poziție excentrică față de termenii de comparație. În această diagramă în care sînt trasate curbele de evoluție a magmelor calcoalcaline, alcaline și tholeiitice după Brown și Schairer (1971), andezitele de tip Măgura sînt plasate în capătul inferior al curbei A—A′.

Individualizarea andezitelor de tip Măgura, ca o entitate bine definită din punct de vedere chimic, ne-o demonstrează și diagrama

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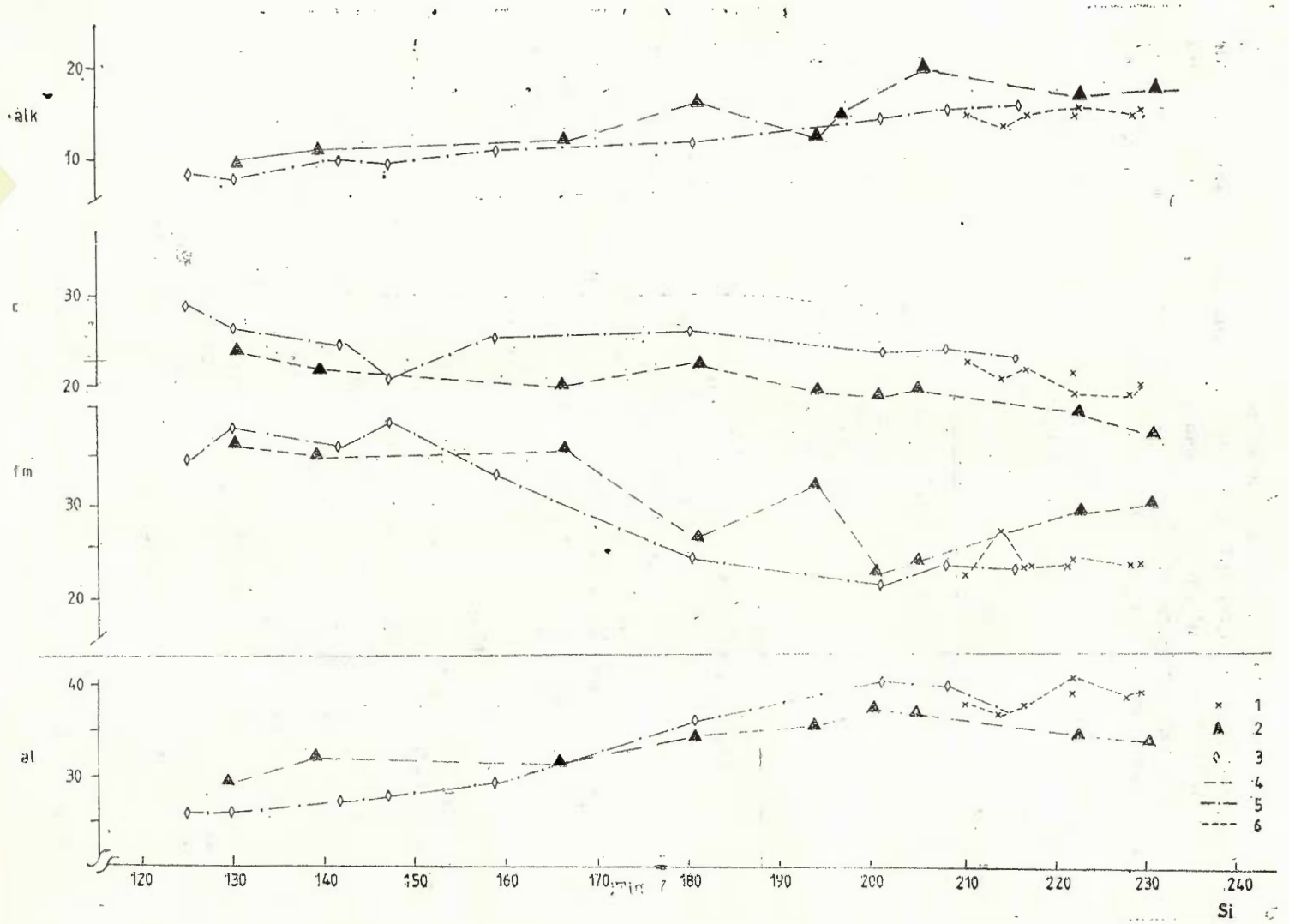

Fig. 7

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K—Na—Ca (fig. 11), unde acestea au o poziție clară, intermediară, pe linia ce marchează sensul de evoluție a suitei calcoalcaline (după Nocholds et al., 1954), între cîmpul rocilor provenite din magma andezitică și cîmpul rocilor acide aparținînd magmei dacitului de Drăgoiasa și al riodacitului de Văcăria.

4. Concluzii

În urma studiului mineralogic și chimic asupra andezitelor cu plagioclazi (± hornblendă opacitizată) de tip Măgura se poate afirma că acestea reprezintă nu numai un tip de rocă cu caractere proprii bine stabilite ci și o fază importantă din succesiunea erupțiilor din munții Călimani, situîndu-se către partea finală a acelora ce au loc în compartimentul inferior C_1. Mineralogia simplă a acestor roci poate fi o consecință a faptului că au erupt printre ultimele produse, cînd variațiile de presiune și temperatură au fost destul de neînsemnate. Caracterele oscilante între o rocă acidă (situarea probelor în cîmpurile rocilor acide, conținutul scăzut în anortit al microlitelor, prezența sticlei recristalizate) și una intermediară par a duce la concluzia unui amestec mecanic dintre o magmă dacitică (în cantitate mică) și o magmă andezitică reinjectată. Acest amestec fiind provocat de variații bruște de condiții termo-dinamice. Ar putea fi o explicație, de ce aceste andezite de tip Măgura apar bine dezvoltate mai ales pe aria de răspîndire a rocilor acide situate la limita dintre zona cristalino-mezozoică și eruptivul andezitic neogen.

Mulțumesc dr. S. Peltz pentru observațiile și sugestiile utile făcute asupra lucrării.

Fig. 7 — Diagrama de variație a parametrilor Niggli.

1, idem fig. 3; 2, andezite din compartimentul superior al munților Călimani; 3, andezite din compartimentul inferior al munților Călimani; 4, sensul de evoluție a andezitelor din compartimentul superior; 5, sensul de evoluție a andezitelor din compartimentul inferior; 6, sensul de evoluție a andezitelor cu plagioclaz (± hornblendă opacitizată) tip Măgura.

Variation diagram of the Niggli parameters.

1, see Fig. 3; 2, andesites from the upper compartment of the Călimani Mts; 3, andesites from the lower compartment of the Călimani Mts; 4, evolution trend of the andesites from the upper compartment; 5, evolution trend of the andesites from the lower compartment; 6, evolution trend of the andesites + plagioclase (± opacitized hornblende) of Măgura type.

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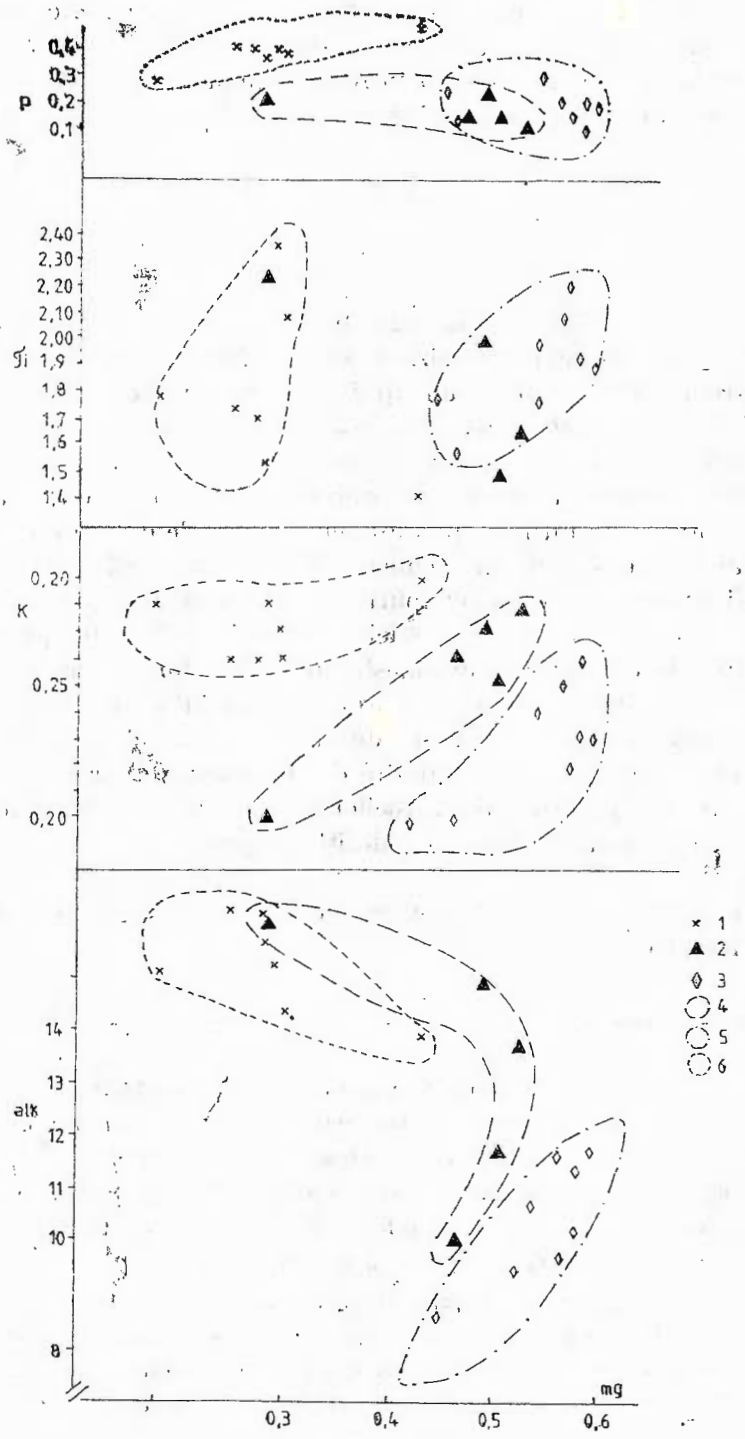

Fig. 8

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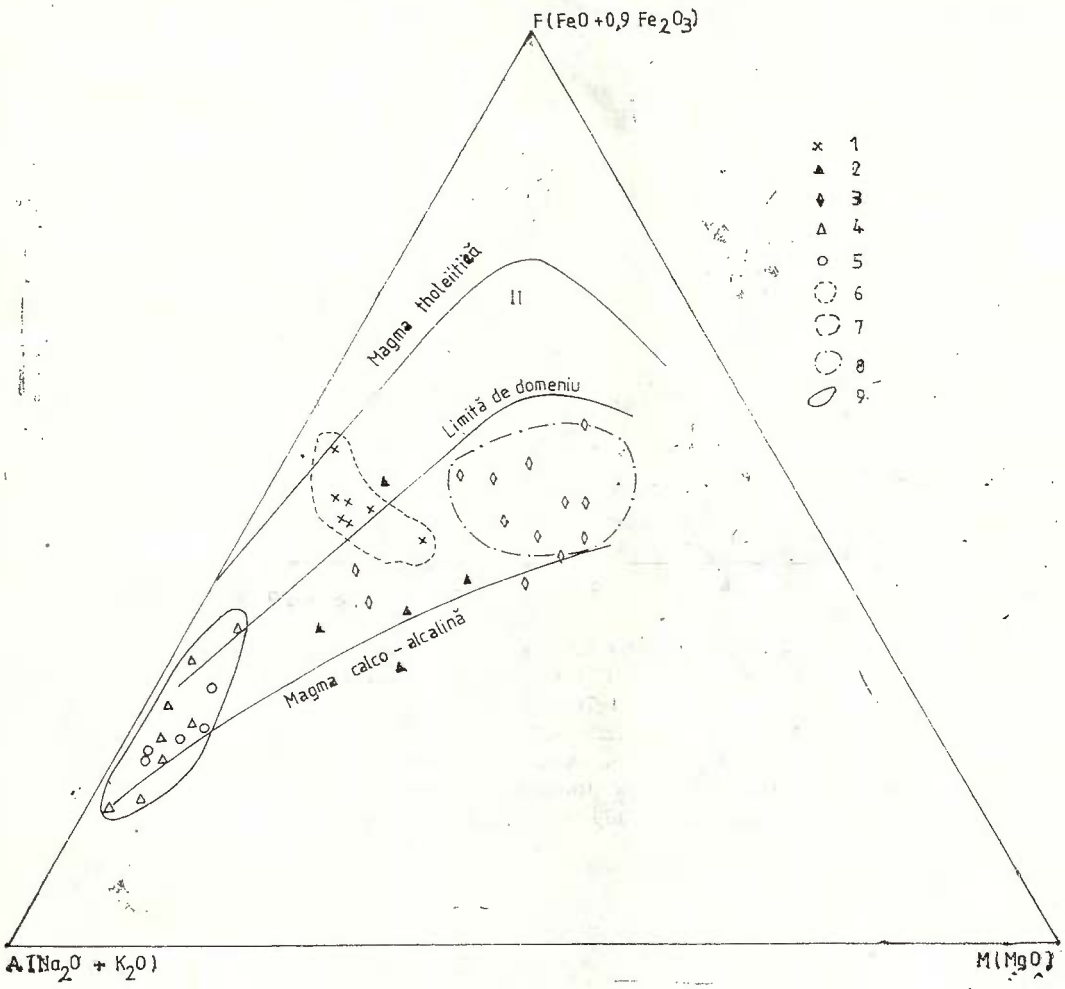

Fig. 9 — Diagrama FAM (F = FeO + 0,9 Fe_2O_3; A = Na_2O + K_2O; M = MgO).
1, 2, 3 idem fig. 7; 4, dacite de Drăgoiasa; 5, riodacite de Văcăria; 6, cîmpul andezitelor de Măgura; 7, cîmpul andezitelor din C_2; 8, cîmpul andezitelor din C_1; 9, cîmpul andezitelor de Drăgoiasa și al riodacitelor de Văcăria.

FAM diagram (F = FeO + 0.9 Fe_2O_3; A = Na_2O + K_2O; M = MgO).
1, 2, 3 see Fig. 7; 4, Drăgoiasa dacites; 5, Văcăria rhyodacites; 6, field of Măgura andesites; 7, field of C_2 andesites; 8, field of C_1 andesites; 9, field of Drăgoiasa andesites and Văcăria rhyodacites.

Fig. 8 — Diagrama parametrilor Niggli *mg* : *alk*, *k*, *ti*, *p*.
1, 2, 3, idem fig. 7; 4, cîmpul andezitelor din C_2; 5, cîmpul andezitelor din C_1; 6, cîmpul andezitelor tip Măgura.

Diagram of the Niggli parameters *mg* : *alk*, *k*, *ti*, *p*.
1, 2, 3 see Fig. 7; 4, andesites field from C_2; 5, andesites field from C_1; 6, field of Măgura-type andesites.

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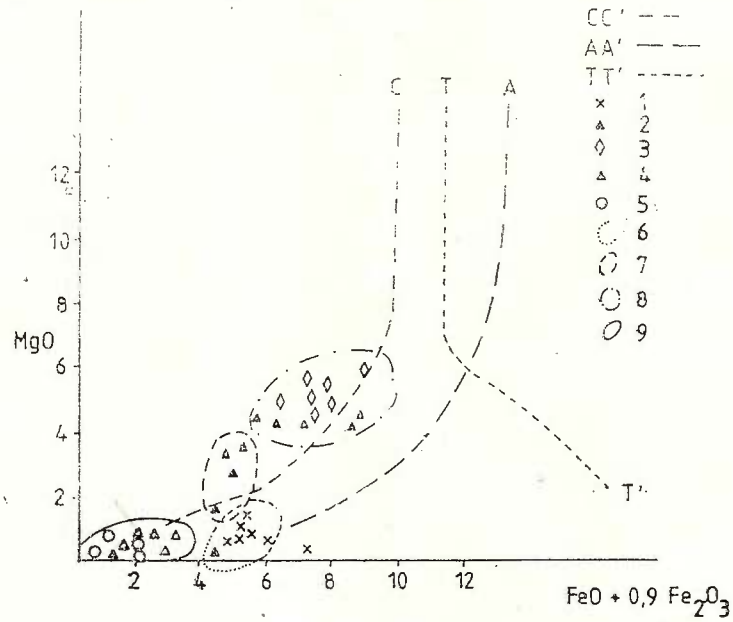

Fig. 10 — Diagrama MgO/FeO + 0,9 Fe_2O_3 (A—A′ serie alcalină Hawai); C—C′ serie calcoalcalină: T—T′ serie tholeiitică (Brown and Schairer, 1971).

1, 2, 3, 4, 5 idem fig. 9 : 6, 7, 8, 9 — idem fig. 9.

MgO/FeO + 0.9 Fe_2O_3 diagram (A—A′ Hawaii alkaline series; C—C′ calc-alkaline series; T—T′ tholeiitic series (Brown and Schairer, 1971).

1, 2, 3, 4, 5, see Fig. 9; 6, 7, 8, 9, see Fig. 9.

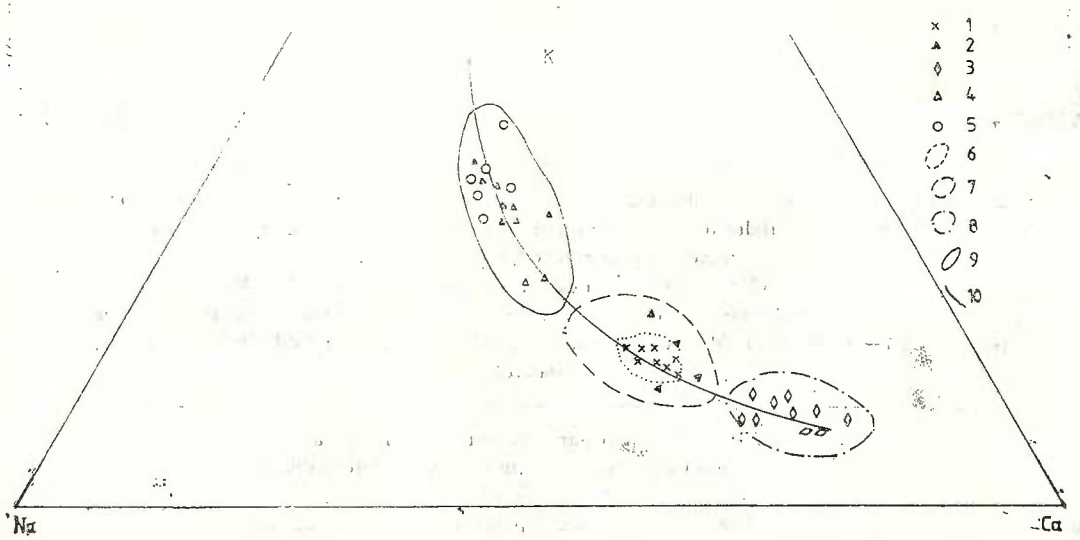

Fig. 11 — Diagrama K—Ca + Na. 1, 2, 3, 4, 5, idem fig. 8; 6, 7, 8, 9, idem fig. 9; 10, sensul de evoluție a suitei calcoalcaline (Nocholds et al., 1954).

K—Ca + Na diagram. 1, 2, 3, 4, 5, see Fig. 8; 6, 7, 8, 9, see Fig. 9; 10, evolution tren d of the calc-alkaline suite (Nocholds et al., 1954).

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MĂGURA PLAGIOCLASE ANDESITE (± OPACITIZED HORNBLENDE) (CĂLIMANI MOUNTAINS)

(Summary)

Măgura plagioclase andesite (± opacitized hornblende) is situated in the Călimani Mts, at the boundary between the Crystalline-Mesozoic Zone and the Neogene Eruptive.

In the study zone such andesites occur as emplaced bodies and lava flows. From the morphological, structural and textural point of view emission centres of some andesitic lavas, quite viscous, spread on a restricted area and very thick (100—150 m), can be observed in the Păltinis and Măgura summits. In the vicinity of these centres the Măgura andesite overlies partially and locally the crystalline basement, the Drăgoiasa dacite and the volcano-sedimentary formation. The main petrographic

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feature of these andesites is the high homogeneity of the texture, structure and mineralogy. Macroscopically, the rock is compact, of a grey-blackish colour; its constituting minerals cannot be observed with the naked eye. The structure is microlitic, microgranular and the texture is fluidal or massive. Phenocrysts are generally absent; when present, only locally, they are represented by plagioclase and opacitized hornblende; in the groundmass, microlites are represented 98 per cent by plagioclase and 2 per cent by rhombic pyroxenes and biotite. Opaque minerals occur especially as fine, black powder. According to the phenocrysts percentage, the Măgura andesites can be aphyric and subaphyric.

A study of several thin sections points out that these andesites with an extremely simple mineralogy came from a magma situated near the surface, when the constituting minerals had not had enough time to crystallize and remained as skeletal microlites, incompletely formed.

The Măgura andesite was minutely studied in chemical respect.

The mineralogical and chemical study emphasizes that the Măgura andesite represents not only a rock with well-established properties but also a significant phase in the eruption succession in the Călimani Mts, being situated towards the final part of those occurring in the lower compartment C_1. The simple mineralogy of these rocks is a consequence of their late eruption when the variations of pressure and temperature were quite unsignificant. The oscillating characters between an acid rock and an intermediary one seem to point to a mechanic mixture between a dacitic magma (the magma of the Drăgoiasa dacite) in a small amount and a reinjected andesitic magma. This mixture was determined by sudden variations of the thermo-dynamic conditions. It might explain why these andesites are well developed particularly in the spreading area of the acid rocks situated at the boundary between the Crystalline-Mesozoic Zone and the Neogene andesitic eruptive.

EXPLICAȚIA PLANȘEI

Fig. 1, 2 — Andezite cu plagioclazi (± hornblendă opacitizată) tip Măgura.
Andesites + plagioclases (± opacitized hornblende) of Măgura type.

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PETROLOGIA ROCILOR MAGMATICE

PRELIMINARY OBSERVATIONS ON THE CO_2—BEARING FLUID INCLUSIONS IN THE OLIVINE AND PYROXENE PERIDOTITIC NODULES AT HOGHIZ (PERŞANI MOUNTAINS)[1]

by

IOAN PINTEA [2], IOAN MÂRZA [3]

Olivine. Pyroxene. Peridotites. Nodules. Fluid inclusions. Lava. Basaltic composition. East. Carpathians — Neogene eruptive-Quaternary — Perşani.

Abstract

This paper deals with fluid inclusions (Co_2 + silicate melts) in the peridotitic nodules (olivine, pyroxene) at Hoghiz (Perşani Mountains). The depth (12—15 km) and pressure (3—4 kb) interval, within which occurred the decrepitation of the CO_2 primary inclusions and the equilibrium of the secondary inclusions by "necking-down", determined on the basis of specific halos, has been established.

Résumé

Observations prélimindires sur des inclusions à CO_2 situées dans les olivines et pyroxènes des nodules péridotitiques de Hoghiz (monts Perşani). On analyse les inclusions fluides (CO_2 ± coulée à silicates) des nodules péridotitiques (olivine, pyroxènes) de Hoghiz (monts Perşani) et on détermine l'intervalle de profondeur (12—15 km) et de pression (3—4 kb) où on a constaté, sur base de halos spécifiques, la décrépitation des inclusions primaires par CO_2, tout comme l'établissement de l'équilibre des inclusions secondaires par „necking-down".

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The upper mantle plays an overwhelming part in the magmatic petrometalogeny and therefore any revealing information on the physical and chemical conditions specific to this geoferric segment are of importance. The study of the fluid inclusions occurring in minerals of the rocks originating in the mentioned zone — with significant results in some countries — represents a novelty for our relevant literature.

The presence of fluid inclusions in the basalt phenocrysts is known since the second half of the last century (Zirkel, 1870); their identification in minerals of the peridotitic nodules is more recent (Roedder, 1965 a, 1983, 1984; Bilal, Touret, 1976, 1977; Belkin et al., 1985).

Fluid inclusions in minerals of the peridotitic nodules identified up till now are grouped according to their content into three categories: inclusions with CO_2, inclusions of silicate melts and inclusions with sulphide melts; from the genetical point of view they belong to the three known inclusion types: primary, pseudosecondary and secondary (Roedder, 1965 a). Their study in numerous occurrences with peridotitic nodules in the world, transported by alkaline basaltic lavas, made possible the understanding and clearing up of problems referring, on the one hand, to the evolution of the basaltic magmas in their origin place (mantle) and, on the other hand, to their physico-chemical state in their ascent way to the surface. The presence of supracritical liquid CO_2 in primary inclusions in the olivine and pyroxene peridotitic nodules attests its participation as free fluid phase in the active processes at the upper part of the mantle, whereas CO_2 in secondary inclusions, formed as a result of the thermometamorphism underwent by enclosed nodules, point to its part played together with the basaltic lavas during the ascent through the crust.

The presence of glass inclusions (silicate melts + sulphide melts) marks moments of the crystallization or recrystallization processes both at depth (when they are primary) and during the lava ascent towards the surface (when they are secondary).

Microthermometric determinations effectuated on CO_2 inclusions made possible the estimation of the corresponding densities and pressures, physical properties characteristic of the conditions under which they were trapped in the mineral structure.

The aim of this paper is to present preliminary data referring to the presence and significance of the fluid inclusions in the olivine and pyroxene of the peridotitic nodules at Hoghiz.

1. Petrological Considerations

The peridotitic nodules at Hoghiz (Braşov District) represent fragments of ultramafic cumulates (lherzolites, saxonites) originating in the upper mantle, brought to the surface by alkaline basaltic magmas. Although they have been known for a long time, there are only few studies on them up till now: petrographical-petrological (Măldărescu et al., 1983), geochemical (Mârza et al., 1986), volcanological (Mârza, 1986). Studies referring to the geology-petrology of the basalts at Racoş and Hoghiz were carried out by Laţiu (1928), Preda (1940), Rădulescu (1964), Peltz et al. (1972), Peltz and Bratosin (1986), Mihăilă et al. (1972), Tövissi (1972), etc.

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The different frequency from one place to another, the variation of the shape (oval shapes prevailing), size (up to Ø 30 cm) and colour (after the mineralogical composition and the alteration degree) can be observed in the field. Olivines (allochromatic) and pyroxenes are the main components; chrome spinels represent accessory minerals. Recrystallization took place at the contact with the basaltic lava under thermometamorphic conditions; it is omnipresent in case of pyroxenes in the contact zone and occurs rarely in case of olivines (Mârza et al., 1986).

2. Fluid Inclusions

2.1. **CO_2 inclusions.** The preliminary study carried out on the olivine and pyroxene crystals in the peridotitic nodules at Hoghiz emphasized the presence of primary and secondary inclusions with CO_2 content in variable amounts. The identification of CO_2 in inclusion was based on the homogenization way of the two-phase inclusions (liquid + gas) as well as on the expansive character of the fluid when released from inclusions by crushing of the minerals under the microscope. The examination of the microthermometric results was made by calculation in case of CO_2 inclusions (liquid + gas) of spherical or tubular shape (cylinders with spherical caps), according to Walther's method (1981). The information given in the papers of Roedder (1965 a, 1983, 1984), Bilal and Touret (1976, 1977), Blekin et al., (1985) was also used for comparison.

2.1.1. *Primary inclusions* are spherical-, ovoidal- or tubular-shaped and randomly spread within the crystals[4]. The reduced size represents (1—5 μm, rarely 10—20 μm) a general characteristic and it reduced much the microthermometric measurements; the only inclusions that could be investigated were the tubular ones (cylindrical with spherical caps), with sizes of 15—20 μm (Pl.I, Fig. 1). In this case the homogenization temperatures in liquid phase are of about 27.5 ± 1.0°C. The corresponding density obtained (provided that CO_2 in inclusions is pure or almost pure) both by calculation and on the temperature-density diagram in the two-phase field of the pure CO_2 system (Roedder, 1965 a) is of ca 0.62 g/cm^3 (Fig. 1). Referring to the occurrence way of the inclusions in the study minerals it is of a great importance the presence of an elliptic halo around them, whose development is in agreement with the size of the inclusion — the average size being of ca 100/60μm — and which resulted from the natural decrepitation, due to the difference between the internal pressure of the inclusions and the external pressure determined by the new conditions of the nodule minerals at different depths in their ascent way. After decrepitation, a part of the inclusion fillings (locally all of them) invades the formed microfissures, their reclosing (reequilibration of the "mother" inclusions) taking place under the new pressure conditions. Similar aspects are given by Roedder (1983), Belkin et al. (1985), in case of the CO_2 inclusions in the minerals of peridotitic nodules, as well as by Leroy (1979), Pécher and Boullier (1984) as regards the experimental decrepitation of the fluid inclusions from crystals of synthetic quartz. This phenomenon implies also the decrease of the CO_2 density, and, consequently, the values of the homogenization temperatures determined by us do not correspond

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to those of initial trapping of the primary inclusions, but to the new conditions created as a result of the pressure decrease[5].

Numerous spherical globules surrounded by halos, which according to the relevant literature (Roedder, 1965 a) can be of secondary origin, too, could be referred to the group of primary inclusions. For the time being we consider them very dense CO_2 drops, both primary and secondary (Pl. I, Fig. 2).

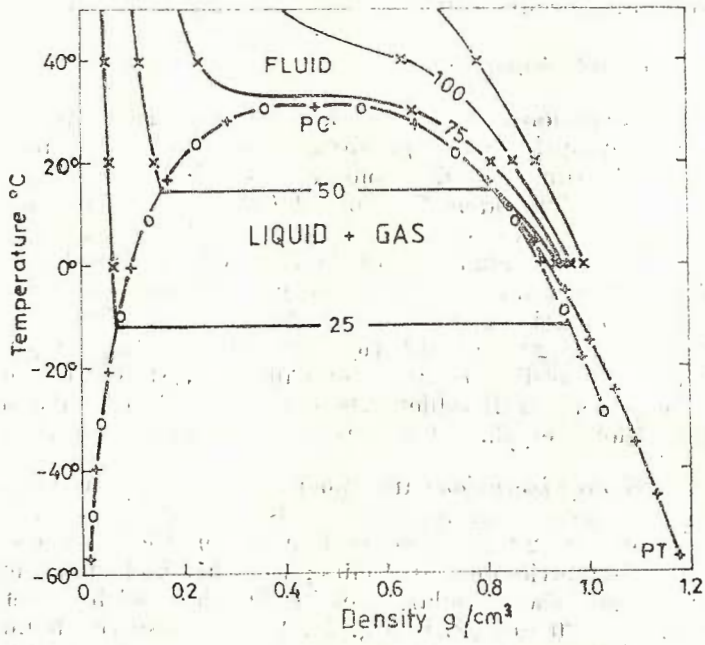

Fig. 1 — Temperature-density diagram for the two-phase field of the pure CO_2 system. Isobars (x) after Kennedy (1954); (+) after Plank and Kuprianoff (1929) and (0) after Hodgman (1954) (in Roedder, 1965 a).

2.1.2. *Secondary inclusions* are usually found as channels with a quite various morphology (irregular, vermicular, ramificated, Y-shaped, etc.) formed as a result of their penetration and closing up with fluids from the lava that drew the nodules towards the surface (Pl. II). They are frequently oriented in relation to crystalline elements (cleavage lines) and, in this case, by the separation determined by the "necking-down" phenomenon, occurred rows of secondary inclusions with shapes of (rhombic) negative crystals and with different compositions (Pl. I, Fig. 5). According to Boivin's opinion (1982) on the opening of the cleavage lines from the nodule minerals under the influence of the thermometamorphism, oneican suppose that subsequently they made possible the penetration of uids from the basaltic lava. Finally, they closed up the open planes and t led to the formation of automorphic inclusions. Locally, these planes are curved suggesting pressure efforts underwent by the minerals of the nodules subsequently to the closing up of these inclusions.

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The filling of the secondary inclusions consists of a silicate phase solidified as glass, accompanied by spherical or ovoidal bubbles of liquid CO_2 (in several cases gaseous CO_2 was observed beside liquid CO_2 and glass, Pl. I, Fig. 4).

In case of an advanced closing up of the fissures isolated inclusions, usually occurring as series marking the old fissures, were separated by "necking-down". According to composition several types of inclusions are observed:

— spherical globules of liquid CO_2 (Pl. II, Fig. 2 a);

— rounded three-phase inclusions with glass, liquid CO_2 and mobile bubbles of liquid and gaseous CO_2 (Pl. II, Fig. 1 a);

— two-phase inclusions with liquid CO_2 and gaseous CO_2 (Pl. II, Fig. 2 b);

— giass inclusions with immobile gas bubble (Pl. II, Fig. 2c).

Although the CO_2 inclusions thus formed do not preserve the values of the physical parameters (pressure, density) since the fluid penetration on the mineral fissures, they can provide information on the conditions of CO_2 at the time of their closing up by "necking-down". The disappearance of the gaseous phase in case of three-phase inclusions takes place at temperatures close to 25°C ± 1.0°C, the corresponding density being of ca 0.70 g/cm³.

Considering that on the last 25 km the temperature of the ascending basaltic lava did not undergone notable fluctuations (it varies about the value of 1200°C — Roedder, 1965 a) the data obtained in the two cases allow the estimation of a depth interval as well as the determination of the corresponding pressure at which the mentioned phenomena took place. Thus, plotting the obtained values on the depth-density diagram (proposed by Belkin et al., 1985) made up on the assumption of a pure CO_2, a trapping temperature of the inclusions of 1200°C, a depth interval ranging between 12 and 15 km is obtained, the values of the pressure on the temperature-density diagram for the pure CO_2 system being of 3—4 kb (Fig. 3).

As the phenomena presented are secondary, the formation of primary inclusions in minerals decrepitated during their transport towards the surface took place at much higher depths and pressures.

Besides the above-mentioned types of inclusions, there are also three-phase glass inclusions within which glass and gas are accompanied by an opaque solid phase probably represented by a sulphide (Pl. I, Fig. 3). The sulphide presence (in microcavities), formed after the closing up of the inclusions, could suggest the primary origin (Roedder, 1984) although more often than not they occur in those parts of the crystals where the closing up is obvious. It is also of note the presence of opaque solid inclusions which, according to the morphological features, could represent spinel microcrysts.

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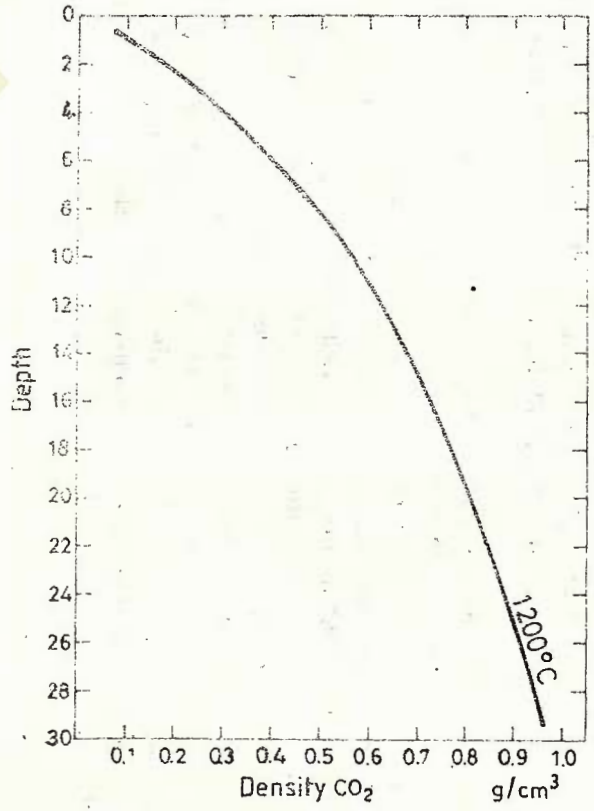

Fig. 2 — Plotting of CO_2 densities (determined on the basis of the homogenization temperatures of the two-phase inclusions with CO_2—liquid + gas) acc. to the formation depth of the inclusions at a temperature of 1200°C. It is supposed that: (1) inclusions contain pure CO_2 and (2) the hydrostatic pressure is given by the weight of the magmatic column with d = 2.7 g/cm^3 (Belkin et al. 1985, simplified).

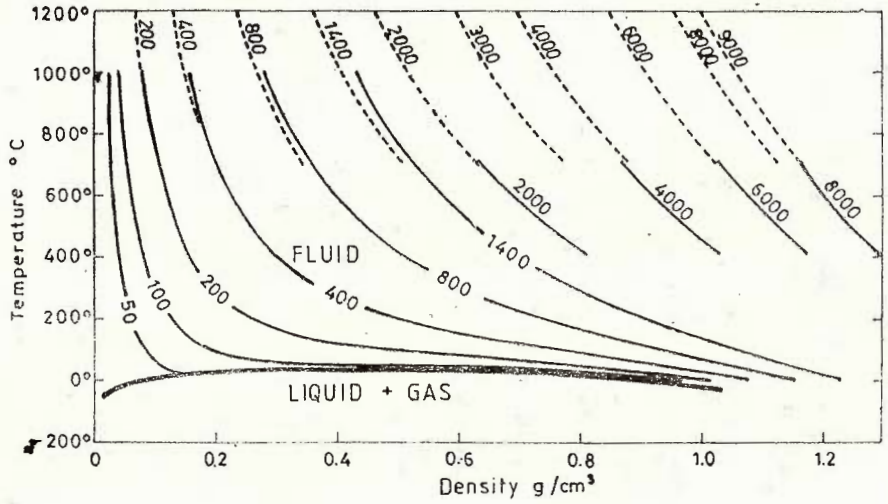

Fig. 3 — Temperature-density diagram for pure CO_2 system. Isobars represented by uninterrupted lines $\leqslant$ 1400 bars are after Kennedy (1954) and Roedder (1965); isobars represented, by continuous lines $\geqslant$ 2000 bars are after Shmonov and Shmulovich (1974); isobars marked with dashed lines are after Kerrik and Jakobs (1981) by the change of the equation Redlich-Kwong (Belkin et al., 1985).

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2.1.3. *CO_2 origin*. The studies carried out up till now on the CO_2 inclusions in minerals of peridotite nodules pointed out its origin mostly in the deep-seated zones of the earth's crust (upper mantle). The petrological and mineralogical observations on the Hoghiz (Mârza et al., 1986) and Racoşul de Jos (Măldărescu et al., 1983) nodules indicate their origin in the upper mantle. The primary character of the CO_2 inclusions, described in the study peridotitic nodules minerals, presumes the existence of this compound as free fluid phase in the peridotitic melt of the mantle.

Its presence also in the secondary inclusions of the same mineral inclusions formed as a result of the thermometamorphism confirms in our case the CO_2 formation by its separation from the basaltic lavas during its shifting to the surface and partly by the decrepitation of the primary inclusions. There is also the hypothesis of the CO_2 formation by the dissociation of carbonatic rocks enclaves, but no evident proofs are available.

3. Conclusions

The olivine and pyroxene crystals in the peridotitic nodules at Hoghiz contain CO_2 primary inclusions. It shows proof of the CO_2 presence as free fluid phase at the level of the upper mantle, as pointed out by several workers who dealt with peridotitic nodules in different occurrences (Roedder, 1965 a; 1983; 1984; Bilal, Touret, 1976, 1977; Belkin et al., 1985, etc.).

The change of the physico-chemical conditions due to the nodule basaltic lava ascent to the surface led to the formation of secondary inclusions with different shapes and compositions. The natural decrepitation of the CO_2 primary inclusions as a result of the difference between the internal pressure of the inclusions and the external pressure of the magmatic column determined the formation of a network of secondary inclusions (halos) with liquid CO_2 of a primary origin. During the enclosing and transport, in the same minerals from nodules the thermometamorphism gave rise to a multitude of channels or fissure networks with a random or oriented disposition, previously closed up by fluids from the upper mantle. The fissures filled with liquid CO_2 and a silicate melt provided several morphological and composition types of inclusions by "necking-down".

Microthermometric measurements effectuated on primary CO_2 inclusions as well as on those formed by "necking-down" from secondary inclusions indicate the depth interval within which their closing up took place (12—15 km). The corresponding hydrostatic pressure given by the magmatic column with a density of 2.73 g/cm³ and the temperature of 1200°C (Roedder, 1983) ranges between 3 and 4 kb for the described phenomena.

The presence of CO_2 in the study inclusions can be due to at least two major sources: the mantle, where it existed as free fluid phase, and the ascending basaltic magma. There is also the possibility that at least a part of CO_2 might be the result of the dissociation of carbonatic rocks enclaves.

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Beside CO_2 inclusions there are also glass inclusions in different percentages, consisting of a silicate melt and an immobile gas bubble locally accompanied by an opaque solid phase (probably sulphide). Moreover, opaque solid inclusions occur too. According to the morphological characters they can represent spinel microcrysts.

[4] Due to the great thickness of the polished sections (on both faces), about 0.3—0.5 mm thick, it is difficult to differentiate the pyroxene crystals from the olivine ones. The occurrence of cleavage lines, locally after two obvious directions, can represent a criterion for the recognition of pyroxene.

[5] The microthermometric measurements were carried out only on those inclusions with a visible closing up and, consequently, we consider their filling not contaminated.

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OBSERVAŢII PRELIMINARE ASUPRA INCLUZIUNILOR CU CO_2 DIN OLIVINELE ŞI PIROXENII NODULILOR PERIDOTITICI DE LA HOGHIZ (MUNŢII PERŞANI)

(Rezumat)

Lucrarea studiază incluziunile de CO_2 din olivinele şi piroxenii nodulilor peridotitici enclavaţi în bazaltele alcaline (cuaternare) de la Hoghiz (munţii Perşani). S-au distins incluziuni primare (mono- şi bifazice cu CO_2) şi incluziuni secundare, rezultate în procesul termometamorfic exercitat de lava bazaltică asupra mineralelor nodulilor.

În acord cu datele obţinute, se apreciază un interval de adîncime (12—15 km) şi presiune (3—4 kb) la care s-a produs decrepitarea (şi reechilibrarea) unui număr însemnat de incluziuni primare cu CO_2, dezvoltîndu-se în jurul lor halouri caracteristice. Paralel cu acest fenomen s-au format incluziuni secundare de CO_2 + topitură silicatică cu o geometrie variată (reţea de canalicule anastomozate, liniare simple şi ramificate, sub formă de Y, rombice izolate sau în şiraguri, sferice sau ovoidale), suportînd în evoluţia lor strangulări („necking-down") ce au condus la apariţia unor noi forme de incluziuni fluide cu compoziţie diferită.

Originea CO_2-ului din incluziuni este dublă : mantaua, la nivelul căreia exista ca fază fluidă liberă, şi topitura bazaltică în ascensiune.

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EXPLANATION OF PLATES

Plate I

Fig. 1 — Tubular primary inclusion with liquid CO_2 (L) and gaseous CO_2 (V); (H) halo formed as a result of the partial decrepitation of the inclusion during the basaltic magma ascent (explanation in the text).

Fig. 2 — Spherical globule of quite dense liquid CO_2, surrounded by the same characteristic halo (explanations in the text).

Fig. 3 — Three-phase primary (?) inclusion in olivine; (V) immobile gas bubble; (G) silicate melt solidified as glass; (K) opaque solid phase probably represented by sulphide.

Fig. 4 — Network of two-and three-phase inclusions (in the right top corner occurs a gaseous CO_2 bubble — V): (L) liquid CO_2 and (G) silicate melt solidified as glass.

Fig. 5 — One- and two-phase secondary inclusions with automorphous contour arranged as raws along the cleavage lines in pyroxene; (L) quite dense liquid CO_2; (G) silicate melt solidified as glass (explanations in the text).

Plate II

Separation by "necking-down" of several inclusion types with a different composition.

Fig. 1 a — Rounded three-phase inclusions of glass (G), liquid CO_2 (L) and highly mobile bubbles of gaseous CO_2 (V).

Fig. 2 a — Spherical bubble of liquid CO_2.

Fig. 2 b — Two-phase inclusions with liquid CO_2 and gaseous CO_2 (mobile bubbles).

Fig. 2 c — Glassy inclusions with an immobile gas bubble.

Figures scale : microns.

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D. S. Inst. Geol. Geofiz. vol. 74/1 (1987), 1989, pag. 117—122

PETROLOGIA ROCILOR MAGMATICE

ASPECTE FIZIOGRAFICE ALE INCLUZIUNILOR STICLOASE ÎN UNELE RIOLITE DIN DOBROGEA DE NORD [1]

de

VASILE POMÂRLEANU [2]

Rhyolites. Quartz. Feldspar. Fluid inclusions. Devitrification. Dobrogea — North Dobrogea — Măcin Zone.

Abstract

Physiographic Aspects of the Glass Inclusions in Some Rhyolites in North Dobrogea. Glass inclusions have been observed in quartz and feldspar phenocrysts in the rhyolites occurring on the Eschibalic-Consul and Sacar Hill-Camena alignments. They belong, in their turn, to several types : monophase, three-phase, polyphase and devitrified. The devitrification of the glass inclusions is due to hydrothermal solutions. The homogenization temperature of the nondevitrified glass inclusions, in comparison with data from the relevant literature, ranges from 860°C to 1200°C.

Résumé

Aspects physicographiques des inclusions vitreuses dans quelques rhyolites de Dobrogea de nord. Dans les phénocristaux de quartz et feldspath des rhyolites des alignaments Eschibalic-Consul et Sacar-Camena on a observé des inclusions vitreuses qui présentent plusieurs types : monophasiques, triphasiques, polyphasique et dévitrifiées. La dévitrification des inclusions vitreuses est déterminée par les solutions hydrothermales. La température d'homogénéisation des inclusions vitreuses non dévitrifiées par comparaison aux données de la littérature de spécialité est estimée entre 860° et 1200°C.

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1. Introducere

În Dobrogea de nord, pe lîngă formaţiuni magmatice intrusive se află şi roci efuzive de tipul riolitelor.

Cele mai multe aflorimente de riolite din Dobrogea de nord se dispun pe aliniamentele : Meidanchioi-Consul şi Sacar-Camena.

Riolitele de pe aceste aliniamente au fost studiate din punct de vedere mineralogic, petrografic şi chimic de către Cădere (1925), Savul (1935), Bălan (1967), Stiopol et al., (1975), Constantinescu et al. (1981, 1985), Ştefan (1982), Ştefan şi Roşu (în Nedelcu et al., 1986, 1987) şi alţii.

Din studiile menţionate rezultă că riolitele de la Camena, spre deosebire de cele de la Eschibalîc-Consul, se caracterizează prin tipuri structurale mai variate. Totuşi sub aspectul compoziţiei mineralogice şi chimice prezintă unele asemănări.

În nota de faţă, spre deosebire de lucrările anterioare, sînt prezentate principalele tipuri de incluziuni sticloase din riolitele de pe aliniamentele Meidanchioi-Consul şi Sacar-Camena. Întrucît incluziunile sticloase conservă picături de topituri — în prezent sticlă — din care s-au format riolitele, studiul lor prezintă interes pentru lămurirea unor aspecte petrogenetice, cristalogenetice şi termodinamice ale acestor roci.

2. Incluziuni sticloase

Primele menţiuni asupra incluziunilor sticloase în riolitele din Dobrogea de nord sînt consemnate de Cădere (1925).

Ca o trăsătură comună a incluziunilor sticloase este forma de cristal negativ şi rareori prezintă forme rotunjite. Dimensiunile variază de la unităţi pînă la 100 de microni. Rareori apar şi forme de cristal negativ deformat ca urmare a unor fenomene de coroziune internă. Ele apar izolate sau în grupe de două pînă la trei incluziuni. Au o distribuţie neuniformă. În cazul cînd incluziunile sticloase sînt dispuse în grupe, după această dispoziţie se pot stabili stadiile de creştere ale feţelor de cristal.

Fenocristalele de cuarţ ale riolitelor din cele două aliniamente prezintă diferite stadii de coroziune magmatică, de la una incipientă şi pînă la o coroziune totală ce şterge orice urmă cristalografică a cuarţului. Totuşi cu ajutorul incluziunilor sticloase sub formă de cristal negativ se pot reconstitui formele cristalografice ale fenocristalelor de cuarţ iniţiale, în scopul interpretării statistice a unor aspecte cristalogenetice.

După distribuţia lor în fenocristalele de cuarţ şi feldspaţi (sanidin) precum şi după conţinutul de faze incluse, incluziunile sticloase sînt monofazice, bifazice, trifazice şi polifazice.

2.1. *Incluziuni sticloase monofazice*

Cel mai simplu tip de incluziuni este reprezentat prin incluziuni ce conţin numai o singură fază solidă, sub formă de sticlă fără bulă de gaz. Astfel de incluziuni au fost întîlnite în unele fenocristale de cuarţ din riolitele microgranitice (Holdurmi) de la Camena (pl. I, fig. 1). Formarea incluziunilor sticloase monofazice mici fără faze de vapori se datoreşte fenomenului de metastabilitate a topiturii incluse. Frecvent bulele de vapori se formează la limitele de separaţie a topiturii cu peretele cristalului

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gazdă. În cazul cînd peretele nu acționează în calitate de germen, faza solidă, în general submicroscopică nu va funcționa ca suport pentru formarea bulelor de gaz.

În cazul riolitelor de la Camena s-au observat și incluziuni monofazice sticloase, care ca mărime sînt similare cu cele bifazice (pl. I, fig. 1). Lipsa bulelor de gaz, în aceste incluziuni, este trecută pe seama vîscozității mai mari a topiturii și a răcirii mai rapide a acesteia (Roedder, 1979). Incluziuni similare sînt menționate în tufuri și roci vulcanice de natură explozivă (Takenouchi, Imai, 1975; Roedder, 1979).

2.2. *Incluziuni sticloase bifazice*

Incluziunile sticloase bifazice, pe lîngă sticlă, conțin și o bulă de gaz. Acestea de obicei sînt mai mari și pot depăși 50—60 de microni. După captarea picăturii de topitură, în incluziune prin răcire treptată are loc formarea unei faze fluide imiscibile. Această fază odată formată ca urmare a contracției la răcire a topiturii incluse, poate ocupa un spațiu destul de neuniform de la 5% și pînă la 55% din volumul total al incluziunii sticloase (pl. I, fig. 3a; pl. II, fig. 1). Valorile diferite ale volumului bulelor de gaz se datoresc în parte concentrării diferite a gazelor în topitura inițială și într-o mică măsură fenomenului de cristalizare parțială a topiturii incluse în cursul răcirii pe pereții microcavității.

Strivirea sub microscop, în balsam de Canada a fenocristalelor de cuarț din riolitele gri de la Eschibalîc (aliniamentul Meidanchioi-Consul) au pus în evidență eliberarea gazului inclus în incluziunile sticloase. S-a constatat că volumul gazului eliberat era cu mult superior celui din bula inițială. Acest fapt arată că gazul respectiv din incluziune se află sub o presiune mai mare decît cea normală (Consul, Holdurmi, Eschibalîc). Unele analize efectuate de Clocchiatti (1970) asupra unor incluziuni sticloase similare din unele porfire din Italia au arătat că faza gazoasă din incluziuni este constituită din H_2O, CO_2 și N_2.

2.3. *Incluziuni tri- și polifazice*

Incluziunile tri- și polifazice sînt constituite din sticlă, bulă de gaz unul sau mai multe cristale (pl. I, fig. 2, 3; pl. II, fig. 3). Fazele cristaline pot fi constituite din minerale preexistente, incluse în timpul formării cristalelor și sub formă de depuneri de cristale de obicei aciculare, transparente în interiorul microcavităților magmatice (Eschibalîc). Acestea din urmă sînt situate la limita dintre bula de gaz și sticlă sau limita sticlă-cristal și s-au format odată cu fenocristalele din magmă.

2.4. *Incluziuni devitrificate*

Incluziunile devitrificate au dimensiuni mai mari de 60 microni. Ele pot fi parțial sau total devitrificate (pl. II, fig. 2, 3). Incluziunile parțial devitrificate, prin observații în lumină polarizantă, s-a constatat că sînt constituite din agregate de cristale, fapt ce ar indica o recristalizare a sticlei fără alterații și fără aport de substanță din exterior (pl. II, fig. 2). Incluziunile total devitrificate s-au format prin acțiunea soluțiilor hidrotermale.

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Sub aspect genetic se cunosc două tipuri de devitrificări (Deicha, 1955 citat de Clocchiatti, 1970) și anume: autodevitrificarea propriu-zisă și devitrificarea hidrotermală. În cazul autodevitrificării, fără pierderi și fără aporturi de substanțe, orice incluziune poate fi considerată ca un sistem izolat ce evoluează în funcție de timp și temperatură. Devitrificarea hidrotermală, dimpotrivă, corespunde unor fisurări ale cuarțului și uneori cu formarea pe fisuri a incluziunilor fluide secundare de salinitate diferită (pl. I, fig. 3; pl. II, fig. 3a).

Tipuri de incluziuni sticloase descrise s-au remarcat numai în unele fenocristale. Lipsa lor în cele mai multe cristale s-ar putea explica prin faptul că acestea — în conformitate cu ipoteza emisă de Ștefan (1982) — ar reprezenta xenolite rezultate prin topirea incompletă a unor granitoide preexistente. De altfel lipsa incluziunilor sticloase în granite a fost semnalată și explicată încă de Tuttle (1952).

În riolitele din ambele aliniamente cu excepția unor particularități locale (prezența uneori a incluziunilor sticloase monofazice la riolitele de pe versantul vestic al dealului Holdurmi (Văcăria) și a incluziunilor trifazice de tipul sticlă + gaz + cristale aciculare transparente la Eschibalîc), s-au remarcat incluziuni sticloase și devitrificate asemănătoare.

Relații de consangvinitate ale riolitelor din ambele aliniamente s-au remarcat și pe baza interpretărilor geochimice ale elementelor majore și pămînturilor rare din aceste roci, fapt care a determinat pe Ștefan și Roșu (Nedelcu et al., 1987) să ajungă la concluzia că un fundament comun ar fi participat la generarea magmelor din care s-au format riolitele.

Tipuri de incluziuni sticloase identice au fost descrise în unele riolite din Japonia indicînd temperaturi de omogenizare între 850°C și mai mari de 1180°C (Takenouchi, Imai, 1975), în Islanda cu un interval de temperatură de 900°C—1250°C (Naumov, Poliakov, 1974), în Transbaicalia de est cu temperaturi cuprinse între 860°C—1260°C (Naumov et al., 1976).

Luînd în considerare cele menționate și tipurile identice de incluziuni sticloase în riolitele de la Camena cu cele din Japonia, Islanda și URSS, se apreciază că temperaturile de omogenizare ale incluziunilor sticloase în cuarț și feldspat din riolitele de Camena ar putea fi cuprinse între 860°C și 1200°C.

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PHYSIOGRAPHIC ASPECTS OF THE GLASS INCLUSIONS IN SOME RHYOLITES IN NORTH DOBROGEA

(Summary)

This paper presents general petrographic aspects and types of glass inclusions in the rhyolites occurring in the Meidanchioi-Consul and Camena (Sacar Hill-Movila Goală) alignments.

The Camena rhyolites, unlike the Meidanchioi-Consul ones, are characterized by complex petrographic aspects.

In both alignments the phenocrysts of quartz and locally of sanidine from the Camena rhyolites contain glass inclusions of several types: monophase, three-phase, polyphase and devitrified. The devitrification of the glass inclusions is due to the circulation of postmagmatic hydrothermal solutions on fissures in phenocrysts. For the study rhyolites, as compared with the data from the relevant literature, a temperature interval of 850°C — 1200°C is estimated.

EXPLICAȚIA PLANȘELOR

Planșa I

Incluziuni sticloase în fenocristalele de cuarț în riolitele de la Camena.
Glass inclusions in quartz phenocrysts in the Camena rhyolites.

Fig. 1 — Incluziune sticloasă monofazică în riolitele microgranitice de la Holdurmi.
Monophase glass inclusion in the Holdurmi microgranitic rhyolites.

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Fig. 2 — Incluziune sticloasă trifazică (sticlă + gaz + minerál opac) în riolitele de la Movila Goală.
Three-phase glass inclusion (glass + gas + opaque mineral) in the Movila Goală rhyolites.

Fig. 3 — Incluziuni sticloase de la Holdurmi : a, incluziune bifazică (sticlă + gaz) în care faza de gaz domină, ocupînd mai mult de jumătate din volumul incluziunii ; b, incluziune trifazică (sticlă + fază de vapori cu meniscul foarte atenuat + mineral opac) ; c, incluziune parţial devitrificată de către soluţii hidrotermale, care prin cicatrizarea fisurilor au înlesnit formarea incluziunilor secundare.
Holdurmi glass inclusions : a, biphase inclusion (glass + gas), the gas phase prevailing (more than half of the inclusion volume) ; b, three-phase inclusion (glass + vapour phase with very attenuated meniscus + opaque mineral) ; c, partly devitrified inclusions by hydrothermal solutions that, by the healing of the fissures, facilitated the formation of secondary inclusions.

Planşa II

Incluziuni sticloase în fenocristalele de cuarţ din riolitele de la Consul.
Glass inclusions in quartz phenocrysts in the Consul rhyolites.

Fig. 1 — Grupe de incluziuni sticloase bifazice (gaz + sticlă) distribuite în plane diferite ale cristalelor de cuarţ. Sub incluziunile sticloase apare un plan orientat NE—SW de incluziuni fluide secundare.
Biphase glass inclusion groups (gas + glass) spread in different planes of the quartz crystals. Under the glass inclusions occurs a NE—SW trending plane of secondary inclusions.

Fig. 2 — Incluziune sticloasă bifazică (sticlă + gaz) sub formă de cristal negativ parţial devitrificată.
Partly devitrified biphase glass inclusions (glass + gas) as negative crystals.

Fig. 3 — Incluziune sticloasă complet devitrificată (incluziune mare închisă la culoare) alături de o incluziune mică sticloasă (trifazică). Aceste incluziuni sînt separate de plane de incluziuni fluide secundare dispuse aproximativ NV—SE. Incluziunea mare complet devitrificată este intersectată marginal de un plan cu incluziuni fluide secundare orientat N—S.
Wholly devitrified glass inclusion (dark-coloured large inclusion) beside a smaller glass inclusion (three-phase). These inclusions are separated by secondary fluid inclusion planes approximately trending NW—SE. The large, wholly devitrified glass inclusion is marginally intersected by a N—S trending plane with secondary fluid inclusions.

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SIGNIFICATION PÉTROGÉNÉTIQUE DU ZIRCON DU MASSIF GRANITOÏDE DE PONEASCA [1]

par

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Granitoid. Zircon. Crystals. Habit. Magmas. Alkalic composition. Magmatic differentiation. South Carpathians — Crystalline Getic and Supragetic realms — Semenic Mountains.

Résumé

Les caractères morphologiques et typologiques des cristaux de zircon indiquent qu'ils proviennent d'un magma qui a évolué en deux directions distinctes : la première alcaline et la seconde plus pauvre en alcalis. Selon le diagramme T.E.T. de distribution des roches granitiques le magma est originaire de la croûte pour la plus grande partie du massif, ou mixte pour une petite partie. L'origine mixte est soutenue par le zircon à double histoire (cristaux à noyau). La température de cristallisation du zircon, selon l'étude statistique des caractères morphologiques des cristaux, varie entre 850°et 600°C.

Abstract

Zircon Petrographic Significance in the Poneasca Granitoid Massif. The morphological and typological features of the zircon crystals point out their origin in a magma which evolved after two different directions : an alkaline and a low-alkaline one. The origin of the magma, deduced on the basis of the TET diagram of the granitic rock distribution, is crustal for most of the massif or mixed for a small part. The mixed origin is showed by the zircons with a double history (crystals with core). The zircons crystallization temperature, inferred on the basis of the statistic study of the zircon morphological features, ranges between 850°and 600°C.

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Le zircon, comme minéral accessoire dans le massif granitoïde de Poneasca, a été observé dans les lames minces par tous ceux qui ont effectué des études pétrogénétiques dans la région. Comme minéral à profondes significations pétrogénétiques il a été présenté par Bîrlea (1976) pour le massif de Sichevița et Robu, Robu (1986) pour quelques massifs granitoïdes de Parîng.

Les implications pétrogénétiques de ce minéral ont été présentées par Poldervaart (1950, in Pupin, 1980), sur base des variations des dimensions des cristaux.

Pupin, Turco (1972), utilisant une méthode nouvelle statistique, fondée sur la variation morphologique des cristaux, fonction des conditions physico-chimiques du milieu de cristallisation, indiquent que le zircon est un minéral qui peut fournir des données concernant le habitus de cristallisation.

Le rôle pétrogénétique du zircon du granitoïde de Poneasca est présenté dans cette note tenant compe de cette méthode statistique même.

Cadre géologique. La zone où se situe le massif granitoïde de Poneasca est inclue dans l'unité tectonique de la nappe gétique ; le périmètre comporte des schistes cristallins, roches magmatiques et dépôts sédimentaires.

Les formations cristallines appartiennent aux séries de Sebeș-Lotru et Miniș-Buceava (Savu, 1979).

Les formations de la série de Sebeș-Lotru constituent la masse principale des métamorphites, qui renferme le pluton granitoïde de Poneasca. Elles appartiennent au complexe des micaschistes (C_4) et sont représentées par des micaschistes, paragneiss, migmatites, amphibolites et quartzites. Ces roches, rencontrées aussi sous forme de xénolites dans la masse des granitoïdes, ont été métamorphisées dans les conditions du faciès des amphibolites à almandine pendant le Précambrien supérieur (Savu, 1979).

Les formations de la série de Miniș-Buceava affleurent uniquement dans le versant gauche de la vallée de Poneasca, entre les confluents avec le ruisseau de Cușec et le ruisseau d'Alibeg ; elles comportent des roches plus faiblement métamorphisées du type des schistes quartzitiques à biotite $\pm$ muscovite, des schistes amphibolitiques, des quartzites à biotite et des schistes carbonatés (Stan et al., 1987).

Les roches magmatiques sont représentées par le granitoïde de Poneasca et sporadiquement par des roches banatitiques disposées sous forme de filons ou apophyses dans le corps du granitoïde ou dans les roches environnantes.

Depuis 1940 le massif granitoïde a formé l'objet de plusieurs études ; dans cette année-là Codarcea a indiqué que les roches magmatiques de la zone constituent un seul corps formé de diorites, granodiorites et aplites granitiques, semblables à celles du massif de Sichevița.

Constantinof (1959) présente ce massif dans la feuille de Reșița-Oravița à 1 : 100.000, scindé en deux parties (Poneasca au sud et Gruiu au nord) qui renferment des granodiorites à biotite, granodiorites à grenats

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et filons d'aplites et pegmatites. Hurduzeu (1962) aussi présente le granitoïde dans la même lumière.

Dès 1961, le massif granitoïde de Poneasca est considéré comme un tout unitaire et est investigué systématiquement par Savu (1963 — 1969) qui nous a fourni de nouvelles données concernant la minéralogie, la pétrochimie et les processus génétiques qui ont contribué à la formation de ce massif. Savu (1969) arrive à la conclusion finale que le magma lithogène qui a engendré les granitoïdes a été généré par des processus de palingenèse et anataxie, controlés par la subduction de l'éperon valaque, au-dessous de la zone mobile des Carpathes Méridionales.

Les déterminations d'âge K/Ar sur les biotites et feldspath potassique (Kräutner et al., 1978) ont donné 300 ± 15 m.a., fait qui indique que la mise en place de l'alignement de granitoïdes de Poneasca-Sichevița a eu lieu pendant l'orogenèse varisque.

Les dernières années on a initié plusieurs études géologiques et géophysiques pour établir les perspectives économiques (Întorsureanu et al., 1981, 1983, in Stan et al., 1987).

Du point de vue pétrographique, le corps granitoïde de Poneasca comporte principalement des granodiorites à biotite et granites à deux micas. On a identifié aussi des diorites quartzifères et tonalites.

Les granodiorites à biotite se développent surtout dans la partie centrale du sud du pluton, dans le bassin de la vallée de Poneasca ; au nord, dans le bassin de la vallée de Bîrzava, celles-ci affleurent sous forme de petits corps allongés et ont un développement plus réduit.

Les granites à deux micas se situent dans la partie sud du granitoïde et occupe presque toute la zone de nord du pluton. Selon la présence en quantité plus grande ou plus réduite d'un mica ou de l'autre, on peut parler des granites à biotite ± muscovite ou granites à muscovite ± biotite (Stan et al., 1987).

Les dépôts sédimentaires mésozoïques appartiennent à la zone de Reșița-Moldova Nouă. Ils affleurent dans la partie sud-ouest du massif granitoïde de Poneasca et sont représentés par des calcaires récifaux de Plopu (Barrémien-Aptien inférieur), calcaires de Minis (Aptien supérieur) et grès de Golumb (Albien).

Le long des principales vallées il y a des dépôts quaternaires, représentés par des alluvions et cônes de déjection.

Méthodologie de préparation des échantillons. Pour l'élaboration d'une étude sur les cristaux de zircon du massif granitoïde de Poneasca, on a prélevé et préparé 7 échantillons, chacun à un poid de 5 kg environ. La localisation de ceux-ci est présentée dans la figure 1.

Les échantillons ont été préparés dans le laboratoire et ont subi les suivantes opérations :

— le broyage et le tamisage à des dimensions inférieures à 0,630 mm ;

— l'obtention du concentré de minéraux lourds à la table de tamisage ;

— l'obtention de la fraction plus lourd 2,9 g/cm³ par le traitement du concentré avec bromoforme ;

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— l'obtention du concentré de zircon au séparateur Frantz (à une intensité de 2A) ;

— le choix et la séparation des cristaux de zircon sous la loupe binoculaire.

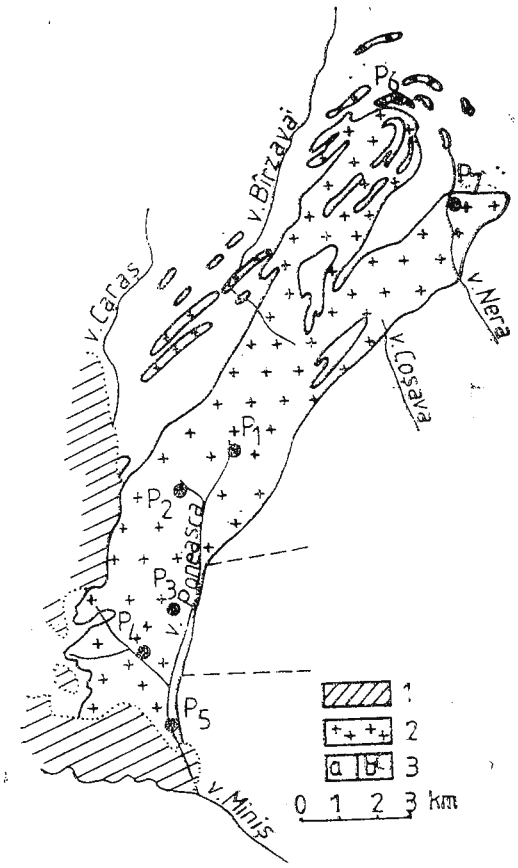

Fig. 1 — Schéma géologique du massif granitoïde de Poneasca (d'après Savu, 1969). Crétacé : 1, calcaires, grès ; 2, granodiorites, granites, diorites, tonalites ; 3, Série de Sebeș-Lotru (a) ; Série de Miniș-Buceava (b).

Etude optique des cristaux de zircon. Pour l'étude des cristaux de zircon on a choisi 100 cristaux entiers pour chaque échantillon, à des couleurs différentes et des faces bien déterminées et variées comme type de raccord, de manière que la multitude des cristaux séparés soit représentative pour tout l'échantillon.

Cristallographiquement, les cristaux de zircon présentent des faces de prisme et de pyramide par le raccord desquelles résultent des cristaux à habitus prismatique (avec la prisme plus ou moins développée). Les prismes peuvent être tétragonaux ou ditétragonaux à faces de type (110) et/ou (100) dont le développement a été déterminé par les variations de température du magma pendant la cristallisation du zircon (Pupin, Turco, 1972). Les cristaux long-prismatiques sont caractéristiques, ceux court-prismatiques étant moins fréquents.

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Les prismes de zircon terminent, aux deux extrémités, en pyramides ditétragonaux ou tétragonaux (plus rare). Parmi les formes les plus fréquentes de faces de pyramide sont (211) et/ou (101). La variation du nombre et du type de faces est en rapport avec les variations du chimisme du magma au moment de la cristallisation de ce minéral (Pupin, Turco, 1972).

Etant données les dimensions très réduites des cristaux observés, on a utilisé pour l'indexation des faces la méthode de l'abaque (Caruba, Turco, 1971).

La couleur des cristaux de zircon varie de l'incoleur (rares) vers des différentes teintes de rose (abondants) et brun (très rares).

Selon certains auteurs (Kuharenco, 1961, in Bîrlea, 1976), le zircon présente plusieurs teintes de rose, jusqu'à brun roujeâtre, sous l'influence des émanations des éléments radioactifs y renfermés; l'intensité de la couleur est en rapport directe avec l'âge du minéral et/ou la teneur en éléments radioactifs.

L'éclat des cristaux de zircon est adamantin à ceux incolores ou roses et gras à ceux bruns.

La plupart des cristaux de zircon présentent des inclusions en nombre variable. Elles sont représentées pas des cristaux à des dimensions très réduites, transparents et opaques, disposés à l'intérieur des cristaux de zircon dans la zone de développement des faces de prisme.

Les inclusions des minéraux transparents sont allongées, ayant la forme de prisme très petite, arrondie aux extrémités, à relief élevé, extinction droite et biréfringence réduite. Elles sont disposées soit parallèlement à l'axe verticale „c", soit obliquement par rapport aux faces prismales, tant dans la partie centrale qu'aux extrémités des cristaux aussi. Selon les propriétés optiques, ce minéral est l'apatite. Mais les prismes de zircon renferment aussi des cristaux plus réduits de zircon, à disposition spatiale, semblable à ceux d'apatite.

Les inclusions opaques sont ovales ou arrondies, plus rares que celles transparentes, disposées dans la partie centrale des cristaux de zircon. Elles ont des dimensions inférieures à celles des inclusions transparentes, fait qui rende difficile la détermination optique avec précision. Le magnétite ne fait pas partie de cette catégorie étant donné le manque des propriétés magnétiques de ces inclusions (elles ne sont pas attirées par l'aimant).

Les inclusions liquides et gazeuses sont allongées, arrondies ou ovales. Les inclusions liquides renferment une autre inclusion petite, gazeuse.

Toutes les inclusions sont primaires puisqu'elles ne sont pas liées de l'apparition et le développement des fissures dans le cristal.

Les cristaux incolores ou rose pale (transparents) sont en général clairs, sans zonation et renfermant de très peu inclusions. Les cristaux dont la couleur varie de rose intense vers brun présentent une structure zonée, développée du centre vers les bords des cristaux, parallèlement aux arêtes, disposée au tour d'un noyau préexistent (quand il existe). Le nombre et l'extension des zones diffère d'un cristal à un autre. Les cristaux zonés sont plus fréquents dans les échantillons P_5, 2319 et 2331 et plus rares dans les échantillons P_1, P_2, P_3, P_4.

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Il y a plusieurs hypothèses concernant les zonations des cristaux de zircon. Certains chercheurs expliquent ces zonations par les variations rythmiques de température (Shubnicov, 1914, in Pupin, Turco, 1970). D'autres (Bruce, Jewitt, 1936, in Pupin, Turco, 1970) considèrent que la disposition régulière des inclusions à des dimensions extrêmement petites, pourrait déterminer l'apparition des zones dans le cristal. L'hypothèse des variations rythmiques de température est plus plausible, étant donné que Ipateva (1960, in Pupin, Turco, 1970) a observé que dans les zones marginales des corps intrusifs il y a des fréquents cristaux zonés, c'est-à-dire dans les zones ou les variations de température pendant le processus de cristallisation sont plus grandes. Les analyses effectuées pour cette étude ont relevé que les échantillons prélevés des zones marginales du massif (P_5, 2319, 2331) comportent, la plupart, des cristaux zonés.

Les cristaux de zircon qui renferment un noyau préexistente sont rares et ont été rencontrés seulement dans les échantillons P_1 et P_5. Le noyau est représenté par un autre cristal, plus vieux, ovale, autour duquel s'est développé un nouveau cristal. Etant donné que ce noyau ne présente pas des propriétés optiques particulières par rapport à tout le cristal (excepté une opacification plus intense) on considère qu'il représente un autre cristal de zircon, formé antérieurement et qui a constitué le support pour la cristallisation d'une nouvelle génération de zircon.

Variation des dimensions des cristaux de zircon. Les dimensions des cristaux ont été déterminées par des mesures effectuées par l'oculaire micrométrique. On a déterminé les trois dimensions maximums $a : b : c$ qui sont réciproquement perpendiculaires. On a constaté que celles-ci varient entre 0,2/0,1/0,1 mm et 0,8/0,2/0,2 mm jusqu'à la longueur de 1,0 mm.

Sur le diagramme de la variation de la longueur des cristaux de zircon on peut observer que les échantillons P_1, P_2, P_3, P_4 présentent le plus grand nombre de cristaux (35 à 50%) ayant la longueur proche de 0,4 mm, tandis que les échantillons P_5, 2319 et 2331 comportent le plus grand nombre de cristaux (45 à 50%) à une longueur de 0,3 mm environ (fig. 2).

Quant à l'élongation des cristaux de zircon, il existe deux maximums de fréquence (fig. 3) : la valeur près de 3 pour les échantillons P_1, P_2, P_3, P_4 et près de 4 pour les échantillons P_5, 2319 et 2331.

Des deux diagrammes résulte la séparation des roches en deux groupes, la première de type granodioritique et granitique (P_5, 2319, 2331) et la deuxième de type dioritique et tonalitique (P_1, P_2, P_3, P_4).

De la corrélation des deux diagrammes avec la distribution typologique des cristaux de zircon (Pupin, 1972), résulte que tant les échantillons avec la valeur de l'indice d'élongation 4, que ceux avec la valeur du même indice 3, présentent le même développement pour les mêmes types de faces de prisme : plus (110) moins (100) et de pyramide (211), moins ou absents (101) ; la seule différence entre les deux groupes est déterminée par la fréquence de leur apparition.

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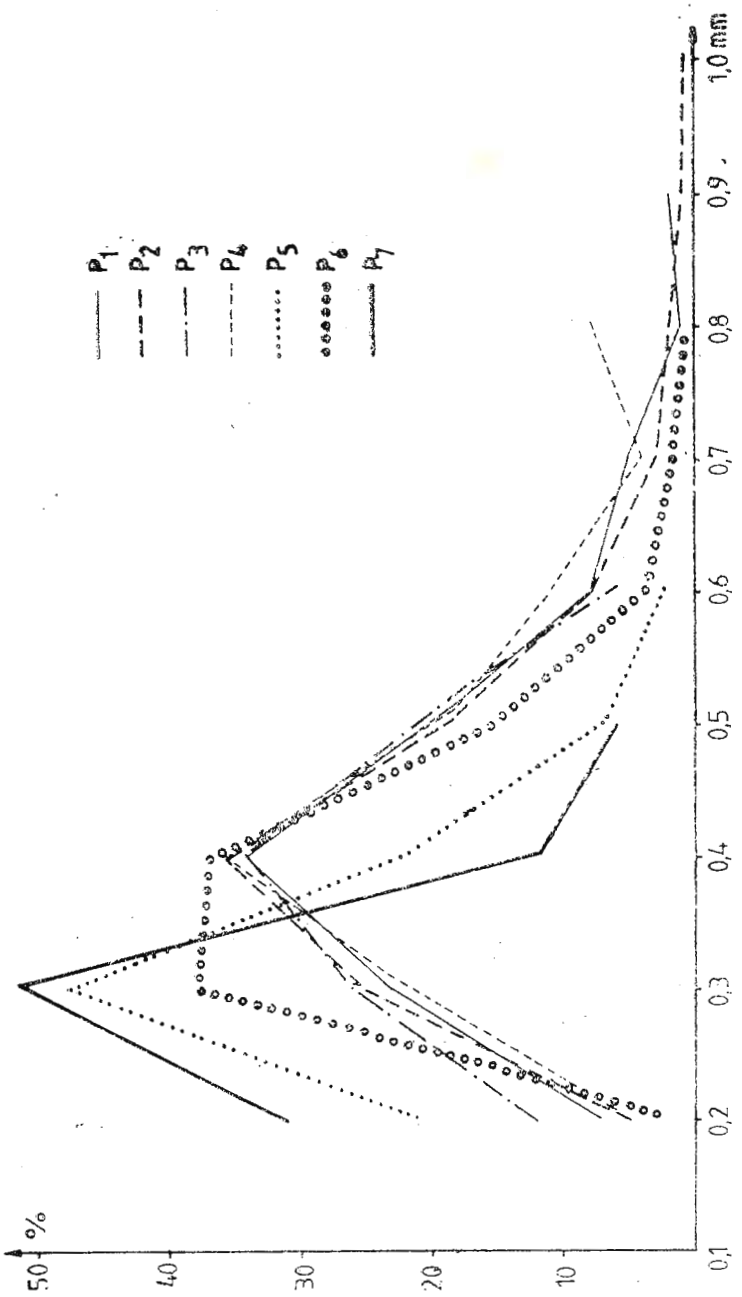

Fig. 3 — Diagramme de variation de la longueur des cristaux de zircon.

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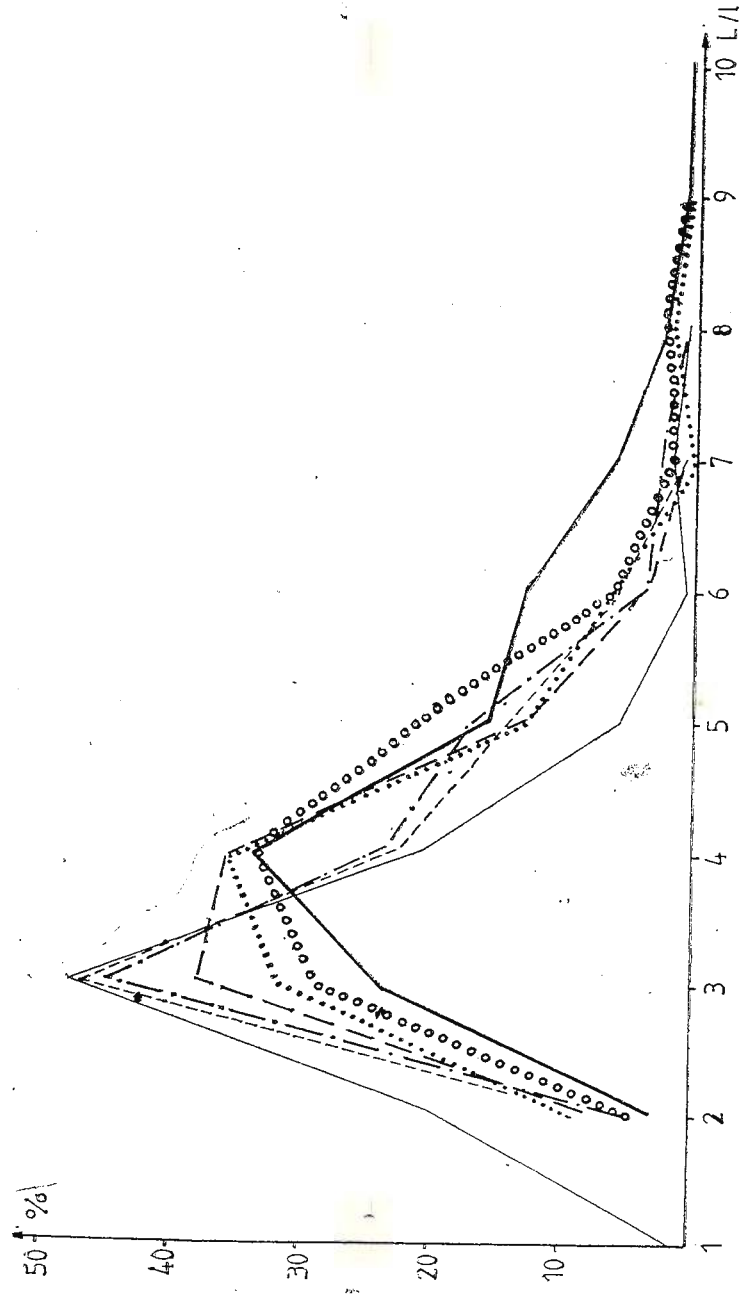

Fig. 3 — Diagramme de variation de l'élongation des cristaux de zircon (voir la légende de la figure 2).

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Méthode typologique. La morphologie des cristaux de zircon a été étudiée en vue de rédiger un diagramme de tendence de l'évolution typologique T.E.T. (Pupin, 1972). On a analysé 100 cristaux de chaque échantillon pour déterminer le type de habitus qui corresponde aux 64 types principaux qui forment la base du diagramme typologique (Pupin, Turco, 1972). On a mis en évidence des cristaux de type Q_3, Q_4, Q_5, S_{11}, S_{12}, S_{13}, S_{16}, S_{17}, S_{18}, S_{19}, G, P_1 et P_2, dont la fréquence varie d'un échantillon à un autre (fig. 4).

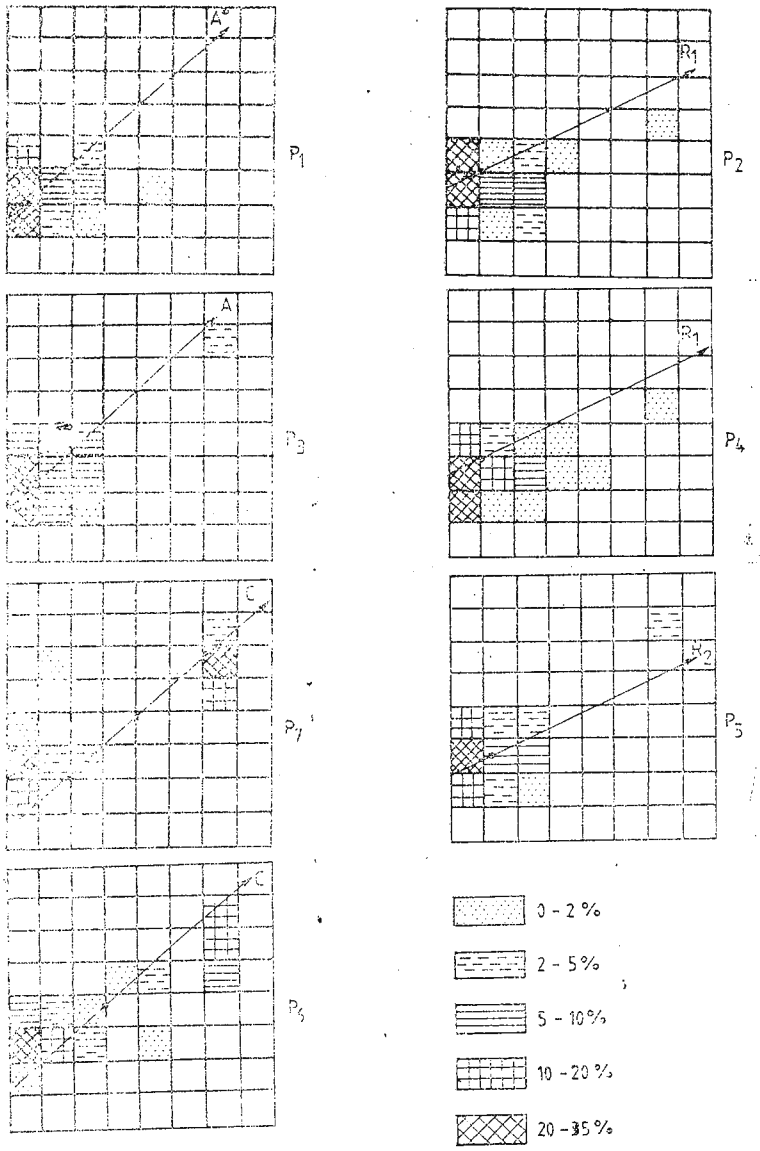

Fig. 4 — Diagramme de fréquence des cristaux de zircon.

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La tendence d'évolution des cristaux de zircon est indiquée dans la figure 5 et a été calculée sur base de la variation des indices IA et IT, en rapport avec la déviation standard du vecteur de tendence (Pupin, Turco, 1980). On peut observer, dans ce diagramme, que la tendence d'évolution est semblable pour tous les échantillons, ayant des tendences vers des faces de pyramide de type (301) et (101):

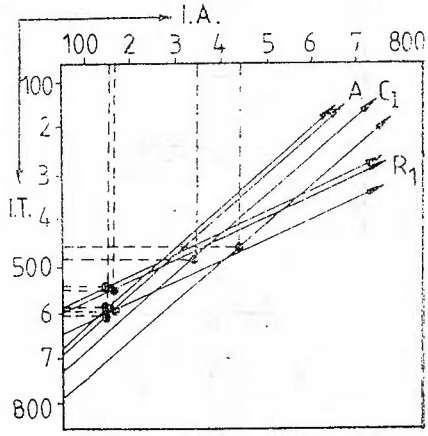

Fig. 5 — Diagramme de tendence de l'évolution typologique (T.E.T.) des cristaux de zircon (selon Pupin, 1980).

Pupin (1980) établit, par la méthode typologique, une étroite interdépendence entre l'aspect extérieur des cristaux de zircon et le chimisme des magmas ou se sont formés. Ainsi, il a observé que la présence des faces pyramidales (211) indique un milieu hypoalcalin ou hyperalumineux, tandis que la présence de la face (101) est déterminée par une alcalinité plus élevée ou par une diminution dans le magma de la teneur en aluminium.

Tenant compte de ces caractères et ayant en vue l'étude morphologique des cristaux de zircon, on peut affirmer que l'évolution du magma du granitoïde de Poneasca a eu lieu dans le sens de l'augmentation de l'alcalinité. Les cristaux originaires des diorites présentent la face (101) très peu dévelopée par rapport aux cristaux provenant des granodiorites et granites qui ont cette face très bien développée. De l'étude sur les cristaux de zircon résulte une transformation du magma initial le long de l'évolution ; on passe d'un magma plus pauvre en alcalis, dioritique-tonalitique où prédomine la face (211) vers un autre, riche en alcalis de type granodioritique-granitique, où apparaît et se développe la face (101).

L'intégration des échantillons étudiés dans le diagramme T.E.T. de distribution des roches plutoniques (fig. 6) indique que les échantillons investigués se projettent dans le champ des granodiorites (2319) ou très proche du champ des diorites (P_1, P_2, P_3, P_4, P_5, 2331), fait qui est en concordance avec les données de chimisme ; on écarte ainsi l'hypothèse de la formation dans un magma à une autre composition chimique et leur assimilation ultérieure.

La projection des données dans le diagramme T.E.T. de distribution des roches granitiques (fig. 7) indique que les roches prélevées du massif

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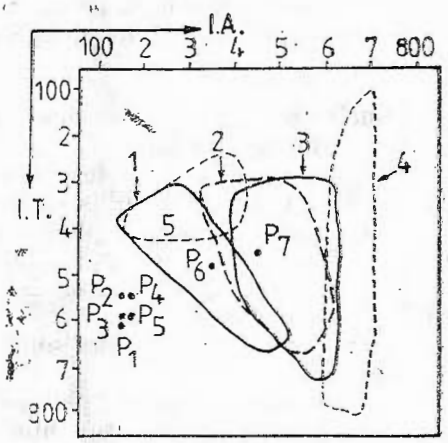

Fig. 6 — Diagramme T.E.T. de distribution des roches plutoniques (selon Pupin, 1980): 1, diorites, gabbros et diorites quartzifères; 2, granodiorites; 3, monzogranites et monzonites; 4, syénites et granites hyperalcalines.

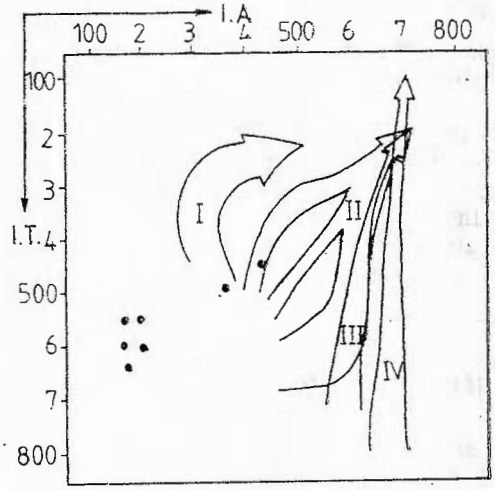

Fig. 7 — Diagramme T.E.T. de distribution des roches granitiques (selon Pupin, 1980): I, granites originaires de la croûte ou à origine principalement dans la croûte; II + III, granites à origine mixte (croûte + manteau); IV, granites à origine dans le manteau ou principalement dans le manteau.

granitoïde de Poneasca peuvent être encadrées dans le champ des granitoïdes de la croûte (P_1, P_2, P_3, P_4, P_5, 2331) et dans le champ des granitoïdes à origine mixte (de la croûte ou du manteau — 2319).

L'apparition de ces quelques cristaux de zircon à noyau atteste la présence d'un zircon à double histoire, fait qui indique l'existence de certains processus d'assimilation pendant l'évolution du magma, mais d'importance réduite, étant donné qu'il n'ont pas déterminé des transformations essentielles dans le chimisme du magma initial.

L'encadrement des types de zircon rencontrés, dans l'échelle géothermométrique (Pupin, Turco, 1972) a conduit à l'établissement de la température de cristallisation du zircon dans le magma et aussi à la détermination de la température de cristallisation du magma. Ainsi, les premiers cristaux de zircon ont cristallisé à des températures de 850°C environ et ont continué de se former tout le long de la cristallisation

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du magma jusqu'à 600°C, avec des petites discontinuités produites par les variations du chimisme du magma et plutôt par les variations de la teneur en zircon.

Conclusions. De l'étude des cristaux de zircon des granitoïdes du massif de Poneasca résulte :

— les faces de prisme les plus fréquentes sont (110) et plus rares (100) ; celles de pyramide plus fréquentes sont (211) et plus rares (101) ;

— quant aux types morphologiques (Pupin, 1972), les plus nombreux sont les cristaux de type Q_5 et Q_4 ;

— la couleur varie de l'incolore jusqu'à brun, les plus abondantes étant les cristaux roses ;

— la plupart des cristaux sont claires ou zonés, ceux à noyau étant très rares ;

— les inclusions sont primaires et sont représentées par des minéraux transparents (zircon, apatite), minéraux opaques, liquides et gaz ;

— les caractères morphologiques indiquent un zircon formé entre 850 et 600°C et deux directions différentes d'évolution du magma : la première alcaline, la deuxième plus pauvre en alcalis ;

— l'encadrement des échantillons étudiés dans le diagramme T.E.T. de distribution des roches plutoniques indique le champ de granodiorites et diorites ;

— dans le diagramme T.E.T. de distribution des roches granitiques, les échantillons analysés se projettent dans le champ des granites originaires de la croûte pour 6 d'entre eux et dans le champ des granites à origine mixte pour un seul echantillon ; ce fait indique une certaine contamination des magmas, situation témoignée aussi par la présence des cristaux à noyau (à double histoire).

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SEMNIFICAȚIA PETROGENETICĂ A ZIRCONULUI DIN MASIVUL GRANITOID DE PONEASCA

(Rezumat)

Studiul cristalelor de zircon s-a efectuat pe un număr de 7 probe a cîte 100 cristale reprezentative. Habitusul cristalelor este lung prismatic, bipiramidat. Modul de îmbinare a fețelor de prismă și de piramidă este determinat de variațiile condițiilor fizico-chimice ale mediului de cristalizare. Fețele de prismă cele mai des întîlnite sînt de tip (100) și/sau (110), iar cele de piramidă sînt de tip (211) și/sau (101).

Cristalele de zircon sînt incolore, roz sau brune în funcție de gradul lor de metamictizare. În interiorul acestor cristale au fost observate incluziuni solide transparente (apatit, zircon), solide opace, lichide și gazoase, dispuse haotic în zona de dezvoltare a fețelor de prismă.

Cristalele pot fi limpezi, zonate sau cu nucleu. Prezența zonelor în interiorul cristalelor este determinată de variațiile de temperatură produse în timpul procesului de cristalizare a magmei. Existența cristalelor cu nucleu atestă posibilitatea preluării unor cristale din zonele limitrofe și regenerarea lor în procesele magmatice ce au dus la punerea în loc a granitoidului de Poneasca.

Din punct de vedere genetic, morfologia cristalelor de zircon indică formarea lor dintr-o magmă cu origine în crustă sau cu origine mixtă (crustă + manta).

Temperatura de formare a acestor cristale, stabilită pe baza încadrării lor în scara geotermometrică, a condus la stabilirea temperaturii magmei în timpul cristalizării zirconului, temperatură cuprinsă între 850—600°C.

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SIGNIFICATION PÉTROGÉNÉTIQUE DU ZIRCON DU GRANITOÏDE DE TISMANA [1]

par

NICOLAE I. OBU [2], LUCIA I. OBU [2]

Granitoid. Zircon. Crystals. Habit. Acidic magmas. Granitic composition. Magmatic differentiation. South Carpathians — Crystalline Danubian Realm — Vilcan Mountains.

Résumé

L'étude morphologique, optique et typologique des cristaux de zircon revèle une similitude de ceux-ci dans tous les échantillons investigués. La température de formation de ces cristaux, résultée des caractéristiques morphologiques, varie entre 900 et 650°C, intervalle marqué aussi par une augmentation permanente de l'alcalinité du magma. Le magma d'où provient le zircon est d'origine crustale ou surtout crustale; ce fait est en faveur de l'hypothèse de Berza (1978) selon laquelle on soupçonne l'existence dans la croûte ou à la base de la croûte, d'un volume de magma qui, par différentiation, a engendré, finalement, un magma résiduel granitique, qui a entrénait aussi dans son ascension une partie de la zone basale, cristallisée ultérieurement comme schlieren ou enclaves dioritiques.

Abstract

Zircons Petrogenetic Significance in the Tismana Granitoid Massif. The morphologic, optic and typologic study of zircon crystals points out a similarity of these crystals in all the study samples. Their formation temperature, deduced on the basis of the morphological features, ranges from 900 to 650°C, interval within which a continuous increase of the magma

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alkalinity is observed. The magma from which zircon crystallized originated in the mantle or mainly in the mantle ; this fact is in favour of the hypothesis formulated by Berza (1978) which presumes the existence, in the mantle or at the crust base, of a magma volume finally generating, by differentiation, a granitic residual magma that in its ascension involved a part of the basic fraction, subsequently crystallized as schlieren or dioritic enclaves.

Le zircon, comme minéral accessoire, a été identifié en sections minces, dans tous les types pétrographiques du granitoïde de Tismana; il se trouve dans la biotite ou forme les bien connues auréoles pléochroïques (Berza, 1978). On a remarqué la fréquence plus élevée du zircon dans les granitoïdes porphyroïdes par rapport à ceux échigranuleux.

En vue d'établir la signification pétrogénétique du zircon, on a procédé à l'identification des cristaux de celui-ci pour obtenir certaines données morphologiques et optiques.

Des données concernant la signification pétrogénétique du zircon ont été présentées par Pupin, Turco pour quelques granitoïdes précambriens de Côte d'Ivoire (1975) ou le massif granitoïde d'Argentera-Mercantour de la France (1969); Bîrlea (1976) pour le massif de Sichevița; Törnroos (1985) pour le zircon des granites pegmatoïdes de Mosambique; Gbelski, Hatar (1982) pour le zircon du massif granitoïde des monts de Mecsek (R.P. de Hongrie); Rozložnik et al. (1982) pour quelques cristaux de zircon des granitoïdes de Stiavnica (R.S. Tchécoslovaquie); Robu, Robu (1987) pour quelques massifs granitoïdes des monts de Parîng.

Cadre géologique. Le massif granitoïde de Tismana est situé dans les schistes cristallins de l'unité danubienne, schistes qui appartiennent à la série de Lainici-Păiuș et sont recouverts par une couverture sédimentaire autochtone, comportant des dépôts paléozoïques et mésozoïques.

La série de Lainici-Păiuș borne le massif granitoïde du nord-est, par le nord et ouest jusqu'au sud-ouest. Berza (1978) individualise, dans le cadre de cette série, deux complexes : le complexe carbonato-graphiteux, comportant des calcaires cristallins, calcschistes, quartzites biotitiques à grenats, gneiss micacés à graphite, amphibolites et le complexe des quartzites et gneiss biotitiques, comportant des quartzites, quartzites feldspathiques, quartzites biotitiques, gneiss biotitiques, gneiss quartzo-feldspathiques.

Le massif granitoïde de Tismana présente une grande variété pétrographique, fait souligné le long des années par tous les chercheurs qui ont investigué ce massif. Berza (1978) y sépare des granites porphyroïdes et granites échigranuleux (syénites, monzonites quartzifères, granodiorites, monzodiorites quartzifères, tonalites, diorites quartzifères et diorites.

Méthodologie d'investigation des échantillons. Pour rédiger cette étude, on a prélevé et préparé 10 échantillons, de 5 kg chacun, du massif granitoïde de Tismana, précisément des granites porphyroïdes de celui-ci. Le lieu de prélèvement est indiqué dans la figure 1.

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Pour obtenir les cristaux de zircon de la roche, on a utilisé la méthodologie de laboratoire concernant la séparation des minéraux lourds : broyage, tamisage, séparation par des liquides lourds (bromoforme), concentration au séparateur Frantz, séparation sous la loupe binoculaire.

De chaque échantillon on a séléctionné 100 cristaux entiers, à caractéristiques morphologiques variées et bien conservées, (faces et arêtes bien individualisées), à des différentes couleurs, de manière que la population de zircon ainsi obtenue soit représentative pour tout l'échantillon.

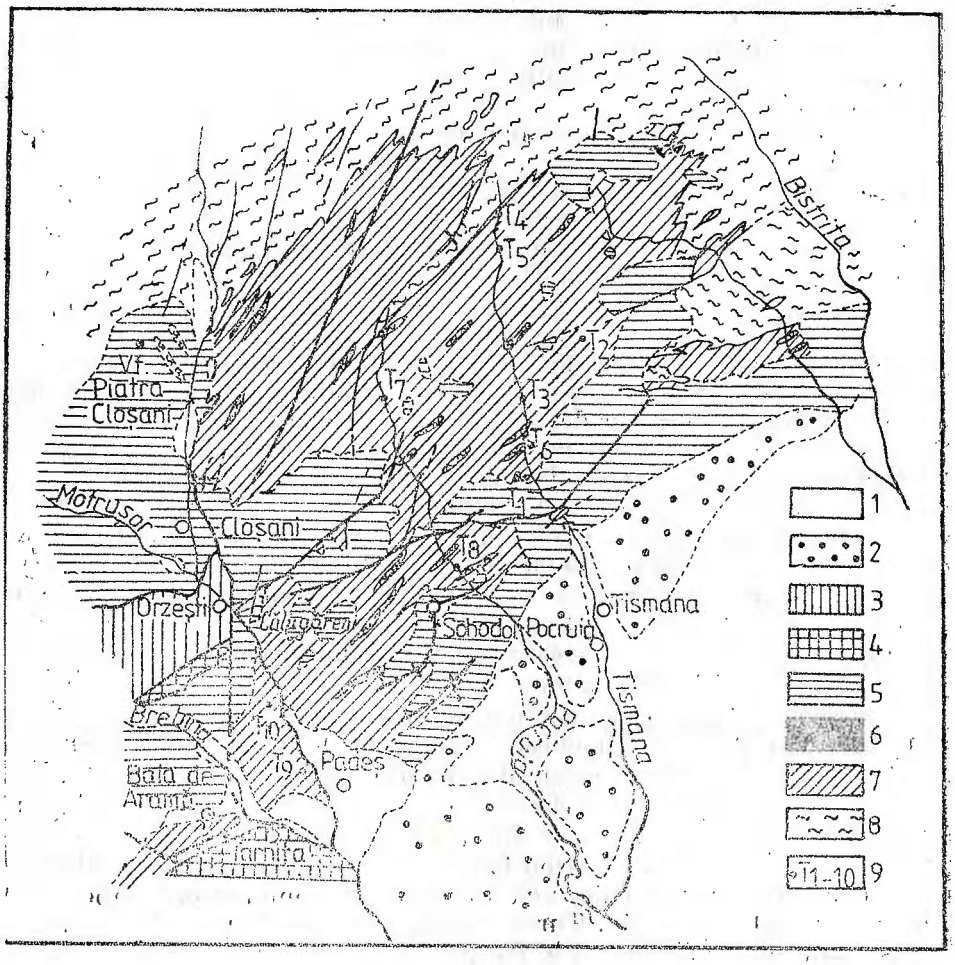

Fig. 1 — Schéma géologique du massif granitoïde de Tismana (selon Berza, 1978).

1, Quaternaire : alluvions ; 2, Tertiaire : sables ; argiles. Nappe Gétique ; 3, Précambrien moyen, Série de Sebeș-Lotru. Nappe de Severin ; 4, Crétacé inférieur, Jurassique supérieur : Couches de Sinaia, Couches d'Azuga et ophiolites. Danubien ; Couverture sédimentaire ; 5, Crétacé supérieur, Jurassique inférieur : arcoses, calcaires, dolomies, argilites ; 6, Permien : arcoses, argilites ; 7, Soubassement, précambrien ; 8, Précambrien supérieur, granitoïde de Tismana, Série de Lainici-Păruș ; 9, T_{1-10} échantillons géologiques.

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Etude optique des cristaux de zircon. L'étude optique des cristaux de zircon a été effectuée à la loupe binoculaire pour obtenir des données concernant la morphologie des cristaux (types de faces, types de habitus, dimensions) et au microscope pour établir les propriétés optiques et les types d'inclusions.

Le habitus des cristaux de zircon est prismatique, résultant de la jonction des faces de type (110) et/ou (100) avec celles pyramidales (211) et/ou (101). Le développement d'une face de ces deux types est déterminé par les variations de température et de chimisme qui ont eu lieu pendant la cristallisation du magma (Pupin, Turco, 1972).

Le granitoïde de Tismana se caractérise par des cristaux lourds prismatiques. Ceux court prismatiques sont plus rares et ceux isométriques sont absents.

Au microscope les cristaux sont incolores, transparents ou de couleur brune, plus obscure ou plus claire, en rapport avec le degré de métamictisation de ceux-ci. La plupart des cristaux sont bruns et opaques ; le taux de cristaux transparents est très bas.

Fréquemment, les cristaux présentent un nombre variable d'inclusions. Les cristaux faiblement métamictisés comportent un nombre plus grand d'inclusions par rapport à ceux incolores et transparents où les inclusions sont très rares ou absents. Elles sont très petites, étant très difficile de les identifier. On a observé des inclusions solides transparentes (zircon, apatite), solides opaques (ilménite, magnétite), fluides et/ou gazeux. Leur dispersion à l'intérieur des cristaux de zircon est chaotique ou parallèle aux faces de la prisme, marquant ou non la zonalité des cristaux. Ces inclusions abondent dans les parties centrales du cristal et dans la zone de développement des faces prismales. Elles présentent un contour arrondi ou ovale pour ceux solides opaques, liquides et gazeux et un contour cristallographique pour l'apatite ou le zircon.

Les fissures des cristaux de zircon ont une position transversale par rapport aux faces prismales, étant déterminées probablement par les variations de pression manifestées pendant l'évolution postmagmatique du massif granitoïde de Tismana.

Etant donné que la pression n'a pas eu des influences sur l'apparition et le développement des inclusions, on peut les considérer primaires, formées pendant la cristallisation du zircon.

Une grande partie des cristaux qui n'ont pas subi des processus de métamictisation (ceux roses ou brun claire) présentent des zonations fréquentes. Le nombre et l'extension des zones diffère d'un cristal à un autre. Leur apparition est liée aux variations rythmiques de température (Shubnicov, 1914 in Pupin, Turco, 1970) qui se sont succédées pendant la cristallisation du zircon. Si le temps de cristallisation est relativement court et précoce, le caractère rythmique de la variation est enregistré au niveau des cristaux par l'apparition des structures zonées. Si la période de cristallisation est longue et la teneur en zircon varie dans le magma, ayant une cristallisation tardive, les structures zonées ont un caractère local (Pupin, Turco, 1970).

Ipateva (1960, in Turco, Pupin, 1970) soutient, elle aussi, l'hypothèse des variations rythmiques de la température. Elle a observé la fréquen-

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ce plus élevée des cristaux zonés dans les parties marginales des corps intrusifs, donc dans les zones où les variations de température sont plus marquantes pendant le processus de cristallisation.

Une autre hypothèse, difficilement à démontrer, soutient que l'apparition des zones est déterminée par la disposition régulière des inclusions à dimensions très réduites.

Tenant compte de ce présenté et du fait que dans les granitoïdes de Tismana la plupart des cristaux de zircon sont zonés, on peut affirmer que pendant la cristallisation du zircon ont eu lieu de nombreuses variations de température.

Exceptant les cristaux zonés et ceux à inclusions, on a observé aussi, dans la majorité des échantillons étudiés (excepté les échantillons T_5 et T_6), des cristaux de zircon à noyau. Ils présentent une partie centrale (noyau) et une zone périphérique qui peut comporter ou non des structures zonées. Le noyau est constitué d'un cristal plus vieux, ovale, plus ou moins allongé, autour duquel se développe un nouveau cristal ayant des propriétés optiques similaires à celle du noyau; résulte donc que le noyau est aussi un cristal de zircon antérieurement formé.

Méthode typologique. Etant donné les dimensions très réduites des cristaux, il a été impossible de les figurer avec l'aide de la chambre claire; l'indexation des faces a été réalisée selon la méthode de l'abaque et par

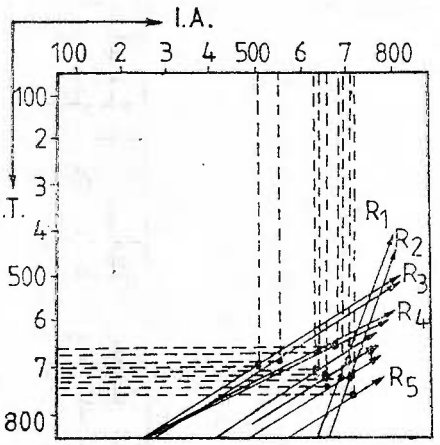

Fig. 2 — Diagramme de tendence de l'évolution typologique (T.E.T.) des cristaux de zircon (selon Pupin, 1980).

l'assimilation des formes rencontrées avec celles présentées par Pupin, Turco (1972) dans le diagramme typologique (T.E.T.).

Les 100 cristaux élus pour chaque échantillon ont été encadrés dans les 64 types principaux établis dans le diagramme susmentionnée et on a obtenu les suivants types de cristaux: Q_3, Q_4, Q_5, S_{12}, S_{16}, S_{17}, S_{19}, S_{20}, S_{21}, S_{25}, P_2, P_3, P_4, P_5, D. La fréquence de chaque type varie d'un échantillon à un autre (fig. 2).

La tendence d'évolution des cristaux de zircon est représentée dans la figure 3 et a été calculée sur la base des indices I.A. et I.T. et de la

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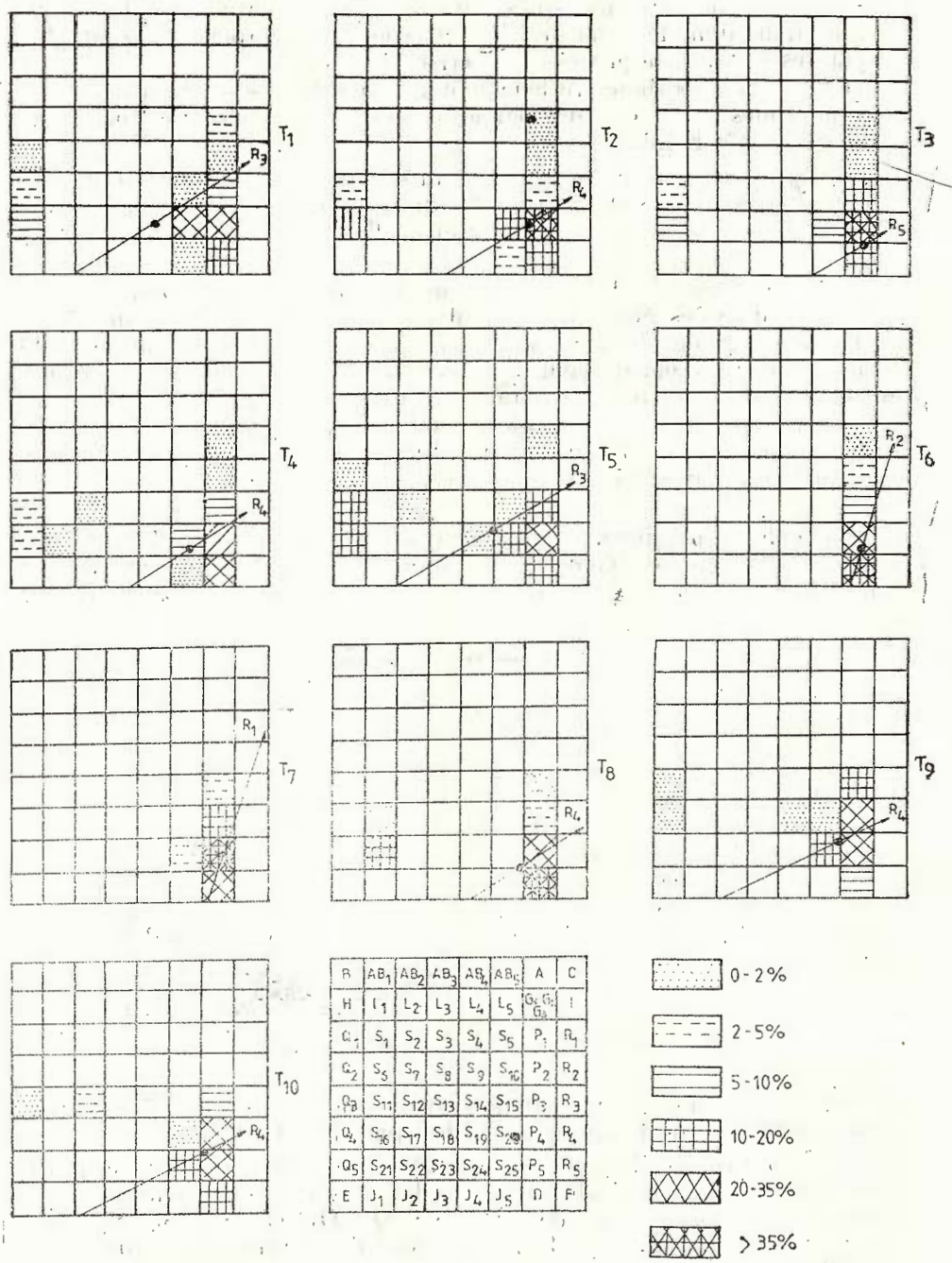

B	AB_1	AB_2	AB_3	AB_4	AB_5	A	C
H	l_1	l_2	l_3	l_4	l_5	G_1 G_2 G_3	I
G_1	S_1	S_2	S_3	S_4	S_5	P_1	R_1
G_2	S_6	S_7	S_8	S_9	S_{10}	P_2	R_2
G_3	S_{11}	S_{12}	S_{13}	S_{14}	S_{15}	P_3	R_3
G_4	S_{16}	S_{17}	S_{18}	S_{19}	S_{20}	P_4	R_4
G_5	S_{21}	S_{22}	S_{23}	S_{24}	S_{25}	P_5	R_5
E	J_1	J_2	J_3	J_4	J_5	D	F

Fig. 3 — Diagramme de fréquence des cristaux de zircon.

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déviation standard du vecteur de tendence (Pupin, Turco, 1980). Résulte du diagramme la tendence d'évolution pareille des 10 échantillons vers des formes avec la face pyramidale (301) prédominante (R_1, R_2, R_3, R_4, R_5). Ce fait atteste l'homogénéité, du point de vue chimique, des magmas, ou ont cristallisé.

Le développement des deux types de faces (101) et/ou (211) reflète une variation du chimisme manifestée dans le sens de l'augmentation

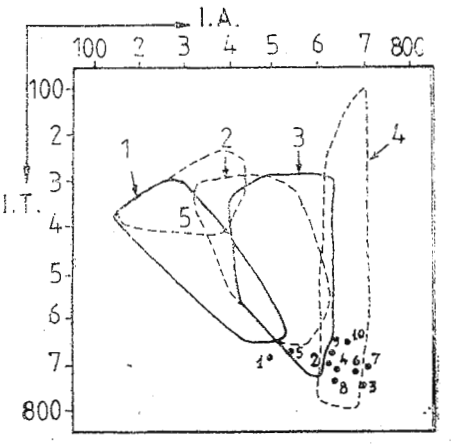

Fig. 4 — Diagramme T.E.T. de distribution des roches plutoniques (selon Pupin, 1980).
1, diorites, gabbros et diorites quartzifères; 2, granodiorites; 3, monzogranites et monzonites; 4, syénites et granites hyperalcalines.

de l'alcalinité du magma — prédominance de la face (101) par rapport à (211) (Pupin, Turco, 1980).

L'intégration des cristaux de zircon étudiés dans le diagramme T.E.T. de distribution des roches plutoniques (Pupin, 1980) indique que la majorité de ceux-ci se projettent dans le champ des granites et syénites alcalines et hyperalcalines (fig. 4), ôté deux échantillons qui sont inclus dans le champ des monzogranites et monzonites (T_5) ou près de celles-ci (T_1).

La projection des cristaux de zircon dans le diagramme T.E.T. de distribution des roches granitiques (fig. 5) indique l'encadrement de ceux-ci dans l'aire des granites originaires du manteau ou principalement du manteau, ôté deux échantillons placés dans le champ des granites d'origine mixte du manteau ou de la croûte (T_5) ou tout près (T_1).

La possibilité que le zircon existente dans le granitoïde de Tismana soit cristallisé d'un magma originaire du manteau, est en faveur de l'hypothèse de Berza (1978), selon laquelle on suppose la formation dans le manteau ou à la base de la croûte, d'un volume de magma basique, qui, par différentiation, a engendré finalement un magma résiduel de composition granitique qui a entraîné dans son ascension une partie de la fraction basique, cristallisée ultérieurement comme schlires ou enclaves dioritiques.

La température de cristallisation du zircon du granitoïde de Tismana varie entre 900 et 650°C et coïncide aux données établies par la cristallisation des magmas, ayant à la base des considérents d'ordre minéralogique. Ainsi, la coexistence d'un plagioclas basique (An 30) avec un autre très

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potassique, pour lequel on a estimé une composition $Or_{70}Ab_{30}$, indique, selon Barth (1962 in Berza, 1978), une température de cristallisation de ces minéraux d'approximativement 700°C. Aussi les données chimiques

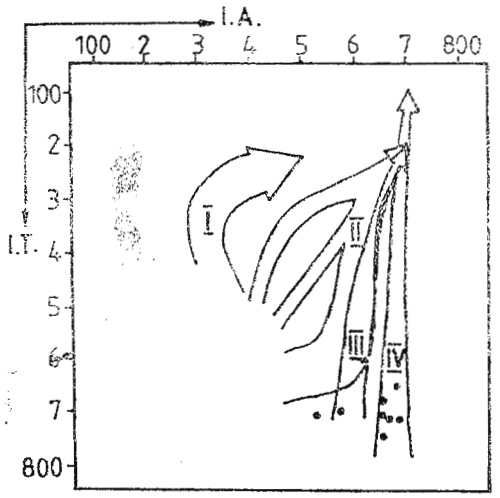

Fig. 5 — Diagramme T.E.T. de distribution des roches granitiques (selon Pupi n, 1980).

I, granites originaires de la croûte, ou principalement de la croûte; II + III, granites à origine mixte (de la croûte + manteau); IV, granites originaires du manteau ou principalement du manteau.

de la biotite des granitoïdes porphyroïdes de Tismana indiquent une température de formation de 900°C environ, conformément aux données expérimentales concernant la cristallisation de la biotite (Deer et al., 1962, in Berza, 1978).

Les dimensions des cristaux de zircon ont des valeurs comprises entre 0,2/0,1–0,1 mm et 1,1/0,3–0,2 mm, présentant, pour la longueur, deux maximums de fréquence : le premier d'approximativement 0,3 mm (30—49%) pour les échantillons T_2, T_5, T_7, T_9 et le second 0,4 mm environ (32—46%) pour les échantillons T_1, T_3, T_6, T_8, T_{10} (fig. 6).

L'indice d'élongation varie entre 1 et 8, à un maximum de fréquence de 3 (36—55%) pour presque tous les échantillons, ôté l'échantillon T_4 dont le maximum de fréquence (40%) a la valeur 4 (fig. 7).

Résulte de l'étude des diagrammes une homogénéité concernant la variation des dimensions des cristaux de zircon (maximums de fréquence de la longueur et de l'élongation très proches pour tous les échantillons), donc une variation relativement uniforme des conditions physico-chimiques du milieu de cristallisation.

Conclusions. L'étude effectuée sur les cristaux de zircon du massif granitoïde de Tismana relève les suivants :

— les faces plus fréquentes sont, pour les prismes (110) et/ou (100) et pour les pyramides (211) et/ou (101);

— les plus abondantes types morphologiques des cristaux de zircon sont P_4, P_5, D, S_{25};

— la couleur des cristaux de zircon est brune claire en général (faiblement opacifiés aussi) et plus rare incolores et transparents;

— la plupart des cristaux présentent des inclusions et une petite partie comportent des noyaux. Les inclusions sont primaires et de plusieurs

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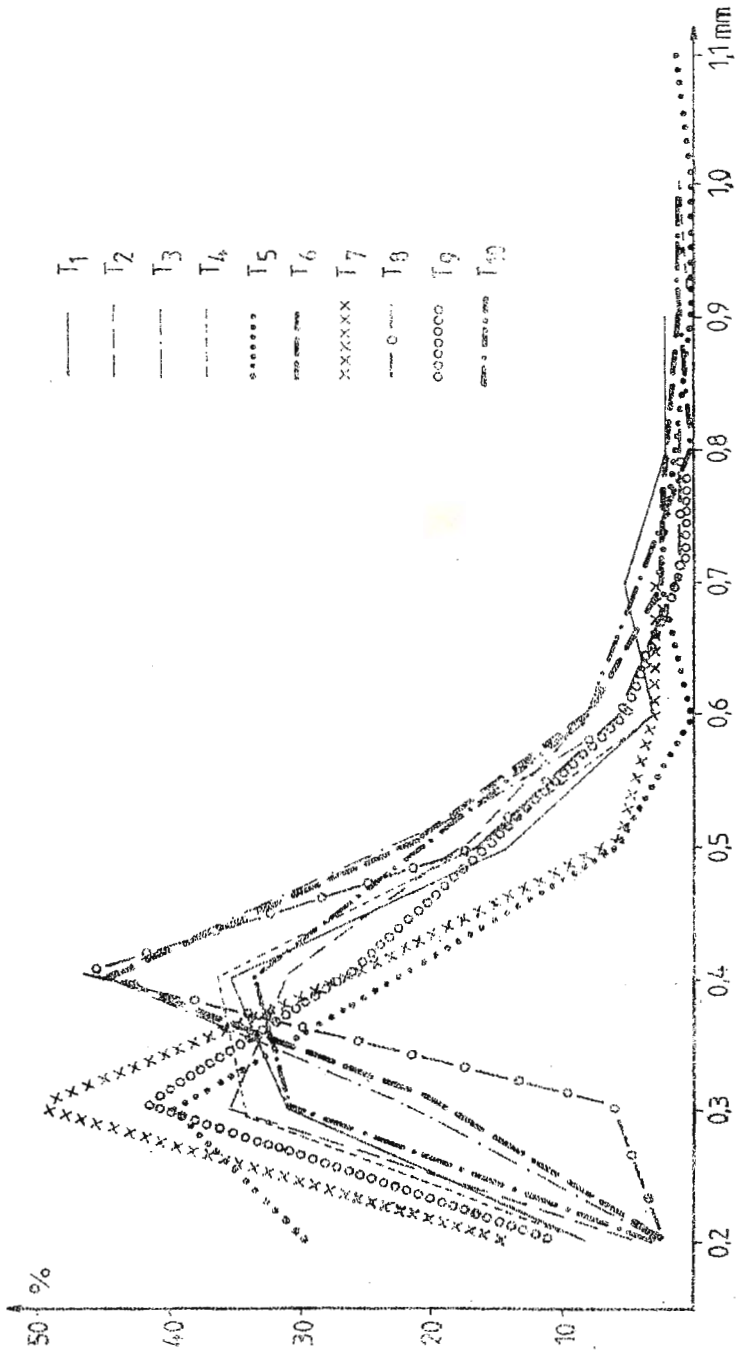

Fig. 6 — Diagramme de variation de la longueur des cristaux de zircon.

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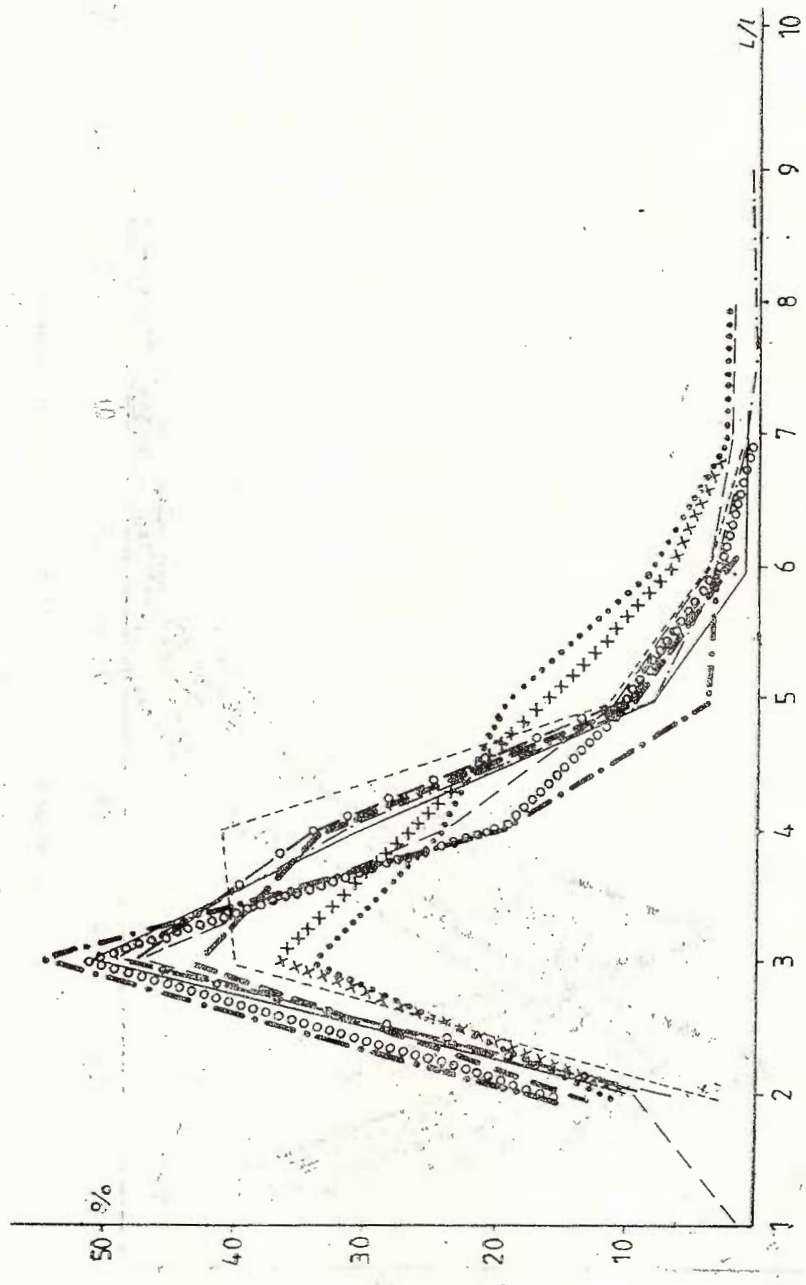

Fig. 7—Diagramme de variation de l'élongation des cristaux de zircon (voir la légende de la figure 6).

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types : solides transparentes (zircon, apatite), solides opaques (ilménite, magnétite), liquides et gazeux ;

— les caractéristiques morphologiques des cristaux indiquent un zircon formé entre 900 et 650° C, intervalle où le degré d'alcalinité du magma a enregistré une ascension permanente ;

— l'encadrement des échantillons étudiés dans le diagramme T.E.T. de distribution des roches plutoniques indique le champ des granites hyperalcalins ;

— on observe, dans le diagramme T.E.T. de distribution des roches granitiques, que les échantillons en discussion se situent dans l'aire des granites originaires du manteau ou principalement dans le manteau.

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SEMNIFICAȚIA PETROGENETICĂ A ZIRCONULUI DIN GRANITOIDUL DE TISMANA

(Rezumat)

Semnificația petrogenetică a zirconului a fost stabilită în urma studiului optic efectuat pe 10 probe a cîte 100 cristale reprezentative. Habitusul cristalelor este, în general, lung prismatic, bipiramidat. Fețele de prismă observate sînt de tip (110) și/sau (100), iar cele de piramidă sînt de tip (211) și/sau (101). Frecvența mai mare a cristalelor cu fețe de piramidă de tip (101) indică o creștere continuă a alcalinității magmei în timpul cristalizării zirconului.

În urma studiului microscopic au fost observate mai multe tipuri de incluziuni : solide transparente (zircon, apatit), solide opace (ilmenit, magnetit), fluide și gazoase. Dispunerea lor în interiorul cristalelor este haotică sau paralelă cu fețele de prismă.

Au fost observate numeroase cristale de zircon zonate. Numărul și extinderea zonelor diferă de la un cristal la altul, fiind determinate de variațiile ritmice de temperatură sau ale conținutului de zircon din magmă.

Unele cristale de zircon prezintă nucleu, dispus în partea centrală și reprezentat printr-un cristal mai mic, format anterior și regenerat în ultimul proces magmatic.

Încadrarea tipurilor morfologice ale cristalelor de zircon studiate în clasificarea tipologică (Pupin, Turco, 1972) a permis plasarea granitoidului de Tismana în grupa granitoidelor cu origine în manta sau în principal în manta.

Temperatura de formare a zirconului, dedusă din scara geotermometrică, este cuprinsă între 900—650°C, suprapunîndu-se cu datele stabilite pentru cristalizarea magmelor pe considerente mineralogice.

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1. MINERALOGIE—PETROLOGIE

PETROLOGIA ROCILOR MAGMATICE

PETROLOGY OF LIASSIC OPHIOLITES AND LATE KIMMERIAN ISLAND ARC VOLCANICS IN THE VAȚA—CĂZĂNEȘTI—ȚEBEA REGION (MUREȘ ZONE) AND THEIR TRACE ELEMENT CONTENTS[1]

by

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Ophiolite. Lower Jurassic. Volcanic rocks. Upper Jurassic, Lower Cretaceous, Island arcs. Oceanic crust. Upper mantle. Subduction. Mafic magmas. Basaltic composition. Magmatic differentiation. Major elements. Minor elements. Apuseni Mountains — South Apuseni Mountains — Drocea Mountains, Metaliferi Mountains; Lower Cretaceous-Paleogene magmatite — South Apuseni Mountains.

Abstract

In the region lying between Vața, Căzănești and Țebea (Mureș Zone) two Mesozoic eruptive series occur : (1) a Liassic ophiolitic series, which forms the basement of the Căpîlnaș—Techereu Unit and (2) a Late Kimmerian island arc (postophiolitic) volcanic series, formed concomitantly with the J_3—Cr_1 flysch, which overlies the ophiolitic rocks. The latter represents a fragment from the Mureș Ocean crust and is represented by basalts, variolites, dolerites and gabbros of the complex O_1. The rocks display petro- and geochemical features of ocean floor tholeiites originating in the upper mantle by melting of at least 15 per cent of it. Island arc volcanics are represented by porphyritic basalts, basaltic andesites and augite- and hornblende-bearing andesites with a mostly porphyritic texture. They are calc-alkaline rocks with a tholeiitic trend, coming from a basaltic magma formed by the melting of the subducted Mureș Ocean crust or from the upper mantle overlying the Benioff plane, strongly influenced by solutions and gases emanated from this subducted oceanic crust. The parent magma of the Late

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Kimmerian volcanics was involved in an intense process of fractional differentiation, as indicated by the negative anomaly of Eu and the low HREE content of the rocks. This process led to the formation of a wide variety of volcanic rocks.

Résumé

Pétrologie des ophiolites liasiques et des vulcanites d'arc insulaire néochimmériennes de la région de Vața—Căzănești—Țebea (zone de Mureș) et la teneur en éléments mineurs. Dans la région située entre Vața, Căzănești et Țebea (zone de Mureș) on rencontre deux séries de roches éruptives mésozoïques : (1) une série de roches ophiolitiques liasiques qui forme le soubassement de l'unité de Căpîlnaș—Techereu et (2) une autre série de vulcanites d'arc insulaire néochimmériennes postophiolitiques, qui se sont formées parallèlement au flysch J_3—Cr_1 et qui reposent sur les roches ophiolitiques. Ces dernières représentent un fragment de la croûte de l'Océan de Mureș et comportent des basaltes, variolites, dolérites et gabbros du complexe (O_1). Les roches présentent des caractères pétro- et géochimiques des tholéiites de fond océanique et sont le résultat de la fusion de plus de 15% du manteau supérieur. Les vulcanites d'arc insulaire sont représentées par les basaltes porphyriques, andésites basaltiques et andésites à augite et hornblende, dont la structure est porphyrique en grande partie. Elles sont des roches calco-alcalines à tendence tholéiitique qui ont résulté d'un magma basaltique formé par la fusion de la croûte de l'Océan de Mureș entraîné en profondeur ou provenant du manteau supérieur situé au-dessus du plan Benioff, fortement influencé par des solutions et gaz émanés par cette croûte océanique. Le magma qui a engendré les vulcanites néochimmériennes a été impliqué dans un processus intense de différentiation fractionnée selon résulte de l'anomalie négative de l'Eu et de l'appauvrissement des roches en HREE. Ce processus a déterminé la formation d'une grande variété de roches volcaniques.

Introduction

The study region (Pl.1) belongs to the Mureș Zone (South Apuseni Mountains). It lies between Vața, Căzănești, Birtin and Țebea, in the northeastern part of the main outcropping area of the Liassic ocean crust fragment in the basement of the Căpîlnaș—Techereu Nappe, including also a part of its cover of Mesozoic volcanic and sedimentary formations.

In the past only a few studies were carried out in the region.

Referring only to the Mesozoic eruptive rocks presented in this paper we observe that Ghițulescu and Socolescu (1941) marked on their map, south of Birtin, the complex of Triassic (?) melaphyres consisting of nonseparated porphyrites, gabbros, diabases, melaphyres, etc. Since then the knowledge on the Mesozoic magmatites in the Mureș Zone developed continuously. Giușcă et al. (1963) referred the volcanics of the above-mentioned melaphyre complex to three different volcanic phases of the same Mesozoic volcanism. Since the beginning of the '60, Savu (1962 a,b) mentioned in the western part of the Mureș Zone two distinct magmatic series : an older series representing an ophiolitic series and a younger (Late Kimmerian)

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calc-alkaline one. In 1974 Hertz et al. assigned the Liassic age (180 MA, Rb/Sr) to the ophiolitic series. In 1975 Savu and Nicolae pointed out that the magma generating the calc-alkaline series had been formed by subduction processes. Later researches (Savu et al., 1978; Savu, 1980, 1983; Savu, Udrescu, 1981) revealed that the two Mesozoic magmatic series are different: the first series represents an ophiolitic slab of ocean floor magmatic rocks obducted from the Mureș Ocean floor, and the second series represents the products of a postophiolitic Late Kimmerian island arc volcanism.

In the western part of the region the basalts and gabbros at Căzănești-Ciungani (west of the annexed map) were studied petrologically by Cioflica (1962). The ophiolitic rocks and the Late Kimmerian island arc volcanics south of this area were studied by Savu (1962 c) and Savu et al. (1984 a,b; 1985; 1986 b). Jude et al. (1973) were concerned with the Mesozoic and Neozoic volcanics at Caraci as well as with the eastern part of the banatitic intrusion at Măgureaua Vaței, an eruptive body investigated especially by Cioflica (1960) and Ștefan et al.(1982). More recently Ștefan (1986) evidenced the petrochemical particularities of the Eocretaceous magmatites in the region. The Mesozoic sedimentary deposits were investigated by Ghițulescu and Socolescu (1941) and more recently by Lupu and Lupu (unpublished data).

Considerations on the Geological Structure of the Area

As already mentioned (Savu, 1983), there are two pre-Laramian Mesozoic magmatic series in the area of study: (1) Liassic ophiolites and (2) Late Kimmerian island arc volcanics. The former constitute the western part of the area and the latter—overlain by Mesozoic and Neozoic sedimentary deposits and by Neogene volcanics — cover the eastern part.

It is of note that older workers (fide Papiu, 1953) considered that somewhere along the Basarabasa-Zam line — within the basic or melaphyre complex — there were diabases westward and melaphyres eastward. That imaginary line represented, in fact, the boundary between the ophiolitic rocks and the island arc rocks, separated later on. The limit between the two series was interpolated by Savu (1983) and Savu et al. (1986 c) and minutely mapped in 1986 on the occasion of the elaboration of the geologic map, scale 1:50 000, Vorța sheet. It is represented by a sinuous line extending westward in the synclinal zones and regressing eastward in the anticlinal zones (Pl. I), indicating the eastward sinking of the geological structures in the area. Thus, the limit between the two magmatic series occurs under the Neogene volcanics at Basarabasa and continues southeastward up to Vața de Sus, where is the axis of a small anticline; then it extends southwestward west of the Măgureaua Peak (905 m) up to Ponor Valley, where is the axis of a synclinal structure (Pl. I) approximately trending E—W. The banatitic (Laramian) intrusion at Măgureaua Vaței, intruding Mesozoic (Senonian inclusively) sedimentary deposits, lies in the northern side of the structure.

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As shown by Savu et al. (1984 a) from the Ponor Valley the limit regresses southeastward up to the axis of the Cerbia-Luncoi-Almaşu Mare anticlinal and then it extends again south-westward, to Zam, parallel to Cărmăzinești-Căprioara syncline lying in the axial zone of the southern island arc in the Mureș Zone (Savu, 1983).

Petrography of Ophiolitic Rocks

The ophiolitic rocks in the basement of the region are represented by the Liassic ocean floor basaltic complex (O_1) in the Mureș Zone, where some small gabbroic bodies and a peridotitic body are intruded. The Liassic age (180 MA) of the ophiolitic rocks was determined by the Rb/Sr method (Herz et al., 1974).

Basalts constitute submarine lava flows mostly in pillow lava facies (Pl. II, Fig.1). The big axis of the pillow lava separations varies between 0.3 and 2 m in length. Hyalobasalts (Pl. II, Fig.2) — especially in the crust of the pillow lava separations — basalts (Pl.II, Fig.3), amygdaloidal basalts, anamesites (Pl.III, Fig.1), variolites and rarely sills and dykes of dolerites can be distinguished among them. These rocks show frequently an intersertal texture, rarely doleritic (ophitic), variolitic and arborescent. They consist of anorthitic or albitic (spilites) plagioclase laths, crystals of augite ($c \wedge Ng = 48°$), magnetite, rarely pyrite and devitrified glass. Small, isolated plagioclase crystals with microphenocryst aspect are locally observed particularly in rocks with a high-glass content.

The basaltic rocks in pillow lava facies locally contain formations of the vesicle-type described by Walker et al. (1980) and by Savu and Manea (1982) in basalts in the Laleşinț area (Mureș Zone). These "vesicles" are filled with basaltic material, finer than that in the surrounding basaltic rock, from which it differs clearly.

Thin levels of agglomerates and volcanic breccias of ocean floor basalts are rarely intercalated between the amygdaloidal basalts in the Cerboaia and Vătișoara valleys.

The small gabbroic bodies of the basaltic complex are located as follows: one on Vața Valley, two in Ponor Valley and three in Strîmbu Brook (Almaș Sălișteț) where also occurs the peridotitic body at the mouth of the Șovaru Brook, minutely studied by Savu et al. (1986 b).

The rocks of the basaltic complex were affected by an ocean floor metamorphism (Coleman, 1977) in the greenschist and zeolitic facies (Savu, 1967), as also indicated by the secondary mineral assemblages from rocks and their amygdals, usually represented by zeolites and by chlorite-calcite paragenesis. A peremptory proof of the ophiolites formation during the Liassic, not concomitantly with flysch J_3—Cr_1, is represented by the fact that while they had been affected by the ocean floor metamorphism, the flysch deposits were not influenced by it, only by diagenesis phenomena that generated calcite veinlets. It is only the Liassic limestones intercalated between the basaltic flows in the south of the Mureș Zone that recrystallized under the influence of the ocean floor metamorphism (Savu, 1984).

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TABLE 1

Chemical composition of the ophiolitic rocks

No *	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
SiO_2	43.62	44.86	46.26	46.88	48.50	48.82	48.92	48.94	49.26	49.45	49.70	49.91	50.00	50.60	51.40
Al_2O_3	15.55	13.89	13.14	14.41	14.79	14.95	13.59	14.56	14.90	15.80	17.51	14.11	14.50	14.30	15.34
Fe_2O_3	9.20	3.52	7.68	2.43	4.58	5.97	6.59	4.99	12.88	4.06	3.41	3.25	4.01	3.74	4.31
FeO	3.75	5.87	2.63	5.42	5.31	5.32	5.51	4.22	1.91	5.53	3.98	6.39	6.00	6.86	6.87
MnO	0.23	0.25	0.15	0.15	0.16	0.24	0.19	0.13	0.22	0.19	0.11	0.16	0.20	0.21	0.20
MgO	5.20	6.52	5.59	7.65	6.70	5.61	5.00	8.55	5.05	6.76	7.37	7.16	7.21	6.56	5.99
CaO	12.40	11.79	12.26	13.89	10.58	8.94	8.51	9.42	6.60	9.76	9.72	10.55	9.32	9.11	7.11
Na_2O	1.18	3.41	3.78	2.84	2.68	3.41	3.86	2.49	4.86	2.49	2.48	3.06	3.54	3.37	7.38
K_2O	0.0	0.27	0.33	0.43	0.10	0.64	2.06	0.20	0.21	0.34	0.19	0.30	0.34	0.30	0.20
TiO_2	1.55	1.06	0.92	0.78	1.33	1.4	1.44	1.16	1.76	1.32	0.78	1.16	1.18	1.62	1.64
P_2O_5	0.15	0.09	0.01	0.04	0.11	0.15	0.10	0.10	0.15	0.06	0.04	0.10	0.07	0.15	0.10
CO_2	3.94	5.08	4.75	2.35	1.05	0.82	2.31	0.62	0	1.21	1.77	1.25	0.66	0	0
S	0.10	0.10	0.10	0.22	0.22	0.10	0.10	0.19	0.11	0.12	0.10	0.24	0.10	2.22	0.11
H_2O	2.67	3.44	2.71	2.58	3.95	3.14	2.55	3.88	2.09	2.56	2.28	2.71	3.49	2.86	3.85
Total	99.58	99.73	100.20	99.82	100.09	99.71	100.75	99.45	100.00	99.65	99.44	100.35	100.62	99.90	99.50
Ni (ppm)		40	36	100	32	20	19	42	15	100	75	50	70	75	70
Co		38	38	42	45	42	36	40	42	38	37	40	59	40	40
Cr		80	130	360	55	8	9	125	3.5	310	140	160	135	130	180
V		320	300	270	320	320	260	280	550	360	250	280	340	400	280
Sc		42	42	34	30	40	40	44	44	36	40	40	38	38	36
Y		26	22	19	27	32	30	23	34	29	18	22	31	43	30
Yb		3.1	3.2	2.4	3.6	3.4	3.5	3.5	3.5	2.8	2.2	2.5	2.9	4.7	3.5
La		<30	<30	<30	<30	<30	<30	<30	<30	<30	<30	<30	<30	<30	<30
Nb		<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10
Zr		60	68	30	80	100	110	62	110	76	56	68	70	135	92
Ba		15	20	18	<10	170	520	<10	<10	32	15	72	34	26	15
Sr		340	520	259	170	750	370	90	100	310	460	310	330	180	170
Pb		< 2	2	< 2	< 2	< 2	3	< 2	< 2	< 2	< 2	< 2	< 2	12	3
Cu		65	95	50	65	48	31	80	31	145	7	70	17	72	56
Ga		12	14	12	16	16	11	14	15	15	12	15	14	19	19
Sn		< 2	2	< 2	< 2	< 2	< 2	2.5	< 2	< 2	< 2	2.5	< 2	< 2	< 2

* The analyses in this table represent : amygdaloidal basalts : 1, Vaţa Valley ; basalts : 14, 15, Vaţa Valley ; 8, Mestecăroaia Valley ; 4, Cerboaia Valley ; 2, Surăneasa Valley ; hyalobasalts, 6, Surăneasa Valley, anamesites ; 10, Vaţa Valley ; spilites, 3, Cerboaia Valley, 7, Surăneasa Valley ; 9, Vaţa Valley ; dolerites ; 5, Cerboaia Valley ; 12, Vaţa Valley ; 13, Căzăneşti Valley ; gabbro ; 11, Căzăneşti Valley.

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Geochemistry and Origin of Ophiolitic Rocks

The chemical composition of the ophiolitic rocks (Tab. 1) indicates SiO_2 between 43 and 52%, values which assign them to the basalts field, one of the rocks being a gabbro. With only one exception the content of K_2O is less than 0.50%, characteristic of the tholeiitic basic rocks (Miyashiro, 1975), as shown on the diagram in Figure 1. Although in most of the rocks Na_2O varies around 2.50%, in some of them (e.g. spilites and partially spilitized rocks) it rises up to 4.84%. For this reason on the diagram in Figure 1 the rocks fall from the tholeiitic basalts field to the calc-alkaline rocks, although they belong to the same tholeiitic series. The appurtenance of these rocks to the tholeiitic series is also confirmed by the average value 0.68 of the CaO/Al_2O_3 ratio, which is very close to the minimum value (0.70) of the average of this ratio established by Werner (1984) for ophiolites.

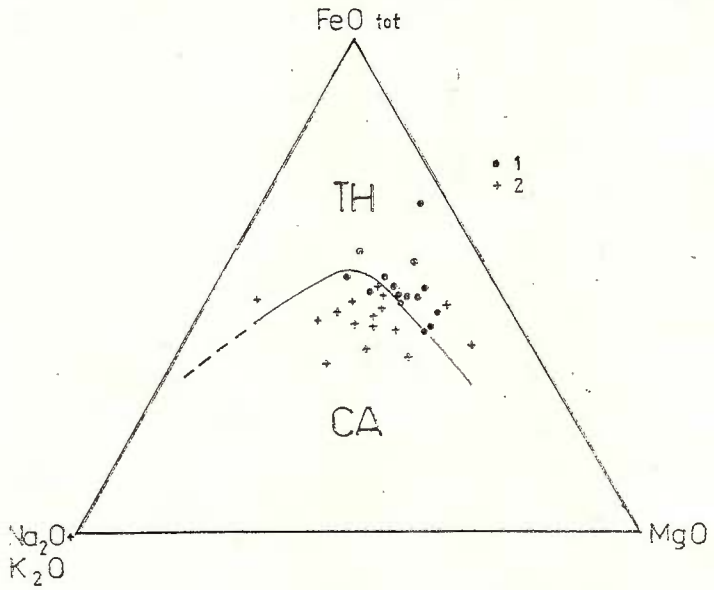

Fig. 1 — FeOtot—MgO—$Na_2O + K_2O$ diagram for ophiolitic rocks and island arc volcanics (Irvine, Baragar, 1971).

1, ophiolitic rocks; 2, island arc volcanics.

The TiO_2 content of the ophiolitic rocks, with a few exceptions, is similar to that of the "high-Ti"-type ophiolites established by Serri and Saitta (1980) for the tholeiites from the middle oceanic ridges (Fig. 2).

The trace element contents in ophiolitic rocks (Tab. 1) also emphasize their character of ocean floor tholeiites. Thus, we observed that the variation domains of Cr, Ni, V, Co and Sc range between the established limits (Savu et al., 1984 b) for normal or "nondifferentiated" ophiolitic rocks from the sheeted dyke complex in the Mureș Zone. An exception are sam-

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ples nos 6, 7 and 9, among which two spilites and one hyalobasalt which, according to the very low content of Cr and Ni, indicates rocks yielded as a result of a liquid-magmatic differentiation process of the parental

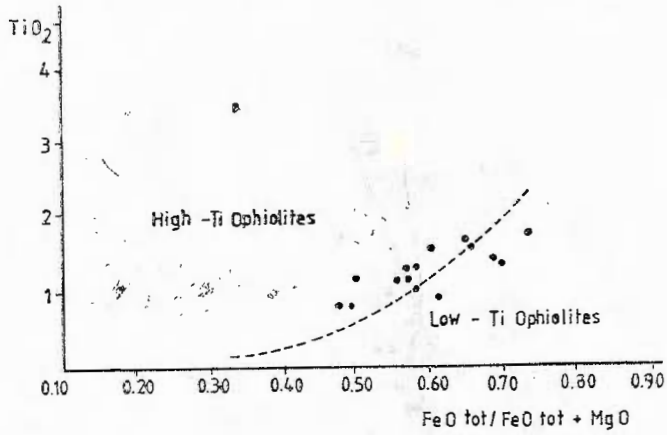

Fig. 2 — TiO_2—FeOtot/FeOtot + MgO diagram (Serri, Saitta, 1980) for ophiolitic rocks.

tholeiitic magma. The lowest-Cr and -Ni rock (no. 9) displays the highest values of V and Ti pointing out its character of differentiated rock, considering the variation tendencies of those elements in the "basic differentiated rocks" of the ophiolitic series in the Mureș Zone (Savu et al., 1984 b). This conclusion is confirmed by diagram V—Ti/1000 (Fig. 3, Shervais, 1982) on which the majority of the basaltic rocks fall in the MORB field.

Zr, Y, Yb and La show variation domains ranging within the variation limits of these elements in the normal basic ophiolitic rocks in the Mureș Zone. On the Ti—Zr diagram (Fig. 4, Pearce, 1980) practically all the basic ophiolitic rocks in the Vața—Căzănești area plot in the MORB field as on the above-mentioned diagram. They also form a quite restricted field on Zr—Ti/Zr diagram (Fig. 5) and it clearly differentiates them from the Late Kimmerian basic island arc volcanics.

The low-Ba values, generally less than 100 ppm, situate the rocks on the diagram in Figure 6 (Miyashiro, 1975) in the abyssal tholeiite field of the North and South Azore Islands. The higher Ba contents in two basaltic rocks (nos 6 and 7) might be due to a supply determined by hydrothermal solutions related to Laramian (banatitic) magmatic rocks that in places affected the ophiolites too, as these intrusions yielded low-baryte deposits in the area of study (Ștefan, Istrate, 1973, unpublished data).

With one exception, the Sr contents vary between 90 and 750 ppm, the average value being 310 ppm. The higher Sr values than in the other ophiolites in the Mureș Zone might be due to the presence of this element in secondary calcium, frequently found in basalt amygdales.

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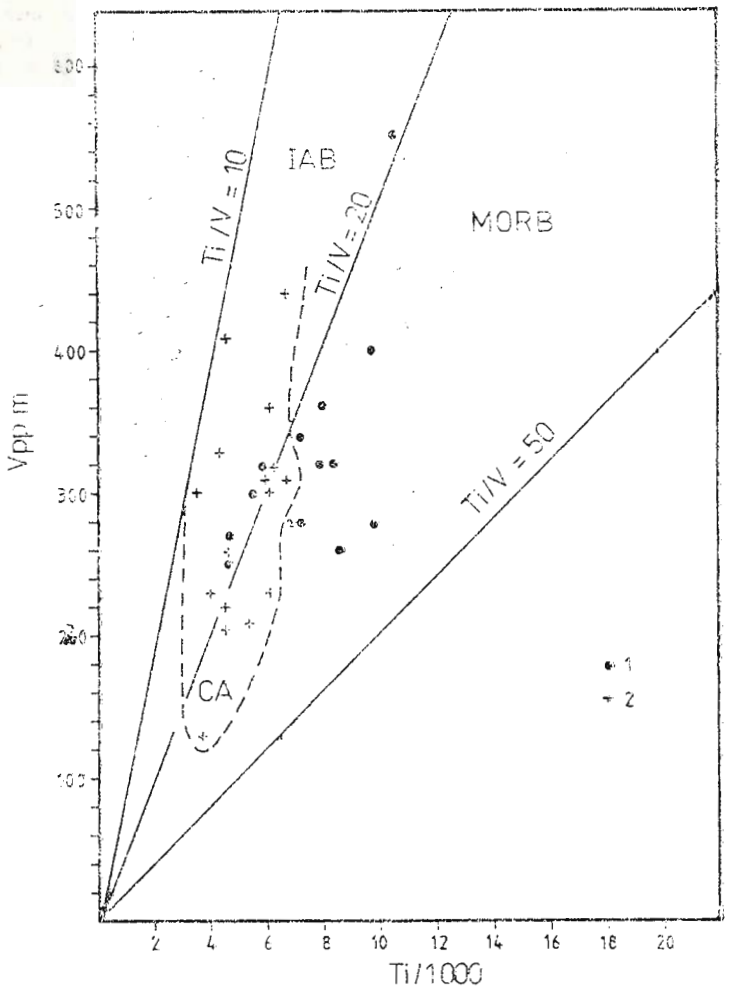

Fig. 3 — V—Ti/1000 diagram (Shervais, 1982).

1, ophiolitic rocks; 2, island arc volcanics; CA, field of island arc volcanics in the New Hebrides islands.

The diagram in Figure 7 shows that the content of P_2O_5 in almost half the Liassic basalts in the Vața—Căzănești area is even lower than the contents of this element in the MORB-type tholeiitic series studied by Hawkins (1980) in the "back arc" basins. This particularity emphasizes the character of ocean floor tholeiites of the basic rocks in the region. Therefore, similar rocks occurring in the west of the Mureș Zone have already been regarded as normal or "nondifferentiated" basalts (Savu et al., 1984 b) and supposed to originate in the parental magma — little differentiated — that yielded the ophiolitic rocks of the Mureș Zone. That was a basaltic magma with the chemical composition similar to that indi-

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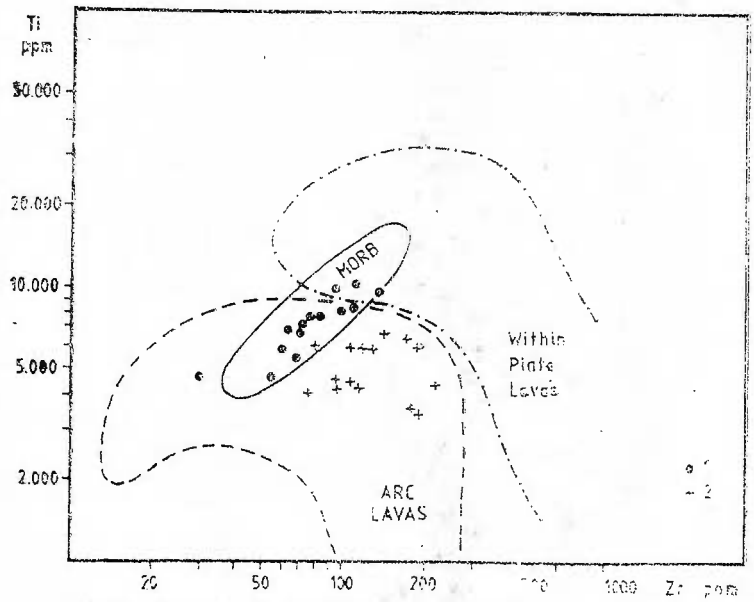

Fig. 4 — Ti—Zr diagram (Pearce, 1980).
1, ophiolitic rocks; 2, island arc volcanics.

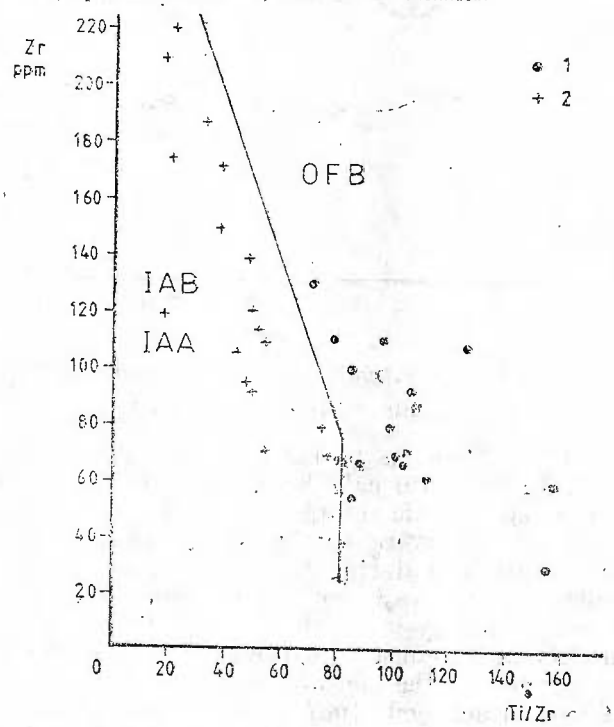

Fig. 5 — Zr—Ti/Zr diagram (Săvu).
1, ophiolitic rocks; 2, island arc volcanics.

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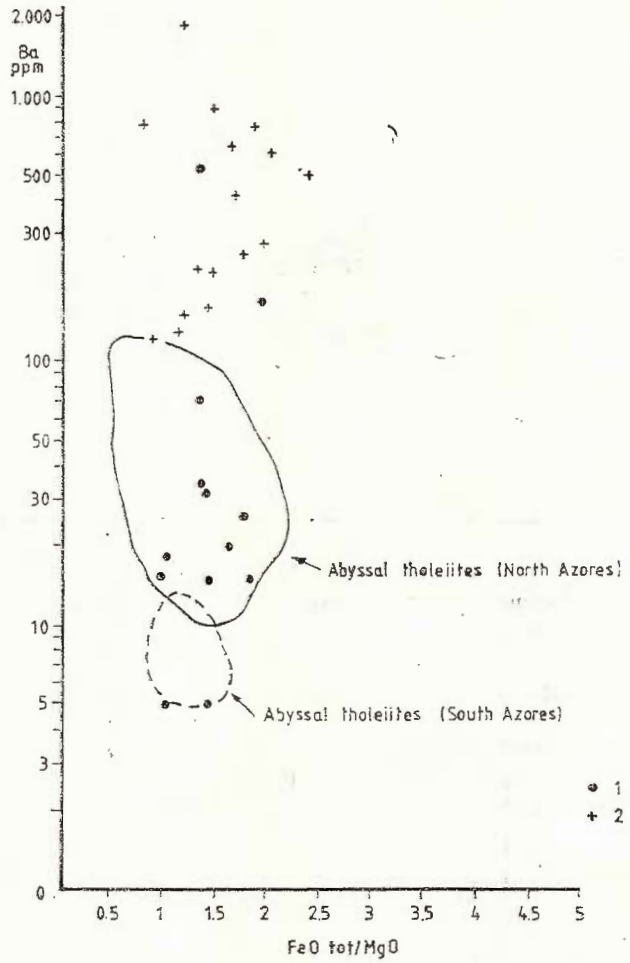

Fig. 6 — Ba—FeOtot/MgO diagram (Miyashiro, 1975).
1, ophiolitic rocks; 2, island arc volcanics.

cated by Fisk (1984; 1986) as parental magma for MORB because the values of the FeO/MgO ratio calculated from the chemical composition of the rocks from the basaltic complex in the Vaţa—Căzăneşti area vary between 0.22 and 1.15, the average value being 0.78. Due to the secondary transformations manifested during the ocean floor metamorphism, that led to the increase of the Na_2O content in some rocks, the value of the CaO/Na_2O ratio is generally lower than that indicated by the mentioned author for the MORB parental tholeiitic magma, its average being 3.70. Such a magma formed by the melting of at least 15% of the source from the mantle (Green, Ringwood, 1967). In two gabbro samples from the Almaş Sălişte body, situated in the ocean floor basalt complex, immediately south of the adjoining map, the $^{87}Sr/^{86}Sr$ ratio shows values of 0.7032

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and 0.7029, respectively, and the average value of the Rb/Sr ratio is 0.020 (Herz et al., 1974), small values similar with those from other ophiolitic rocks in the Mureș Zone, which correspond to the ocean-floor tholeiitic rocks originating in the upper mantle. The same significance have the very

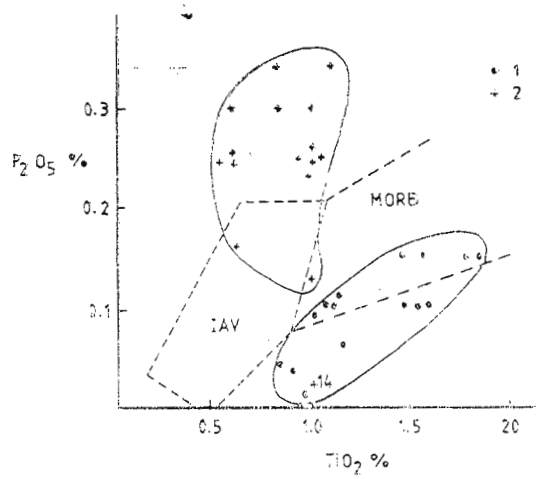

Fig. 7 — P_2O_5—TiO_2 diagram (Hawkins, 1980).
1, ophiolitic rocks; 2, island arc volcanics.

low U and Th contents from the basaltic rocks nearby the study area (Savu et al., 1986 c).

Petrography and Age of the Island Arc Volcanics

The Late Kimmerian island arc volcanics, unconformably overlying the ophiolitic basement (Pl. I), are represented by rocks of the lower complex (IAV_1) of the calc-alkaline volcanic series in the Mureș Zone: agglomerates, volcanic breccia (Pl. III, Fig. 2) and rarely flows of porphyritic basalts and basaltic andesites, often amygdaloidal (Tab. 2) and their tuffs (Pl. III, Fig. 4). Between them occur (Vățișoara Valley, Birtin Valley, Măgura Peak) intercalations of andesite agglomerates with large phenocrysts of augite and hornblende + augite (Pl. III, Fig. 3), the texture of some of them being almost similar with the texture of the Visca ankaramites. The texture of these rocks is mostly porphyritic, typical of the calc-alkaline volcanic rocks. They consist usually of a hyalopilitic matrix, more rarely pilotaxitic or fluidal, within which plagioclase phenocrysts are floating: poorly zoned labrador-andesine in basalts and andesine with obvious zonation in andesitic rocks; large phenocrysts of augite and hornblende (10—15 mm long) occur as well. In the Birtin Valley basin and especially in Arsurii Valley occurs a typical banded formation consisting of thin bands of basic tuffs with different granulation as thick alternations intercalated with scarce bands of coarser agglomerates.

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TABLE 2

Chemical composition of the Late Kimmerian island arc volcanics

No. *	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
SiO_2	44.86	44.90	45.00	45.27	47.59	48.79	48 89	48 94	49 24	50 97	51.68	52 96	53.75	55.10	55.46	57.8(
Al_2O_3	14.51	14.71	15 50	12.29	15 79	13.79	16.91	9 95	17.90	15 91	16 91	15 76	14.61	15.94	15.88	17.60
Fe_2O_3	3.82	3.36	3.43	2.81	5.49	3.32	3.58	2.54	3.48	4 29	2 41	2.89	3.71	4.87	3.63	3 9
FeO	5.49	5.20	3.65	4.31	3.15	4.42	4.43	2.59	4 47	3 49	5 92	3 54	2.47	3.84	2 38	2 1
MnO	0.16	0.15	0.16	0.17	0.15	0.12	0.14	0.19	0.27	0.13	0 15	0 14	0.12	0.15	0 07	0.0
MgO	11.95	7.53	4.77	7.90	4.28	5.19	5.62	4 26	4.90	4 54	4 83	4 95	3.10	4.60	1 22	5 1
CaO	7.96	11.66	13.50	10.08	13.55	13.60	9.93	17.87	9.47	8 93	9.25	5.83	8.89	7.68	8 88	5 6
Na_2O	2.30	1.89	3.01	2.38	3.84	2.90	4.39	1.62	3.07	3.26	2.72	6.26	3.06	3.19	3.89	3.1
K_2O	0.44	0.18	0.84	1.82	1.95	0.68	0.21	0.99	0.36	1.57	0 58	0.76	1.81	1.73	1.49	1 1
P_2O_5	0.25	0.24	0.12	0.25	0.24	0.26	0.20	0.24	0.34	0.23	0.16	0.27	0.30	0.02	0 30	0.3
TiO_2	1.12	0.62	1.02	0.90	1.02	0.76	0.76	0.68	1.12	1 02	0 76	1,00	0.76	1.02	0 58	0.7
CO_2	1.22	4.31	6.31	6.51	0.87	4.13	1.82	8.36	0.49	2.38	1 35	3.18	5.02	0 42	3.11	0
S	0.10	0.19	0.10	0.09	0.09	0.10	0.11	0.09	0.10	0.08	0.14	0.08	0.09	0.10	0.24	0.1
H_2O+	5.35	4.51	2.47	5.46	1.46	2.15	2.49	1.64	5.25	3.09	3.60	2.39	1.75	1.37	2 36	2.4
Total	99.53	99.45	99.88	100.24	99.47	100.21	99.48	99.96	100.43	99.89	100.49	100.01	99.44	100.03	99.49	100.4
Ni(ppm)	82	12	95	175	90	44	52	44	46	34	100	35	60	14	21	23
Co	38	13	34	36	31	26	32	22	27	27	48	24	22	26	28	40
Cr	70	6.5	420	330	300	160	150	410	145	10	320	52	195	14	50	10
V	240	130	230	210	300	220	410	230	310	360	260	310	205	320	300	330
Sc	39	14	37	31	41	32	37	30	37	36	37	29	20	31	43	43
Y	30	20	22	19	20	17	21	42	28	23	20	24	25	34	20	35
Yb	3.0	2,8	2.2	2.7	2.2	2.6	2.8	5.0	3.4	3 1	3.0	3.2	2.4	3 5	2.2	4.2
La	34	58	<30	<30	40	<30	<30	46	35	33	<30	30	33	46	40	<30
Nb	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10
Zr	140	175	80	70	115	95	105	75	170	110	95	120	220	190	190	115
Ba	780	130	220	120	280	900	160	150	650	420	250	220	610	760	950	1800
Sr	420	800	620	240	650	>1000	550	400	1000	650	600	430	>1000	>1000	>1000	>100
Pb	6.5	13	3	6	8	3 5	3	5	4	9	5	4	4	17	14	<2
Cu	60	31	18	12	50	31	85	52	42	30	50	75	16	24	34	30
Ga	15	21	10	12	20	20	19	12	20	30	16	16	15	18	18	11
Sn	<2	<2	<2	<2	<2	<2	2.5	<2	<2	<2	<2	<2	<2	2.5	<2	<2

* The analyses in this table represent: porphyritic basalts with augite phenocrysts; 2, Strimbu Brook; 3, 9, Vățișoara Valley; 7, 8, Măgureaua-Vaței; 1, 6, Birtin Valley; 4, Arsurii Brook; 10, Țebea Valley; amygdaloidal porphyritic basalts; 11, Măgureaua Vaței; 12, Birtin Valley; basalt xenoliths in banatitic (Laramian) granodiorite; 5, Birtin Valley; basic tuff; 13, Birtin Valley; oligophyres; 14, Surăcasa Valley; hornblende andesites; 15, Birtin Valley; 16, Vățișoara Valley.

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Around the dyke-like Laramian (banatitic) intrusion in the southern part of the Măgureaua Vaței area these rocks have been metamorphosed at the contact (Pl. I) and changed into basic hornfelses with diopside, hornblende and biotite (Cioflica, 1960; Jude et al., 1973; Ștefan et al., 1982; Savu et al., 1987).

Numerous veins, dykes and small bodies of oligophyres, orthophyres (paleotrachytes) and andesites, rocks belonging to the upper complex (IAV_2) of the Late Kimmerian volcanic series, are crossing both the ophiolitic basement and the mass of the lower complex volcanics (IAV_1).

South of the study region, between the lower complex pyroclastics (IAV_1) occur jaspers in which Dumitrică (oral communication) determined a radiolarian assemblage indicating the end of the Callovian. It results that the Late Kimmerian volcanism is postophiolitic. It started towards the end of the Middle Jurassic and lasted till the end of the Neocomian, as pointed out by the age of a basaltic vein intruding the Almaș Săliște gabbroic body of ca 120 MA (Herz et al., 1974). The lower complex (IAV_1) formed mainly during the Upper Jurassic as it often underlies Tithonian limestones, e.g. those in the study area (Pl. I). It is of note that outside Mureș Zone the activity of the Late Kimmerian (postophiolitic) calc-alkaline volcanism was also emphasized within the Vardar Zone (Tethys area; Rocci et al., 1980) and the Himalaya (Dietrich et al., 1983).

Geochemistry and Origin of Island Arc Volcanics

The chemical composition of the island arc volcanics (Tab. 2) characterizes them as calc-alkaline rocks. The SiO_2 content (45—58%) points out that the rocks analysed are specific to the lower complex (IAV_1), being represented by porphyritic basalts, basaltic andesites and hornblende-augite andesites.

On the diagram in Figure 1 [4] all the Late Kimmerian volcanic rocks plot in the field CA. Nevertheless, the value of the FeO_{tot}/MgO ratio which is considered to indicate the magma types that yielded the rocks (Miyashiro, 1975) is, with one exception, less than 2. It shows that the Late Kimmerian volcanics of the lower complex (IAV_1) display features close to the island arc tholeiites. In these rocks the value of the CaO/Al_2O_3 ratio varies from 0.44 to 0.93, the average value being 0.68. Generally their sodium contents is higher than the potassium one as pointed out by the value of the K_2O/Na_2O ratio, always smaller than 1, the average value being 0.35. The FeO/Fe_2O_3 ratio presents values varying within quite large limits (0.55—2.43), the average value being 1.14. All the mentioned characteristics point (acc. to Jakeš and White, 1972) to volcanic rocks formed in subduction zones.

The trace elements (Tab. 2) of the Late Kimmerian volcanics also show wide variation limits, although this paper deals with the rocks in the first part of the volcanic series, e.g. porphyritic basalts and intermediary rocks (partially). This trend is already illustrated for V on the diagram in Figure 3 on which Ti ranges within close limits and V within wider limits. Consequently the plotting fields extend from IAB field to the lower part of the MORB field, a situation similar to that of the por-

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phyritic basalts in the northern island arc of the Mureş Zone (Savu et al., 1986 a), a zone in which also plot the island arc tholeiitic basalts of the New Hebrides.

Zirconium susceptible to the magma differentiation (Savu, Udrescu, 1975) displays contents varying between 70 and 220 ppm. On the diagram in Figure 4 (Pearce, 1980) the rocks plot in the upper part of the island arc lava field. On the Zr—Ti/Zr diagram in Figure 5 the volcanic rocks plot in the IAB + IAA field, in a way that points out the differentiation trend of the parental calc-alkaline magma parallel to the increase of the Zr content. It is of note that the andesites with augite and hornblende phenocrysts in the Văţişoara Valley display Th contents much higher than the other Late Kimmerian volcanics in the area (Savu et al., 1986 c). Likewise La occurs in most of the Late Kimmerian volcanic rocks both in more basic terms and in intermediary ones, usually displaying higher values than those specific to the island arc tholeiites whose average value has been calculated by Werner (1984).

Barium presents contents higher than 100 ppm, characteristic of the island arc volcanics. They vary within wide limits (120—1800 ppm) as shown on the diagram in Figure 6.

Strontium shows contents varying from 400 to 1000 ppm, much higher than those in the volcanics yielded by the southern island arc of the Mureş Zone (105—430 ppm) (Savu et al., 1984 a). Mention should be made that some of the island arc basalts in the west of the Mureş Zone (Savu et al., 1986 a) display higher Sr contents, close to those in the Căzăneşti-Ţebea region. In a dyke intruding the Almaş Sălişte gabbroic body the value of the $^{87}Sr/^{86}Sr$ ratio is 0.7043 pointing to its origin in an island arc calc-alkaline magma, like the other Late Kimmerian volcanics.

The parental magma of the Late Kimmerian island arc volcanics was a calc-alkaline one with a tholeiitic tendency and showed a more alkaline character than the parental magma of the island arc volcanic series studied by Hawkins (1980) (see the diagram in Figure 7 on which most of the rocks fall above the IAV field, at higher contents of P_2O_5). According to the characteristics of the Late Kimmerian volcanics the parental magma formed either as a result of the partial melting of the Mureş Ocean crust (Savu, 1983) subducted on the Benioff plane — whose tholeiitic rocks were first transformed into eclogites — or from the upper mantle situated above the Benioff plane, strongly influenced by solutions and gases released from the subducted oceanic crust (Roden, Murthy, 1985) that was being transformed into eclogite. This magma was involved in an intense process of fractional differentiation, as pointed out by the Eu negative anomaly and the decrease of the content of HREE of the volcanic series (Savu et al., 1986 c), whose REE pattern is rather similar with that of the high-K orogenic andesites (Gill, 1981).

Conclusions

In the study region in the north of the Mureş Zone, lying between Vaţa, Căzăneşti and Ţebea, occur two series of Mesozoic eruptive rocks: (1) the series of Liassic ophiolitic rocks constituting the basement of the Căpîlnaş-Techereu Unit and (2) the series of Late Kimmerian island arc

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volcanics — postophiolitic — formed concomitantly with the J_3—Cr_1 flysch, which overlies the ophiolitic rocks.

Ophiolitic rocks represent a fragment of the obducted Mureș Ocean crust, formed of basalts, anamesites, rocks with an intersertal texture, variolites, dolerites and gabbros. These rocks display petro- and geochemical features of ocean floor tholeiitic rocks.

Ophiolites originate in a tholeiitic magma formed in the upper mantle by the melting of at least 15 per cent of the ultramafic source from the mantle.

Island arc volcanics are represented by porphyritic basalts, basaltic andesites and augite + hornblende andesites, rocks with a mostly porphyritic texture.

From the petro- and geochemical point of view they are calc-alkaline rocks with a tholeiitic tendency. They come from a basaltic calc-alkaline magma formed as a result of the partial melting of the subducted Liassic oceanic crust or from the upper mantle situated above the Benioff plane, strongly influenced by solutions and gases released from the subducted oceanic crust.

The parental magma of the Late Kimmerian volcanics was involved in an intense process of fractional differentiation that led to the formation of a wide variety of volcanic rocks.

[4] For economy of printing space island arc volcanics occur on the same diagrams with ophiolites.

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PETROLOGIA OFIOLITELOR LIASICE ȘI A VULCANITELOR DE ARC INSULAR NEOKIMMERICE DIN REGIUNEA VAȚA — CĂZĂNEȘTI—ȚEBEA (ZONA MUREȘ) ȘI CONȚINUTUL LOR DE ELEMENTE MINORE

(Rezumat)

Regiunea prezentată în lucrare se situează în partea de nord a zonei Mureș, între Vața, Căzănești și Țebea (pl. I). În cuprinsul ei se deosebesc două serii principale de roci eruptive mezozoice și anume: (1)

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o serie de roci ofiolitice liasice care formează fundamentul unității de Căpîlnaș-Techereu și (2) o serie de vulcanite de arc insular neokimmerice — postofiolitice — care s-au format concomitent cu flyschul J_3—Cr_1 și care acoperă discordant rocile ofiolitice.

Rocile ofiolitice aparțin complexului (O_1) și se dezvoltă în partea de vest a regiunii, între Vața de Sus și Căzănești. Ele reprezintă un fragment din crusta Oceanului Mureș (Savu, 1983) — care a fost dezmembrată — și constau din bazalte și variolite, de obicei în facies de pillow-lava la care se adaugă rar dolerite și gabbrouri.

Rocile ofiolitice au caracteristici petro- și geochimice de tholeiite (fig. 1) de fund oceanic (fig. 3, 4, 5, 6), ce au provenit dintr-o magmă parentală formată în mantaua superioară prin topirea a cel puțin 15% din materialele ultrabazice ale acesteia. Ele aparțin tipului de ofiolite „high-Ti" (fig. 2).

După cum rezultă din figura 7, rocile ofiolitice din regiunea Vața-Căzănești sînt chiar mai sărace în P_2O_5 decît seriile tholeiitice de tip MORB din bazinele de „back arc", fapt care subliniază proveniența lor dintr-o magmă tholeiitică de fund oceanic.

Vulcanitele de arc insular neokimmerice se dezvoltă mai ales în partea de est a regiunii (pl. I) între Vața de Sus și Țebea. Ele sînt reprezentate în special prin roci din complexul (IAV_1), cum sînt: bazalte porfirice, andezite bazaltice și andezite cu hornblendă și piroxeni, roci a căror structură este prin excelență porfirică. Ele sînt roci calco-alkaline (fig. 1) cu tendință tholeiitică (vezi valoarea raportului FeOtot/MgO).

Vulcanitele neokimmerice sînt roci tipice de arc insular, cum rezultă din figurile 3, 4, 5 și 6. Ele sînt mai bogate în P_2O_5 decît alte vulcanite de arc insular cercetate de Hawkins (1980, fig. 7), ceea ce indică o tendință alcalină.

Vulcanitele de arc insular neokimmerice au provenit dintr-o magmă bazaltică calco-alcalină formată prin topirea crustei Oceanului Mureș, subdusă, sau din mantaua superioară situată deasupra planului Benioff, care a fost puternic influențată de soluții și gaze emanate din crusta oceanică subdusă. Această magmă parentală a fost implicată într-un intens proces de diferențiere fracționată, după cum rezultă din anomalia negativă a Eu și din sărăcirea rocilor în HREE. Acest proces a condus la formarea unei mari varietăți de roci vulcanice.

EXPLANATION OF PLATES

Plate II

Fig. 1 — Pillow lava separations in ocean floor basalts in the Vața Valley, at Vața de Sus.
Fig. 2 — Hyalobasalt. Surăneasa Valley, parallel nicols, × 30.
Fig. 3 — Basalt. Măgureaua Summit, parallel nicols, × 30.

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Plate III

Fig. 1 — Anamesite with intersertal structure. Căzănești, crossed nicols, × 25.

Fig. 2 — Volcanic breccia in the base of the island arc volcanics of the lower complex (IAV_1), nearby the boundary with the basement ophiolites. Vața Valley.

Fig. 3 — Porphyritic basalt with augite phenocrysts. Birtin Valley, parallel nicols, × 17.

Fig. 4 — Basaltic tuff. Măgureaua Summit, parallel nicols, × 25.

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H. SAVU

SIMPLIFIED GEOLOGICAL MAP OF THE VAȚA–CĂZĂNEȘTI–ȚEBEA REGION

(AFTER VORȚA AND HĂLMAGIU GEOLOGICAL MAPS SC. 1:50.000 AND AUTHOR'S DATA)

0 1 2 km

Basarabasa
Vața de Jos
Vața
Crișu Alb
Vața de Sus
Birtin
Cărăstău
Baia de Criș
Tătărăștii de Criș
Țebea
Căzănești
v. Cerbaia
Vf. Măgureaua 809
Dl. Urzica
Vf. Corac 833
Vf. Matu 859
v. Pantofului
Vf. Pitiorii 760
Vf. Sirenu 809
v. Strîmbu
Părosa
Deleni

LEGEND

- Alluvial and terrace deposits
- Neogene volcanics
- Neozoic sedimentary deposits
- a, Banatitic intrusive rocks; b, thermic contact zone.
- Upper Aptian, Albian and Senonian sedimentary formations; a, olistoliths of Neokimmerian volcanic rocks
- Neojurassic limestones (J_3)
- Neokimmerian island arc volkanics; a, agglomerates of porphyritic basalts and basaltic andesites; b, piroxene- and hornblende bearing andesites; c, tuffs
- Jurassic ophiolites (J_1)-ocean floor rocks: a, ocean floor basalts complex; b, gabbro intrusions; c, picritic peridotite intrusions
- Fault
- Unconformity
- Normal geological boundary
- Synclinal axis
- Anticlinal axis
- O_4 Chemical and spectral analysis

NW I — SE I'

v. Vața — vf. Măgureaua — Dl. Pitiorii

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AGE DES GRANITOÏDES DE CODRU (MONTS APUSENI) [1]

par

NICOLAE STAN [2]

Granitoid. Upper Proterozoic. Lower Proterozoic. Absolute age. K/Ar. Sandstone. Permian. Apuseni Mountains — North Apuseni Mountains — Codru-Moma Mountains.

Résumé

La notion de „granitoïdes de Codru" a désigné tant les roches granitoïdes des monts de Codru que les granitoïdes migmatiques associés aux roches basiques des monts de Bihor et Muntele Mare. Etant donné que les granitoïdes des monts de Codru traversent des formations permiennes, on propose pour ceux-ci le nom de granitoïdes de Codru, tandis que pour les intrusions migmatiques des monts de Bihor et Muntele Mare, d'âge précambrien-cambrien (mesures radiométriques K—Ar) le nom de granitoïdes d'Arieș-Iara.

Abstract

Age of the Codru Granitoids (Apuseni Mountains). The notion "Codru granitoids" referred both to granitoid rocks in the Codru Mts and to the migmatic granitoids associated with basic rocks in the Bihor and Muntele Mare Mts. As in the Codru Mts, for these granitoids it is proposed the denomination of Codru granitoids and for the migmatic intrusions in the Bihor and Muntele Mare Mts of Precambrian-Cambrian age (K—Ar radiometric age) the denomination Arieș-Iara granitoids.

Le concept de granitöides de Codru renferme tant les magmatites intrusives des monts de Codru que les intrusions acide-migmatitiques associées aux schistes cristallins, parfois basiques des monts de Bihor et

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Muntele Mare (Bleahu et al., 1984; Dimitrescu et al., 1974; Hârtopanu et al., 1982).

Les nouvelles recherches montrent que les granitoïdes des monts de Codru ne peuvent pas être corrélés avec les granitoïdes des monts de Bihor et Muntele Mare, ayant des âges différents.

Les granitoïdes des monts de Codru, qui affleurent au nord-ouest de Moneasa sont connus dans la littérature géologique de la première moitié du XX-ème siècle. Palfy et Rozlozsnik (1939) les a attribué l'âge précarbonifère à cause de l'idée que les formations sédimentaires les plus vieilles, permien-inférieures, reposent transgressivement tant sur les schistes cristallins que sur les roches granitoïdes aussi.

L'idée de la position discordante des formations sédimentaires par rapport au soubassement cristallin-granitique a été maintenue tant sur la carte géologique à 1 : 100.000, feuille de Moneasa (Bleahu, 1965) que sur la carte géologique à 1 : 200.000, feuille de Brad (Bleahu et al., 1967) où il était précisé que les roches granitoïdes et les formations cristallophylliennes associées sont d'âge protérozoïque. Cette interprétation a résulté de la corrélation avec les granitoïdes migmatitiques des monts de Bihor et Muntele Mare pour lesquels on a obtenu, en 1966, des âges radiométriques K—Ar de 547 m.a. (Ianovici et al., 1969; analystes : Ionescu et Tănăsescu). Plus tard, en 1975, on a récalculé cet âge et on a obtenu 524 m.a. (Ianovici et al., 1976). D'autre 14 déterminations effectuées aussi sur les granitoïdes des monts de Bihor et Muntele Mare ont indiqué des valeurs comprises entre 344 et 158 m.a. (Ianovici et al., 1976). On a arrivé à la conclusion que les intrusions des monts de Codru aussi, synorogènes avec les schistes cristallins associés sont d'âge précambrien supérieur et plus probable paléozoïque inférieur. Les schistes cristallins et les roches granitoïdes des monts de Codru apparaissent sur la carte géologique à 1 : 50.000, feuille de Dumbraviţa (Bleahu et al., 1984) comme ayant l'âge paléozoïque inférieur.

Pendant l'été de 1985, travaillant pour la feuille de Sebiş à 1 : 50.000 j'ai découvert un affleurement dans le versant droit de la vallée de Teuz, approximativement 4 km (nord) en amont de Susani (fig. 1). Ici les granitoïdes des monts de Codru traversent les formations sédimentaires permiennes représentées par les grès feldspathiques (fig. 2) [3].

Les grès feldspathiques jaunâtres, pséphitiques, stratifiés parfois, comportent du quartz détritique (5 à 7%), feldspath détritique séricitisé (70 a 80%), biotite et biotite séricitisée ou opacifiée (5 à 10%). Ce type de roches peut apparaitre aussi dans les dépôts permiens tant au niveau de la série vermiculaire qu'à celui des grès feldspathiques aussi.

Le granitoïde blanc-grisâtre a une structure holocristalline, est fréquement altéré supergène. Il présente la suivante composition minéralogique : quartz (20 à 35%), albite-oligoclase partialement séricitisé (30 à 40%), microcline-perthite (10 à 15%), muscovite et biotite (5 à 7%). C'est un leucogranite. Au contact entre les roches granitoïdes et celles sédimentaires font leur apparition quelques minéraux de néoformation : la biotite à pléochroïsme brun-rougeâtre qui forme souvent des auréoles millimétriques de réaction et l'épidote sporadiquement diffusée dans la masse de la roche.

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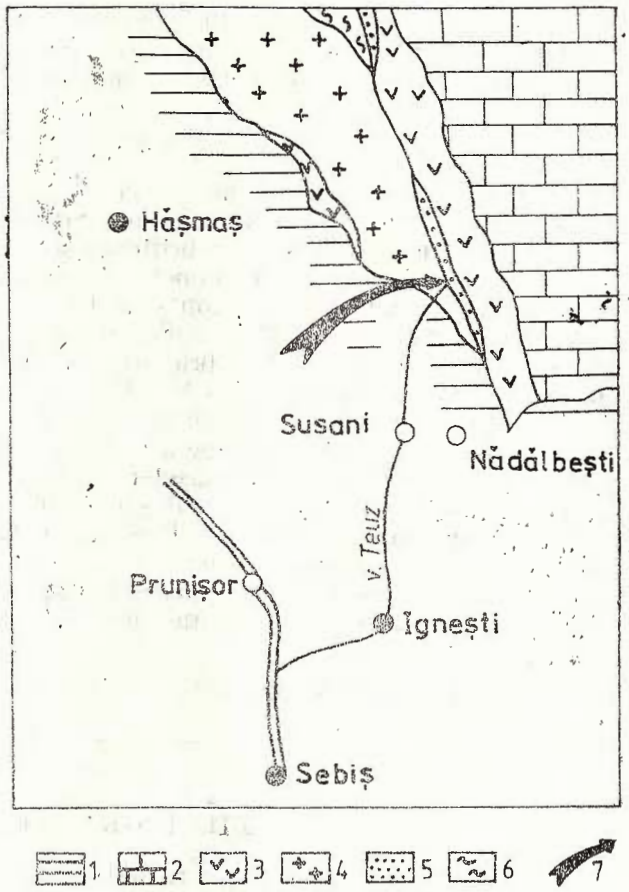

Fig. 1 — Schéma géologique de la zone de Susani-Nădălbești.

1, Quaternaire et Néogène ; 2, Mésozoïque ; 3, rhyolites permiennes ; 4, granitoïdes permiens ; 5, grès feldspathiques permiens ; 6, cristallin ; 7, localisation de l'affleurement.

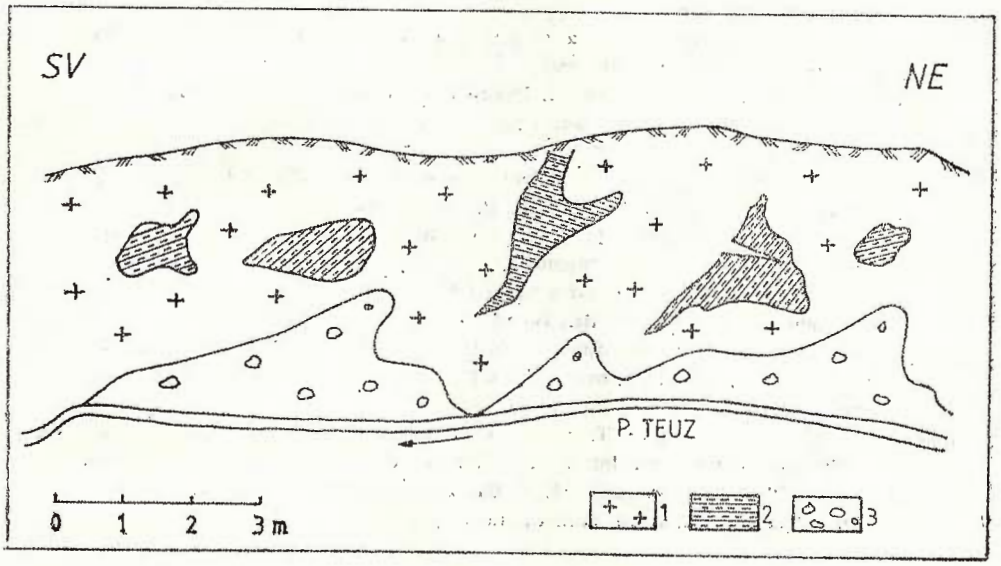

Fig. 2 — Relations entre les roches granitoïdes et les grès feldspathiques permiens.

1, granitoïdes de Çodru ; 2, grès feldspathiques permiens ; 3, éboulis de roches.

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Rarement on observe un mélange intime entre les minéraux sédimentaires détritiques et les minéraux magmatiques représentés par le quartz, albite, microcline-perthite, muscovite et biotite.

Conclusions

a) Les granitoïdes des monts de Codru ont l'âge permien. Ils sont plus vieux que les filons rhyolitiques par lesquels sont traversés et les écoulements de rhyolites ignimbritiques qui les recouvrent; ces derniers sont aussi intercalés par des dépôts sédimentaires permiens.

b) Les granitoïdes des monts de Codru ne peuvent pas être corrélés avec les intrusions et les migmatites des monts de Bihor et Muntele Mare, plus anciennes, précambrien-cambriennes, dont l'âge est déterminé par les mesures radiométriques K—Ar.

c) Etant donné que le concept de granitoïdes de Codru renferme tant les roches granitiques d'âge permien que les granitoïdes migmatiques précambrien-cambriens, on considère nécessaire de faire la description des deux groupes de roches sous des noms différents. Nous proposons que pour toutes les intrusions semblables aux granitoïdes de leur zone typique d'affleurement on emploie le nom de granitoïdes de Codru, en tant que pour les intrusions acides migmatiques des monts de Bihor et Muntele Mare on emploie le nom de granitoïdes d'Arieș-Iara.

[3] Cette affleurement a été ultérieurement visité avec dr. M. Ștefănescu aussi.

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VÎRSTA GRANITOIDELOR DE CODRU (MUNȚII APUSENI)

(Rezumat)

Noțiunea de granitoide de Codru cuprinde atît magmatitele intrusive din munții Codru cît și intruziunile acide-migmatice asociate cu șisturi cristaline, uneori bazice, din munții Bihor și Muntele Mare. Ele au fost considerate de aceeași vîrstă.

Deoarece granitoidele din munții Codru străbat formațiunile permiene, ele nu mai pot fi paralelizate cu acelea din munții Bihor și Muntele Mare, pentru care măsurătorile radiometrice K-Ar arată vîrste precambrian-cambriene.

Se propune denumirea de granitoide de Codru pentru cele din munții Codru (permiene) și denumirea de granitoide de Arieș-Iara pentru cele din munții Bihor și Muntele Mare (precambrian-cambriene).

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PETROLOGIA ROCILOR MAGMATICE

BASE SURGE DEPOSITS IN THE CIOMADUL MASSIF (SOUTH HARGHITA MOUNTAINS) [1]

by

ALEXANDRU SZAKÁCS [2], IOAN SEGHEDI [2]

Pyroclastics. Base surge. Lapilli. Neogene. Dacitic composition. Eruptions. Bedding. East Carpathians — Neogene eruptive-Quaternary — Harghita Mountains.

Abstract

Base surge deposits with accretionary lapilli occurring in the Ciomadul massif in the south Harghita Mts are first described in Romania. These deposits are the result of phreatomagmatic eruptions in the first phase of the dacitic volcanic activity in the study zone.

Résumé

Dépôts pyroclastiques de type „base surge" dans le massif de Ciomadul (monts de Harghita de sud). On décrit, pour la première fois dans la Roumanie, les dépôts pyroclastiques de type „base surge" à lapillis accrétionnaires situés dans le massif de Ciomadul, au sud des monts de Harghita. Ces dépôts sont engendrés par des certaines éruptions phréatomagmatiques de la première phase de développement du volcanisme dacitique de la zone.

Introduction

Significant results obtained in volcanology during the last years made possible a better understanding of the complex phenomena related to the explosive volcanic activity. It opened the way to the reconstitution of the explosive volcanism on the basis of the characteristics of its products as well as the way to a more accurate genetic systematization

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of the pyroclastic deposits. An excellent synthesis of these results is given in Fisher and Schmincke's "Pyroclastic Rocks" (1984).

At present, the following genetic classification of the pyroclastic deposits is admitted: (1) pyroclastic fall deposits; (2) pyroclastic flow deposits and (3) pyroclastic surge deposits.

The recognition of the base surge deposits (a category of the pyroclastic surge deposits), of the volcanic phenomenon that has generated them, respectively, is relatively recent, brought about by the 1965 eruption of the Taal volcano (Philippines) (Moore et al., 1966 in Fisher, Schmincke, 1984). In volcanic eruptions base surge represents the blast carrying ash particles as diluted suspension skimming the ground around the eruption centre, usually occurring during the phreatomagmatic eruptions. The sedimentation of the solid charge generates pyroclastic deposits displaying some characteristics of eolian deposits. Such types of deposits were identified in numerous zones with recent or older volcanism. Their occurrence in the Ciomadul massif was evidenced on the occasion of the Second Geological Symposium held at Gheorghieni (Szakács et al., 1984, unpublished paper).

Geological Setting

The dacitic explosive-extrusive volcanism in the Ciomadul massif developed on a basement represented by the Cretaceous flysch and by older Pliocene pyroxene andesites. During the first phase the pyroclastic cone was built as a result of the explosive activity of two known eruptive centres — Sfînta Ana and Mogoş. The pumice-rich deposits constituting the cone belong to all genetic types, the surge-type ones being among the last ones. The loose tephras were redistributed towards the base of the cone and outside it, through lahars and by erosional reworking. During the second stage domes and short lava flows occurred on the flanks of the pyroclastic cone and outside it, under tectonic control.

Volcanics consist of hornblende and biotite dacites and hornblende biotite, pyroxene and quartz dacites, displaying a high-K character (Szakács, Seghedi, 1986).

Base Surge Deposits

The base surge deposits crop out in an artificial exposure, nearby the road leading to the Sfînta Ana Lake, in the vicinity of the north-eastern border of the eruption centre in the Mohoş swamp zone. They are fine-grained unconsolidated tephras, consisting of lithic fragments of dacites, pumice, volcanic glass and crystals (mainly plagioclase, hornblende and biotite).

This deposit is characterized by a dune-type bedding (Pl. I, Fig. 1). The stoss side displays fine laminas dipping 5—10° towards the source. The lee side laminas are thicker, generally plan-parallel, locally oblique, with dips of 18—20° (Fig. 1). The visible height of the dune-type structure is of about 1.5 m.

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The outcrop displays, in fact, a longitudinal section through the "dune" so that one can observe the shifting of the crest concurrently with the deposition of newer beds, giving rise to a dip axial plane of about 40°. According to Fisher and Schmincke the mentioned characteristics point out a low flow regime unlike the deposits emplaced in high flow regime, characterized by antidune-type sedimentary structures.

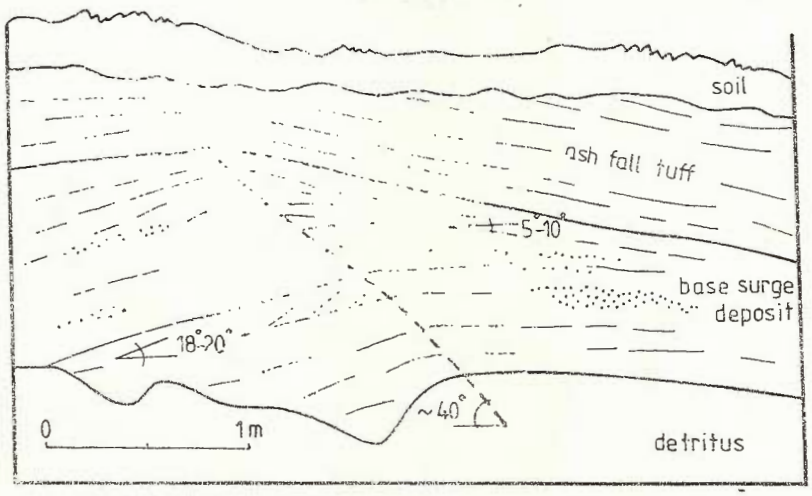

Fig. 1 — Sketch of the outcrop with base surge deposits (after the photo in Plate I, Fig. 1).

The deposit presents a certain grading of the pyroclastic material. The coarser material, rich in lithic fragments, forms thin levels or elongated lenticular "trains" (Fig. 1). The granulometric features of the deposit are shown in Figure 2 as a cumulative curve. The average diameter (in Ø units) is +0.4 and the sorting coefficient (~Ø) is 1.7. These parameters are in agreement with the closeness to the source (less than 200 m); similar deposits of the Taal volcano (Philippines) point out a more reduced granulation and a better grading at gradually increasing distances from the source (Waters, Fisher, 1971, mentioned by Fisher, Schmincke, 1984).

The presence of accretionary lapilli [3] represents another feature of the base surge deposits in the Ciomadul massif. A large number of accretionary lapilli is found within a well-developed level of ca 15—20 cm thick as well as in some thinner levels (1—3 cm). Although they occur also in the stoss side their abundance is characteristic of the lee side of the dune structure. Their size varies from some millimetres to about 1 cm. They display a strong spherical shape (Pl. II, Fig. 1). They often show concentrical internal structure (Pl. II, Fig. 2). Accretionary lapilli represent small agglomerations of fine moist ash, formed during the explosive volcanic eruptions in conditions of liquid or steam water abundance.

The mere presence of the accretionary lapilli does not have diagnostic value for phreatomagmatic eruptions; however, their association with base surge deposits is quite relevant for this type of volcanic activity.

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The hypothesis of a phreatomagmatic eruptive process for the formation of base surge deposits with accretionary lapilli in the Ciomadul massif represents the most plausible explanation of the field evidence, indicating the water abundance in the eruption column and a massive production fine tephras, deposited in the source vicinity.

In the examined outcrops the base surge deposit is overlain by air-fall tuffs some decimetres thick (Fig. 1).

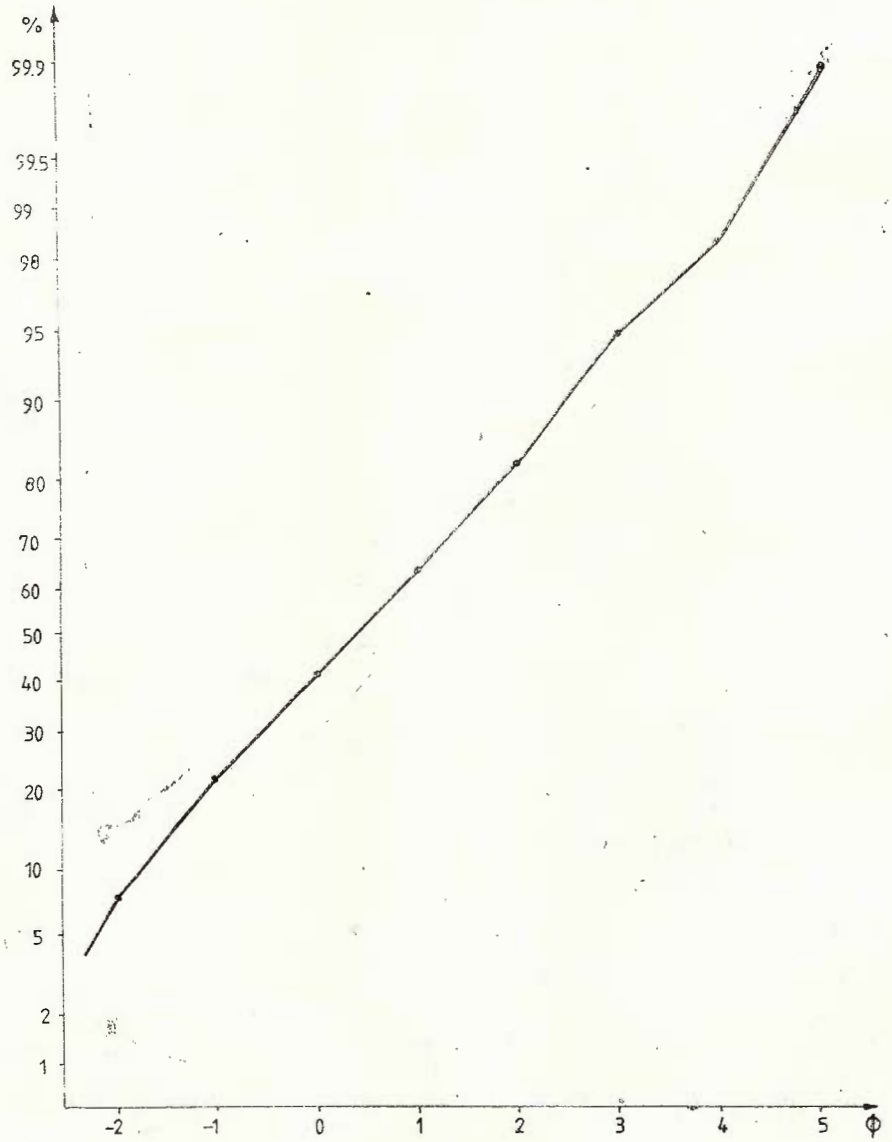

Fig. 2 — Granulometric cumulative curve of the base surge deposits; MdØ=0.4; ~Ø=1.7.

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Our observations indicate a possible wider development of the base surge deposits in the Ciomadul massif, showing more and more attenuated characters (wave height and length of the dune-type structures, grain-size, etc.) as we move away from the source. Likewise, the presence of base surge deposits in other areas of the Călimani-Gurghiu-Harghita eruptive chain, especially in connection with the first stage of volcanism, is also likely.

[3] The presence of accretionary lapilli was mentioned in Romania only in the Carboniferous (?) acid volcanics in the Carapelit Formation of North Dobrogea (Seghedi, Szakács in Seghedi et al., 1987).

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DEPOZITE PIROCLASTICE DE TIP „BASE SURGE" ÎN MASIVUL CIOMADUL (MUNȚII HARGHITA DE SUD)

(Rezumat)

Într-o deschidere din imediata apropiere a centrului de erupție Mohoș aflorează depozite piroclastice dacitice de tip „base surge" cu caractere tipice : granulație redusă (Md Ø = +0,4), sortare moderată (~Ø = 1,7), structuri de tip dună cu lamine paralele și oblice la unghi mic (5—10° pe partea „în curent", 18—20° pe partea „sub curent"). Este caracteristică prezența abundentă a lapililor acreționari, mai ales pe partea „sub curent" a structurii de dună. Aceste caractere tipice denotă proveniența depozitelor de „base surge" cu lapili acreționari în urma unor procese hidroclastice (erupții freatomagmatice).

Depozite de val piroclastic (pyroclastic surge deposits) sînt descrise pentru prima dată, cu această ocazie, pe teritoriul României.

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EXPLANATION OF PLATES

Plate I

Fig. 1 — Outcrop in base surge deposits near Mohoș swamp.
Fig. 2 — Accretionary lapilli tuff in base surge deposit (lee side).

Plate II

Fig. 1 — Accretionary lapilli detached from base surge deposit.
Fig. 2 — Concentrical structure in accretionary lapillus.

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PETROLOGIA ROCILOR MAGMATICE

VOLCANIC BOMBS AND BLOCKS IN HARGHITA MTS [1]

by

ALEXANDRU SZAKÁCS[2], CSABA JÁNOSI[3]

Volcanic rocks. Dacites. Andesites. Volcanoclastics. Neogene. East Carpathians — Neogene eruptive-Quaternary — Harghita Mountains.

Abstract

Morphology and structure of dacitic and andesitic polygonal volcanic bombs and blocks from occurrences in two areas of Harghita Mts (Ciomadul massif in South Harghita and west Dănești in North Harghita) are described. A great number of bombs and blocks show bread-crust cracks and chilled margin. Bombs with radial impact cracks for the first time are pointed out.

Résumé

Bombes volcaniques des monts de Harghita. On fait, pour la première fois dans notre pays, une description morphologique et structo-texturale des bombes volcaniques polygonaux à composition dacitique et andésitique localisées en deux occurrences des monts Harghita (massif de Ciomadul et la zone située à l'ouest de la localité de Dănești). Une grande partie des bombes présente des fissures de type „croûte de pain" et un bord vitreux. On signale, en première, la présence de quelques bombes à fissures radiaires d'impact.

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Introduction

Bread-crust type volcanic bombs were for a long time known in Ciomadul massif, south-east of South Harghita Mts (Cholnoky, 1922 cited by Bányai, 1964). There are some specimens in the Geology-Mineralogy Department's collection within Babeș-Bolyai University in Cluj-Napoca, originating from this region. They did not subject to any volcanological study until now.

Basaltic volcanic bombs were described by Peltz (in Mihăilă et al., 1972 and Mihăilă, Peltz, 1977) within Pleistocene volcanic products in Perșani Mts.

In the neogene volcanic areas in Romania, Ciomadul massif and Perșani Mts constitute the only known occurrences of volcanic bombs [4].

This paper proposes to describe dacitic volcanic bombs and blocks in Ciomadul massif and andesitic ones from an unknown occurrence in North Harghita Mts.

Geological Setting

Neogene volcanism in Harghita Mts as well as in the whole Călimani-Gurghiu-Harghita volcanic chain took place in two main stages: (1) a prevailing explosive stage during which the lower volcanic compartment (the volcano-sedimentary formation in its peripherical parts) was constituted and (2) another mainly effusive stage building the actual volcanic edifices (upper volcanic compartment) of the axial zones (Rădulescu et al., 1981). Products of the first stage are represented by andesitic and dacitic fallout tuffs, pyroclastic flow deposits, mud flow (lahar) deposits and subordinately lava flows. Lava flows and domes with minor interbedded pyroclastics, mainly andesitic in composition, constitute volcanic structures resulted during the second stage of volcanism.[3]Locally (Ciomadul massif) explosive volcanism prevailed.

Typically calc-alkaline chemism of generating magmas in the most part of volcanic chain turn to high potassic tendency in South Harghita and to shoshonitic one in their southern extremity (Seghedi et al., 1986).

Occurrences

Volcanic bombs and blocks in Harghita Mts occur in two zones: (1) Ciomadul massif and (2) eastern part of North Harghita Mts, west of Dănești (Fig. 1). In the first case, they occur within the upper volcanic compartment and in the second one within the lower volcanic compartment.

1. In Ciomadul massif known bombs originated from Komlosárok brook — Olt Valley tributary downstream of Tușnad Băi — where they were found in the valley bottom detached from weakly consolidated pyroclastic material in which they were buried. Today there are still found some specimens or fragments along the valley.

Volcanic bombs and blocks are largely found in new occurrences in Porcului and Lacului brooks on the southern slope of the massif. Here

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they are met either detached from its matrix along the valley bottom and engulfed in a clayey block and ash deposit of a likely laharic nature. Small specimens and bomb fragments are also found on northern slope in Benevize and Valea Roșie brooks.

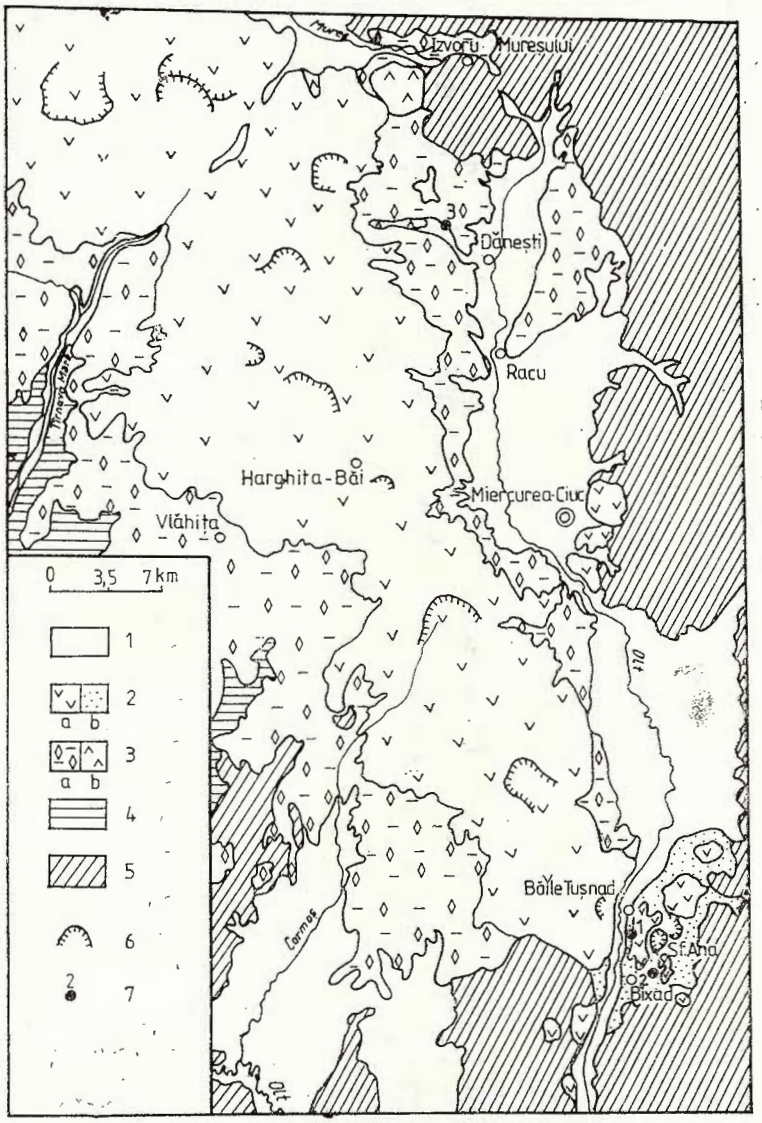

Fig. 1 — Geological sketch of Harghita Mts with volcanic bomb and block occurrences.

1, Quaternary deposits; 2, Upper volcanic compartment: massive volcanics (a), pyroclastics (b); 3, Lower volcanic compartment: volcano-sedimentary formation (a), massive volcanics (b); 4, Neogene sedimentary deposits of Transylvanian Depression; 5, Crystalline-Mesozoic zone and Internal Flysch zone; 6, Crater; 7, Occurrences of volcanic bombs and blocks; (1) Komlosárok brook, (2) Porcului brook (Ciomadul massif), (3) West Dănești.

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In fact in all these occurrences, bombs and blocks represent coarse elements in the mass of volcanoclastic deposits which form the pyroclastic cone around the well known Sfinta Ana and Mohoş craters, built in a first explosive eruption phase. The only situation in which the relation of bomb and block bearing deposit is exposed (Porcului brook) their chaotic structure, lack of bedding and sorting is evident. They consist of centimetric and decimetric angular and subangular dacitic lithic clasts in an ashy-clayey unconsolidated matrix. Volcanic bombs and blocks representing some of these clasts are chaotically included. No bomb-sags are found in the subjacent fines. These observations point to the flow origin of deposit, rather laharic than pyroclastic, and suggest a secondary inclusion for bombs.

2. In the west Dănești zone, volcanic bombs and blocks are found on an about 1 km² area along Groapa Apei Valley, upstream the confluence with Urasag brook and on their slopes (Bățului and Surduc hills). They occur exclusively as isolated coarse fragments along the valley or on the volcano-sedimentary plateau on an outcrop lacking area. Most of them are gathered by people in hummocks for releasing hayfields.

Local occurrence and large number of volcanic bombs and blocks, some of them of great volume, suggest the relatively closeness to source (an eruption center respectively) likely buried and situated inside the lower volcanic compartment.

Volcanic Bombs and Blocks

1. In Ciomadul massif, volcanic bombs and blocks as most of volcanics in this area are dacitic in composition (amphibole, biotite±pyroxene ±quartz dacites) (Szakács, Seghedi, 1986).

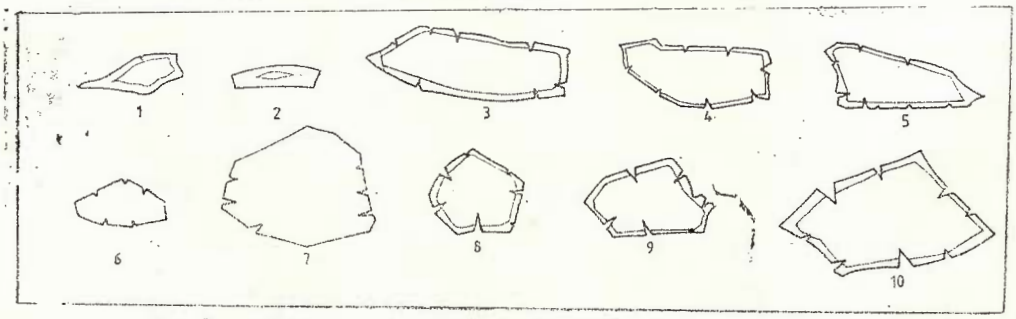

Fig. 2 — Morphology of volcanic bombs and blocks in Ciomadul massif.

Their size varies from 0.1 — 0.5 m to 0.8 — 1.0 m in diameter, most of them having 0.2 — 0.3 m. No significant size variation with distance from presumed source was observed. This proves their subsequent inclusion in volcanoclastic deposit.

Shape is also variable. Most are characterized by polygonal shape. Isometric polygonal forms are relatively rare (Pl. I, Figs. 1, 2). Most frequent flattened polygonal forms are found. Highly typical for Ciomadul massif, are weakly flattened polygonal specimens with pentagonal or hexagonal contour, usually having an asymmetric transversal section with a flatter lower half and more curved upper half (Fig. 2). Resulted

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polygonal surfaces are weakly curved, more frequent convex, rare concave or flat. There were also found very flattened forms, some of them with oblong polygonal section (Pl. III, Fig. 2), others quasirectangular (Fig. 2).

A number of bombs and blocks have chilled margin consisting of smooth glassy crust, from several mm to 2 — 3 cm thick. Their interior is always porous and has a lighter colour (grey) contrasting to dark-grey margins (Pl. III, Fig. 2). Specimens lacking chilled margin are light in colour. Their internal structure is usually massive, but oriented fluidal structures, parallel to flattened base, are also found (Pl. II, Fig. 2).

The spectacular characteristic feature of volcanic bombs and blocks in Ciomadul massif is bread-crust cracking. Cracks are large, deep and less close-spaced in specimens with chilled margin (Pl. I, Fig. 1) and they form a fine close-spaced shallow network in those lacking chilled margins (Pl. III, Fig. 1). Sometimes cracks show a subparallel development to flattened base.

Some small bombs have characteristic star-like impact cracks situated exclusively on the more curved surface (Pl. III, Figs. 3, 4).

2. In the west Dănești area, volcanic bombs and blocks are andesitic in composition (pyroxene andesites, with $SiO_2 \approx 61\%$; Szakács, Seghedi

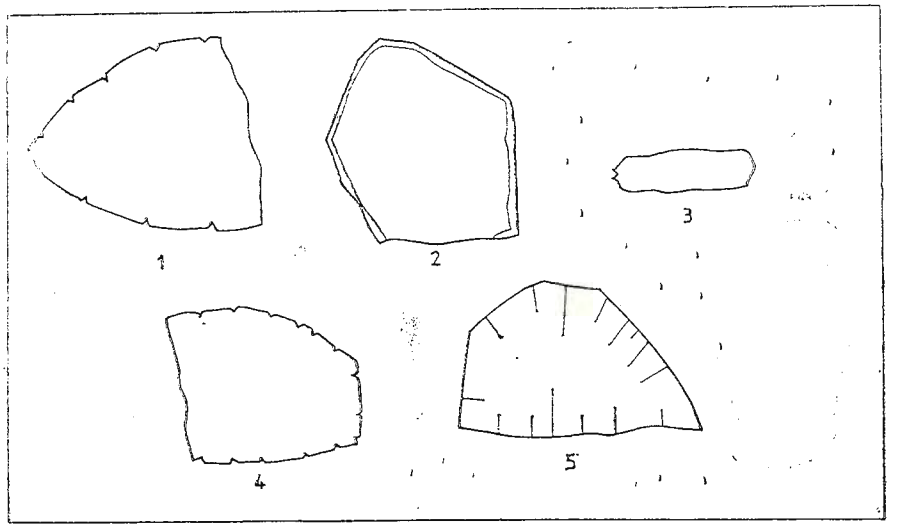

Fig. 3 — Morphology of volcanic bombs and blocks in west Dănești area.

in Mureșan et al., 1986). Except typical specimens, they cannot be easily recognized. Most of them may be confused at first sight with ordinary subangular andesitic fragments which usually compose coarse volcaniclasts in this area. Their polygonal shape often with curved surfaces and edges, chilled margin or internal radial jointing, represent recognising features at a more careful examination.

Size range is between few decimenters to about 1.5 m.

Shapes are polygonal, limited by plane or curved surfaces (Fig. 3). Some tend to be sharper at one end and truncated at others resembling to a shell (Pl. VII, Fig. 2). It was found a smaller elongated specimen,

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pentagonal in section, with „pinch" at one end, the point where the bomb detaches from lava. Its aspect suggests a torpedo (Pl. V, Figs. 2, 3). Generally prevail quasiisometrical and elongated shapes, flattened forms being rare.

Some bombs and blocks show bread-crust cracks. Crack network is more close-spaced and more superficial than in typical bread-crust cracked dacitic bombs from Ciomadul massif (Pl. VII, Figs. 1, 2). Sometimes crack system is „virtual" when polygonal andesitic block breakes at hammer bits along a zig-zag surface. Several specimen show internal radial jointing (Pl. VI, Fig. 2).

Rarely are found chilled margins, 0.2—1 cm thick, showing reduced crystallinity with respect to the interior (Pl. VI, Fig. 1). In most cases there are no structural heterogeneities.

Discussions

In relevant literature, generally a clear conceptual distinction between volcanic bombs and blocks is made (Mac Donald, 1972; Fisher, Schmincke, 1984). The former represents material ejected in molten state (essential ejecta), their shape resulting through solidification during aerian trajectory and shortly later. On the contrary, blocks are ejected in solid state with definitive shape in the ejection moment. Therefore, generally they represent accesory or accidental material, but blocks can also originate as essential ejecta (MacDonald, 1972). In this last case, distinction between bombs and blocks is not always possible due to feature convergencies.

Bread-crust cracking may occur both in bombs and blocks. Large and deep cracks are more frequent in bombs and they are the result of interior degasing and the gase pressure caused breaking of initially formed crust. Such bombs are characteristic for more silicic, mainly dacitic magmas.

Blocks can also have bread-crust type cracks, shallower and more close-spaced but there are all intermediate types between this two aspects (MacDonald, 1972). For blocks, cracking would result by reheating, sometimes up to incandescence, of preexistent consolidated fragments.

In Ciomadul massif, all these aspects are found. Specimens with chilled margin and deep bread-crust cracks most likely are bombs. Those of fine cracking and without chilled margin likely belong to blocks. Bomb versus block nature of several specimens with fluidal structure pointing to a previous consolidation is highly doubtful. The same petrographical constitution shows essential or at most accesory nature of fragments. Different structural aspects suggest that expulsed material was in various consolidation states in the eruption moment. Disruption through a crust covered by growing lava dome is a likely process for originating the described bombs and blocks. Alternatively part of them may be not balistically emplaced and subsequently removed ejecta, but simply thermally craked lava dome crust fragments included in flow deposits. Impact cracked specimens are at least ballistic explosion products.

In west Dănești zone where massive varieties without chilled margin predominate, most of described coarse polygonal fragments seem to

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be blocks. Bread-crust features in several cases may originate in thermal cracking of previously consolidated clasts. Same petrographical feature and evidence of detaching from partial unconsolidated material suggest that they are essential and/or accessory fragments, and presence of bombs is not excluded.

Because an accurate discrimination between volcanic bombs and blocks is impossible, we prefer to refer to described coarse volcanic fragments from Harghita Mts as "volcanic bombs and blocks".

Acknowledgements

The authors are indebted to dr. S. Peltz and dr. I. Seghedi for critical lecture of the manuscript and for pertinent suggestions in the text improvement.

[4] Although the presence of volcanic bombs in South Călimani Mts was mentioned (Peltz, 1969), there are no observational data (occurrence, description) about them. Long time the volcano-sedimentary formation in Călimani-Gurghiu-Harghita volcanic chain was erroneously considered to be mainly constituted of agglomerates, i.e. volcanic bomb deposits, without presenting real cases.

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BOMBE VULCANICE DIN MUNȚII HARGHITA

(Rezumat)

Pe lîngă ocurența cunoscută din pîrîul Komlosárok, bombe vulcanice apar, în număr mare, în valea pîrîului Porcului de pe versantul sudic al masivului Ciomadul. Bombele sînt înglobate secundar într-un depozit vulcanoclastic haotic cu blocuri și liant cineritic-argilos, de natură laharică. Compoziția lor este dacitică. Bombele, de dimensiuni variabile, în general decimetrică, au forme variate, de regulă poligonale, cele mai caracteristice fiind ușor aplatizate, cu o parte (inferioară) mai teșită și alta mai bombată. Majoritatea au o margine hialină și crăpături de tip „coajă de pîine". S-au întîlnit și mici bombe cu crăpături radiar-divergente, de impact.

În munții Harghita de Nord, într-o zonă la vest de comuna Dănești, au fost identificate la suprafața terenului bombe vulcanice andezitice în aria de dezvoltare a formațiunii vulcano-sedimentare. Ele sînt poligonale, adesea cu fețe și muchii curbate. Unele au o margine cu cristalinitate mai redusă. Cîteodată prezintă o crustă cu crăpături de tip „coajă de pîine".

În literatura de specialitate se face distincție între bombe și blocuri, ultimele reprezentînd ejecții în stare solidă. Majoritatea exemplarelor din masivul Ciomadul se încadrează la bombe, dar în cazul unui număr de exemplare atribuirea la bombe sau blocuri este incertă, datorită existenței formelor de trecere. Aspectele morfologice și structural-texturale observate denotă stări variate de consolidare a magmei în momentul erupției. Originea bombelor prin disrupție printr-un dom sau lac de lavă cu o crustă în curs de solidificare este cea mai probabilă.

În zona de la vest de Dănești, majoritatea exemplarelor, prin caracterele lor, par să aparțină blocurilor, deși unele prezintă evidența faptului că în momentul expulzării nu se aflau în stare solidă. Din acest motiv și pentru a evita unele confuzii, s-a preferat folosirea termenului de "bombe vulcanice" pentru toate ejecțiile grosiere descrise.

EXPLANATION OF PLATES

Plate I

Fig. 1 — Quasiisometric polygonal volcanic bomb, with chilled margin and bread-crust cracks. Porcului brook, Ciomadul massif. Collection of Harghita District Museum, Miercurea Ciuc.

Fig. 2 — Isometric polygonal volcanic bomb with chilled margin and bread-crust cracks. Porcului brook, Ciomadul massif. Authors' collection.

Plate II

Fig. 1 — Flatty polygonal volcanic bomb with chilled margin and bread-crust cracks. Porcului brook, Ciomadul massif. Collection of Mineralogy Department in Biology Geology and Geography Faculty, University of Bucharest.

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Fig. 2 — Fragment of flattened polygonal volcanic bomb or block with bread-crust cracks and fluidal structure. Porcului brook, Ciomadul massif. Collection of the Geology and Geophysics Institute, Bucharest.

Plate III

Fig. 1 — Flattened volcanic block with a fine network of shallow bread-crust type cracks. Porcului brook, Ciomadul massif, Authors' collection.

Fig. 2 — Sectioned flattened polygonal volcanic bomb with chilled margin and bread-crust cracks. Note the porous structure of interior. Porcului brook, Ciomadul massif. Authors' collection.

Figs. 3, 4 — Small volcanic bombs with impact cracks. Porcului brook, Ciomadul massif. Authors' collection.

Plate IV

Fig. 1 — Small volcanic bombs and blocks and fragments. Porcului brook, Ciomadul massif. Collection of the Geology and Geophysics Institute, Bucharest.

Fig. 2 — Flattened polygonal volcanic bomb with bread-crust cracks. Porcului brook, Ciomadul massif. Authors' collection.

Plate V

Fig. 1 — Lateral view of bomb in plate IV, fig. 2.

Figs. 2, 3 — Elongated polygonal volcanic bomb; note the detaching "pinch" at one end. Groapa Apei brook, west Dănești. Authors' collection.

Plate VI

Fig. 1 — Naturally sectioned compact polygonal volcanic block. Note less crystalline marginal band. Bățului hill, west Dănești.

Fig. 2 — Naturally sectioned compact, curved faced volcanic block with radial internal jointing. Groapa Apei brook, west Dănești.

Plate VII

Fig. 1 — Polygonal volcanic block or bomb with fine bread-crust cracks. Groapa Apei brook, west Dănești.

Fig. 2 — Shell shaped volcanic bomb with bread-crust cracks. Groapa Apei brook, west Dănești. Collection of the Geology and Geophysics Institute, Bucharest.

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D. S. Inst. Geol. Geofiz. vol. 74/1 (1987), 1989, pag. 191—207

CONTRIBUTIONS TO THE GEOLOGICAL EVOLUTION OF THE CRYSTALLINE FORMATIONS IN THE SOUTH BANAT (DANUBIAN REALM)[1]

by

IOAN DINICĂ[2]

Crystalline rocks. Dynamic metamorphism. Regional metamorphism. Retrograde metamorphism. Alpine Orogeny. Strike-slip faults. Right-lateral faults. Relics. Andalusite. Staurolite. Sillimanite. Garnet. South Carpathians — Crystalline Danubian Realm — Almăj Mountains.

Abstract

In the Danubian Realm of the South Banat the crystalline schists previously ascribed to the Corbu and Vodna epimetamorphic formations contain relics of index minerals — e.g. staurolite, sillimanite, andalusite, garnet, diopside, green hornblende, and plagioclase — pointing to a medium-grade initial metamorphism. Their low-grade-schists aspect is due to a Caledonian dynamic retromorphism, which turned the rocks into blastomylonites and phyllonites, as well as to a Hercynian regional retromorphism in the greenschists facies. The Alpine tectonics occurs in the study zone as dextral strike-slip faults along the Cerna graben. A branch of this tectonic zone — Tișoviţa-Lucidol Fault — forms, by clogging up the eastern compartment in the Baia Nouă-Iuţi zone, a synform with a north-westward axial sinking in whose axis occurs the southern extremity of the Poiana Mraconia Series.

Résumé

Contributions concernant l'évolution géologique des formations cristallinnes du Banat de sud (domaine danubien). Dans le domaine danubien du Banat de sud les schistes cristallins, considérés épimétamorphiques, des formations de Corbu et de Vodna comportent des reliquats de minéraux index tels staurotide, sillimanite, andalusite, grenat, diopside, hornblende verte et plagioclase qui indiquent un métamorphisme initial de degré moyen. L'aspect des schistes faiblement métamorphosés est dû à un rétromorphisme dynamique d'âge calédonien qui a transformé les

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roches en blastomylonites et phyllonites, tout comme à un rétromorphisme régional hercinien dans le faciès des schistes verts. La tectonique alpine se manifeste dans la zone par des décrochements dextres le long du graben de Cerna. Une ramification de cette zone tectonique — faille de Tișovița-Lucidol — détermine, par le blocage du mouvement du compartiment de l'est dans la zone de Baia Nouă-Iuți, une synforme à effondrement axiale nord-ouest, dans l'axe de laquelle est située l'extrémité sud de la série de Poiana Mraconia.

Introduction

This paper deals with the results of the researches carried out with a view to elaborate the Svinița and Orșova sheets of the Geological Map of the S.R. of Romania, sc. 1 : 50 000. It is also based on the observations on the metamorphic sequences known as Poiana Mraconia, Neamțu, Corbu and Vodna in the Danubian Realm. This geological map includes also other geological formations which are in direct relationships with the study formations and have been studied by other researchers. They have been taken over from unpublished sketch maps of the Orșova (Hârtopanu et al., 1986) and Svinița (Năstăseanu et al., 1982) sheets. For the northernmost part of the map (not entirely mapped), in the zone between Teiul Moșului Summit, Mișelep Summit, confluence between Lucidol Valley and Secărștița Valley, and northwards in the Iardașița Valley we used data according to Stan et al. (1986) (with the author's permission).

1. Evolution of Ideas

The crystalline schists in the South Banat have been studied by several geologists. However, modern and systematic researches have been carried out since the first half of the 20th century.

Streckeisen (1934) divided for the first time the Almaj crystalline into several zones with a NNE—SSW trending, among which mention should be made of Cătrămăț zone, Predeal zone (including Corbu phyllites) and Ogradena zone (including the Ogradena granite and the present-day Neamțu and Vodna Series).

Codarcea (1937) distinguished a western crystalline complex, comprising Toronița zone, Ielova zone and Poiana Mraconia zone (=Cătrămăț zone) and an eastern crystalline complex including the serpentinite zone, Iuți gabbro, Plavișevița epigabbro zone, Corbu phyllite zone, Cherbelezu granite, Neamțu zone, Vodna phyllite zone and Ogradena massif. The above-mentioned author (Codarcea, 1940) considered Corbu zone as an "incontestable median zone" between the western (internal) complex and the eastern (external) one. Later on (Codarcea et al., 1961) the internal complex was called Almaj Unit and the external complex (including Corbu zone) Retezat-Ogradena Unit.

The complex geology of the South Banat, previously presented in several reports and papers of the period 1954 — 1970, was synthetized by Bercia and Bercia (1962, 1975, 1980). According to composition, the space relationships and the metamorphic degree the researchers assert that there are some significant differences between Neamțu Series and

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Corbu and Vodna Series, i.e. premetamorphic terrigene rocks are mainly psamitic in Neamţu Series and predominantly pelitic in Corbu and Vodna Series. The age of the Corbu Series — Cambrian according to Dessila-Codarcea and Codarcea (1968) — was proved to be Upper Precambrian-Lower Cambrian on the basis of the correlation with the diabase-phyllite formation in Yugoslavia (Stara Planina — Poreč zone), its metamorphism in the greenschist facies being of Baikalian age. Neamţu and Poiana Mraconia Series are referred to the Middle Precambrian and their metamorphism in the amphibolite facies has been achieved during the pre-Baikalian orogenesis, later on specified as Dalslandian orogenesis (Bercia, Bercia, 1980; Kräutner, Savu, 1978).

According to Savu et al. (1978) Poiana Mraconia and Neamţu Series are of Upper Precambrian B age, with a Cadomian metamorphism and Corbu and Vodna Series are Lower and Middle Cambrian in age, with a Late Assyntic metamorphism.

Within the lithostratigraphic correlation of the Precambrian in Romania, Kräutner (1980) referred Poiana Mraconia and Neamţu Series to the Carpian Supergroup (Upper Precambrian A), and Corbu and Vodna Series to the Marisian Supergroup (Upper Precambrian B — Cambrian). The same ages are assigned to the series mentioned by Năstăseanu et al. (1981).

Up to now, most of the authors considered that the Corbu and Vodna Series are characterized by a low-grade metamorphism (Codarcea, 1940; Dessila-Codarcea, Codarcea, 1968; Bercia, Bercia, 1975, 1980; Gheruci, Serafimovici, 1965, 1966; Anastasiu, 1976; Năstăseanu et al., 1981). On the basis of the instability of some mineralogical associations Gunnesch, Gunnesch (1978) mentioned for the first time the polymetamorphic character of the Corbu Series. Based on the identity of lithology, age, degree of metamorphism and space relationships the above-mentioned authors regarded the Vodna Series as a component of the Corbu Series. This last correlation — the appurtenance of the Vodna phyllites to the Corbu zone — was previously mentioned by Gheruci and Serafimovici (1965, 1966) on the basis of petrographical similarities.

The identification of staurolite and andalusite in the rocks of the Corbu Series (Stan et al., 1981; Stan, 1984; Mărunţiu, Seghedi, 1983 a, 1983 b) is a significant argument in favour of the polymetamorphism. Andalusite was interpreted by Stan et al. (1981) and Stan (1984) as a thermal contact mineral, generated by the Cherbelezu Granite intrusion, staurolite, garnet and amphibole representing relics of a primary metamorphism (Assyntic or Early Caledonian); later on (Stan, 1985) staurolite was regarded as thermal contact mineral.

Mărunţiu and Seghedi (1983 a, b) consider staurolite and andalusite as relict minerals of the regional metamorphism. They point out that the Corbu Series represents a shear zone affecting the western part of the Neamţu Series and partly the Plavişeviţa metagabbros. The deformation, represented by mylonitic structures, is accompanied by recrystallization and neomineralization, thus providing mylonites, phyllonites and blastomylonites. These deposits are Early Caledonian in age, Lower Paleozoic grey schists and limestones being involved into the tectonic mélange thus achieved.

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2. New Data on the Crystalline Formations

The geological mapping effectuated in recent years revealed new aspects of the geology, petrography and metamorphic evolution of the crystalline formations, which will be pointed out further on. The above-mentioned series and formations will not be minutely described; we shall present only those data which are of interest for this paper. Thus, Neamțu Series will be presented in so far as a comparison with Corbu and Vodna Series is necessary. Likewise only certain aspects, especially tectonic ones, characterizing the sedimentary formations or the magmatic series will be described.

2.1. *Poiana Mraconia Series*

Poiana Mraconia Series (Bercia, Bercia, 1975; = Cătrămăț Series — Streckeisen, 1931; Poiana Mraconia zone — Codarcea, 1937) was described in details by Bercia, Bercia (1975). It consists of quartzitic paragneisses, garnet ± kyanite micaceous paragneisses, quartz-feldspathic gneisses, feldspathic quartzites and amphibolic rocks. The Barrovian-type regional metamorphism occurred during the Dalslandian Orogenesis (Bercia, Bercia, 1980; Kräutner, Savu, 1978) under conditions of almandine-amphibolite facies, kyanite-almandine-muscovite subfacies (Bercia, Bercia, 1975).

In the territory of the Svinița geological map (1 : 50 000) two formations were separated:

— formation of micaceous paragneisses in association with amphibolic rocks, consisting of a close alternation of paragneisses and micaschists ± kyanite ± amphibolite garnets and amphibolitic gneisses;

— formation of quartz-feldspathic gneisses and amphibolites, geometrically in an upper position, in which the petrographic background is represented by quartz-feldspathic gneisses with muscovite ± biotite; paragneisses display a coarser granulation than in the lower formation, and amphibolites form more rare, well individualized levels or lenses, with a variable granulation (Bercia, Bercia, Dinică in Năstăseanu et al., 1982).

2.2. *Corbu and Vodna Formations*

Corbu and Vodna formations are equivalents of the Predeal zone (Streckeisen, 1934), Corbu phyllites zone and Vodna zone (Codarcea, 1937) or of the Corbu and Vodna series (Bercia, Bercia, 1975).

The two mentioned series will be presented together as there are strong arguments which make us regard them as equivalents.

The cartographic outline of the Vodna Formation between Mraconia Valley and Iardașița Valley and of the Corbu Formation from the Danube up to Lucidol Valley revealed that all the marbles referred up till now to the Neamțu Series belong to the Corbu and Vodna formations. Moreover, the same relict minerals mentioned for the Corbu Formation — staurolite and andalusite — were identified in the Vodna Formation; sillimanite was found as well. The lithological composition of the Corbu

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and Vodna "series" is the same : sericite ± chlorite quartz schists, amphibolite schists, quartz-feldspar schists, feldspar quartzites, retromorphic micaschists and micaceous paragneisses ± staurolite ± sillimanite ± andalusite ± garnet, rare lenses of marbles, serpentinites and metagabbros. Although their aspect resembles low-metamorphosed schists, their retrograde character, determined by a marked dynamic metamorphism, is revealed only under the microscope. The rock structure and texture are that described by Mărunțiu and Seghedi (1983 a, b) for the rocks of the Corbu Series. The authors described rocks affected by the dynamic metamorphism and transformed into blastomylonites, ultramylonites and phyllonites. They consider that this "series" represents in fact a shear zone of the western part of Neamțu Series, this tectonic mélange involving both Plavișevița-type uralitized and saussuritized gabbros and Lower Paleozoic deposits. Consequently the mentioned authors proposed the denomination of "Corbu Zone". A similar idea is maintained by Stan (1982, 1985) who also admits that a part of Poiana Mraconia Series belongs to this zone.

In our opinion there are differences between Corbu and Vodna formations, on the one hand, and Neamțu Series, on the other hand, marked both by the presence of marbles and serpentinites in the Corbu and Vodna formations and by the predominance, especially in the Corbu Formation (more completely represented), of metapelitic terrigene rocks (Bercia, Bercia, 1975). Neamțu Series is mostly amphibolic and metapsamitic (Bercia, Bercia, 1956, 1975). In the Corbu Formation the amphibole schists are represented particularly in the lower half of the pile but over a mostly metapsamitic and metapelitic level. A confirmation of these lithological differences is that, although metamorphosed in the amphibolite facies, staurolite-almandine subfacies, staurolite, sillimanite or andalusite are not to be found in the Neamțu Series.

The dynamic metamorphism which intensely affected the Corbu and Vodna formations made possible an intimate mélange of the involved material. Therefore, the two formations were tectonically delimited. They have experienced much stronger dynamic efforts in comparison with the Neamțu Series. This fact could be interpreted either as an ante-Hercynian nappe plane or as the limits of a metamorphosed tectonic mélange, or, less probable, as lithostratigraphic limits within the Neamțu Series superposed by a dynamic metamorphism. In all the three situations the term formation is suitable.

It is of note that along Vodna Formation, developed as a 100 — 400 m wide strip, there is a variation of the lithological content due to the existence of zones with amphibolic schists, serpentinites, calcareous shales, marbles, quartz-sericite schists and retromorphic paragneisses (Mraconia Valley basin) and of zones in which only some of the terms are to be found. For example, on the ridge between Eșelnița Valley and Crivița Valley Vodna Formation is represented almost exclusively by quartz-feldspar schists with an extremely sheared aspect, finely banded; northwards it is represented by the same quartz-feldspar schists which grade into feldspar quartzites, intercalated with amphibole schists and mica paragneisses with a submillimetric banding. In Seracova Mică Valley

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and rather northwards amphibolites and marbles predominate; however, other petrographic varieties are also to be found: sericite-chlorite schists, fine paragneisses, etc.

2.3. *Physiography of Index Minerals of the Corbu and Vodna Formations. Succession Relationships*

Staurolite, andalusite and sillimanite occur in three places in the Vodna Formation:

— left side of Mraconia Valley — garnet-, staurolite- and andalusite-bearing paragneiss;

— Pițigui ravine — garnet-, sillimanite- and andalusite-bearing micaschists;

— right side of Cerna Valley, between Iardașița Valley and Secărștița Valley — garnet-, sillimanite- and andalusite-bearing micaschist.

Stan et al. (1981), Stan (1984), Mărunțiu, Seghedi (1983 a, b) described the presentation mode of staurolite and andalusite in the rocks of the Corbu "Series".

Staurolite in the Vodna Formation is identical with that in the Corbu Formation. It displays more or less crushed grains, locally prism-like, with the long axis oriented along the foliation, subidiomorphic or slightly curved prisms (Pl. II, Fig. 2); however, it often occurs as xenomorphic grains with marked deformations (Pl. II, Fig. 1). Staurolite contains opaque mineral inclusions, often as straight or curved parallel rows. The rows of inclusions are conformable or unconformable in relation to the blastomylonitic foliation. Frequently staurolite includes tourmaline crystals. Staurolite is a mineral with the most obvious alteration, all the possible stages of pinitization being visible, up to the complete replacement.

Sillimanite is found as well-developed crystals (3 — 4 mm thick and 10 mm long), aligned with the shear cleavage, partially deformed and pinitized marginally or on fissures. One of the sections displays an incomplete rim of andalusite which seems to substitute the deformed sillimanite, intruding slightly its fissures (Pl. III, Figs. 1, 3). In other sections andalusite forms continous films at the fasciculate ends of the sillimanite and intrudes the cleavages as bands alternating with sillimanite bands (Pl. IV, Fig. 1). Andalusite is pinitized marginally. Sillimanite also shows opaque mineral inclusions. Between the deformed sillimanite grains on certain shear cleavages or on breaks inside the sillimanite, a fibrolite is observed as needles arranged in divergent sheaves or clusters. Considering that fibrolite also occurs in albite crystals with relics of basic oligoclase and in association with quartz or micas, locally in close connection with the appearance of a lamination plane on which muscovite and chlorite are formed, it obviously represents a retrograde process.

Andalusite is found in several sections as less deformed crystals, with inclusions of opaque minerals or of titaniferous biotite. No helicytic inclusions of opaque minerals have been observed in the study thin sections, as pointed out by Mărunțiu and Seghedi (1983 a, b), but only straight

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or slightly curved parallel inclusions. Even when it is not deformed, andalusite is pinitized, in places locally or even wholly, or occurs as isolated andalusite masses preserving common extinction. Opaque minerals with the same disposition as in fresh crystals remain in places in the pinite masses with an andalusite crystal contour. In such cases it is to be mentioned that the rock preserved biotite with a brown-greenish pleochroism, nondeformed and fresh, developed as small leaflets, more or less oriented, forming parallel bands that give the rock foliation. Between andalusite and staurolite are clear succession relationships, andalusite including staurolite as small xenomorphic grains. The andalusite pinitization does not affect the staurolite relics included in it (Pl. IV, Fig. 4).

Diopside (Pl. III, Fig. 2) was found in a single section, on a grey-greenish, plane-parallel phyllite-like blastomylonite of the Vodna Formation. Under the microscope the rock is fine-grained, with a blastomylonitic structure. It consists of small grains of epidote-clinozoisite, sericite, quartz, chlorite and opaque minerals, within which diopside occurs as deformed rounded porphyroclast in association with small quartz grains filling the fissures.

Garnet is found as:

— garnet I — deformed and chloritized. The chlorite mass included small grains of quartz, muscovite with a slightly greenish pleochroism and opaque minerals. In places chlorite displays anomalous interference colours typical of pennine. Relics of fringed forms of garnet, bordered by a powder of opaque minerals, can be observed in the chloritic mass (Pl. V, Fig. 1, 3);

— garnet II — often idiomorphic, as small crystals, with a more turbid central zone, unaltered even when deformed. Small fissures occur on which sericite and chlorite are formed. It is of note that the second generation of garnet is contemporaneous with brown-greenish biotite occurring in the pressure shadow of the undeformed fresh garnet beside quartz and, in places, muscovite. This biotite participates in the formation of the schists as bands alternating with quartzfeldspathic ones. It is altered into chlorite only affected by a new deformation, when muscovite is formed on the new S planes (Pl. VI, Figs. 1, 2).

Green hornblende occurs as porphyroclasts in basic rocks. It is found as broken up crystals (Pl. V, Fig. 3), surrounded by an oriented mass formed of small crystals of actinote, biotite, chlorite, quartz and opaque minerals. Locally the biotite formation is observed on the margin of the hornblende crystals.

Plagioclase represents a basic oligoclase in micaschists and paragneisses and an andesine with 34 — 36 per cent anorthite in amphibolites. It occurs as porphyroblasts elongated along the foliation, locally polysynthetically twinned, usually sericitized and kaolinized or saussuritized, with a turbid aspect (Pl. VI, Fig. 2). In retromorphosed rocks, chlorite and muscovite occur in association with albite, sometimes very limpid, other times with inclusions of epidote, quartz and fresh biotite. Locally epidote is separated on the plagioclase cleavage, by its decalcifiation, as well as on the biotite cleavage.

Biotite is usually found as small rectangular lamellas which statistically parallel the schistosity. It forms bands alternating with bands

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within which quartz and plagioclase feldspar predominate. In most cases biotite is not deformed, giving the impression of static crystallization, after the rock blastomylonitization. However, the arrangement in parallel bands and the appearance in the pressure shadows of the relict minerals, deformed and altered (staurolite, sillimanite, chloritized garnet, locally andalusite), or of the contemporaneous minerals, e.g. fresh, idiomorphic garnet or plagioclase (oligoclase, andesine), points out its syndeformational blastesis, probably with repeated crushings and recrystallizations. Its pleochroism is brown-dark green (Ng) and yellow-greenish (Np). Biotite displaying the above-mentioned aspects occurs on the whole thicknesses of the Vodna and Corbu formations, where it has not been preserved unaltered into chlorite at a new deformation. The chloritization process can be observed in all its stages, up to the complete replacement by chlorite. In some cases zircon was observed, forming pleochroic aureoles in biotite.

Besides the brown-greenish biotite there is also a biotite with a brown-reddish (Ng) and brown-yellowish (Np) pleochroism, included into andalusite as very small crystals, locally with blurred, torn edges.

Nearby the Cherbelezu granite occurs a biotite which differs from the brown-greenish biotite only by the random orientation of the small-sized crystals, in places crossing the quartz-feldspathic bands. It seems to be statically recrystallized under the thermal influence of granite. Also in this zone muscovite is found as well-developed films, unoriented, or as muscovite replacing pinite formed on andalusite (Pl. IV, Figs. 3, 4).

Muscovite occurs frequently in the schists of the Corbu and Vodna formations in relation to the latest deformational moment. It is found between the bands of minerals with a granular habitus, giving the rock foliation. Muscovite is altered into sericite. Although it seems connected especially to this last deformation moment, it is worth mentioning that it often occurs in association with fresh unchloritized biotite and that it has been found in a blastomylonite of the Vodna Formation as a porphyroclast surrounded by a fine-grained mesostasis with phyllonitic structure formed of epidote, sericite and opaque minerals (Pl. V, Figs. 3, 4). In this case, it is obviously rounded, with a slightly curved cleavage. Another crystal from the same section displays obvious deformations and tearings of the cleavage after certain planes parallel with the phyllonitic foliation, outlining the tearing off of a piece from the crystal and its joining to the fine-grained mass. Such aspects are of importance, on the one hand, because they make evidence of the existence of at least two generations of muscovite and, on the other hand, because it proves the possibility of preservation of relict minerals in rocks strongly affected dynamically, within which the mineral mass is destroyed and displays a typical fluxion texture.

Chlorite is also frequently found; it is formed especially on biotite or actinote. It occurs also in the pinitic mass that replaces staurolite or andalusite, locally forming radiating crystals. Very often chlorite is prevailing in retromorphosed rocks.

Quartz forms bands alternating with micas, in association with plagioclase feldspar. Its outline is more often than not fringed, interfin-

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gered, seldom with a granoblastic polygonal texture. Its extinction is undulatory.

Potash feldspar practically does not occur in rocks of the Corbu and Vodna formations except for feldspathic quartzites in which oligoclase, biotite (in places chlorite) and muscovite occur beside rare crystals of microcline-perthite.

Two polished sections carried out on rocks with staurolite, andalusite and garnet from Corbu and Vodna formations have been analysed chalcographically by dr. Udubaşa, who informed us of his observations.

Opaque minerals are represented by ilmenite, rutile and seldom by magnetite. Ilmenite is predominant in staurolite and garnet relics. It is partially altered into rutile. The alteration into rutile ± hematite ± sphene is more advanced in phylosilicates. Free independent grains of large-sized rutile occur also beside rutile in association with ilmenite. In places the alteration into rutile is more advanced, preserving only ilmenite relics. "Clean" lamellae of ilmenite occur, as well.

Leucoxenic aggregates certainly formed at the expense of ilmenite prisms are observed, almost all possible transition being visible. These indications of ilmenite instability concomitantly with the existence of a "clean" ilmenite point to the existence of two generations of ilmenite.

Taking into account the possibility of the existence of two generations of rutile, the situation is similar to that in crystalline areas where dr. Udubaşa proved the existence of two generations of ilmenite and of rutile : the former typical of the medium-grade initial metamorphism (M_1) (kyanite zone), the latter of the M_2 retromorphism (Udubaşa et al., 1984 ; Udubaşa, 1986).

2.4. *Considerations on the Metamorphism of the Corbu and Vodna Formations*

The identification of staurolite and andalusite in the Corbu Formation brought about the problem of their origin : are they thermal contact minerals, generated by the intrusion of the Cherbelezu granitoid, or minerals pointing to a previous higher grade regional metamorphism ? The identification of these minerals and of sillimanite in the Vodna Formation does not solve the original problem, as in the vicinity of the alignment where Vodna Formation is exposed occurs the Orgadena granitoid.

Streckeisen (1934), Codarcea (1937, 1940), Bercia, Bercia (1956, 1975) described thermal contact phenomena in the enclaves of the Ogradena granitoid, acicular sillimanite (fibrolite) being mentioned among the contact minerals. However, the above-mentioned authors also pointed out that these phenomena are to be found in the crystalline schists enclaves of the granitoid not in its western contact with the Neamţu Series (zone) where only migmatization and injection phenomena occur (Codarcea, 1937 ; Bercia, Bercia, 1975). Moreover, Bercia and Bercia (1975) brought arguments in favour of the thermal metamorphism of the Ogradena granitoid in the phyllite enclaves and also mentioned that in Sohodol Valley (Ogradena Nouă) 20 — 30 cm thick enclaves are transformed at the contact into garnet-bearing biotitic schists, preserving inside relics of sericite-chlorite phyllite schists. That fact had also been stated by

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Codarcea (1940) who pointed out that in different places still occur phyllites, quartzites, crystalline schists and amphibolites preserving their character, e.g. Vodna band at Ogradena.

All this emphasizes that the thermal effect of the granitoids is felt on small distances. Therefore, we should consider that minerals such as staurolite, andalusite and garnet, occurring on the whole thickness of the Corbu Formation, are not due to the thermal effect and they represent relics of a higher-grade metamorphism.

The disposition at the surface of the staurolite and andalusite occurrences points out that within the Corbu Formation they appear not only nearby the Cherbelezu granitoid but also at the boundary with Neamțu Series or, from the Mraconia Valley up to the Danube, zone in which the Cherbelezu granitoid is not exposed, especially close to the eastern boundary.

Anastasiu (1976) pointed out that there are no assimilation phenomena of the enclaves by the Ogradena granitoid; he mentioned, especially for biotite, weak recrystallization phenomena in the schists in enclaves.

Vișoiu et al. (1975, 1981) agreed with Anastasiu. Mărunțiu and Seghedi (1983 a) mentioned a poor thermal effect of the Cherbelezu granitoid manifested by the biotite recrystallization.

The aspect and the relationships of the minerals presented in the previous chapter lead to the following observations:

a — staurolite is the oldest mineral, considering its marked deformation and its pinitization, locally total;

b — sillimanite is also deformed and torn, partially pinitized;

c — there are no direct relationships between staurolite and sillimanite, staurolite being absent in rocks where sillimanite occurs;

d — garnet I, chloritized and deformed, coexists with staurolite and andalusite in the same sections;

e — andalusite is younger than staurolite and sillimanite. It includes staurolite and surrounds with a rim or intrudes on cleavages or breaks sillimanite. The relationship seems to be of substitution, but it can be only of superposition. It is less deformed than sillimanite; locally it seems undeformed, although it often displays a rotation indicated by the opaque minerals inclusions, with a different orientation as against the blastomylonitic foliation. Mărunțiu and Seghedi (1983 a) mentioned andalusite with a syntectonic blastesis. Andalusite includes crystals of brown-reddish biotite;

f — andalusite is pinitized, often totally. Muscovite is locally formed on pinite;

g — in the pressure shadow of staurolite, sillimanite and andalusite, deformed and pinitized, a fresh and undeformed brown-greenish biotite is developed. It coexists in blastomylonite with garnet II, in places with muscovite, too. Blastomylonitization of the rocks occurs, therefore, in the greenschists facies, biotite and garnet zone (Stan et al., 1981, 1985). It is the moment of the Caledonian orogenesis (Stan et al., 1981; Stan, 1984; Stan et al., 1985; Mărunțiu, Seghedi, 1983 a, b) or even an older moment;

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h — a new deformation moment generates S planes on which muscovite (sericite), chlorite and actinote are formed. Garnet II is partially deformed and albite occurs. This moment is the result of the Hercynian orogenesis.

All the observations mentioned up to point g (exclusive) refer to deformations and alteration processes previous or contemporaneous with the rock blastomylonitization, emphasizing that they existed at that time.

There follows the intrusion of the Cherbelezu and Ogradena granitoids, which include already blastomylonitized rocks. The age of these intrusions is of 372—393 MA (K/Ar) for the Cherbelezu granitoid and of 345 MA (Rb/Sr) for the Ogradena one (Tiepac in Mînzatu et al., 1970, fide Bercia, Bercia, 1980).

Staurolite and andalusite relics in the Corbu Formation and sillimanite relics in the Vodna Formation are proof of a medium-grade metamorphism. It could represent an intermediary low-pressure metamorphism that affected, before the blastomylonitization, the two formations. The andalusite association with sillimanite in the northernmost part of the Vodna Formation and with staurolite in the southernmost part would be due to an initial metamorphic zonation similar to that in the Bahna Outlier (Hârtopanu, 1975) but, unfortunately, the available data are not sufficient to prove it. In places, staurolite is included in andalusite, indicating the formation of the latter at the expense of the staurolite which, in this case, could belong to a Barrovian-type metamorphism.

3. New Data in the Clearing up of the Tectonic Evolution

The structure and tectonics of the crystalline schists in the South Banat have been presented in several papers. Streckeisen (1931) was the first who mentioned the thrust of the crystalline over serpentinites, the age of this overthrust being Hercynian.

Bercia, Bercia (1975) regarded it as a deep-seated fracture which acquired a thrust character during the Sudete phase (Middle Carboniferous).

Specifying the Caledonian age of the Cherbelezu granitoid — which comprises Corbu phyllites — and considering the dynamic metamorphism of the Corbu Series as a result of the thrust of the Poiana Mraconia Series over the Retezat—Ogradena Unit, Stan et al. (1981) and Stan (1984) consider that the overthrust occurred during the Ordovician-Silurian time-span.

Mraconia line is unconformably and transgressively overlain by the Upper Carboniferous and Permian deposits (Codarcea, 1937). However, Bercia, Bercia (1957, 1975) mentioned cataclasis phenomena, close shear folds and diaphthoresis processes in the crystalline schists of the Poiana Mraconia Series, at the boundary with the Permian deposit in the west, from the spring zone of the Mraconia Valley, considering that the Alpine, Austrian or Laramian movements could have yielded a dislocation of a part of the sedimentary cover of the Svinița—Svinecea Mare zone from the crystalline basement. Codarcea (1937) and Streckeisen (1934) drew the attention on some "strange" cataclasation phenomena of the gabbros at the

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Permian contact, of crushing of the Upper Carboniferous between Permian and crystalline.

Our mapping emphasized a NE—SW trending tectonic line, crossing obliquely the pre-Alpine units from the spring zone of the Tişovița Valley up to the spring zone of the Crivița Valley, extending with an overthrust line (Stan et al., 1986) in the left side of the Lucidol Valley, called Tişovița—Lucidol Fault. This fault is marked by cataclasations, laminations, mylonites and ultramylonites, their foliation pointing to a NW dipping.

There is a dextral strike-slip fault with a relative shifting of ca 6 km of the Upper Carboniferous and Lower Permian sedimentary deposits, a marked mylonitization of the Tişovița serpentinites, affected up to the Perilor Valley, and a slip of the Poiana Mraconia Formation of ca 8—9 km. Further on, in the Perilor and Corbu valleys, it limits the Cherbelezu granitoid from the Corbu Formation, crossed this formation and north of the Predelu Summit it determined the thrust of the Corbu Formation over the Neamțu crystalline. Between Lucidol and Iardașița valleys occurs a system of NW-SE trending faults, mainly displaying a senestral strike-slip character.

The tectonic efforts that yielded the Tişovița—Lucidol Fault determined the folding of the edifice of pre-Alpine Nappes forming a large synform, with a NW trending. The quartz-feldspathic gneisses formation of the Poiana Mraconia Series crops out in the axis of the synform, towards the spring zone of the Sirinia Valley. Northwards, this formation is well represented in the Mraconia Valley, Morilor Valley and Perilor Valley; however, our mapping evidence does not allow yet its cartographic separation.

On the synform flanks, the Poiana Mraconia Series was detached over serpentinites, marking the preexistent overthrust and forming mylonites along it, described by former workers particularly in this zone.

At Baia Nouă, from the Tişovița Valley towards NW, the thrust plane of the crystalline over serpentinites is overturned.

The Tişovița-Lucidol Fault reduces its detachment up to its extinction in the south-westernmost part, by the obstruction of the shifting of the eastern compartment, a process which led to the outlining of the synform. In this case a certain displacement of the sedimentary cover from the crystalline basement is possible, which explains the phenomena pointed out by Streckeisen (1931), Codarcea (1937) and Bercia, Bercia (1957, 1975).

A branch of the Tişovița—Lucidol Fault affects the western boundary of the serpentinites in the southwesternmost part of the Cherbelezu granitoid.

The genesis and working of the Tişovița—Lucidol Fault can be connected with the tectonic processes which led to the appearance and evolution of the Cerna Graben. The dextral slip of its compartments was mentioned by Bercia (1975) and its ampleness of about 35 km was specified by Berza, Drăgănescu (1985). Săndulescu (1984) pointed out that Cerna Graben, connected with a deep-seated crustal fracture, had a long history; its functioning has started at least since the Mesozoic and continued till the Upper Neogene or even up till now.

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The Tişoviţa—Lucidol tectonic line is a secondary fault, a branch of the tectonic system of the Cerna Graben which, by the tectonic effect, belongs to its relatively right slips.

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CONTRIBUȚII LA DESCIFRAREA EVOLUȚIEI GEOLOGICE A FORMAȚIUNILOR CRISTALINE DIN BANATUL DE SUD (DOMENIUL DANUBIAN)

(Rezumat)

Șisturile cristaline din domeniul danubian, separate de diferiți autori sub numele de seria (zona) de Corbu și considerate pînă recent ca fiind metamorfozate în condițiile faciesului șisturilor verzi, s-au dove-

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dit a fi roci care au suferit un puternic retromorfism dinamic în timpul orogenezei caledoniene, care le-a transformat în milonite, blastomilonite, ultramilonite sau filonite (Mărunțiu, Seghedi, 1983 a, b). Identificarea în aceste roci a staurolitului și a andaluzitului (Stan et al., 1981; Stan, 1984; Mărunțiu, Seghedi, 1983 a, b) demonstrează existența unui metamorfism inițial de grad ridicat.

În lucrarea noastră se arată că în rocile cuprinse, pînă în prezent, sub denumirea de seria de Vodna au fost identificate aceleași minerale — staurolitul și andaluzitul și, în plus, sillimanitul. Staurolitul și sillimanitul nu apar împreună, fiecare putînd fi, însă, însoțit de andaluzit. Relațiile dintre ele arată că staurolitul și sillimanitul preced apariția andaluzitului, primele fiind deformate și parțial sau total incluse în andaluzit. Andaluzitul nu este întotdeauna deformat dar, ca și celelalte două, este pinitizat uneori total.

Pe lîngă aceste minerale au mai fost identificate relicte de granat I, cloritizat și deformat, diopsid, hornblendă verde, plagioclaz și biotit cu pleocroism în tonuri brune-roșietice.

Mineralele relicte au fost identificate în roci care arată un metamorfism dinamic accentuat, asemănător cu cel descris anterior în seria de Corbu. Studiul microscopic relevă faptul că blastomilonitele formate în orogeneza caledoniană se caracterizează prin prezența în parageneza mineralogică a biotitului brun-verzui, a muscovitului și a granatului idiomorf și proaspăt. Biotitul și muscovitul apar în umbra de presiune a relictelor de minerale deformate și alterate, sau le învăluie, ceea ce presupune preexistența lor în roci.

Relictele de staurolit, sillimanit și andaluzit sînt interpretate ca fiind expresia unui metamorfism de grad mediu anterior blastomilonitizării. Ele pot reprezenta fie un metamorfism intermediar de presiune scăzută, cu o zonalitate inițială care nu poate fi recunoscută în prezent, fie două evenimente metamorfice suprapuse.

Deformările ulterioare (hercinice) au determinat formarea a noi plane S, pe care se formează clorit, muscovit (sericit), actinot, pe lîngă alte minerale.

Asemănarea proceselor petrogenetice, dinamica accentuată și conținutul litologic asemănător (blastomilonite, calcare cristaline, serpentinite) al „seriilor" de Vodna și de Corbu, diferite de cele ale seriei de Neamțu, conduce la ideea apartenenței lor la o entitate aparte, pentru care motive sînt descrise împreună sub numele de „formațiunile" de Corbu și de Vodna. Pe hartă sînt delimitate tectonic față de seria de Neamțu, delimitare care semnifică fie o pînză antehercinică, fie un melanj tectonic metamorfozat, fie o diviziune litostratigrafică a seriei de Neamțu, dar afectată de un metamorfism dinamic.

Cartarea geologică din ultimii ani a pus în evidență o linie tectonică orientată NV—SE care traversează oblic unitățile tectonice prealpine. Această falie pe care am numit-o falia Tișovița—Lucidol se caracterizează printr-o decroșare dextră de cîțiva kilometri a formațiunilor, afectînd și depozitele sedimentare carbonifer-superioare și permian-inferioare. Fiind în general înclinată spre NV ea se prelungește cu o falie inversă de aceeași vergență, semnalată de Stan (Stan et al., 1986) în bazinul văii Lucidol. Deplasarea compartimentului estic, blocată în zona Baia Nouă-

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Iuți, a condus la apariția unei sinforme în care formațiunea de Poiana Mr.aconia apare în ax. Falia Tișovița—Lucidol este un plan tectonic secundar, ramificat din grabenul Cernei, și prin efectul său tectonic se înscrie în ansamblul decroșărilor dextre ale acestuia.

EXPLANATION OF PLATES

Plate II

Fig. 1 — Staurolite (St) with opaque mineral inclusions, partly altered into pinite (P). Tourmaline (T) occurs locally in the sericitic masses. Corbu Formation. N II, × 60.

Fig. 2 — Plagiogneisses with biotite, staurolite and garnet. Vodna Formation. N II, × 30. Staurolite (St); Biotite (Bi II); Muscovite (M); Chlorite (Cl).

Plate III

Fig. 1 — Sillimanite (sil) porphyroclast with a partial andalusite (a) rim in a pinitic mass (P). Vodna Formation. N +, × 10.

Fig. 2 — Diopside (d) porphyroclast within a blastomylonite with actinote (Ac), epidote, quartz (q) and albite. Vodna Formation. N II, × 10.

Fig. 3 — Detail from Figure 1. Broken sillimanite with opaque minerals. At the margin of the sillimanite porphyroclast, andalusite (a) forms a rim. N II, × 70. Muscovite — M; Quartz — q; Pinite — P.

Plate IV

Fig. 1 — Sillimanite (sil) on whose cleavage penetrates andalusite (a). Marginally it alters into pinite. Vodna Formation. N II, × 10.

Fig. 2 — Deformed sillimanite (sil), substituted by andalusite (a) in a pinitic mass on which muscovite (M) is subsequently formed. Vodna Formation. N II, × 10.

Fig. 3 — Andalusite (a) formed at the expense of staurolite (st) from which nondigested relics remain. Muscovite (M) as undeformed crystals occurs on pinite. Corbu Formation, Eșelnița Valley, NII, × 23.

Fig. 4 — The advance of the andalusite (a) alteration releases partially or totally staurolite (st) and opaque minerals included. Muscovite (M) is subsequently formed on pinite. Corbu Formation. N II, × 23.

Plate V

Fig. 1 — Mylonite on micaschist with staurolite and garnet. Garnet I (G I) deformed and chloritized. Staurolite (st) altered up to complete substitution by pinite (P). Corbu Formation. N II, × 10.

Fig. 2 — Blastomylonite with garnet II (G II) and staurolite (st). Staurolite is distinguished by the pinite rim. Corbu Formation. N +, × 10.

Fig. 3 — Porphyroclast broken by common hornblende (Hb) in a blastomylonite with actinote and epidote. Vodna Formation. N II, × 10.

Fig. 4 — Porphyroclast of garnet I deformed and chloritized. Corbu Formation. N II, × 20.

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Plate VI

Fig. 1 — Blastomylonitic (Caledonian) foliation, marked by biotite (Bi), is intersected by a new (Hercynian) foliation on which chlorite and muscovite are formed. Garnet II (G II) relatioship with the older foliation is observed, biotite occurring in the garnet pressure shadow. Vodna Formation. N II, × 26.

Fig. 2 — In the oligoclase (Ol) pressure shadow biotite (Bi) is chloritized (cl) concomitantly with the appearance of more recent planes S. Vodna Formation. N II, × 26.

Fig. 3,4 — M porphyroclasts in a phylonite of the Vodna Formation. N II, × 60.

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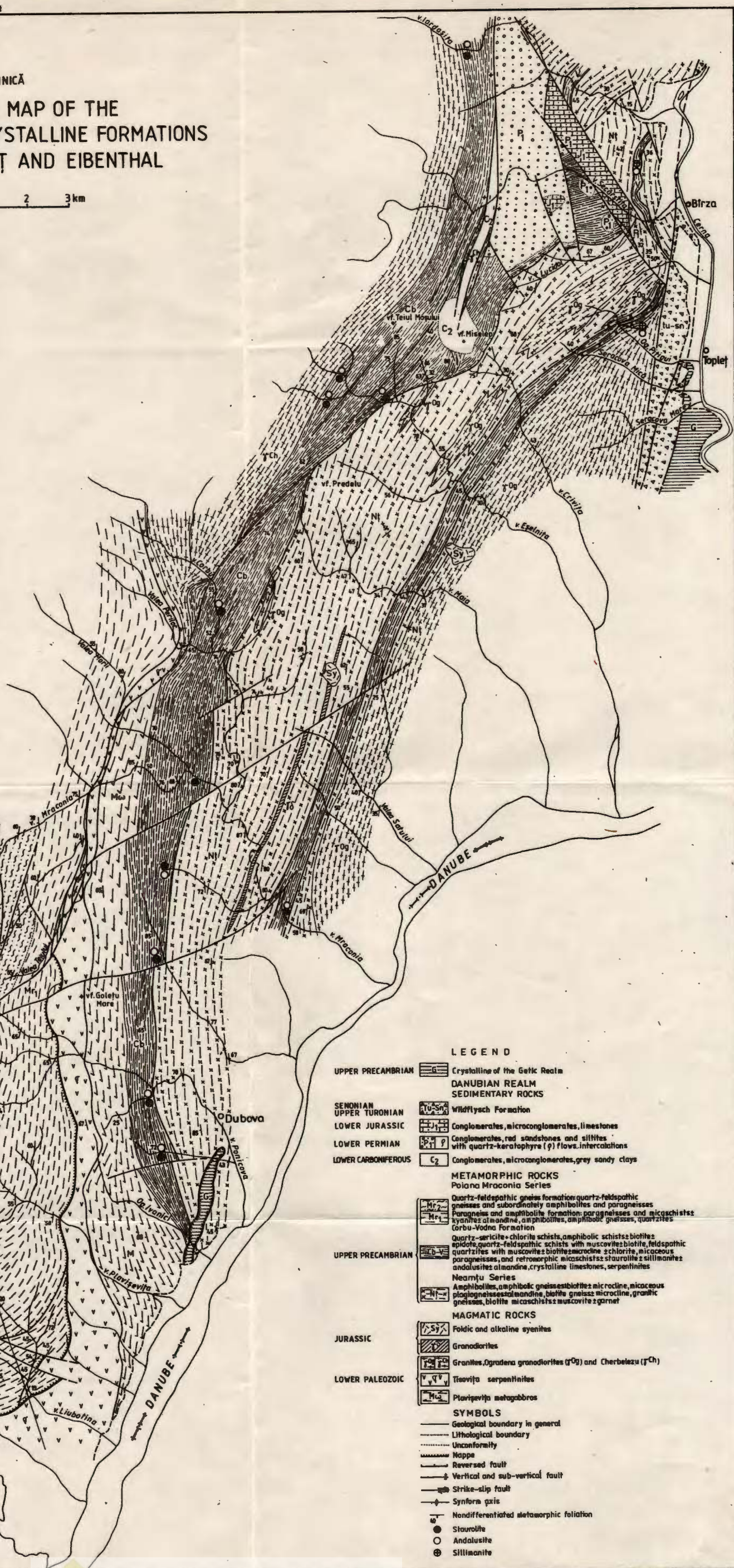
I. DINICĂ

GEOLOGICAL MAP OF THE DANUBIAN REALM CRYSTALLINE FORMATIONS BETWEEN TOPLEȚ AND EIBENTHAL

0 1 2 3 km



LEGEND

UPPER PRECAMBRIAN — G — Crystalline of the Getic Realm

DANUBIAN REALM SEDIMENTARY ROCKS

SENONIAN UPPER TURONIAN — Tu-Sn — Wildflysch Formation

LOWER JURASSIC — Cj-T — Conglomerates, microconglomerates, limestones

LOWER PERMIAN — P1 — Conglomerates, red sandstones and siltites with quartz-keratophyre (φ) flows, intercalations

LOWER CARBONIFEROUS — C2 — Conglomerates, microconglomerates, grey sandy clays

METAMORPHIC ROCKS

Poiana Mraconia Series

Mr2 / Mr1 — Quartz-feldspathic gneiss formation: quartz-feldspathic gneisses and subordinately amphibolites and paragneisses; Paragneiss and amphibolite formation: paragneisses and micaschists ± kyanite ± almandine, amphibolites, amphibolic gneisses, quartzites

Corbu-Vodna Formation

UPPER PRECAMBRIAN — Cb-V — Quartz-sericite ± chlorite schists, amphibolic schists ± biotite ± epidote, quartz-feldspathic schists with muscovite ± biotite, feldspathic quartzites with muscovite ± biotite ± microcline ± chlorite, micaceous paragneisses, and retromorphic micaschists ± staurolite ± sillimanite ± andalusite ± almandine, crystalline limestones, serpentinites

Nearmțu Series

Nt — Amphibolites, amphibolic gneisses ± biotite ± microcline, micaceous plagiogneisses ± almandine, biotite gneiss ± microcline, granitic gneisses, biotite micaschists ± muscovite ± garnet

MAGMATIC ROCKS

JURASSIC — Sy — Foidic and alkaline syenites

Granodiorites

γOg / γCh — Granites, Ogradena granodiorites (γOg) and Cherbelezu (γCh)

LOWER PALEOZOIC — V — Tisovița serpentinites

Mω — Plavișevița metagabbros

SYMBOLS

- Geological boundary in general
- Lithological boundary
- Unconformity
- Nappe
- Reversed fault
- Vertical and sub-vertical fault
- Strike-slip fault
- Synform axis
- Nondifferentiated metamorphic foliation
- Staurolite
- Andalusite
- Sillimanite

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CHLORITOID-BEARING BLASTOMYLONITES IN THE CIBIN MOUNTAINS: GENETICAL AND STRUCTURAL SIGNIFICANCES [1]

by

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Blastomylonite. Relic. Staurolite. Kyanite. Garnet. Secondary minerals. Chloritoid. Sillimanite. Polyphase processes. Retrograde metamorphism. Thrust fault. Sebeș-Lotru Series. South Carpathians — Crystalline Getic and Suprogetic realms — Cibin Mountains.

Abstract

In the Sebeș and Cibin Mts the chloritoid ± sillimanite-bearing blastomylonites extend along a regional lineamentar array. This consists of laminated rocks with relics of kyanite, staurolite and garnet, within which chloritoid grows statically under postkinematic conditions. Sillimanite is formed during the incipient retromorphic phases due to external retromorphic processes. It may stably coexist with chloritoid on a relatively restricted area on a P/T diagram. The blastomylonite plane is an ante-Alpine thrust plane.

Résumé

Blastomylonites à chloritoïde des monts de Cibin: significations génétiques et structurales. Dans les monts de Sebeș et Cibin, les roches blastomylonitiques à chloritoïde ± sillimanite sont diffusées sur un aréal linéaire à caractère régional. Celui-ci comporte des roches laminées à reliquats de disthène, staurotide, grenat, où le chloritoïde a un développement statique en des conditions postcinématiques. La sillimanite est un minéral issu pendant les premières phases de rétromorphisme, de quelques processus rétromorphes externes. L'association entre la sillimanite et le chloritoïde est stable sur un aréal assez réstreint d'un diagramme P/T. Le plan des blastomylonites représente un plan de chevauchement préalpin.

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Introduction

The deep-seated dynamometamorphic rocks have been studied from different points of view, very often resulting incomplete images or different understandings of the terms designating them. The term blastomylonite was introduced by Sander in 1911 to designate "mylonites which experienced subsequent complete recrystallization". This term was also used by Bellière (1958) and Antonioz (1971) for chemically reactivated assemblages consisting of relics affected by brittle and ductile deformations included into a recrystallized matrix. According to the above-mentioned authors, these rocks developed during three successive stages : ruptural deformation, plastic deformation and recrystallization. This conception is particularly based on the interpretation of the physiographic features and less on data on the geotectonic background ; so ambiguities still persist in discriminating blastomylonites as against polycyclic metamorphites.

Higgins (1971) and Sibson (1977) underlined the conditioning of blastomylonites by deep-seated shear zones, where extensive processes of recrystallization and neomineralization took place parallel to deformation.

Due to their characters, blastomylonites fall in the convergence field between mylonites and polycycle metamorphites ; there are, however, differentiation criteria between them.

As against mylonites, the distinction is simple ; mylonites display only minor recrystallizations, whereas blastomylonites show proofs of intense recrystallization and neomineralization. As regards the polymetamorphic rocks the differences concern the nature, extent and space distribution of the deformation. The deformation mechanism in the blastomylonite zones produces a planary form for them which on the map corresponds to generally narrow alignments.

In the Sebeș—Cibin Mts, the chloritoid-bearing rocks were mentioned in different places by Savu (1958).

Chloritoid-bearing blastomylonites have an areal or linear spreading covering relatively wide zones in the Sebeș and Cibin Mts. Their linear disposition was described by Hârtopanu and Hârtopanu (1976) first as chloritoid-bearing rocks and then as blastomylonites (Hârtopanu, 1982).

This paper deals with chloritoid-bearing rocks spread in an area lying south of Păltiniș resort or west—south—west of it, in the Cibin and Sadu rivers basins.

The study blastomylonites are found in the vicinity of the kyanite ± staurolite micaschists or more rarely in sillimanite-bearing rocks. They display the most characteristic spatial relationships with the kyanite-bearing rocks. These occur in the study area in the highest zones being included, from the lithostratigraphic point of view, in the metapelitic complex at the upper part of the Sebeș—Lotru Series (Fig. 1). As for the area with kyanite-bearing rocks south of Păltiniș (Surdu Summit, Bătrîna Summit, Oncești Summit) the chloritoid-bearing rocks border the kyanite-bearing rocks, approximately following their footwall along a sinuous alignment. Only in the spring zone of the Dăneasa Valley the spreading of the chloritoid-bearing rocks (blastomylonites) is areal.

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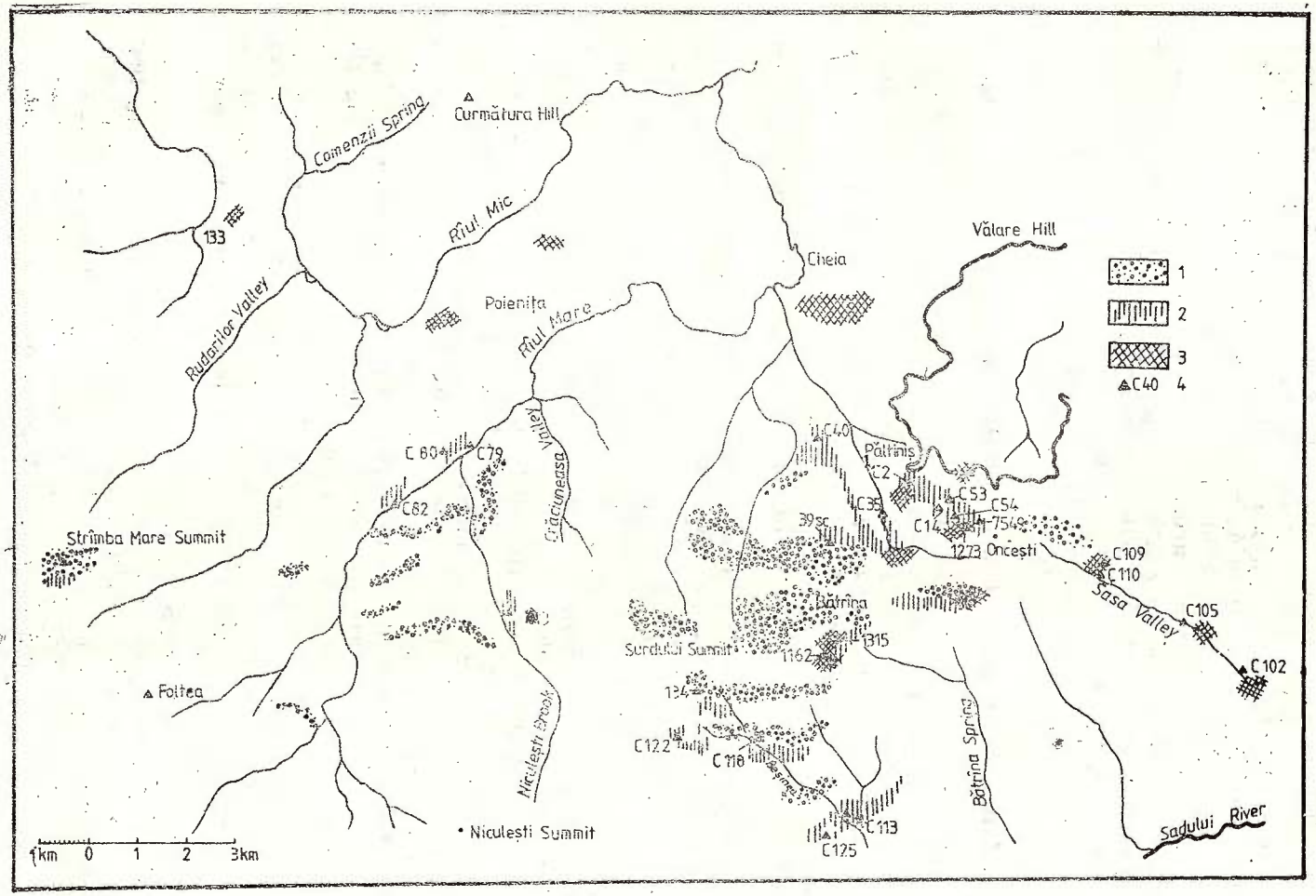

Fig. 1 — Sketch map of teh chloritoid-bearing blastomylonites in the Cibin Mountains.

1, kyanite-bearing rocks; 2, chloritoid-bearing blastomylonites; 3', sillimanite-bearing laminated rocks; 4, analysed samples.

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Their setting in the footwall of the kyanite-bearing rocks is found again in apparently disparate lenses of kyanite-bearing rocks situated in zones with less elevation, as in case of those upstream the Rîul Mare—Niculești confluence or in the spring zone of the Beșineu Valley.

The association of the sillimanite-bearing rocks or of the sillimanite mineral with the chloritoid-bearing rocks is common and from the point of view of the metamorphic zonality it represents a complication difficult to overpass. Blastomylonites put in association deformational elements which determine specific textures (generally laminated) as well as a neoformation mineral assemblage within which chloritoid is the most characteristic one, with relict minerals such as kyanite, staurolite and garnet.

1. Lithostratigraphy and Metamorphism of the Sebeș-Lotru Series

The Sebeș—Lotru Series was subject to some notable lithostratigraphic attempts in some massifs; this fact, besides the advantage of the systematization and ordering of several previous researches of the respective authors, has the disadvantage of validity in relatively limited areas. Thus, we mention Savu (1970) for the Semenic Mts, Bercia (1975) for the Godeanu Mts, Hârtopanu in Pop et al. (1985) for the Sebeș Mts, Hârtopanu, Iancu in Stănoiu et al. (1979) for the Mehedinți Mts, Savu, Schuster (1977) for the Lotru Mts, etc.

The major difficulties in tackling the lithostratigraphic problem first consisted in postulating the schistosity foliation as a unique structural element on the mesoscopic scale. This determined the neglect of the possibility of the existence of older cryptic S planes and, on the other hand, facilitated the merging of schistosity planes of different ages but with a differentiated penetrative character from one zone to another or from one rock to another (according to the degree of tectonic competence). In conclusion, the metamorphic study always assumed a monocycle character; there were no concerns as regards the differentiation on groups of synchronous minerals (parageneses) separated between them by deformational episodes. The proving of the metamorphism polycyclity is difficult in general and especially in case of the Cibin Mts where the last metamorphic phase obliterated almost entirely the traces of the previous phase. The cause which determined this mineralogical-textural "memory loss" originates in the advanced transposition phenomenon accompanying the deformations of the last metamorphic phase. Recent researches revealed relict mineralogical assemblages (parageneses) attesting that before the metamorphism visible nowadays and accompanied by deformations which resulted in foliation planes (S_2), folds, etc., the rocks in the study area had undergone another metamorphism of a similar intensity with different structural elements.

The relict foliation S_1 is hard to be preserved on a mesoscopic scale, surviving especially in zones with reiterated alternations of competent and incompetent rocks. Finally, a third metamorphic cycle with a retrograde character accomplished a deformation D_3 occurring in some areas (foliations and crenulation folds), materialized or not by chlorite or micas. The static chlorite is widespread, but its position versus the deformational cycle 3 is uncertain.

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The lithostratigraphic problem of polydeformed terrains implies a previous study of the metamorphism and deformations materialized by S planes of different types.

As a consequence of the fact that in some cases the metamorphic foliation coincides or is quasiparallel to the lithological limits and of the minimalization of the possibilities of the existence of other S planes than the primary ones on the maps with lithostratigraphic aims in the Cibin Mts effectuated up to now the foliation plane S_2 is the unique geometric element.

The most recent lithostratigraphic correlation of the South Carpathians (Sebeș—Lotru Series) has been made by Savu et al. (1973). These authors admitted that in lithostratigraphic respect the lower complexes of the series are developed in the Lotru Mts and the upper complexes occur in the Sebeș—Lotru Mts.

The Sebeș—Lotru Series was divided into five complexes among which the first three constitute the lower division — Upper Precambrian A — (Pc III A_1) and the last two the middle division (Pc III A_2). In the above-mentioned order the complexes are :

— C_1 complex — sillimanite-bearing paragneisses and crystalline limestones
— C_2 complex — mica paragneisses and orthoamphibolites
— C_3 complex — quartz-feldspar gneisses
— C_4 complex — micaschists and paragneisses
— C_5 complex — manganiferous schists.

According to the above-mentioned authors, complexes C_3, C_4 and C_5 occur in the area of the Cibin Mts ; complexes C_4 and C_5, included in an E—W trending syncline, are widely spread in the Păltiniș zone.

In lithostratigraphic respect, Hârtopanu (1982) considers that the Central South Carpathians consist of a lower granitoid complex and a high-alumina upper metapelitic complex. Between them there is a median complex mostly plagiogneissic. Numerous manganiferous rocks are intercalated in the upper complex. Amphibolic rocks are intercalated especially in the lower and middle complex. A similar lithostratigraphy has been established in neighbouring areas (Central and South Sebeș Mts), too (see Pui and Lupeni geologic map sheets, sc. 1 : 50 000).

2. Petrographic Features of the Chloritoid Blastomylonites

The mineral assemblage of the chloritoid blastomylonites is : sericite, kyanite, staurolite, garnet, sillimanite (± fibrolitic), quartz, muscovite, chlorite. The general textural aspect of these rocks is that given by the alternation of lamina of granular quartz interfingered with sericite (or pinite) spangles. Besides this banded aspect, texture can also be complicated by the porphyroclastic aspect, determined by the presence of relics of kyanite, staurolite or garnet, or by the porphyroblastic one, determined by the chloritoid occurrence. Muscovite and chlorite complicate more the textural aspects because it is difficult to establish in all cases whether these two minerals are pre-, syn- or post-kynematic with respect to the shear that generated the blastomylonitic rock.

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It is obvious that in some cases lamination manifested differentiatedly as regards the intensity on some plane sets ; it made possible the partial preservation of certain more resistant minerals (e.g. kyanite, garnet, staurolite and the afferent matrix) in the less deformed interspaces. On the other hand a "maturation" of the sericite laminas can be observed ; it displays an ever coarser granulation up to the developemnt of large muscovite crystals. At the same time the margin of the quartz crystals presents a lot of fringes for the fine-sericite stage and it is linear in the wide-developed muscovite stage (laminas and nodules).

The relationships that are or might be of a genetic significance are those between chloritoid, on the one hand, and sericite, kyanite, staurolite, garnet, on the other hand. The chloritoid/sericite relationship is the most frequent one. The chloritoid idioblasts are developed mostly in the sericite bands or nests, displaying a random orientation versus the rock schistosity, very rarely parallel to it. Very frequently chloritoid grows strictly within the sericite bands, the segments of these bands being now filled up by chloritoid or preserving their initial contour. More rarely chloritoid occurs as porphyroblasts within sericite. In both cases chloritoid displays a typical static growth under post-kynematic conditions. The position of the trails of inclusions with an undeterminable composition or consisting of sericite (probably sillimanite pseudomorph) also indicates the static character of the chloritoid blastesis. If the sericite spangles are oriented outside the chloritoid crystal area, they maintain this orientation inside it, too. This fact is also valid for other kinds of inclusions occurring in rows which do not change their array through the chloritoid crystal.

The relationships with kyanite, garnet and staurolite are mostly ambiguous and therefore it is difficult to state whether there is a substitution of one of these minerals or only their inclusion by chloritoid. In some cases, kyanite included within chloritoid as islands with the same crystallographic orientation seems to suggest a direct substitution (Pl. I Fig. 1).

However, it is not certain whether this is a previous selective substitution of kyanite by sericite and then a blastesis of chloritoid spatially superposed over the kyanite islands included into sericite. Both staurolite and garnet have been observed included into chloritoid or beside it (Pl. I, Fig. 2 ; Pl. II, Fig. 1), and it suggests that at least in some cases the two minerals can provide the substance and could represent the basis on which chloritoid nucleates.

Sillimanite, identified in the chloritoid-bearing blastomylonitic rocks, is in most cases a fibrolitic sillimanite. It could be identified in two situations : included into chloritoid and included into quartz. Certain pseudomorph probably of sericite after sillimanite, could prove the association of sillimanite with the chloritoid from laminas. Sillimanite included into chloritoid is quite rare, the relationships between the two minerals being hardly estimated. However, in some cases we can state that there is more than a simple inclusion. Thus, in Figure 1, Plate IV, within a lamina of chloritoid-bearing sericite, sillimanite needles trending along the lamina and included into chloritoid are extending outside it being, however, pseudomorphosed by sericite. In the same section sillimanite included into quartz occurs, too.

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The case of sillimanite included into quartz is more diversified than the previous one. Thus, two categories of sillimanite can be distinguished :

— inside the quartz crystal, more often as whiskers randomly or preferentially oriented after one or two directions ;

— sillimanite grown at the limit between two quartz grains, more widely developed after the crystallographic axes *a* and *b* and less after *c* (Pl. V, Fig. 1). It displays a general tendency to occur perpendicular to the quartz-quartz interfaces.

In comparison with chloritoid, this quartz seems to be younger as it substitutes the former one (Pl. V, Fig. 2 ; Pl. VI, Fig. 1).

3. Genetical Considerations

The petrographic analysis effectuated in a previous chapter points out the tectonite character of the chloritoid-bearing rocks as well as its metamorphite character. A concordance is observed between the schistosity planes of the metapelitic rocks generally occurring in the hanging-wall of the chloritoid rocks or at their level and a significant difference of dip between the foliations S_2 of the more competent rocks in the lower zones. In less competent rocks, a crenulation cleavage plane S_3, quasihorizontal or (very frequently) gently dipping, similar, as regards the size of the dipping angle, with the schistosity foliations from the above-mentioned metapelites and the chloritoid-bearing blastomylonites, can also be measured.

Considering these findings, we suppose that the deformational moment M_2 occurred on wide areas in the Central South Carpathians but the resulting transposition degree is very different according to the competence of the affected rocks. Thus, the kyanite-bearing metapelites from the upper lithostratigraphic complex have entirely been transposed after S_3, this plane being the only foliation visible nowadays. The more competent rocks but micaceous enough maintain the foliation planes S_2 and the crenulation lineation l_3 (obviously an intersecting lineation).

The gneissic rocks, more or less banded, the amphibolic rocks in the large-sized bodies, etc. underwent only the deformation D_2. Consequently, we can infer that the foliation of the chloritoid-bearing blastomylonites coincides with the foliation S_3 and therefore they might have been formed concomitantly. Our assumption is that at the level of the chloritoid-bearing blastomylonites the movement was much more intense in

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the mentioned S_3 planes, outside the blastomylonite level. This fact was determined by a strong destruction of the old paragenesis and an obvious neoformation of chloritoid and phyllosilicate minerals.

The existence of two superposed parageneses in the blastomylonitic-type rocks attests two moments of mineral blastesis. The former, with kyanite, staurolite, garnet and, probably, biotite, represents the old paragenesis, being in disequilibrium relationships with the new paragenesis represented by muscovite, chlorite, sericite, chloritoid. The latter could be subdivided into a previous paragenesis consisting of muscovite-chlorite, and a new one, with sericite and chloritoid.

The sillimanite place in the mentioned mineral parageneses is hardly specified particularly because of its acicular habitus which makes difficult the estimation of its relationships with other more coarse-grained minerals. However, two problems are significant :

— the sillimanite appurtenance to one of the above-mentioned parageneses ;

— the explanation of its existence in equilibrium relationships with the associated minerals, especially for the recent parageneses.

Although a high-grade mineral, sillimanite seems, in many cases to postdate minerals characteristic of lower grades. Its relationships of metamorphic compatibility with other minerals raised a particular interest towards the study of the microstructural relationships of sillimanite with the other minerals from the rock, one of the most specialized study belonging to Vernon (1975, 1977). According to him, these microstructural relationships can indicate whether the fibrolitic sillimanite grew before or after the final position of the limits between other mineral grains. In this respect he proposed two terms : "harmonious" used when the limits of other minerals seem to have been adjusted to the fibrolitic sillimanite, and "disharmonious" used when no such adjustment is obvious. The relationships in case of the "harmonious" sillimanite support the hypothesis that the sillimanite grains were in their place before the final position of the limits between other minerals of the aggregate. The relationships in case of the "disharmonious" sillimanite are in favour of the hypothesis that sillimanite grew after the final position of the grain boundaries. The microscopic observation on the chloritoid-bearing blastomylonites points out that the sillimanite from these rocks is the "disharmonious" type. In fact, one can notice that the interfaces between different grains, other than silimanite, meet the prismatic faces of sillimanite at random angles and orientations. In numerous cases sillimanite is concentrated at grain boundaries, displaying almost a rectangular position of the axis *c* versus this limit.

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In some cases radial sillimanite aggregates are developed as sheaves at the limit between the quartz/quartz grains especially when it is curved and has a very small radius of curvature.

All these possibilities point to a sillimanite growth after the establishing of the final position of the limits between other mineral grains, especially quartz, which in any case is one of the most mobile and, consequently, most newly formed minerals. Therefore, sillimanite is quite recently formed, in a fair agreement with the supposition that is belongs to the late parageneses of the blastomylonitic rocks.

The fact that sillimanite occurs in association with chlorite, sericite and chloritoid, epizonal-type minerals that outline a retrograde event at the expense of the initial rocks with kyanite, staurolite and garnet, cannot be integrated in the classic sequence of crystallization.

Therefore we have to accept the existence of a retrograded rock with sillimanite, a mineral whose existence in an environment outside its stability field must be explained.

The inclusion of sillimanite into chloritoid, quartz and probably beside muscovite makes us suppose its formation at the beginning of the retromorphic period.

The presence of sillimanite in the blastomylonitic rocks might be explained by the formation of these rocks in close connection with considerable deformations occurring at the level of some sets of parallel planes. Likewise, the finding of retromorphic parageneses in association with sillimanite determined us to state that these two phenomena — the laminar-type movement and the retromorphism — are the basis of the main effects that determined the appearance of this mineral.

According to Şeclăman and Hârtopanu (1987), the retromorphic reactions are exothermal, leading to the increase of the rock temperature (in case of hydrations up to 100 cal/g can be yielded). The heat of friction due to movements accompanying the rock deformations in case of intercrystalline slidings yields, according to the above-mentioned authors, 15 cal/g at a depth of 5 km and 40 cal/g at 15 km deep. These cummulated thermal effects result, on the one hand, in the inhibition of the retromorphic reactions (caused generally by the temperature decrease) and, on the other hand, the mineral neoformation that can culminate, in our case, with the formations of relatively lower pressure and higher temperature minerals, e.g. sillimanite.

Sillimanite occurs almost permanently beside blastomylonites (similar to those studied by us) pointing out that the phenomenon which determined the blastesis developed almost under the same conditions.

The coexistence of sillimanite with chloritoid within a relatively small area, as shown on the equilibrium diagrams of the Al_2SiO_5 polymor-

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phs and of some reactions including chloritoid, has to be taken into account (Fig. 2).

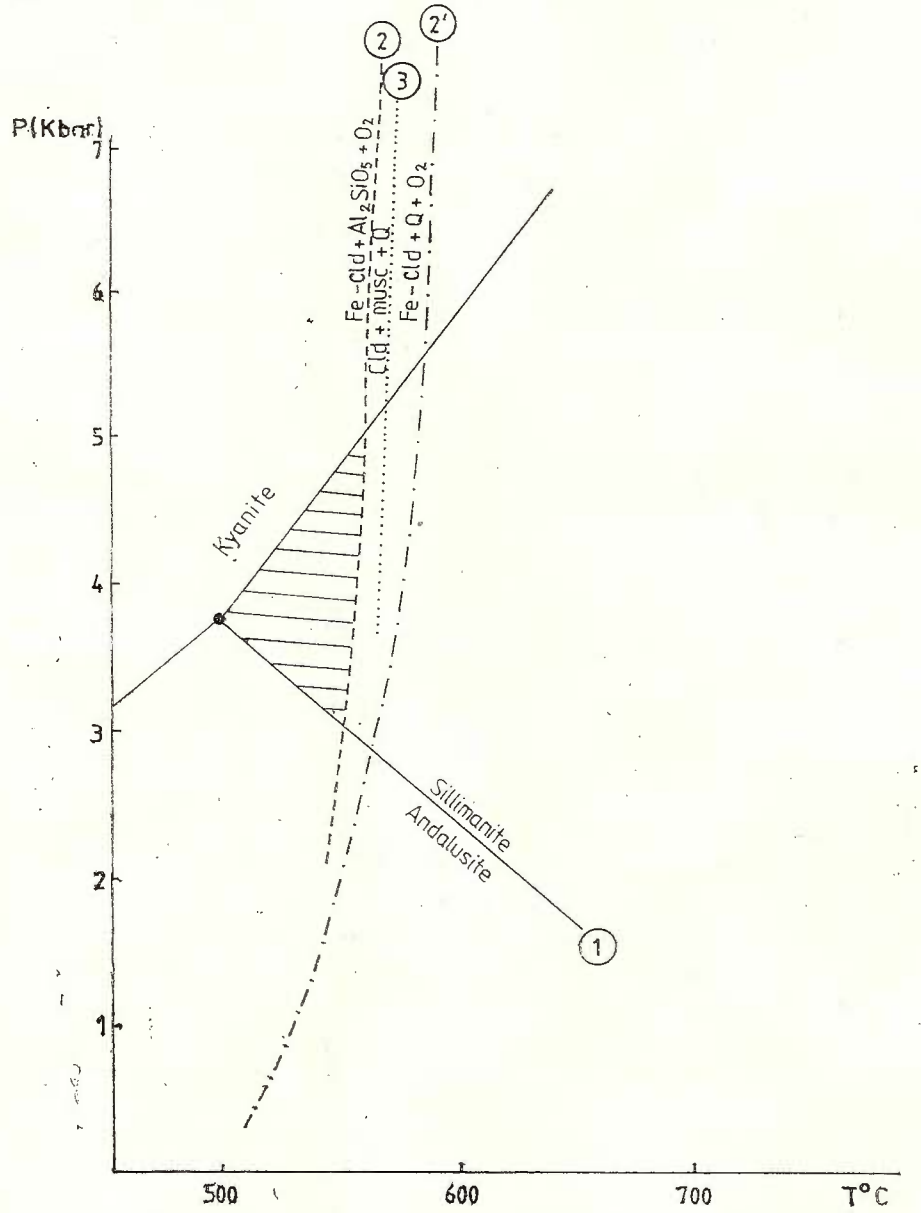

Fig. 2 — Stability curves of the aluminosilicates and chloritoid in the P—T field.
1, Holdaway (1971); 2, Richardson (1968): Fe—Cld + Al_2SiO_5 + O_2 = Fe — Staur + magnetite + Q; 2', Richardson (1968): Fe — Cld + Q + O_2 = Fe — Staur + Alm + magnetite + H_2O; 3, Ganguly (1968, 1969, 1972); Hoschek (1969): Cld + Ms + Q = St + Bi + magnetite.
Hatched: domain of stable coexistence of sillimanite with chloritoid.

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4. Structural Considerations

The geometric position of the chloritoid-bearing blastomylonites and the physical phenomena considered to have occurred and that determined their formation as such, make us infer that they materialize an old movement plane with obvious deformation effects. The plane could be identical as regards the position and simultaneous with plane S_3 or a bit younger but resembling it.

The competence degree of the affected rocks was of great importance as to the efficiency of the deformation caused by tangential movements at the level of this plane. The mineral blastesis concomitant or subsequent to the deformational movements determined the obliteration of the mechanical effects of the deformations or in any case the materialization of the shear zone. It is hard to estimate mesoscopically whether a rock from the "blastomylonitic zone" is intensely deformed and later on or concomitantly affected by a strong mineral neoformation or it is not affected by deformations following the selection principle according to competence. This ambiguity is often difficult to solve in case of the study rocks which often display a banded structure caused by the alternation of the typical chloritoid-bearing blastomylonitic rocks with those of common metapelites type.

In the area investigated by us the plane or surface of blastomylonites seems to outline the footwall of the upper level of metapelites as it results from the lithostratigraphy accepted up till now in the Sebeș-Lotru Series. The chemical composition, lithology and, finally, the mineralogy of the chloritoid-bearing blastomylonitic rocks do not differ essentially from those of the mentioned metapelitic rocks. Therefore one can admit that the blastomylouitic rocks represent the lubricant substratum of the upper metapelites of the Sebeș-Lotru Series in the area of study, at whose level the latter slid as against a more competent basement, involved only to a less extent in deformations. The moment of the shifting corresponds to a deep enough position of the deformed zone so that a notable mineral neoformation could be possible. A consequence of depth could be the particular way of development of blastesis and deformation not on a trenchant surface but rather on a "front" of a certain thickness, such as the spring area of the Dăneasa Valley where blastomylonites are found on an elevation span of ca 250 m. Consequently it results that in the area of study the upper metapelitic lithostratigraphic complex is in fact in a tectonic position as against the lower complexes, being separated from them by an old overthrust plane.

A possible consequence of these relationships could be the outcropping of small parts from the metamorphic zone of sillimanite belonging to the basement from under the upper metapelitic complex in the close vicinity of the blastomylonitic rocks. In the area of the Sebeș Mts, west-southwest of the study area, the kyanite + staurolite zone is separated on a length of several kilometres from the sillimanite zone by an uninterrupted lineament of blastomylonites + chloritoid. Therefore it is possible that the Sebeș-Lotru Series belonging to the Getic Nappe might be less monotonous as well as more complicated as regards the old disjuncti-

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ve tectonics and it might be, as a matter of fact, made up of different domains partly "uniformized" recently, being subjected to penetrative deformations subsequent to the mentioned tectonics.

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BLASTOMILONITELE CU CLORITOID DIN MUNȚII CIBIN: SEMNIFICAȚII GENETICE ȘI STRUCTURALE

152912

(Rezumat)

Rocile blastomilonitice se situează în domeniul de convergență între milonite și metamorfite policiclice. Față de rocile polimetamorfice, blastomilonitele se diferențiază prin amploarea și distribuția în spațiu a deformării.

În munții Sebeș și Cibin au fost întîlnite roci blastomilonitice cu cloritoid răspîndite pe un areal linear cu dezvoltare regională. Blastomilonitele se găsesc în vecinătatea rocilor cu disten sau mai rar în roci cu sillimanit. Asociate în ele apar elemente deformaționale ce determină texturi specifice, în general laminate, la care se adaugă o asociație de minerale de neoformare în care cloritoidul este cel mai caracteristic precum și minerale relicte cum ar fi staurolitul, distenul, granatul.

Asocierea mineralului sillimanit cu rocile ce conțin cloritoid este o situație frecventă, care din punct de vedere al zonalității metamorfice, reprezintă o complicație dificil de explicat. Metamorfitele getice ce alcătuiesc munții Sebeș și Cibin sînt metamorfite policiclice unde ultima fază metamorfică a șters aproape radical urmele fazelor anterioare. Cauza care a determinat această obliterare mineralogică-texturală rezidă în fenomenul de transpoziție foarte avansat, care a însoțit deformările ultimei faze metamorfice. De aceea, abordarea problematicii litostratigrafiei în aceste terenuri polideformate și polimetamorfozate implică dificultăți majore. Blastomilonitele sînt formate în timpul unor deformări ulterioare ultimului moment metamorfic. Cloritoidul este mai nou format și prezintă numeroase relații de includere și substituție față de granat. Față de sericit, cloritoidul arată o creștere statică tipică în condiții postcinematice. Sillimanitul este asociat atît cu cloritoidul cît și cu cuarțul. Cel mai frecvent tip de asociere este cel de tip „disharmonious", în care interfețele dintre diferite minerale întîlnesc fețele prismatice ale sillimanitului într-o poziție rectangulară a axului „c" față de această limită, ceea ce pledează pentru o creștere a sillimanitului după stabilirea poziției finale a limitelor dintre granulele de minerale. Deci sillimanitul este neoformat foarte recent și face parte din parageneza rocilor blastomilonitice. El s-a format la începutul perioadei de retromorfism. Relațiile retromorfe, care au loc, sînt apreciate ca fiind de tip exoterm, conducînd la creșterea temperaturii rocii.

Aceste efecte calorice pot produce inițial o inhibare a proceselor retromorfe și formarea în prima fază a unor minerale de temperatură ridicată, cum este sillimanitul. Acesta mai poate coexista dealtfel cu cloritoidul pe un spațiu relativ restrîns, într-o diagramă de P/T.

În aria cercetată de noi, planul de blastomilonite este plasat în baza nivelului superior de metapelite distenifere și materializează un plan de încălecare vechi, antealpin.

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EXPLANATION OF PLATES

Plate I

Fig. 1 — Kyanite substituted by chloritoid. N+, × 100.
Fig. 2 — Staurolite included into chloritoid. N+, × 100.

Plate II

Fig. 1 — Garnet substituted by chloritoid. N+, × 100.
Fig. 2 — Kyanite substituted by chloritoid. Series of inclusions in kyanite. N+, × 100.

Plate III

Fig. 1 — Sillimanite needles included into quartz and chloritoid formed exclusively in the sericitic matrix. N+, × 100.
Fig. 2 — Chloritoid oriented parallel to the sericitic matrix, on account of which it formed. N+, × 100.

Plate IV

Fig. 1 — Chloritoid intergrown perpendicular to the sericitic schistosity nondeviated at the contact with coarse-grained sillimanite, intergrown parallel to schistosity. N+, × 100.
Fig. 2 — Sillimanite associated with schlieren of ore and chloritoid. N+, × 100.

Plate V

Fig. 1 — Sillimanite intergrown at the limit between two quartz grains showing a tendency to develop perpendicular to the quartz/quartz interfaces. N+, × 100.
Fig. 2 — Recent quartz substituting chloritoid. N+, × 100.

Plate VI

Fig. 1 — Chloritoid substituted by quartz. N+, × 100.
Fig. 2 — Chloritoid schlieren in continuation of sericite schlieren. N+, × 100.

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TORONIȚA ZONE AND PROBLEMS OF THE PRE—ALPINE METAMORPHIC BASEMENT OF THE GETIC AND DANUBIAN REALMS[1]

by

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Abstract

This paper brings arguments on the basis of complex criteria in favour of the appurtenance of the Toroniţa "Zone" or "Series" to the Getic Realm. It belongs to the Ravensca Formation of Proterozoic age and is individualized by the mostly basic lithological assemblage (amphibolites, gabbros ± peridotites) in relation to the mostly gneiss-micaceous sequences of the Sebeș-Lotru Group. These features as well as the tectonic relationship, in a constant lower geometrical position, marked by an ante-Paleozoic unconformity, make possible its appurtenance to the Luchiţa Group (including also the Jidoștiţa and Ivanu formations in the Porţile de Fier and Godeanu outliers, belonging to the Getic Nappe).

Résumé

Zone de Toroniţa et la problématique du soubassement métamorphique préalpin des domaines gétique et danubien. L'étude vient d'argumenter, sur base des critaires complexes, l'appartenance de la „zone" ou de la „série" de Toroniţa au domaine gétique, elle étant englobée dans la formation de Ravensca, protérozoïque. L'association lithologique, basique en grande partie

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(amphibolites, gabbros ± péridotites), individualise cette „série" par rapport aux séquences gneissiques-micacées du groupe de Sebeș-Lotru. Ces caractères, tout comme la relation tectonique en position géométrique invariablement inférieure et marquée par une discontinuité anté-paléozoïque indiquent son appartenance au groupe de Luchița (qui renferme aussi les formations de Jidoștița et Ivanu, des lambeaux de Porțile de Fier et Godeanu appartenant à la nappe gétique).

1. Introduction

"Toronița Zone", situated in the area of the Danubian Realm (Codarcea, 1940), in the western part of the Ielova Series, has been later on individualized as the "Toronița Series" (Dessila-Codarcea, Codarcea, 1968; Bercia, Bercia, 1975) and regarded as a Riphean-Cambrian epimetamorphic sequence normally overlying the Precambrian metamorphites of the Ielova Series. The Toronița Series in the Danubian Realm initially consisted only of the formations in the left side of the Orevița Valley, situated east of the "Rudăria Fault" (Bercia, Bercia, 1975; Gunnesch, Gunnesch, 1976), following the lower course of the Orevița Valley (Codarcea, 1940).

West of the Rudăria Fault, in the Getic Realm, in relation to the Sebeș-Lotru Series, the same authors separated the "Miniș Series" considered of a similar age and lithological content with the Toronița Series. Recently, on the geological map of the Socialist Republic of Romania, scale 1 : 50 000, Sichevița Sheet (Maier, Mărunțiu in Năstăseanu et al., 1981), the "Toronița Series" is delimited eastwards by the Rudăria Fault whose course has been changed (as compared with Codarcea's interpretation) in the southern part. Westwards, the Toronița Series extends up to the vicinity of the Sichevița Valley including also what previous authors attributed to the Miniș Series. The relationships of the Toronița Series (acc. to this interpretation) with the formations of the suprajacent Sebeș-Lotru Series are tectonic and consequently it results that the Sebeș-Lotru Series belonging to the Getic Nappe thrusts over the Toronița Series belonging to the Danubian Realm. Thus, the Upper Cretaceous-Paleocene banatitic rocks occur both in the Getic and in the Danubian Realm.

As one will see from the petrographic data of the pre-Alpine basement, the relationships between different lithostratigraphic units and the features of the Mesozoic cover, the Toronița "Zone" or "Series" belongs to the Getic Realm and is part of the pre-Alpine unit of the Getic Nappe.

The recognition of the petrographic assemblages and of the lithostratigraphic sequences of the Proterozoic and Paleozoic metamorphites implies an accurate clearing up of the major Alpine and pre-Alpine tectonic discontinuities. At the same time, the correct clearing up of the normal or tectonic relationships in the pre-Alpine stages depends on the degree of knowledge of the Mesozoic formations, the delimitation of the two realms, included into basement nappes, requiring the tackling of an interdisciplinary problem. Thus, the problem of the eastward delimitation of the Getic Crystalline and of the pre-Alpine entities and structures in both domains cannot be solved by the "linking" of "lines" representing either faults

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or traces of the thrust planes or by their "shifting" and it requires complex arguments referring to different aspects :

— position and evolution of the cover nappes between the two realms and the recognition of the Mesozoic covers specific to the basement nappes ;

— clearing up of the Alpine structures that, from all the three realms — Getic, Severin-Arjana and Danubian (Codarcea, 1940)—had a polyphase character, with spatial superpositions during the major Austrian and Laramian tectogeneses ;

— the recognition within the Alpine nappes of the petrographic assemblages and of the Paleozoic lithostratigraphic sequences involved in the tectogeneses of the Variscan (Hercynian) and Caledonian cycles ;

— the reconstitution of the Proterozoic lithostratigraphic sequences as well as of the older contacts or unconformities, cropping out discontinuously and fragmentary in the Alpine and Paleozoic nappes in the area of the Getic and Danubian realms (Pl. II, IV).

2. "Toroniţa Zone" — Area of the Getic Realm

2.1. General Geological Data

The zone lying between Şopotul Nou and Liubcova, within which the Toroniţa "Zone" is structurally included, is represented by several entities (Pl. I) :

— Middle Miocene sedimentary deposits of the post-tectogenetic cover (post-Laramian) ;

— Upper Cretaceous-Palcocene banatitic magmatites ;

— formations constituting the Getic Nappe :

a) formations of the Mesozoic cover ;

b) Sichevița-Poneasca granitoids of Carboniferous age ;

c) Proterozoic and Paleozoic metamorphites in association with magmatites included into Variscan and older structures.

Further on we shall briefly deal with the first entites, stress being laid on the Mesozoic cover and the metamorphic formations from a structural point of view.

— The sedimentary deposits of the posttectonic cover belong to the Bozovici and Sichevița basins (Năstăseanu et al., 1981) and are represented by conglomerates, gravels and sands with discontinuous levels of limestones, tuffs and coal of Badenian age.

— The banatitic magmatites in the Liubcova-Şopotul Nou alignment (Şopot Zone inclusively) are found in the Getic Realm. A recent study (Russo-Săndulescu et al., 1987) points out their Paleocene-Lower Eocene age as well as their correlation in time with the second intrusion phase of the plutonic banatites in the west of Banat (Russo-Săndulescu et al., 1984). In relation to the Laramian paroxismal tectogenesis they can be regarded as post-Laramian, unlike the pre-Laramian extrusive and intrusive produces in the Timoc Zone where the formation synchronous to the Şopot Zone (Turonian-Lower Senonian) is volcano-sedimentary (Iankovič et al., 1981).

In comparison with the External Dacides paleorift (Rădulescu, Săndulescu, 1973 ; Săndulescu, 1984), the banatites in the Şopotul Nou-

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-Liubcova Zone apparently occur in the extension of the Timoc Zone; they could be considered more recent and with a more western position.

2.2. Formations Constituting the Getic Nappe

A) Mesozoic Cover. It is represented by: a) Carbonatic rocks in Urgonian facies (from the Jurassic-Lower Cretaceous sedimentary cycle), of Lower Cretaceous age (Codarcea, Pop, 1963); b) "Șopot Zone"-type sedimentary formations of the post-Mesocretaceous Cover (Codarcea, Pop, 1963; Pop, 1968).

The Lower Jurassic mentioned by Andrei et al. (1982) in the Bozovici Zone, also found south of Tîlva Poloamelor, in the area of the "Ielova Series", is represented by small patches of quartzites and quartzitic sandstones with minor coal interlayers preserved along fractures. These exposures, found in the Getic Realm, do not differ from those synchronous in the Danubian Realm, being typical of the Lower Liassic in Gresten facies, marginal or ridge subfacies (cf. Năstăseanu et al., 1974, 1977).

a) The Lower Cretaceous cropping out in the area is represented by isolated occurrences of reef limestones in Urgonian facies similar to the Barremian-Aptian limestones in the Reșița Zone, being of the same age (Codarcea, Pop, 1963; Pop, 1968). According to the mentioned authors they represent remnants of a transgressive carbonatic sequence, formed under shallow marine conditions, that occupied a large part of the Getic Realm as a result of the Austro-Alpine transgression (Pop, 1968), before the outlining of the Șopot "Zone" or "Trough". Urgonian limestone occurrences were found at Tîlva Toronița and in the right side of the Orevița Valley, NW of Liubcova (Tietze, 1870, 1872; Böck, 1881; Schafarzik, 1903; Schréter, 1913, in Codarcea, Pop, 1963).

The mappings also evidenced the preservation of small occurrences of white massive limestones of Urgonian type (a diagnosis confirmed by Pop, oral information) in the area of the "Toronița Series" considered by the authors of this paper to belong to the Getic Realm. The limestone occurrences in Urgonian facies crop out in both units separated by us in the Toronița Zone (Pl. I), which point to a pre-Alpine age of the contact between them.

b) Upper Cretaceous. In this area of the Getic Realm the Upper Cretaceous belongs to the succession described as the "Șopot Zone" by Codarcea (1940). It is represented by deposits constituting complexes of sandy-clayey rocks and marly-limestones partly similar with deposits of the same age found in the Hațeg Zone and the southern part of the Poiana Ruscă massif (Pop, 1968), covering the Cenomanian-Lower Senonian time span. The outlining of the sedimentary trough in the Șopot Zone took place after the ceasing of the Austrian movements (Albian) resulting in the Vraconian-Cenomanian transgression in the Getic Realm.

Detrital rock sequences representing the Upper Cretaceous in the Șopot Zone were separated and described by Codarcea (1940), Codarcea,

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Pop (1963) and Pop (1968) in the Sicheviţa Crest, west of Tîlva Nucului and Tîlva Toroniţa, where numerous bodies of banatitic magmatites are exposed, accompanied by thermal and metasomatic contact phenomena.

The researches carried out south of Tîlva Toroniţa pointed out that formations of this type, belonging to the Upper Cretaceous in the Şopot Zone, also crop out as isolated occurrences in the area of the Toroniţa "Zone" or "Series" (Codarcea, 1940), subsequently assigned to the Danubian Realm (Pl. I). These occurrences, preserved either as small erosion bodies or as lenticular bodies along post-Laramian fractures, are represented by :

— sandy limestones, limestones, micaceous carbonatic sandstones and marly limestones corresponding to the lower part (Cenomanian-Turonian) of the Şopot-type sequence (both sides of the Oreviţa Valley, Ogaşul Morii) ;

— conglomerates, sandstones and breccias that seem to correspond to the transgressive Senonian sequence (Ciocaru Summit, left side of the Oreviţa Valley, the crest between the Oreviţa Valley and the Brestenic Brook). The breccias in the Ciocaru Summit (Mărgeanu breccias) contain numerous elements of retromorphic green "crystalline". Blocks of limestones in Urgonian facies occur in the base (or reworked).

An occurrence of marly limestones situated in a secondary crest in the left side of the Oreviţa Valley contains pelagic microorganisms — globigerines, globotruncanes, rotalipores and calcispherulides (determined by Pop) — specific to the Turonian in the Şopot Zone.

In most of the exposures the described rocks are strongly folded, cleaved, lentilized and sheared as a result of the deformations accompanying the Laramian phase. The disjunctive tectonizations (breccifiations, cataclazites, slikensides) accompany several distension or compression fractures, approximately with a N—S trending. The oldest ones are accompanied by filling material and by thick breccias and can occur beside the banatite occurrences. The faults of the Rudăria system of Miocene age delimit eastwards the Bozovici basin.

In conclusion, the characters of the Mesozoic cover attest the appurtenance of the subjacent formations to the Getic Realm and the disposing of the Mesozoic formations in both units as well as on the contact between them prove its pre-Alpine age.

The characters specific to the post-Austrian Cover in the Şopot "Trough" are recognized towards NNW, in the south of the Poiana Ruscă massif (Pop, 1968) where they lie conformably over the Lower Supragetic Nappes (Moniom, Tîlva Drenii and Bocşa, cf. Iancu, 1985 ; Iancu in Ghenea et al., 1985) as well as the pre-Austrian deformed cover of the Getic Nappe. South of the Danube, in the Timoc area, the synchronous sequence is a volcano-sedimentary formation with associated magmatites (Iancovič et al., 1981).

2.3. Relationships of the Getic Crystalline with the Danubian Units

The extension and relationships of the "Toroniţa Series" (in fact it represents a sequence of the pre-Alpine basement of the Getic Realm), with the adjacent metamorphites and the Mesozoic formations occurring

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in the "Infragetic" position (Iancu, 1985), do not represent only a mere mapping problem but also a clearing up and an understanding of the evolution of the two realms yielding the basement nappes : Median and Marginal Dacides (cf. Săndulescu, 1984), corresponding to the Alpine, (Getic and Danubian) realms, as well as their relationships with the formations of the Severin Realm (sensu Codarcea, 1940) yielding the Infragetic Nappes belonging to the External Dacides (Săndulescu, 1984). In this case the authors distinguished (Pl. II) :

— Getic and Supragetic Nappes, in the Getic Realm (Median Dacides) ;

— Infragetic Nappes originating in the Arjana-Severin Realm (External Dacides) ;

— Danubian Nappes, in the Danubian Realm (Marginal Dacides).

The problem of the delimitation of the metamorphites belonging to the two Alpine realms (individualized since the Lower Jurassic) within which pre-Alpine structures are exposed (Pl. II) as well as the explanation of the location and genesis of the Alpine metamorphites (in the Infragetic Units) require a complex study from various points of views :

— The diversification of the facies of the Getic and Danubian realms (that constituted one body at the end of the Variscan orogenesis) from the Lower Jurassic, concomitantly with the "opening" of the Severin trough (Codarcea, 1940), for which we consider as an argument the allochthonous position of the Arjana Nappe (sensu Codarcea, 1940) represented by the Feneș and Căleanu Austrian Digitations, with associated Mesocretaceous wildflysch and post-Austrian cover (Iancu et al., 1989, Pl. I), as well as the alkali character of the associated initial magmatites of intraplate type (Russo-Săndulescu in Conovici et al., 1987). In this context, the evolution in time and space of the basement units is in direct correlation with the evolution of the Tethysian major suture and of the paleorift within the European continental margin (Rădulescu, Săndulescu, 1973 ; Săndulescu, 1984).

— The recognition of the typical features of the formations of the Jurassic-Lower Cretaceous sedimentary cycle in the three realms (Codarcea, 1940) as well as of the lithofacial affinities of the sequences in the marginal zones of the rift (marginal subfacies, cf. Năstăseanu et al., 1974 ; 1977) with the Getic Realm in the west (Săndulescu et al., 1986, Brașov area) and the Danubian Realm in the east, provided that one accepted the allochthony of the Arjana Nappe over the Internal Danubian Units : Sirinia Nappe (Năstăseanu et al., 1988) an equivalent of the Poiana Mărului (Stănoiu, 1973 ; Kräutner et al., 1981), and the Cornereva Nappe (Năstăseanu et al., 1988). This opinion is uphold by Iancu et al. (1989, Pl. I). At present, Năstăseanu and Negrea (in Conovici et al., 1987) consider that the facies in the Arjana-Feneș-Căleanu belong to the cover of a large basement unit — Muroniu Nappe — to which they include the Brustur Formation belonging to the basement Danubian Nappes. In this zone the external limit of outcropping of the Getic Crystalline is represented by the Rudăria Fault (Năstăseanu et al., 1981), except for small occurrences.

The recognition of the same petrographic assemblages of the Getic Nappe (to which the "Toroniţa Series" belongs) also in the "Cozla Nappe",

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(Năstăseanu et al., 1981) makes possible the extension of the "Getic Crystalline" up to the Cozla-Cîrșia Ielovii zone where Codarcea (1940) mentioned rocks identical with those in the Arjana Nappe (Radiolarite-bearing red limestones and an Upper Cretaceous cover).

The acceptance, on large distances, of the allochthony of the units yielded by the Arjana-Severin Realm as well as of their composite character explain their emplacement over the Danubian structural assemblage in the Laramian phase, concomitantly with the Lower Getic and Supragetic Nappes (Iancu, 1986) and makes possible a unitary interpretation and solves field evidence apparently contradictory.

2.4. Pre-Alpine Basement of the Getic Realm

In lithologic and petrographic respect, the metamorphites of the "Toroniț̦a Series" can be correlated with those of the Ravensca amphibolitic formation, represented on the Sichevița Sheet (Maier, Mărunțiu, in Năstăseanu et al., 1981).

The individualization of the "Orevița Unit" and of the "Ravensca Formation" in relation to the micaceous gneissic rocks of the "Bîrzu Formation", tectonically in an upper position, led also to the partial revision of the structural sketch of the Pre-Alpine Units (Iancu et al., 1988) by the separation of the Orevița Unit over the Miniș Unit and under the Nera Unit.

These units are crossed by the Sichevița-Ponească granitoid of Carboniferous age (Streckeisen, 1934; Codarcea, 1940; Bîrlea, 1976; Stan et al., 1986, 1987) that generated a thermal metamorphism superposed to the mineral assemblages specific to these units (Iancu, in Stan et al., 1986).

The direct correlation of the units already separated determined the estimation of the petrographic content of the following units: Bozovici Unit, in a lower geometric position, Miniș Unit, Orevița Unit, Nera Unit.

a) *Bozovici Unit* consists of:

— Proterozoic polymetamorphic formations of the Sebeș-Lotru Group;

— Buceava Formation of Middle Paleozoic age.

The Sebeș-Lotru Group, cropping out on small areas (Pl. I), is represented by an assemblage of quartz-feldspathic gneisses and amphibolites with plagiogneiss and micaschists intercalations belonging, in the Bozovici Zone, to a middle sequence of the Sebeș-Lotru Group: the leptino-amphibolitic horizon or formation (Savu, 1973; Întorsureanu et al., 1982).

The Buceava Formation (Streckeisen, 1934, redefined by Iancu et al., 1987) crops out between the Sichevița and Găvojdia valleys and is represented by: metaconglomerates, green and cherry-coloured metasandstones (locally "spotted"), argillites and carbonatic rocks with vacuolar metabasalts and bodies of metadiorites and metagabbros (Pl. I). Conglomerates, unconformably overlying Proterozoic retromorphites, mark the basal part of the sequence that forms a NS-trending syncline.

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Northwards, especially in the Miniș Valley basin, magmatites predominate and are represented by : basalts, diorites, gabbros, in association with black or green siltites and with carbonatic rocks.

This sequence, considered by Savu (1973) of Proterozoic age, has been separated by Andrei et al. (1982) as "volcano-sedimentary formation" of Paleozoic age.

The lithological characters and the simple deformations, accompanied by a low-grade metamorphism, ungeneralized (all rocks preserve premetamorphic inherited characters), make possible the comparison with the Valea Satului Formation of the Moniom Nappe (Iancu, 1985), of Devonian-Lower Carboniferous age (Visarion, Iancu, 1985).

The association of terrigene and carbonatic rocks with basalts, peridotites (occurring only in the Valea Satului Formation) and minor intrusions of gabbros and diorites points to the existence of a basin with an oceanic-type crust in the Middle Paleozoic (Devonian), which yielded the mentioned petrographic assemblage involved in the Variscan orogenesis (Breton? and Sudete phases). The low-grade regional metamorphism was followed by the emplacement of the Variscan Nappes and the intrusion of the Sichevița granitoid during the Carboniferous (Bîrlea, 1976). The same relationships (post-overthrust intrusion) are also observed north of the Miniș Valley, the Poneasca granite representing the extension of the Sichevița one (Streckeisen, 1934; Codarcea, 1940).

b) Miniș Unit, well exposed on the eastern flank of the Sichevița-Bozovici post-nappe anticline, consists only of rocks of the Miniș Formation (Codarcea, 1940; redefined by Iancu et al., 1988). This formation is monotonous lithologically and is constituted of quartzitic schists, more or less micaceous, locally coarse-grained (resembling metasandstones) or fine-grained (pelitic), and rare metarhyolite lenses with inherited structures and minerals. The polydeformed character, the more intense metamorphism at the level of the greenschist facies (the sedimentary structures are difficult to recognize) make us consider it older than the Buceava Formation, possibly Lower Paleozoic in age, with an initial prograde Caledonian metamophism by comparison with similar rocks in the Locva massif.

The Miniș Formation is in tectonic relationships with the Buceava Formation and its Proterozoic basement, and it determined us to regard this contact as Variscan. The tectonic contact with the upper Orevița unit can be Variscan or Caledonian taking into account the intense retromorphism of the older Ravensca Formation.

c) Orevița Unit is represented in this area by a sequence considered to belong to the Ravensca Formation, of Proterozoic age (Maier, Mărunțiu, in Năstăseanu et al., 1981). The petrographic types participating in the formation of the Orevița Unit, well preserved in the eastern flank of the Bozovici-Sichevița anticlinorium (Pl. I), are hard to recognize in the intensely retromorphosed sequences and therefore they have been assigned by previous authors either to the "Buceava Series" or to the "Miniș Series". In the area referred to the Danubian Realm they have been called the Toronița "Zone" or "Series".

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Our arguments concerning the Mesozoic cover, specific to the Getic Realm, and the direct correlation with the pre-Alpine structural units in the Bozovici-Șopotul Nou Zone (Iancu et al., 1987) allow us to assign the whole sequence exposed in the Orevița Unit to the Ravensca Formation, partially separated by Maier, Mărunțiu (in Năstăseanu et al., 1981) within the Sebeș-Lotru Series.

The Ravensca Formation (Pl. I) consists of an assemblage of amphibolites, quartzites and quartzitic or micaceous plagiogneisses with rare limestones lenses. Some amphibolites are of magmatic origin, being generated at the expense of gabbroic rocks from which massive nuclei with inherited (pre-metamorphic) structures and minerals are preserved.

Such polymetamorphic metagabbros, in places with blastomylonitic-type structures or strongly amphibolitized marginally occur in this formation both southwards (including the Toronița Zone) and northwards, in the Bozovici Zone, in the retromorphite sequence assigned by Iancu et al. (1987) to the Sebeș-Lotru Group, Nera Unit. In this zone, the amphybolitic sequence also included lenticular bodies of serpentinized metaperidotites.

Sometimes the whole assemblage referred by us to the Ravensca Formation is retromorphosed, other times it preserves thick packets of welded microblastic rocks, with banded or plastic flow structures, with a biotite or green-bluish amphibole blastesis representing old, ante-Caledonian blastomylonites (as products of the dynamic metamorphism). This discontinuity marks a Proterozoic tectonic contact between the visible lithofacial and petrographic assemblages; it has also been emphasized outside the Paleozoic reactivation areas in the Semenic Mts, Pîrvova-Nera Valley zone, in the area referred by Kräutner et al. (1981) to the Cibin Group, Marisian Supergroup.

Both the quartzitic-amphibolitic lithologic assemblage and the tectonic discontinuity tectonically marked by deep-seated blastomylonites (with biotite and amphibole) and the constant lower geometrical position are identical to those in the Porțile de Fier and Godeanu Outliers (eastern zone), where the Jidoștița-Ivanu Formation (Iancu et al., 1982; Iancu, 1985) occurs as islands from under formations of the Sebeș-Lotru "Group" (an equivalent of the Sebeș-Lotru "Series"), without referring it nowadays to one of the known "groups" of the Getic Crystalline.

The possible lithostratigraphic assigning of the probable equivalent Ravensca and Jidoștița-Ivanu formations will be tackled later on.

d) Nera Unit, the uppermost unit exposed in this part of the Getic Nappe, consists of a formation or subgroup (as part of a "group" of formations found in normal sequence with primary petrogenetic affinities evidenced by the associated magmatites, by the petrochemical features), of the Sebeș-Lotru Group *(senso lato)*. The sequence exposed in this area, in the Nera Nappe, has been separated by Maier, Mărunțiu (in Năstăseanu et al., 1981) as "Bîrzu Formation" and is represented by mostly micaceous and quartz-feldspathic rocks: micaschists, plagiogneisses, leucocrate gneisses, with frequent migmatites and pegmatites.

The Bîrzu Formation, like other sequences of the Sebeș-Lotru Group (Godeanu massif, Mehedinți Plateau), shows structural and mineralogical

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proofs indicating the superposition of ante-Paleozoic dynamothermal events, typical of facial series with a medium up to high (M_1) or medium to low pressure (M_2) at the level of the amphibolitic facies with possible transitions towards the granulitic facies (e.g. Mehedinți Plateau, where pyroxene-bearing rocks are preserved as relics from M_1).

2.5. "Ielova Series" — Getic or Danubian Crystalline (?)

The eastern and northeastern prolongation of the Ravensca Formation (to which belongs the "Toroniţa") and its appurtenance to one of the "groups" or "series" known as representing pre-Caledonian, Proterozoic retromorphites raise the problems related to:

— the lower limit of the Getic Units (Austrian and Laramian or Austrian reactivated by the Laramian paroxismal tectogenesis);

— the recognition of the pre-Alpine structural units (Paleozoic and ante-Paleozoic);

— the reconstitution of the lithostratigraphic sequences and of the premetamorphic primary petrogenetic assemblages, allowing the separation of "groups" of formations genetically and evolutionary affiliated, even if they had been subdivided during the younger orogenic cycles: Caledonian, Variscan (Hercynian) and Alpine.

In this sense, the crystalline of the Ielova "Series", cropping out east and northeast of the mentioned formations belonging to the Getic Nappe and described in all papers as representing a series within the Danubian Realm (Cozla-Svinecea Nappe or Nappes, cf. Năstăseanu et al., 1981; Năstăseanu et al., 1988), can represent:

a) An Alpine unit belonging to the Getic Nappes Group in direct relationship with formations identical with those from the Arjana Nappe (Codarcea, 1940), preserved in what Codarcea called Cozla-Cameniţa-Cîrşia Ielovii "syncline" and proved by Năstăseanu et al. (1988) to represent an allochthonous wildflysch with Ielova-type crystalline blocks. Our recent researches carried out east of the Rudăria Fault, in the Toroniţa--Recica-Cozla Zone, evidenced the prolongation of the "crystalline" of the same type as the "Getic" one. Here one could recognize both metabasites and metaultrabasites, similar to those in the Ravensca Formation, eclogites with relict garnet and granulitic rim and matrix with brown hornblende, and amphibolites in association with leucocrate gneisses, resembling the leptino-amphibolitic formation. In this case the Ielova Series, within which Mărunţiu (1987) recognized ultramafic cumulates with characters of an ophiolitic-type assemblage, dismembered and metamorphosed, raises the problem of a Proterozoic ophiolitic suture originating in an oceanic-type crust area that must be reconstituted on the basis of the lithofacial assemblages and of the "ocean-floor" mafic-ultramafic rock assemblages. Such assemblages are to be found both in the Getic Realm (Ravensca and (?) Jidoştiţa Formations) and in the Danubian Realm (at present with an uncertain position, Ielova and Măru Series, "Zeicani" crystalline, cf. Kräutner et al., 1981, Drăgşan Series).

b) A pre-Alpine Nappe within the Danubian Units, lying under the Infragetic Nappes, in which case a part of the Ielova Series (from the Svinecea Nappe, cf. Năstăseanu et al., 1981) can be lithostratigraphically

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correlated with the Proterozoic mafic-ultramafic assemblages of the Danubian Realm.

These uncertainties are to be solved in the future.

2.6. Ravensca and Jidoștița-Ivanu Formations in Relation to the Sebeș-Lotru "Series"

The appurtenance of these formations and the lithostratigraphic correlation with sequences of the Sebeș-Lotru Group or Series (to which they were previously assigned) implicates a complex analysis concerning : the known lithostratigraphic successions ; origin of the primary petrographic assemblages and of the geochemical affinities ; origin of the contacts between the major lithostratigraphic entities and of the subsequent processes underwent : metamorphism, deformation, migmatization, magmatism.

Plates II and IV broadly present the outcropping areas of the formations with mafic and ultramafic rock assemblages and their position in relation to the mostly terrigene silico-aluminous sequences of the Sebeș-Lotru Series, within which the following succession can be recognized in the Semenic (Savu, 1973), Godeanu (Bercia, 1975) and Mehedinți (Hârtopanu, Iancu, models of the Nadanova and Orșova sheets ; Iancu et al., models of the Bala and Bîlvănești sheets) :

— lower plagiogneisses formation : plagiogneisses, quartzites, micaschists, paraamphibolites, limestones, quartz-feldspathic gneisses ;

— gneisses and amphibolites formation (leptino-amphibolitic) ;

— micaschists and plagiogneisses formation (in association with quartz-feldspathic gneisses).

In comparison with this sequence, locally incomplete, the Ravensca and Jidoștița-Ivanu formations occur in a lower geometric position, the contact being always marked by deep-seated blastomylonites and has the shape of a small-angled plane. The origin of the lithofacial assemblages and of the contacts with sequences belonging to the Sebeș-Lotru Group makes us to assign them to a distinct lithostratigraphic entity for which we propose the denomination (possibly temporary) "Luchița Group" introduced by Codarcea (1940) in order to designate a "zone" with specific features in the area of the Lotru crystalline (Streckeisen, 1934 ; Codarcea, 1940).

A special problem is represented by the peridotite-eclogite-granulite assemblage of an infracrustal origin, marked by the spatial distribution of these rocks within the Sebeș-Lotru "Group" in the Sebeș-Cibin (Mărunțiu, 1987), Mehedinți (Iancu et al., 1988), Căpățîna (Hann et al., 1988 ; Săbău et al., 1987). The microstructural and mineralogical data attest the mechanic inclusion of these rocks before the first dynamothermal metamorphism (M_1) which affected the sequences of the Sebeș-Lotru Group, indicating the oldest crustal discontinuity within the Getic Crystalline. The assemblage of "exotic" mafic-ultramafic rocks, with a high-grade metamorphism (eclogitic and granulitic facies), is represented by allochthonous, lenticular bodies, distributed along a "band" (1—5 km wide) enscribed on large surfaces in a curved plane (Pl. IV) situated inside the

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Sebeș-Lotru Group, above the "Ursu Formation" in the Căpățîna massif (Hann et al., 1988; Săbău et al., 1987). The mentioned Ursu Formation considered to be in tectonic relationship with the formations of the Sebeș-Lotru Group (with uncertain position) consists of metaperidotites in association with mafic and felsic rocks with parageneses superposed with cordierite and sillimanite, later on retrograded at the level of the andalusite zone (Săbău et al., 1987), the retrograde transformations being considered continuous in time.

The presence of primary (pre-metamorphic) assemblages of mafic and ultramafic rocks within the Sebeș-Lotru Group raises special problems concerning their appurtenance to a unitary major entity; the evolution of the metamorphic and deformational events undergone in time; origin of the relationships with the polymetamorphic sequences made up of sialic material.

3. Conclusions

The formations in the Toronița "Zone" or "Series" belong to the "Getic Realm", due to the features of the Mesozoic cover, specific to the Șopot "Zone". Tectonically the Toronița "Zone" belongs to the pre-Alpine structures that can be traced, in this part of the Getic Nappe, between the Danube and Bozovici.

The petrological features and the relationships with the low-metamorphic Paleozoic formations made possible its assigning to the "Ravensca Formation" referred to the Sebeș-Lotru "Series" or "Group", of Proterozoic age. The ante-Paleozoic relationships of this formation (and those similar petrographically to the group called by us "Luchița") with other sequences of the Sebeș-Lotru Group are tectonic and their assigning to one of the known groups (regarded as primary, cogenetic assemblages) implies a complex study of the pre-Alpine basement in the Getic and Danubian realms.

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ZONA TORONIȚA ȘI PROBLEMATICA FUNDAMENTULUI METAMORFIC PREALPIN AL DOMENIILOR GETIC ȘI DANUBIAN

(Rezumat)

În această lucrare se argumentează, cu date privind caracterele secvențelor de cuvertură mezozoică, apartenența „zonei" sau „seriei" de Toroniţa la domeniul getic.

Relațiile tectonice cu formațiunile proterozoice și paleozoice din cadrul pînzei getice permit integrarea ei în ansamblul structural varistic, ce se recunoaște pînă la nord de bazinul neogen Bozovici.

Secvența atribuită „seriei" de Toroniţa este reprezentată printr-o asociație de roci amfibolice și plagiognaise (cuarțoase sau micacee), în care sînt prezente corpuri de metagabbrouri (și metaperidotite, în zona de la nord de Șopotul Nou). Întreaga secvență, atribuită formațiunii de Ravensca (de către noi) prezintă dovezi petrologice ce atestă polimetamorfism de grad mediu, antepaleozoic. Unele pachete sînt intens retromorfozate la nivelul faciesului de șisturi verzi, ca urmare a reactivărilor paleozoice ce au însoțit metamorfismul prograd al formațiunii de Miniș (Paleozoic inferior) și, subsecvent, al formațiunii vulcano-sedimentare de Buceava (Paleozoic mediu), în cursul ciclurilor orogenice caledonian și respectiv varistic (hercinic).

Unitățile structurale varistice în care sînt implicate formațiunile proterozoice și paleozoice sînt: Nera (în poziție geometrică superioară), Oreviţa, Miniș și Buceava.

Atît în ariile de reactivare varistică cît și în afara acestor arii (masivele Semenic, Godeanu, Mehedinți), formațiunile grupate de noi sub numele de „grupul Luchiţa", pe criterii petrologice-litofaciale, se află în relații tectonice cu secvențele terigene polimetamorfe ale grupului Sebeș-Lotru. Contactele dintre cele două tipuri de succesiuni sînt marcate de blastomilonite antepaleozoice, cu blasteză de biotit și amfibol (metamorfism dinamic, la nivelul zonei cu biotit, ce însoțește o discontinuitate structurală și litologică majoră).

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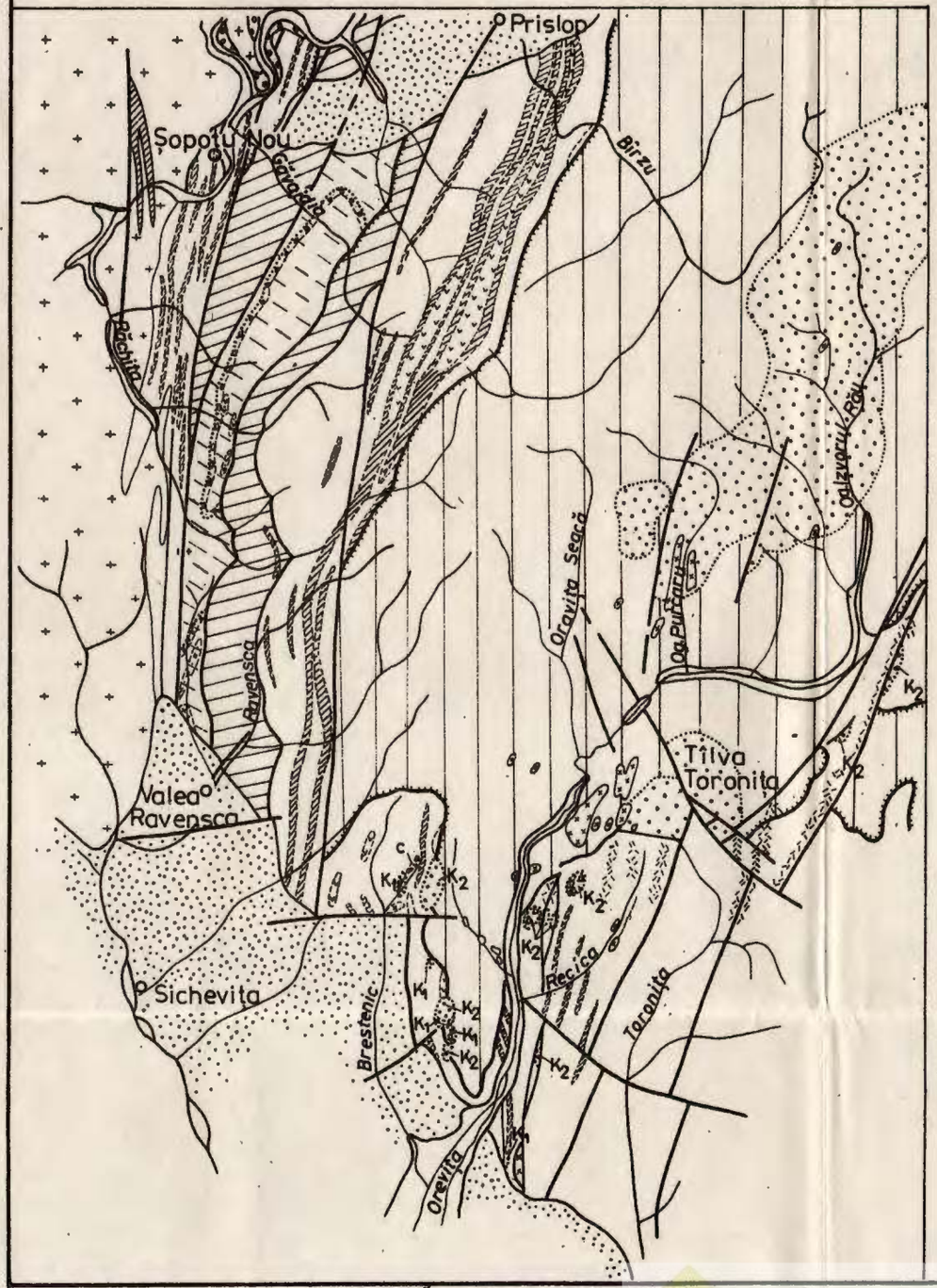
VIORICA IANCU, M.MĂRUNȚIU

GEOLOGICAL SKETCH OF THE TORONIȚA - ȘOPOTU NOU ZONE

0 1 2 km



LEGEND

QUATERNARY — Alluvial deposits; a, terraces

MIDDLE MIOCENE — Conglomerates, gravels, sands

PALEOCENE–UPPER CRETACEOUS — Banatitic magmatites

GETIC DOMAIN

Getic Nappe

UPPER CRETACEOUS (CENOMANIAN-CAMPANIAN) (K_2) — Conglomerates, micaceous sandstones, limestones, marls, (Șopot zone); a, sedimentary breccia

LOWER CRETACEOUS (K_1) — Skeletal limestones (Urgonian facies)

CARBONIFEROUS — Sichevița Granitoids (Poneasca)

VARISCAN UNITS

Nera Unit

MIDDLE PROTEROZOIC — Sebeș - Lotru Group
Bîrzu Gneiss - Micaceous Formation: plagiogneisses, micaschists, quartz - feldspathic gneisses, migmatites

Oreviță Unit

MIDDLE PROTEROZOIC — Ravensca Amphibolitic (retromorphic) Formation
Amphibolites (ortho- and para-amphibolites)
Metagabbros ± metaperidotites
Carbonatic rocks
Amphibolites, quartz plagiogneisses ± micaceous assemblage

Miniș Unit

LOWER PALEOZOIC — Miniș Formation
Metarhyolites
Quartz ± sericite ± chlorite schists

Bozovici Unit

MIDDLE PALEOZOIC — Buceava Volcano-Sedimentary Formation
Conglomerates (a); sandstones, siltites, limestones in association with basalts (b); granodiorite bodies (c); and gabbro (d)
Low-grade metamorphism

Sebeș - Lotru Group

MIDDLE PROTEROZOIC — Gneissic Amphibolitic Formation (= Leptino-amphibolitic) with Anatectic Granitoids

Variscan overthrust plane

Fault

Brecciation

Discordance

Based on: Năstăseanu et al. (1981) - Sichevița sheet ; Bîrlea (1976) - Ph. D. thesis

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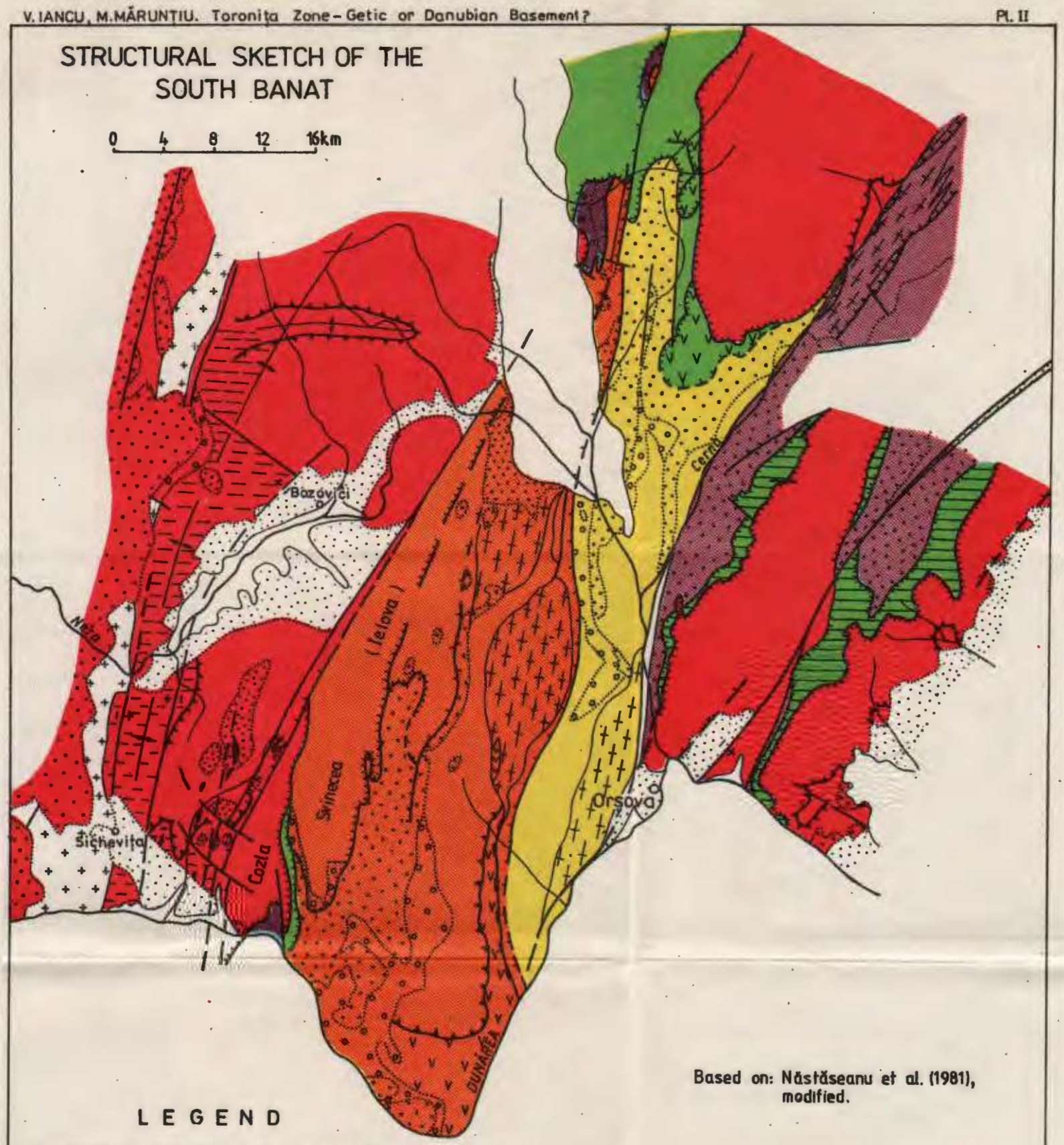
STRUCTURAL SKETCH OF THE SOUTH BANAT

0 4 8 12 16km



Based on: Năstăseanu et al. (1981), modified.

LEGEND

MEDIAN DACIDES - GETIC AND SUPRAGETIC NAPPES
Getic Domain

- Getic Nappes; a - Variscan nappes zone

EXTERNAL DACIDES - INFRAGETIC NAPPES
Arjana - Severin Domain: Severine nappe - a;

- Arjana Nappe { Căleanu Digitation - b / Fenes Digitation - c

MARGINAL DACIDES - DANUBIAN NAPPES
Danubian Domain

A. Internal Danubian

- Rîul Rece Nappe
- Sirinia Nappe
- Cornereva Nappe

B. External Danubian

- Lower Danubian Nappes
- Pliocene - Quaternary
- Miocene
- Banatitic magmatites
- Mesozoic cover: a, nondifferentiated; (J-K_1); b, K_2 - Getic Domain
- Variscan molasse formations (C_3-P)
- Variscan (Hercynian) granitoids
- Caledonian granitoids
- Caledonian ophiolitic complex
- Fault
- Reversed fault
- Trace of the Alpine overthrust plane
- Pre-Alpine tectonic contact (nondifferentiated)
- Obduction nappe
- Discordance

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SEQUENTIAL SYNTHETIC COLUMNS OF THE SEBEȘ-LOTRU „SERIES"

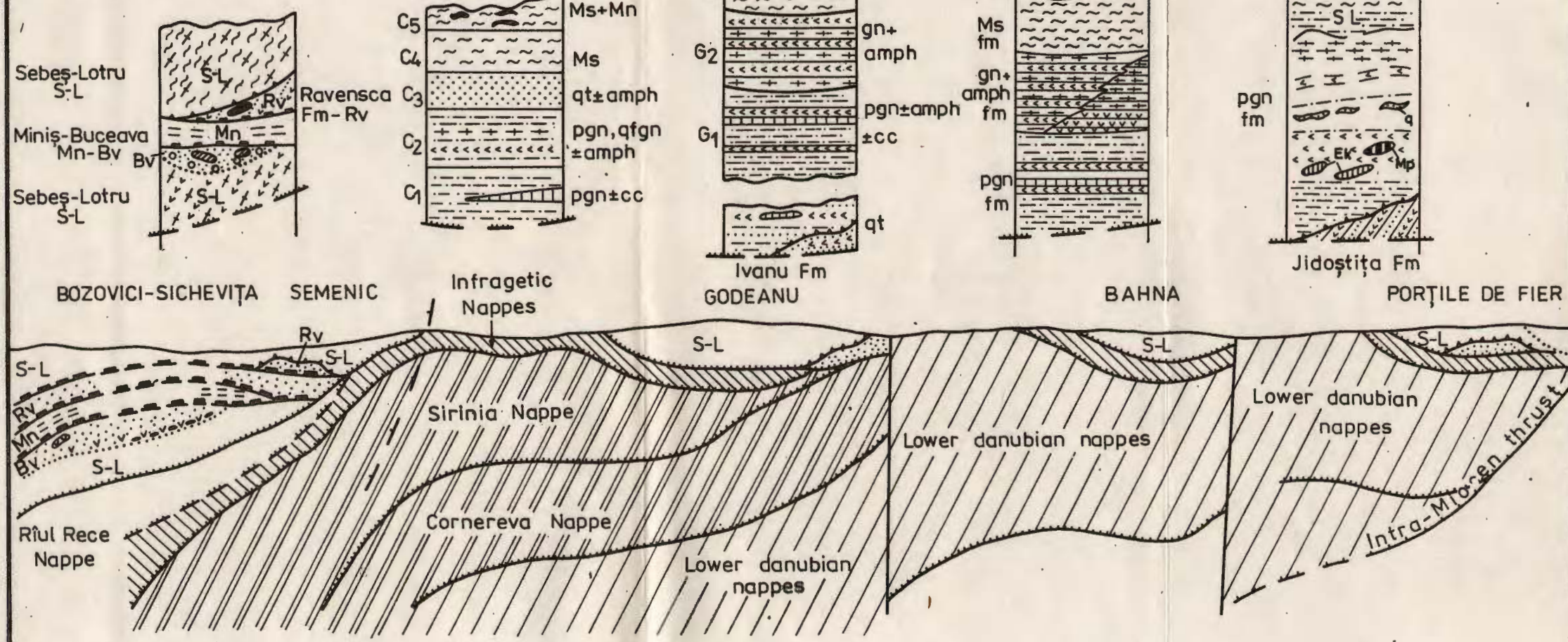
BOZOVICI-SICHEVIȚA
(Iancu et al. 1988)
(Iancu, Mărunțiu)

SEMENIC
(Savu, 1973)

GODEANU
(Bercia, 1975)

BAHNA
(Hârtopanu, Iancu)
(1979, 1975 unp)

PORȚILE DE FIER
(Iancu et al. in press)



LEGEND

- Terrigene formation (Miniș)
- Volcano-sedimentary formation (Buceava)
- Amphibolite-quartzitic formations (Ravensca, Jidoștița, Ivanu)
- Sebeș-Lotru Group
- Micaschists: a, manganiferous rocks (Ms, Mn)
- Plagiogneisses (pgn)
- Amphibolitic rocks (amph)
- Quartz-feldspathic gneisses (qfgn)
- Quartzites: a, quartz segregations (qt, q)
- Carbonatic rocks
- Eclogitic and granulitic rocks (Ek); metaperidotitic rocks (Mp)
- Infragetic Nappes
- Danubian Nappes: a, Upper; b, Lower
- Alpine nappe
- Variscan nappe
- Precaledonian nappe
- Thrust line
- Discordance

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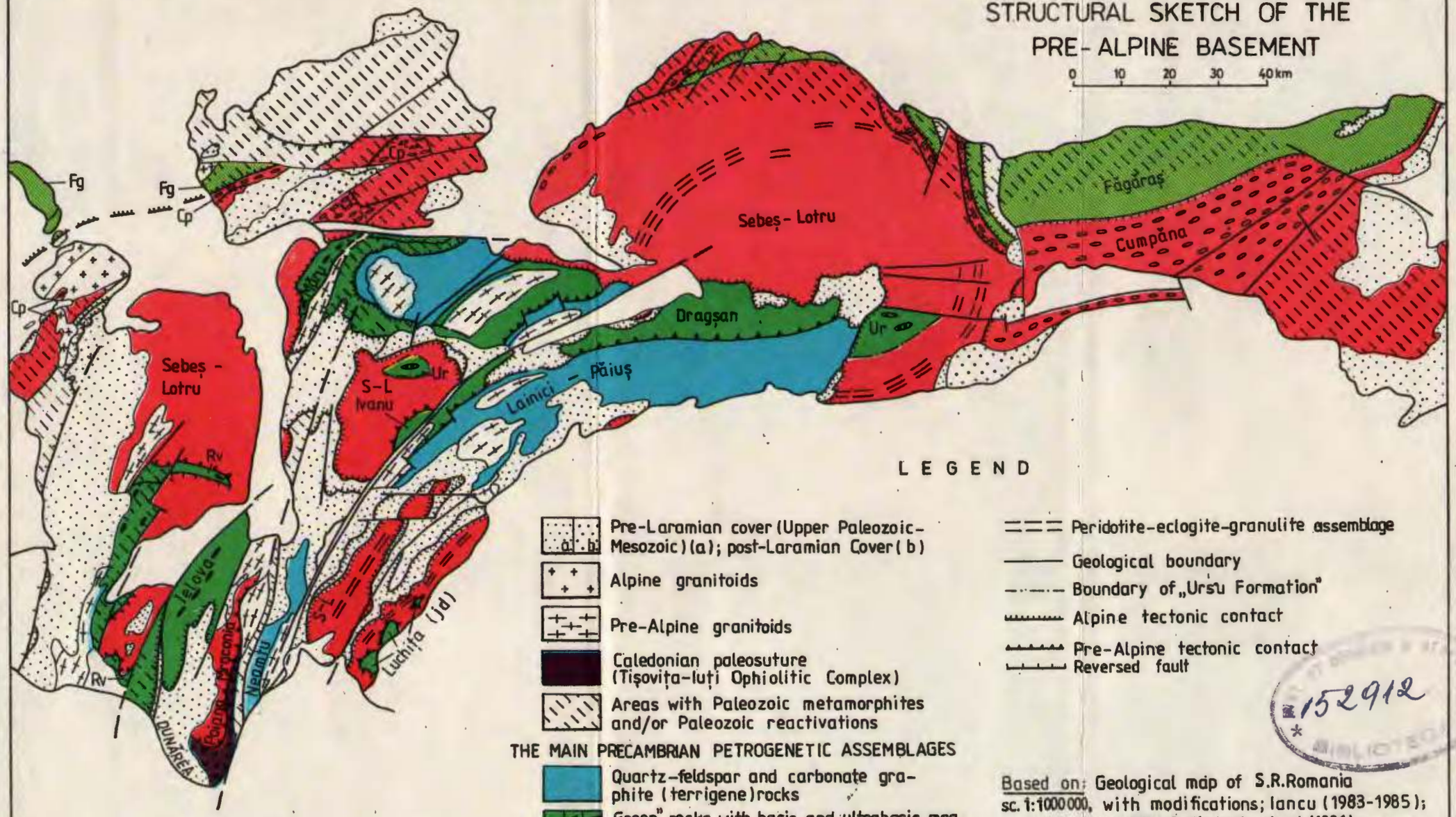
STRUCTURAL SKETCH OF THE PRE-ALPINE BASEMENT

0 10 20 30 40 km



LEGEND

- Pre-Laramian cover (Upper Paleozoic-Mesozoic) (a); post-Laramian Cover (b)
- Alpine granitoids
- Pre-Alpine granitoids
- Caledonian paleosuture (Tișovița-Iuți Ophiolitic Complex)
- Areas with Paleozoic metamorphites and/or Paleozoic reactivations

THE MAIN PRECAMBRIAN PETROGENETIC ASSEMBLAGES

- Quartz-feldspar and carbonate graphite (terrigene) rocks
- „Green" rocks with basic and ultrabasic magmatites: ophiolitic type complex (a); „Ursu Formation" (b)
- Silico-aluminous and quartz-feldspathic rocks; a, potash-augen gneisses
- „Green" rocks with carbonate-graphite and pelito-psamitic rocks

- Peridotite-eclogite-granulite assemblage
- Geological boundary
- Boundary of „Ursu Formation"
- Alpine tectonic contact
- Pre-Alpine tectonic contact
- Reversed fault

152912
BIBLIOTECA

Based on: Geological map of S.R.Romania sc. 1:1000000, with modifications; Iancu (1983-1985); Hann, Szasz (1980); Balintoni et al. (1986); Săndulescu et al. (1986); Hann, Balintoni (1988); Berza et al. (1983); Balintoni, Iancu (1986); Kräutner et al. (1981); Dimitrescu (1986); Conovici, Săbău (unpublished data)

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POST-NAPPE FOLDING OF METAMORPHICS IN NW FĂGĂRAȘ MASSIF [1]

by

DINU PANĂ [2], CĂLIN BICMAN [2]

Metamorphic rocks. Limestone. Amphibolites. Paragneiss. Schists. Metapelite. Retrograde metamorphism. Stratigraphic units. Thrust fault. Blastomylonite. Lithologic controls. Alpine Orogeny. South Carpathians — Crystalline Getic and Supragetic realms — Făgăraș Mountains.

Abstract

It is noticed and described a folded tectonic plane, which is constituted of a continuous level of mylonitic rocks. This level frames the west-wards closing of a E—W trending nappe antiform. There are separated and defined the component formations of these two tectonic units. It is pointed out the importance of discontinuous retrogression on middle grade metamorphics in NW Făgăraș massif. Common post-nappe deformation is shown by the interference of structural elements of some overlapped folds, Alpine in age, commonly oriented in both units. It is proposed a new structural image of the Făgăraș massif northern slope.

Résumé

Plissement post-charriage des métamorphites du NW du massif de Făgăraș. On signale et on décrit un plan tectonique plissé, matérialisé par un niveau continu de roches mylonitiques. Ce niveau marque la clôture vers l'ouest d'une antiforme de nappe orientée E—W. On sépare et on définit les formations composant les deux unités tectoniques. On a relevé l'importance des conséquences du rétromorphisme sur les métamorphites de degré moyen du NW de Făgăraș. L'évolution des déformations post-charriage, commune pour les deux unités, est mar-

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quée par l'interférence des éléments structuraux de certains plis superposés d'âge alpin. On propose une nouvelle image structurale du versant septentrional de Făgăraș.

This paper contains new data on the structure and retromorphic character of metamorphics in NW Făgăraș, between Moașei brook and Olt river. These data are gathered into a first stage of mapping within Suru sheet 1 : 50.000. First of all we dealt with considerations of structural point of view because, on one hand, plotting of some isograde in Făgăraș massif proved to be impossible (Giușcă et al., 1977) and on the other hand, tectonic (alpine) planes obvious in the eastern and western extremities have been only supposed inside the massif.

1. Previous Data

This region is constituted of metamorphics which were described and classified for the first time by Ghika-Budești (1940). He tried a stratonomic classification with Cumpăna and Făgăraș series, then subdividing them in zones and claiming that in the Făgăraș massif, about from north to south, it can be followed a somehow regular succession of index minerals, which would express the classical zonality of a prograde metamorphism of Barrovian type.

Dimitrescu (1963) classification is very much similar to Ghika-Budești's, having the metamorphic grade as "unique basic criterium". But, in 1967 Dimitrescu also drew attention for the first time that differences of metamorphic grade have no decisive stratigraphic role. This idea was later sustained by Balintoni et al. (1984).

Then these classifications have been used by all investigators prospecting this region.

Dimitrescu (1963), then Dimofte (1967), Balintoni (1969), Chivu (1985) showed that locally in the Făgăraș also took place a regressive metamorphism, but on the preexisting background of the Barrovian zonality.

Giușcă et al. (1977) asserted that at least on the Transfăgărășan route, mapping of some isograde is practically impossible and that the retromorphism in the Făgăraș northern slope settled on some metamorphics of comparative grade with the metamorphics in the southern slope. These observations are wholly confirmed by Balintoni et al. (1984).

In the papers of more general correlation, names of series and zones described by Ghika-Budești are maintained even if they partly change their content : Savu et al. (1978 a), Kräutner (1980).

From structural point of view, it is worth to remind the following : in the eastern part of the Făgăraș massif, there were pointed out several overthrust planes, Streckeisen (1934), Codarcea et al. (1967), Săndulescu (1976 a), Nedelcu, Anton (1984) ; in the western extremity of the massif, in the Olt right slope, there were also pointed out several overthrust planes : Dessila-Codarcea (1962), Codarcea et al. (1967), Hann, Szász (1984). Nevertheless, great amount of metamorphics in the Făgăraș massif remained an alpine homogeneous unity from structural point of view,

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without major tectonic discontinuities, with a plicative prealpine synmetamorphic structure.

Chivu (1984) asserted the northwards overthrust of Cumpăna series on Făgăraş series (sensu Dimitrescu, 1963, with modifications).

Balintoni et al. (1984) much more involved Făgăraş metamorphics in alpine tectonic events; in Mesocretaceous paroxysm, metamorphics of Cumpăna series (sensu Ghika-Budeşti, 1940) individualize themselves as tectonic unit — Argeş nappe, and those of Făgăraş series constitute Moldoveanu nappe and Bîrsa Fierului nappe, overlapping Argeş nappe; during the Mediterranean stage, Argeş nappe overlaps through folding Moldoveanu nappe; all this complex structure would be also remobilized within Boia Laramian nappe.

2. Lithology and Structure

Our observations pointed out an important folded tectonic plane, inside Moldoveanu nappe (Balintoni et al., 1984), of a continuous level of mylonitic rocks with tens of meters thicknesses. This level represents a cartographic landmark obvious within metamorphics in the region, very clearly describing the westwards periclinal closing of an antiform structure about east-west oriented (Pl. I).

Since the most detailed and practically unmodified division of crystalline schists in this zone belongs to Ghika-Budeşti (1940) it will be also used in this paper, namely:

— above the mylonitic level — upper unit — of the antiform northern flank, there are rocks attributed to "the zone of Valea Satului-Porceşti micaschists" and some of them to "the zone of Poiana Neamţului chlorite schists", and on the southern flank, metamorphics are attributed to "the zone of Suru limestones and paramphibolites" and to "the zone of Vemeşoaia micaschists";

— under the level of mylonitic rocks — lower unit — on a great part of mapped antiform axis, there are rocks attributed to "the zone of Şerbota paragneisses with garnet" and some of them to "the zone of Poiana Neamţului chlorite schists".

2.1 Upper Unit

As it could be noticed, all "zones" above the mylonitic level are constituted of the same association of rocks (paragneisses + amphibolites + crystalline limestones), composing one uniform lithologic formation. Passage from one "zone" to another is gradually done, sometimes even those megalithons can be noticed on the direction, passing from one "zone" to another. Differences between zones are the following: a) different relation between paragneisses content and intercalations of limestones and amphibolites and b) different intensities with which retromorphism influences the same preretromorphic petrographic content.

Because within the Făgăraş metamorphics, limestones and amphibolites are the most distinctive occurrence, prevailing in „Suru zone" (Ghika-Budeşti, 1940), the association paragneisses + limestones + amphibolites will be called the Suru formation. In the basin of Valea Lupului

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we noticed a level of ocular gneisses with microcline and mirmekitic structures (Pl. III, Figs. 1, 2) in a structural indefinite position, which could represent either "lenticular gneisses" or "porphyrogene gneisses", which Manea (1962) and Dimitrescu respectively (1964) described on northern slope of the Făgăraș, or "microcline gneisses" which Nedelcu, Anton (1984) described in Strîmba unit.

Metamorphics in the upper unit are frequently cut by lamination planes subparallel with the main nappe plane, so that it is impossible, on our opinion, to recognize any pre-nappe mappable structure.

Preretromorphic content of the upper unit formation is very well preserved in the "Suru zone" which is mapped in upper basins of Moașa, Tătaru, Fătu valleys, where are met parageneses of middle grade previously ascribed to the "almandine amphibolites facies".

"Valea Satului-Porcești zone" mapped on Transylvanian basin border, between Valea Satului and Rîul Mare represents in fact an incipient stage of retromorphism in Suru formation, in which paragneisses prevail besides limestones and amphibolites. Middle grade parageneses are preserved; large garnet, muscovite, biotite are often present. "Vemeșoaia zone" occurs in our region only on the main ridge of the Făgăraș, between Fătu peak and Chica Fedeleșului and contains a lithological association constituted of chlorite-sericitous schists, amphibolic schists and tremolite limestones. This "zone" would represent an area intensely affected by the retromorphism in which middle grade parageneses of "Suru zone" have been almost totally rebalanced under low grade conditions.

2.2. Lower Unit

Under the mylonitic level occur the metamorphics which Ghika-Budești described at the "Șerbota garnet paragneisses zone". Its characteristic element is represented by the horizon of alluminous metapellites, constituted of an alternance of micaschists, garnet, staurolite, kyanite paragneisses, of a phaneroblastic aspect; thin intercalations of amphibolites occasionally occur. This horizon has an apparent thickness of about 1 km. We followed it between Ionelu Valley westwards and Suru saddle eastwards, but it seems to continue without important discontinuities up to Șerbota peak. Northwards to alluminous metapellitic horizon is added a sequence of garnet paragneisses — micaschists in which thin levels of amphibolites occur and a lens of crystalline limestones as well (Mînii brook).

Northwards, this sequence is weakly retromorphic; but the retromorphism is more accentuated NE, on Rîul Mare Avrig the rocks getting the aspect of some phylites, that is why Ghika-Budești attributed them to "Poiana Neamțului chlorito-schists zone".

2.3. Description of Mylonites

The level of mylonitic rocks indicates a major tectonic plane formed at least in the mapped region at the limit between two formations with distinct lithological association. This plane is the cause of sudden passage from one lithological association to another. Here are some spectacular

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aspects : a) westwards Ionelu Valley disappearance of paragneisses micaschists with staurolite, kyanite horizon, of lower unit under mylonitic level and amphibolic rocks of upper unit and b) apparition of mylonitic level in the erosion window on Valea Satului. The way of tectonic plane is shown by outcrops and isolated blocks of crystalline limestones intensely tectonically affected.

Mylonites are black-green in colour with superficial limonitizations and here and there have been already mapped (Arion, 1969 ; Apostoloiu et al., 1977) as discontinuous levels of graphitous schists. Under microscope, they are noticed to be mostly constituted of micaschists-paragneisses displaying typical mylonitic and protomylonitic textures with flamboyant fusiform quartz ("ribbon") and sigmoids of opaque minerals and mica, rarely tiny garnets with graphitous inclusions (Pl. III, Figs. 3, 4).

Opaque minerals are generally metallic minerals deposited on fissures, graphite occurring accidentally, especially as graphitous pigment. Pyrrhotine is very widely-spread, being remobilized on fissures, sometimes including quartz and very often occurring as relicts in coloform masses of limonite-goethite. It can also appear peripherically marcatized forming "bird eye" structures ; growth twins sometimes present suggest a syn- or post-tectonic character. Pyrite is also oxidated, calcopyrite is remobilized on fissures and has no anisotropy specific to metamorphosed calcopyrite (Pl. IV, Figs. 1—4). Titan oxides (titanite, ilmenite) either fresh or peripherically replaced by rutile, brought to parallelism with mylonitic layering often occur. Hematite polysynthetically twinned can occur sometimes. Mylonites on amphibolites give characteristic aspects, with subgrained amphiboles and opaque minerals into an obvious mylonitic foliation (Pl. III, Fig. 3).

2.4. Metallogenetical Considerations

Besides mylonitic main level which separates the two formations, within that one in the upper unit, occur frequent concordant levels of laminated, mylonitized rocks of metric thicknesses which represent secondary planes of movement synchronous and subparallel with the nappe main plane.

These levels contain sulphides as well, have superficial limonitizations and locally gave weak anomalies of Zn, Cu, Fe (Tomescu, 1966).

The context in which occur these accumulations (concordant levels of mylonitic rocks) as well as the lack of any obvious magmatic source, suggests these accumulations belonging to tectogene type.

There can also be noticed in NW Făgăraș the two component of the metallotect of Valea lui Stan type (Udubașa et al., 1985) :

— a geochemical one with passive role represented by favorable lithological geochemical premises : a) metallic elements in premetamorphic pile (limestones with syngenetical Zn, Cu, Fe mineralization), b) mobilization and eventually concentration through metamorphic differentiation during the retromorphic processus ;

— another one, tectonic with active role — pseudohydrothermal remobilizations caused by dynamic retromorphism in the vicinity of tectonic planes.

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Pressure gradients from the shearing zone can activate some relatively acide solutions (H_2O — inclusively meteoric + CO_2 + SiO_2) of 200—300° temperature which can extract or transport sulphides; shearing planes (S_4 mylonitic) constituted the discontinuities on which migrated these mineralizing solutions.

From this point of view, of syn- and post-deformation blastesis, rocks which materialize the tectonic plane get a blastomylonitic character.

Because, on northern flank of the mapped antiform, weakly retromorphic sequences of the upper unit are displayed by means of mylonitic level on intensely retromorphic sequences of the lower unit, the age of the plane indicating the mylonitic level can be but post-retromorphic; since the retromorphism in the Getic domain is generally accepted to be of Hercynian age, the plane could form during a Late Hercynian stage or any paroxysmal stages of alpine tectogenesis, the most probable Mesocretaceous. Anyhow, the common folding of the two formations (= tectonic units) together with the mylonitic level between them, has to be admitted to be of Alpine age.

3. Structural Analysis

Besides these general observations, we tried to explain the succession of events which influenced the rock pile and on account of the relations among planary and lineary structural elements at mesoscopic and microscopic scale. Sometimes, these relations are clearly expressed, other times they are screened, either because of the outcrop position in relation with structure or because of different degree of penetration of structural elements in relation with lithology and intensity of deformation.

The mapped region is characterized by the interference of structural elements of some overlapped folds. In the same outcrop there can be measured very often families of planes differently oriented. Their statistical rally, without considering qualitative criteria, is presented in three diagrams: a global one (Fig. 1 a) and two sectorial ones (Fig. 1 b and 1 c).

In the western sector (west of Moaşei Valley) maxima describe an „AC" girdle (Fig. 1 b), which suggests an opened fold E—W trended having a western sinking ($\beta = 243/22$) which corresponds to major mapped post-nappe fold and another "ac" girdle weakly drawn, obliquely displayed to the first ($\beta' = 130/40$) to which correspond only micro- and mesoscopic folds unsignificant at the map scale.

In the eastern sector (east of Moaşei Valley) maxima individualize (Fig. 1 c) suggesting tight folds, with symmetrical limbs subhorizontal hinges E—W trending ($\beta = 75/2$). Here it is also drawn an "ac" girdle obliquely displayed ($\beta' = 120/60$). According to field relations, foliations which give the oblique girdle, weakly drawn in all three diagrams, affect E—W foliations therefore they are the youngest in the region.

Major deformation is E—W trending, on the whole northern Făgăraş slope, mapping, on account of divergent foliations, a system of tight folds in relay which respects truly enough a E—W structural trending. Deformation on NW—SE direction has a much lower intensity, so that the

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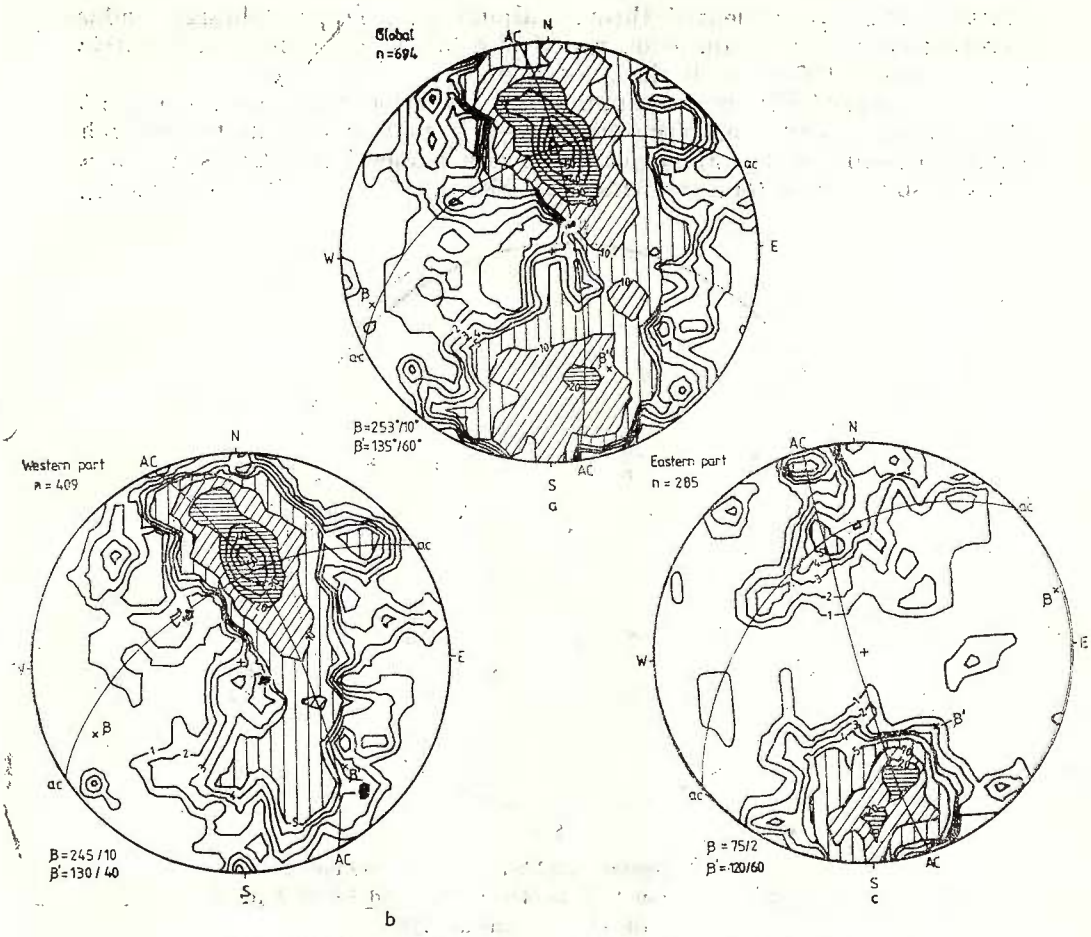

Fig. 1 — π diagrams : 1 a, global ; 1 b, western sector ; 1 c, eastern sector.
n = number of plotted planes. 1...20...60 — frequency isolines. β, β′ = graphically determined theoretical axes of the system of overlapped folds suggested by „AC" and „ac" girdle, respectively.

mapped structure does not come to the stage of "domes and basins" (Whitten, 1966).

Sectorial diagrams suggest a post-nappe deformation history common to both units in the whole region mapped and the intensity increase of E—W trending deformation in the eastern sector.

In the zone of perianticlinal weswards closing, it can be accepted a concentric type folding having as deformation mechanism the flexural slip (eventually flow). This would mean that inside folded lithological sequences (at least those competent) do not occur distorsions and that foliations which give "ac" girdle in the diagram π Figure 2 are metamorphic foliations S_2 (for the sequences of Şerbota and Suru formations — nonretromorphic), S_3 (for retromorphic sequences) and S_4 (mylonitic which trans-

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poses S_2 and S_3) reoriented through alpine post-nappe folding. Under stereoscopic projection lineation would plot in this case on a small circle (Hobbs et al., 1977) but it does not happen.

Eastwards folds become tighter and it is foreseen the surpass of limit of lithological sequences plasticity, the deformation continuing through simple shearing on S_5 alpine planes. These planes have the aspect of a "strain slip" type foliation (fracture cleavage) about E—W striking,

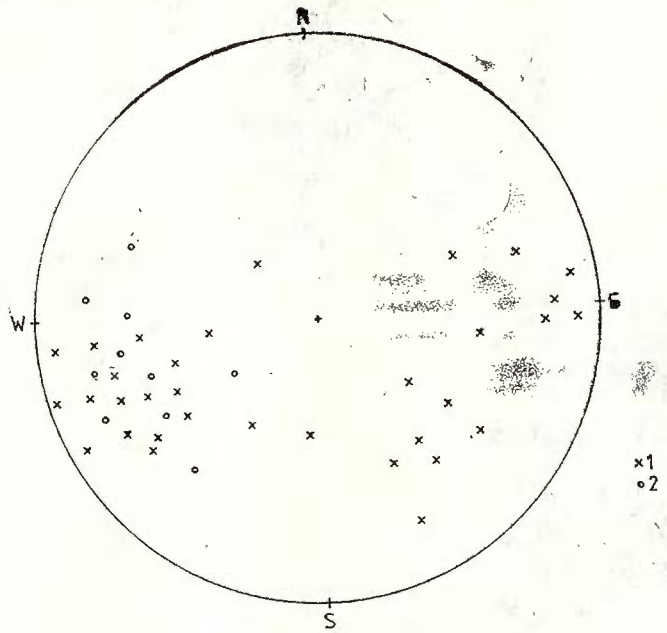

Fig. 2 — Plot of mesoscopic folds axes of the two units.
1, tight folds of chevron type (alpine); 2, opened isoclinal folds (prealpine-synmetamorphic).

sometimes closely, other times widely spaced. This foliation is planely axial to some isoclinal folds tight with very sharp hinges of "chevron" type and with centimetric to metric sizes. Sometimes in this foliation plane, it is drawn a lineation parallel to "a" which is characteristic to folds of simple axial plane shearing (Bedgley, 1965). Under microscope, the alpine S_5 foliation is refound :

A) without neoformation :

a) within an incipient stage as crenulation cleavage, microfolding the S_3 retromorphic foliation both in the lower unit (Pl. V, Fig. 1) and in the upper unit (Pl. V, Fig. 2) and S_4 mylonitic foliation (Pl. III, Fig. 4).

b) within an advanced stage becoming disjunctive and totally transposing S_2, S_3.

B) with neoformation : sericite directions oblique in respect to S_3 (Pl. V, Fig. 4).

There where the S_5 foliation is less penetrative, there are kept quartz + feldspar + biotite + garnet lithons, sometimes as intrafolial similar

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isoclinal folds with opened hinges. They contain an axial plane foliation with chlorite (S_3) in which are refound S_2 relicts; biotite partly chloritized, broken muscovite, both minerals, being incompletely reoriented, corroded garnet, big clasts of undulatory quartz.

In the outcrop, isoclinal folds of "chevron" type tend to hide older opened isoclinal folds.

Anyhow, the two types of folds cannot be geometrically discriminated having the B axes coaxial with β (S_5) (Fig. 2).

So, similar opened folds with the axial plane foliation materialized by chlorite neoformation represent S_2 relicts in S_3 without correspondent at the map scale, and folds of "chevron" type which are younger and have the axial plane foliation of "strain slip" type represent at mesoscopic scale ("parasitic folds") the alpine folds which we map in northern Făgăraș massif.

4. Conclusions

a) In NW Făgăraș there can be distinguished two formations with relatively uniform lithological contents:

— Suru formation: association limestones-amphibolites in garnet paragneisses-micaschists background;

— Șerbota formation — alluminous metapellites in paragneisses, micaschists with garnet, subordinately amphibolites and totally isolated crystalline limestones.

Differences among lithological contents of the two formations would represent the main criterium of the future correlations.

b) Both formations have been regionaly more or less intensely downgraded, wipping off the distinctive preretromorphic characters, so that there where come into contact sequences which were affected by retromorphism with the same intensity, but belonging to different formations, their separation is extremely difficult. That is why the metamorphic grade cannot represent by itself a criterium in N Făgăraș; on the contrary it can lead to confusions.

c) The two formations are at least in the investigated region in relation of tectonic superposition; Suru formation lies on Șerbota formation, limit between them being given by a mylonitic level of variable thicknesses. Mylonites could constitute the mapping landmark of the limit between the two units in the case if they would be accompanied by lithological association change.

d) Intense postnappe common folding with a "strain slip" alpine foliation creates a complex structure of nappe anticlinals and synclinals, which suggest the probability to appear in windows and halfwindows of the formation in the lower unit (Șerbota) and in filling of nappe synclinals of the formation in the upper unit (Suru).

On account of these considerations and some previous observations corroborated with issued data (Dimitrescu, 1963, 1967 a) and nonpublished data (Arion et al., 1965, 1966, 1967 and Balintoni et al., 1984) we present a new image of Făgăraș massif northern slope (Pl. II).

It is noticed that the tectonic plane firstly mentioned in this paper is often at the base of Suru formation limestones, therefore at a certain level inside a lithostratigraphical succession; Suru formation, part of

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Făgăraş series (sensu Ghika-Budeşti, 1940) moved on the relative autochthon represented by Şerbota formation (=Măgura Ciinenilor formation ?).

The level of this plane formation seems to be determined by different competence between limestones and subjacent metapellites.

Through the lithological control of shearing and lack of sediments under the overthrust plane, Suru unit resembles a lot to intrafolded nappes in East Carpathians (Săndulescu, 1984) whose prealpine age is still discussed: Lower pre-Carboniferous — Kräutner (1980), Sudete — Săndulescu (1984), Saalic — Balintoni (1981).

It also can be noticed that on the direction of here mentioned tectonic plane occur all stratiform Pb—Zn—Fe accumulations from Făgăraş crystalline limestones attributed to syngenetical metamorphosed type. Significant accumulations (Porumbacu, Arpaşu, Doamna, Capra, Mesteacănu) occur in the vicinity of the tectonic plane, possible in the same level of crystalline limestones of Suru formation, there where it is affected by regional retromorphism as well.

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CUTAREA POSTȘARIAJ A METAMORFITELOR DIN NORD-VESTUL MASIVULUI FĂGĂRAȘ

(Rezumat)

Lucrarea prezintă date noi asupra structurii și caracterului retromorf al metamorfitelor din nord-vestul masivului Făgăraș. Este semnalat și descris un plan tectonic cutat în interiorul pînzei de Moldoveanu (sensu Balintoni et al., 1984), materializat printr-un nivel continuu de roci milonitice. Acest nivel constituie un reper cartografic evident, ce descrie fidel închiderea spre vest a unei antiforme de pînză orientată est-vest.

Milonitele sînt formate pe micașisturi și paragnaise, mai rar pe amfibolite ; conțin sigmoide de minerale opace : pirotină, pirită, subordonat

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calcopirită de obicei ca relicte în mase colomorfe de limonit-goethit; în proporție relativ ridicată apar oxizi de titan, fie în stare proaspătă, fie transformați periferic; milonitele mai conțin și grafit în cantități infime, sub formă de pigment fin.

Nivelul de roci milonitice separă o unitate tectonică superioară alcătuită din formațiunea de Suru de una inferioară alcătuită din formațiunea de Șerbota. În aceste două formațiuni sînt incluse mai multe „zone" de metamorfism separate anterior. Se arată că diferențele dintre aceste zone sînt date de fapt de intensitatea variabilă cu care fenomenul de retromorfism afectează două asociații litologice relativ uniforme: formațiunea de Suru (paragnaise, amfibolite, calcare cristaline) și formațiunea de Șerbota (metapelite aluminoase cu intercalații subțiri de amfibolite).

Analiza structurală efectuată relevă o istorie deformațională postșariaj comună ambelor unități litologice și tectonice, marcată de interferența elementelor structurale ale unor cute suprapuse, de vîrstă alpină.

Observațiile microscopice sînt coroborate cu cele mezoscopice și macroscopice, insistîndu-se pe importanța cutării postșariaj după o direcție est-vest.

Extrapolînd spre est relațiile stabilite în regiunea cercetată este prezentată o schiță tectonică originală a versantului nordic al Făgărașului.

EXPLANATION OF PLATES

Plate III

Fig. 1 — Microcline with globular quartz inclusions in ocular gneissss on Valea Lupului-Suru formation. N +, 25 ×.

Fig. 2 — Mirmekitic structures in oftalmitic rocks on Valea Moașei-Suru formation. N+, × 50.

Fig. 3 — Opaque minerals, broken mica, flamboyant quartz, in mylonitic foliation S_4-Valea Lupului-Suru formation. N ||, × 100.

Fig. 4 — Subgrained amphiboles and opaque minerals in mylonitic foliation S_4, microfolded. Fătu Valley-Suru formation. N ||, × 100.

Plate IV

Fig. 1 — Oxidated peripherically pyrrhotine. Valea Satului-Turnu Roșu, × 100.

Fig. 2 — Pseudomorphosis of limonite-goethite on pyrite. Valea Lupului. × 100.

Fig. 3 — Oxidated peripherically calcopyrite. Ionelu Valley. × 100.

Fig. 4 — Collomorphic masses of limonite-goethite. Ionelu Valley. × 50.

Plate V

Fig. 1 — Microfolds of S_3 retromorphic foliation with chlorite pseudomorphoses on biotite and garnet, in the upper unit, Suru formation. N ||, × 50.

Fig. 2 — Microfolds of S_3 retromorphic foliation in the lower unit, Șerbota formation. N ||, × 50.

Fig. 3 — Microfolds of S_3 foliation with S_2 relicts, Șerbota formation. N ||, × 100.

Fig. 4 — Incipient sericite neoformation on S_5 alpine foliation, obliquely displayed in relation with retromorphic foliation S_3. N+, × 50.

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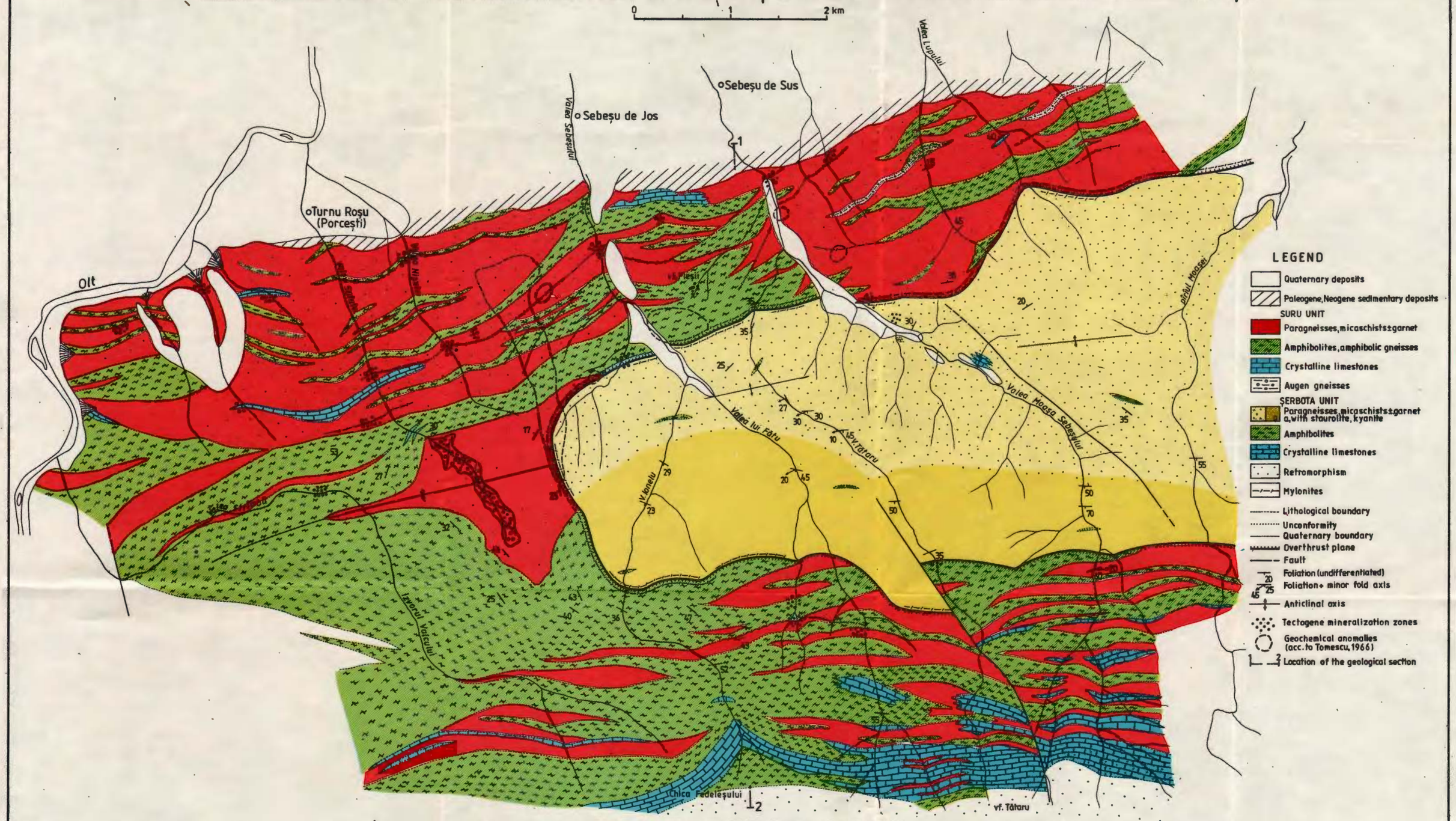
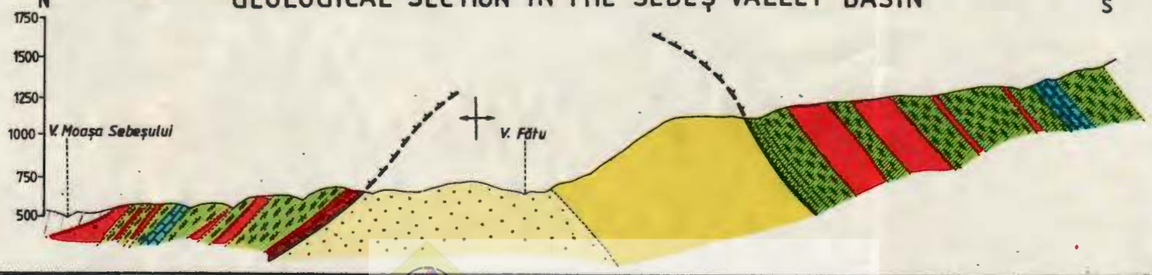
GEOLOGICAL MAP OF THE NORTHERN PART OF FĂGĂRAȘ BETWEEN VALEA STRÎMBĂ AND PÎRÎUL MOAȘEI



GEOLOGICAL SECTION IN THE SEBEȘ VALLEY BASIN



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TECTONIC SKETCH OF NORTHWEST FĂGĂRAȘ



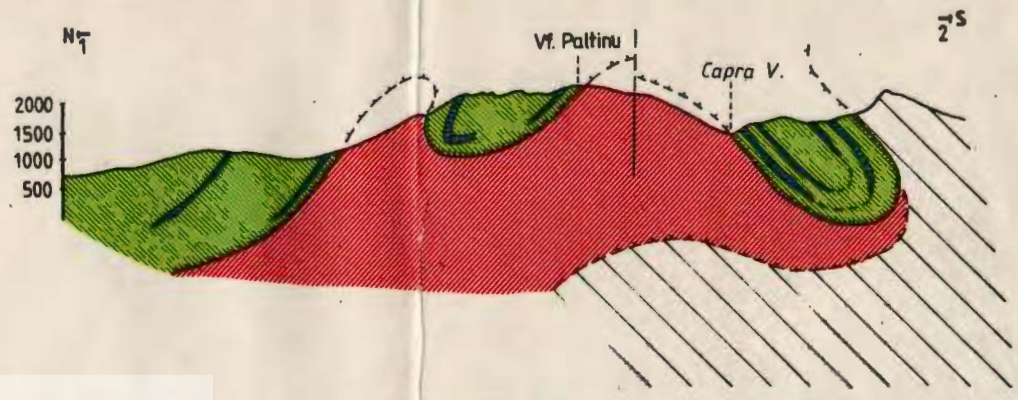
LEGEND

- Sedimentary deposits
- Upper tectonic unit - Suru Formation a, limestones (Făgăraș Group)
- Lower tectonic unit - Șerbota Formation (Făgăraș Group)
- Argeș Nappe - Măgura Cîineni Formation (Cumpăna Group)

ALPINE STRUCTURAL ELEMENTS

- Overthrust plane
- Presumed overthrust plane
- Fault
- Normal antiform
- Overturned antiform
- Overturned synform
- Line of the main crest
- Location of the geological section

Geological section between Cîrțișoara and Piscu Negru



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MINERAL ASSEMBLAGES WITHIN THE LITHIAN PEGMATITES AT CONȚU, LOTRU MTS (CENTRAL SOUTH CARPATHIANS) AND THEIR GENETICAL BEARING[1]

by

GAVRIL SĂBĂU[2], AUREL APOSTOLOIU[3], TIBERIU URCAN[3]

Pegmatite. Lithium. Schists. Gneisses. Lotru Group. Proterozoic. Mineral assemblages. Metasomatism. Mineral data. X-ray data. Amblygonite. Purpurite. Tavorite. Montebrasite. South Carpathians — Crystalline Getic and Supragetic realms — Lotrului and Căpățînei Mountains.

Abstract

Lithian pegmatites from Conțu (Lotru Mts, Getic Nappe) have a complex mineralogical composition, containing albite, quartz, spodumene, muscovite, microcline, triphyline-lithiophilite, amblygonite, montebrasite, purpurite, cassiterite, uraniferous magnocolumbite, topaz, apatite, alluaudite (?), huréaulite, tavorite, sphalerite, garnets. For the first time in Romania, there are mentioned amblygonite, purpurite, magnocolumbite, huréaulite and tavorite. On account of the relations with host rocks (Lotru group) and textural relations, the authors propose a genesis from a metamorphic fluid in absence of important anatectic phenomena at 610—625° C and about 5.5 Kbar.

Résumé

Associations minéralogiques dans les pegmatites lithifères de Conțu, monts de Lotru (Carpathes Méridionales Centrales) et leurs implications génétiques. Les pegmatites lithifères de Conțu (monts de Lotru, nappe gétique) présentent une composition minéralogique complexe

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à albite, quartz, spodumène, muscovite, microcline, triphylline-lithiophyllite, amblygonite, montébrasite, purpurite, cassitérite, magnocolumbite uranifère, topase, apatite, alluaudite (?), huréaulite, tavorite, sphalérite, grenats. On mentionne, pour la première fois dans la Roumanie, l'amblygonite, la purpurite, la magnocolumbite, l'huréaulite et la tavorite. Ayant en vue les rapports avec les roches hôtes (groupe de Lotru) et les relations texturaux, les auteurs proposent une genèse d'un fluide métamorphique en absence des phénomènes anatectiques à 610—625° C. et approximativement 5,5 Kb.

Introduction

By its peculiar mineralogic composition, the Conțu occurrence represents a rather unusual one among the Getic Crystalline pegmatites. On account of this fact, it was commonly regarded to be of magmatic filiation (Maieru et al., 1968; Pomârleanu et al., 1967), unlike other widely-developed metamorphic pegmatites within the mesometamorphic rocks of the Getic Nappe.

The Conțu pegmatite field is located on the slopes of the similarly-named brook, a tributary to the Sadu Valley and extends southwards to the Balindrul Mic saddle in the main range of the Lotru Mts (Fig. 1). There have been encountered seven tabular to lens-shaped pegmatite bodies (up to 200—300 × 2—12 m) hosted in a kyanite-staurolite-bearing micaschists and mica gneisses formation, with manganiferous intercalations, belonging to the Lotru Precambrian Group (Sebeș-Lotru Group, Sebeș-Lotru Series). Discovered as a result of geological prospecting by Apostoloiu in 1963, they have been described as having the typical paragenesis spodumene-albite-quartz-lepidolite and containing subordinate amounts of microcline, biotite, garnet, zircon, apatite, kyanite, titanomagnetite, rutile, pyrite, graphite, triphyline, epidote, samarskite (Maieru et al., 1969). Pomârleanu et al. (1967) relying upon spodumene fluid inclusions study, estimate their formation at 410—530°C, the inferred pressure being of 1.7—2.5 Kbar.

The aim of this paper is to provide more complex data regarding the mineral phases and their relationships, in order to allow a new genetic insight.

Mineral composition

The Conțu pegmatites are mainly feldspathic (albite 30—43%, K feldspar 0—5%, ± oligoclase) containing large prismatic spodumene crystals (15—20%), quartz 25—30%, muscovite 3—10%, other minerals occurring subordinatedly. Nevertheless, a high scale inhomogeneity is conspicuous, without leading however to a manifest zonality, so one can note frequent changes regarding the minerals number and their quantitative ratios. Among mineral assemblages there are replacement relationships which make evidence for a complex evolution.

Occasionally, within the border zone of the pegmatitic bodies, quartz-oligoclase assemblage may appear, containing also reduced amounts of biotite, almandine, kyanite and staurolite, very similar to the quartz ±

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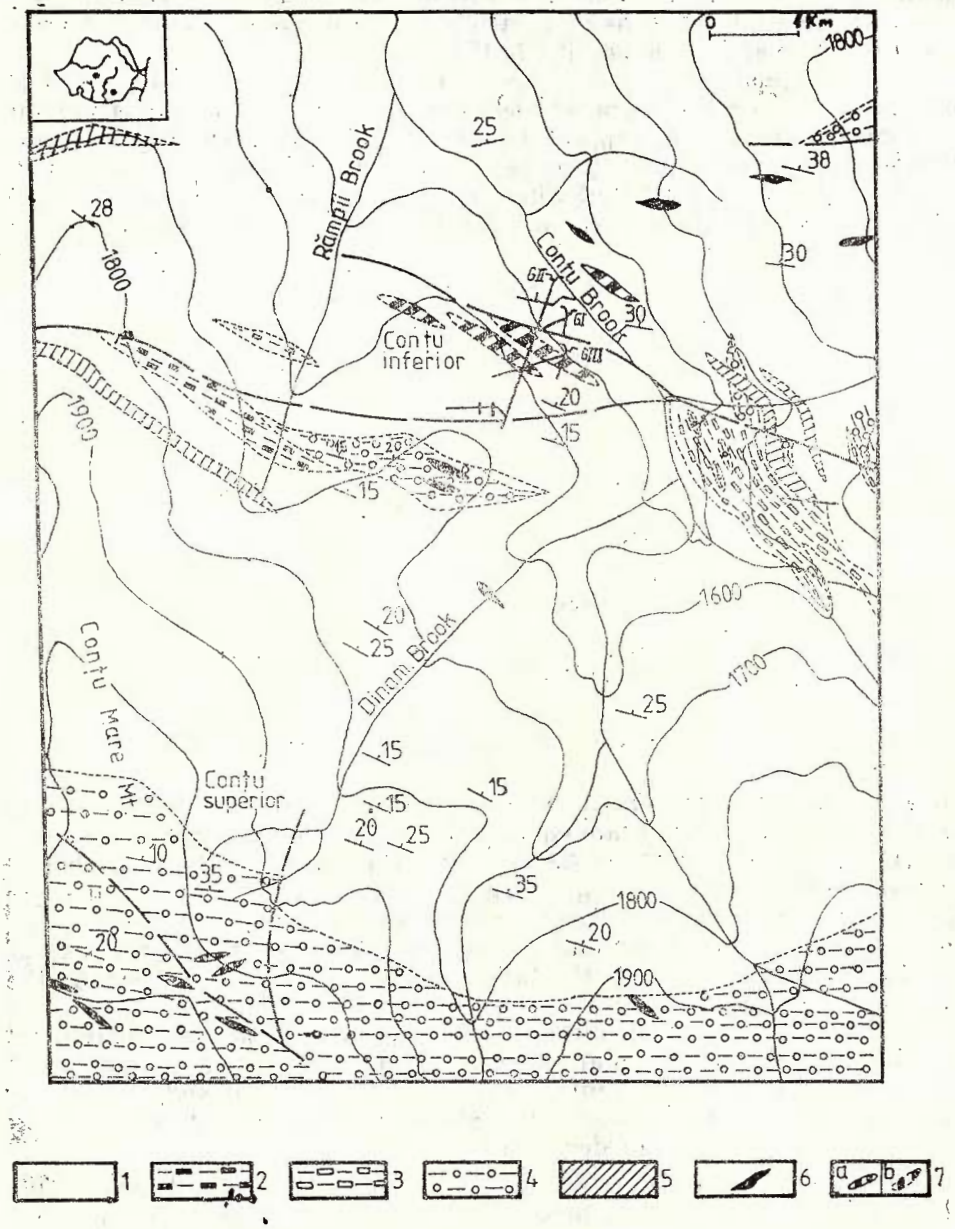

Fig. 1 — Geological map of the Conțu area (A. Apostoloiu). Conțu pegmatitic field. Lotru Group : 1, mica gneisses, micaschists ; 2, kyanite-bearing micaschists ; 3, staurolite-bearing micaschists ; 4, garnet micaschists ; 5, amphibolites, amphibole gneisses and schists ; 6, pegmatites ; 7, spodumene-bearing pegmatites : a, outcropping ; b, found in the underground works.

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oligoclase pegmatoid veins in the near-by rocks which may contain kyanite-staurolite intergrowths and garnet too. The above mentioned association is corroded later on (Pl. I, Fig. 1).

The pegmatites contain frequent local concentrations of phosphate minerals, the most widespread belonging to the lithiophilite-triphyline isomorphic series. They appear mainly as anhedral grains up to 3 cm, yellowish-green to pinkish in colour, sometimes enclosed in spodumene, with brown-black superficial alterations. Euhedral crystals are scarce. The chemical analysis of a green grain is reproduced in table 1, pointing

TABLE 1

Partial chemical analyses

		Triphyline	Purpurite
wt %	FeO	26.78	0.36
	Fe_2O_3	7.68	38.86
	MnO	7.28	—
	Mn_2O_3	—	7.75
mol	FeO	0.84	—
	FeO + MnO		
	Fe_2O_3	—	0.84
	$Fe_2O_3 + Mn_2O_3$		

Analyst: I. Stroilă I.P.G.G., Bucharest.

to an iron-rich term (Fe_{84}). X-ray diffraction measurements (tab. 2) carried out on the same grain equally stands for a ferrous term, as stated by the $d_{200,\ 121}$ maximum at 3.005 Å and the intensity of this peak and of those corresponding to $d_{111} = 3.469$ Å, $d_{131} = 2.509$ Å. The cell edges dimensions, each calculated on account of six independent d_{hkl} pairs range between the variation limits of reference data (JCPDS, 1974), except a_0 which is less, according to the $LiFe^{2+} \leftrightarrow Fe^{3+}$ substitution shown by the chemical analysis.

Under the microscope, lithiophilite-triphyline is always marginally corroded by cleavelandite and sometimes rimmed by younger phosphates. It may be overgrown by radiating laths of tavorite, strongly refringent (n = 1.8), pale yellow, lacking apparent polychroism, with inclined extinction (30°), $+2\ V = 75°$, moderate dispersion, $\Delta_{max} = 0.017$ (Pl. I, Fig. 2) or rimmed by huréaulite ($-2\ V = 75°$, $\Delta = 0.010$) and small indeterminable grains of a green polychroic mineral with high refractive index and birefringence (alluaudite ?) (Pl. I, Fig. 3). Triphyline is seldom mechanically twinned after (001) (parallel hemitropy) (Pl. I, Fig. 3). Investigations with the universal stage have shown the occurrence of several terms of the series, some having the optic axial section within (100), $+2V \leqslant \leqslant 30°$ ($Mn_{<42}$ and some within (001), $+2V = 5-20°$ ($Mn_{>42}$).

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TABLE 2

X-ray diffraction data for phosphate minerals from Conțu

No.	Triphyline Varutrask (JCPDS, 1974)			Lithiophilite Custer (JCPDS, 1974)			Triphyline* Conțu			Heterosite Varutrask (JCPDS, 1974)			Purpurite Feiteira (JCPDS. 1974)			Purpurite Conțu		
	I	d	hkl	I	d	hkl	I	d	hkl	I	d	hkl	I	d	hkl	I	d	hkl
1	30	5.22	020	10	5.12	020	10	5 081	020	40	6.38							
2	90	4.29	011	70	4.26	011	10	4 258	011	10	5 51							
3	30	3.95	120	40	3.92	120	10	3.889	120				30	4.96	020	40	4.898	020
4				10	3.69	101				25	4.85	023				15	4.41	
5	90	3.51	111	80	3.47	111	90	3.468	111	75	4 29	011	70	4.37	011			
6	10	3.33								40	3 87	120	15	3.777	120	25	3 697	
7	90	3.03	200				80	3.005	200	100	3.48	111	40	3.469	111	35	3.446	111
			121						121									
8	40	2.79	031	50	2.779	031	25	3 06										
9	100	2.541	131	100	2.531	131	100	2 509	131	50	2 93	200	100	2.952	121	100	2.896	121
			201									121						
10	20	2.47	211	50	2 469	211	40	2 458	211	75	2 73	031	10	2.702	031	15	2 668	031
11	10	2.39	140	40	2.382	002	40	2.361	140	10	2 51	220						
						140												
12				50	2.292					40	2.43	040	100	2 448	131	50	2.398	131
												131						
13	20	2.29	221	50	2 275	221	40	2 267	221	10	2.32		10	2.339	140			
14	10	2.15	112	40	2.148	112	30	2 200					10	2.337	221			
						022												
15				10	2.046	231				10	2.16	230	20	2.174	230	20	2 148	230
												112						
16				10	2 020	122							10	2.023	122			
17	10	1.86	132	40	1.839	132	40	1 839	132				10	1.974	231			
			202			202												
18				30	1.828	212	40	1.818	212				10	1.859	202			
19	10	1.81	151	10	1.813	151							20	1.82	212	15	1.816	212
			241															
20	50	1.75	922	80	1.750	222	60	1 739					10	1.778	311			
21	10	1.67	042	50	1.666		30	1 668					15	1.733	222	10	1.71	222
22	10	1.63		60	1 628		30	1.648					20	1 697	321	15	1.671	321
23	10	1.59		60	1 597		30	1.627					10	1.64	142			
24	40	1.51		50	1 51		50	1 51					20	1.575	013	20	1.570	013
25													20	1.558	160	20	1 559	160
26													10	1.523		10	1.506	
27										25	1.345		50	1.496		40	1.452	
28													20	1.412		20	1.411	
	a_0	b_0	c_0	a_0	b_0	c_0	a_0	b_0	c_0		b_0	c_0	a_0	b_0	c_0	a_0	b_0	$>c_0$
	6.06	10.32	4.71	6.05	10.42	4.41	6.00	10.35	4.69	5.83	9.70	4.77	5.83	9.70	4.77	5.74	9.42	4.77

* Diffractometer TUR-M-01 rad CuK_{α}, analyst : T. Urcan.

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In a similar context, brown-black or violet heterosite-purpurite, almost opaque, strongly polychroic (brown-reddish-violet) grains are often found associated with albitized K feldspar (Pl. I, Fig. 4). The textural relationships stand for its primary character.

They do not contain relict inclusions or rim triphyline and may react to secondary products as triphyline does. The diffractogram (tab. 2) suggests, by the high intensity of the $d_{121} = 2.896$ reflection, a term closer to purpurite, at variance with the Fe_{84} term resulted from the partial chemical analysis (tab. 1), probably due to the sample contamination. The unit cell parameters as calculated each from five independent d_{hkl} pairs are less than reference values, except c_0 which perfectly fits them.

Apatite is present as an accessory mineral.

Minerals of the amblygonite-montebrasite series occur subordinatedly as subhedral crystals or corroded granular aggregates (Pl. II, Fig. 5). Polysynthetic twins after (101) are frequent, sometimes lamellar twins after (101) are encountered too. On the universal stage, the measured $-2V$ values range between $52°(!) - 85°$ (amblygonite), sometimes exceeding $90°$ ($+2V = 88°$ — montebrasite, with $N_\alpha \perp (010)$. Birefringence measurements made with the Berek compensator, using quartz as a standard, led to the values $\Delta_{\alpha\beta} = 0.0085 - 0.010$, $\Delta_{\beta\gamma} = 0.0105 - 0.012$, $\Delta_{\beta\gamma} = 0.019 - 0.022$. Montebrasite appears to be younger than amblygonite.

Quartz is seemingly an early constituent during the pegmatite formation. Locally quartz rich portions develop, overprinted by later assemblages. The succession relationships are microscopically visible. Nevertheless, rather frequently quartz may recrystallize or accompany younger minerals.

The above mentioned association is corroded by the albite $\pm$ muscovite assemblage. Albite is present as lamellar polysynthetically twinned cleavelandite, sometimes graphically intergrown with muscovite (Pl. II, Fig. 1). Cleavelandite often cross-cuts an earlier quartz generation with scalloped texture, or grows starting from grain boundaries. Muscovite forms greenish subhedral books a few centimeters across or appears as radiating aggregates and graphic intergrowths with cleavelandite. It contains trace amounts of Li (Maieru et al., 1968). The K available after K-feldspar replacement by albite may represent a source for muscovite formation. Topas may seldom be present in this association as millimetric grains with a perfect basal cleavage (Pl. II, Fig. 2), $+2V = 80°$, $\Delta_{\alpha\gamma} = 0.0103 - 0.011$, $\Delta_{\beta\gamma} = 0.006$, $\Delta_{\alpha\beta} = 0.0043 - 0.005$.

However, cleavelandite aggregates are more often monomineralic. Sometimes they have an aplitic texture exhibiting a rather strong prefered orientation, which parallels that in the adjoining schists. When albite corrodes triphyline, the latter frequently recrystallizes and besides the earlier optically positive strained grains with anomalous interference colours one may encounter a new Fe-rich generation ($Fe_{>82}$) with normal interference, $-2V = 26° - 78°$, the optic axial plane lying within (100) (Pl. II, Fig. 3).

Spodumene commonly appears as subhedral, centimetric to decimetric crystals. Crystal faces with rough longitudinal striations, belonging to the (hk0) zone, are seldom seen. On account of textural relation-

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ships it is a late mineral, bearing triphyline, purpurite and quartz inclusions and being frequently intergrown with cleavelandite, muscovite and quartz. The crystals may form divergent aggregates or are perpendicular the pegmatite lens boundaries, making up a comb texture. Its properties, except the lack of polychroism, are those mentioned by Maieru et al. (1968). In a few instances, strained spodumene grains are partly replaced by cleavelandite, accounting for an earlier crystallization of the former.

In subordinate quantities, a reddish spessartinic garnet, coated by a thin alteration film is found in centimetric nests associated with quartz, cleavelandite-spodumene-muscovite assemblage.

Cassiterite builds up bipyramidal crystals up to 3 mm trending to concentrate near triphyline grains (Pl. II, Fig. 4). Some form „bec d'étain" twins. Usually cassiterite is clouded with tiny opaque inclusions.

Sphalerite is intergrown with late quartz and muscovite, sometimes having graphic aspects. It is black and shows abundant reddish internal reflections in reflected light.

In albite-muscovite rich and (not necessarily) finer grained portions, lamellar to tabular black, longitudinally striated crystals up to 4 cm appear, building up local concentrations. The X-ray diffraction pattern (tab. 3) allows the identification of a rare niobate, magnocolumbite. Diagnostic is the spacing at d = 7.12 which does not appear with other niobotantalates and the position and relative intensities of the triplets d = = 2.520 — 2.378 ; 1.702 — 1.720 ; 1.243 — 1.113. U is also present as an isomorphic substituent. A spectrometric investigation using an 0.52% U standard (analyst : I. Tiepac) yielded 0.126% U.

The crystallization succession as indicated by the textural relationships is represented in table 3. The order follows the relative abundances.

Genetical Features

The occurrence of rare elements-bearing minerals within the Conțu pegmatites raises the question of their source and seems to hint their filiation to plutonic bodies. However, this hypothesis can be easily ruled out by bearing in mind that granitoid bodies do not appear in the proximity of these pegmatites ; moreover, the Getic granitoids do not generate pegmatite bodies and contain rare elements only in trace amounts.

By the other hand, rare elements, except lithium, do not strongly concentrate in these pegmatites and their occurrence should be considered rather in connection with the prevailing physical conditions (temperature, vectors, etc.), than starting from a genetic schema (postmagmatic processes) which may fit more or less.

The appearance of index minerals typical of the schists within the pegmatite body constrains both the timing and the conditions of their formation. The host rocks have a polycyclic metamorphic history during the Precambrian, the last event having taken place within the kyanite-staurolite zone, close to the sillimanite isograd, i.e. about 5.5 Kbar and 610° — 625°C (Hârtopanu, 1982).

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TABLE 3

X-ray diffraction data for the magnocolumbite from Contu

No.	Magnocolumbite Pamir (JCPDS, 1974)		Magnocolumbite Contu	
1	I	d	I	d
2	3	7.08	3	7.12
3	1	4.01		
4	7	3.63	7	3.667
5	2	3.54		
6	6	3.26	2	3.32
7	10	2.955	10	2.939
8	2	2.841	4	2.850
9	1	2.736		
10	1	2.621		
11	5	2.516	5	2.520
12	6	2.480	6	2.487
13	6	2.382	5	2.378
14	1	2.280	1	2.240
15	1	2.230	1	2.215
16	1	2.206		
17	4	2.073	4	2.083
18	1	2.043		
19	1	1.985		
20	1	1.952		
21	7	1.897	3	1.897
22	6	1.830	4	1.832
23	2	1.805	1	1.801
24	8	1	5	1.772
25	8	1.738	6	1.726
26	9	1.723	7	1.720
27	1	1.673	1	1.670
28	1	1.618		
29	2	1.603		
30	9	1.535	4	1.532
31	3	1.486	3	1.483
32	9	1.470	7	1.462
33	9	1.454	7	1.444
34	1	1.397		
35	5	1.380	5	1.378
36	3	1.349		
37	4	1.319	2	1.313
38	1	1.289		
39	2	1.272	1	1.265
40	5	1.246	3	1.243
41	6	1.226	4	1.221
42	9	1.197	6	1.193

Diffractomer TUR-M 01 rad Cuk$_\alpha$, analyst: T. Urcan.

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These conditions are at variance with those estimated by Pomârleanu et al. (1967) from fluid inclusions. However, the last ones seem inconvenient as they plot outside the spodumene stability field (London, Burt, 1982).

The pegmatite formation accomplished during the last metamorphic event as the main minerals have grown either synkinematically (cleavelandite with prefered orientation, strained older grains) or statically (randomly oriented cleavelandite + quartz + spodumene).

The high lithium content needs an explanation about its source and concentration mechanism. Li fractionation by the evolution of a magmatic chamber implies large amounts of magma because of the low abundance of this element in melts ; such magmatic bodies are absent in the zone. At least as reasonable an alternative could be the anatectical separation of an Li-rich melt (Stewart, 1983). Experimental investigations (ibid.) show that the eutectic temperature can be depressed by at least 75° C when Li is present, providing an explanation for the frequent situations when lithian pegmatites do not associate with granitoids and appear in the vicinity of the sillimanite isograd.

However, the Conțu pegmatites do not have eutectic composition, containing only 0.8 — 1.2% Li_2O. Apart from this, modal compositions and textures do not argue for the origin from a melt. The large sized crystals suggest a great mobility of the components, being inconsistent with the high viscosity of melts stable at low temperatures. Radiating and comb spodumene textures, marginal replacements favour a metasomatic origin. The quartz and quartz-spodumene concentrations, the presence of fluid inclusions strongly suggest crystallization from a fluid. Aplitic structures, commonly considered to have originated from a melt by a rapid solidification during a fluid-phase escape (Jahns, Burnham, 1969) are the youngest in the Conțu pegmatite and are frequently of monomineralic constitution (albite), an unexpected situation for the final stages of a melt consolidation.

So, the Conțu pegmatites seemingly have originated mainly if not solely as a result of segregation into a fluid under preanatectic conditions and subsequent deposition, the minerals which could have been formed from a melt (however, evidence is lacking) are quantitatively subordinated.

Separation into a fluid phase with high dissolving character before the beginning of anatexis seems quite probable in the light of available theoretical and experimental data. By raising temperature, solubility increases as the dielectric constant of water rapidly falls, favouring the solution of neutral or low-charged complexes containing strong ligands (OH^-, Cl^-, F^-) ; as a consequence of density increase with pressure, solubility also increases (Crerar et al., 1985). At elevated temperatures, solutions contain molecular species of feldspar stoechiometry (Anderson, Burnham, 1983). Actually, the pegmatite formation reduces to the equilibrium between a fluid solution and solid phases under elevated TP conditions irrespective to the origin of the fluid (distillation from a magma or metamorphic nature).

During the sequence of crystallization (Fig. 2) the first minerals which were formed were those characteristic to the host schists, followed by Fe-Mn bearing phosphates (lithiophilite-triphyline), sometimes lac-

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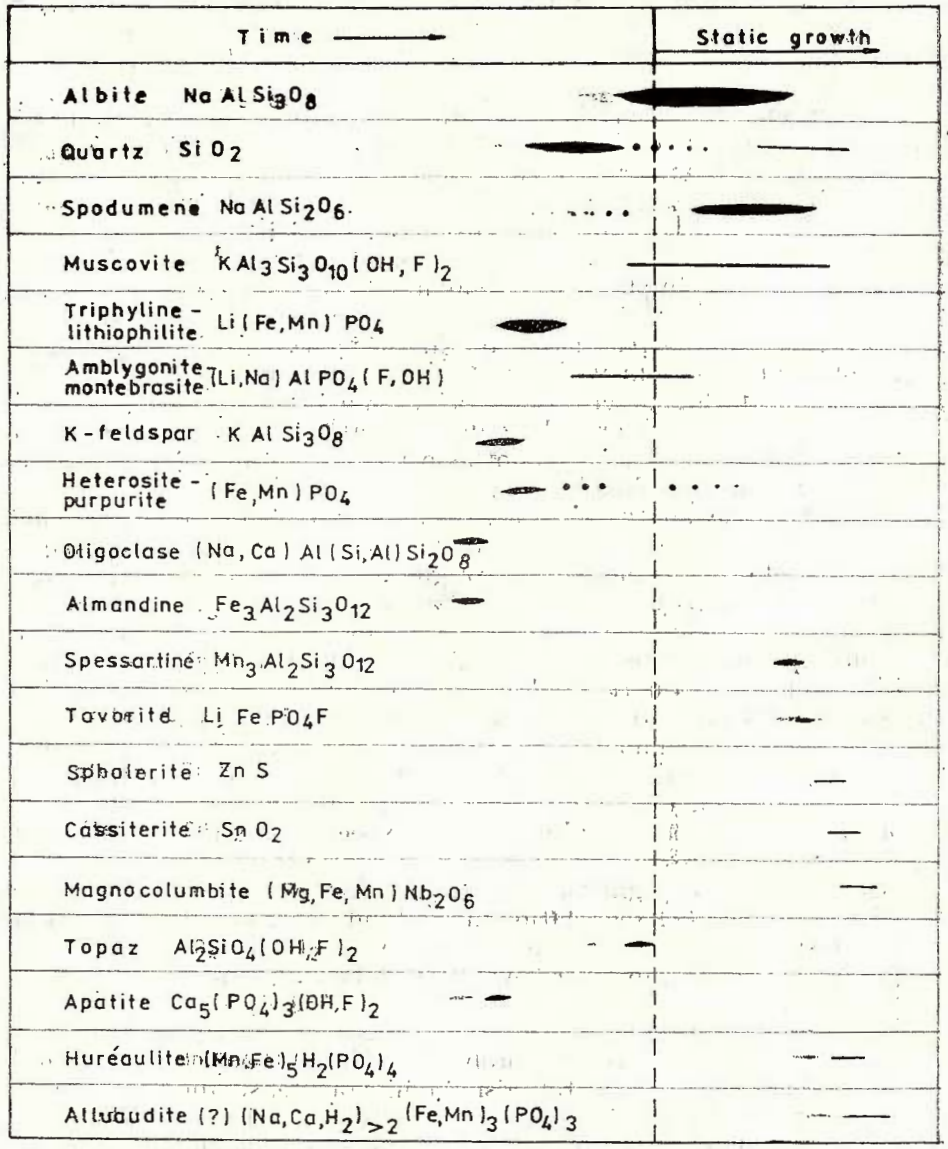

Fig. 2 — Succession of mineral crystallization in the Conțu pegmatites.

king Li (heterosite-purpurite). This corresponds to a Fe, Mn, P enriched and relatively Li-poor composition of the fluid (Černy et al., 1985). After the exhaustion of available Fe and Mn, amblygonite-montebrasite was first to crystallize in a P and F rich medium, Li aluminosilicates being unstable under those conditions (ibid).

The bulk separation of albite and quartz and strong fractionation of lithium into the resideral fluid result in an Li-enrichment followed by

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the deposition of the quartz-albite-spodumene assemblage. Volatile components — bearing minerals or those containing elements which use volatiles as vectors are subsequently formed contemporaneously with the appearance of secondary phosphates as alteration products of earlier phases. The appearance of Mg and U bearing niobates correlates with the early intense fractionation of Fe and Mn, elements which as a rule concentrate in their network, and offers a strong argument for a closed system evolution of the pegmatites.

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ASOCIAȚII MINERALE ÎN PEGMATITELE LITIFERE DE LA CONȚU, MUNȚII LOTRULUI (CARPAȚII MERIDIONALI CENTRALI) ȘI IMPLICAȚIILE LOR GENETICE

(Rezumat)

Pegmatitele litifere de la Conțu (munții Lotrului) sînt cantonate în formațiunea micașisturilor și a gnaiselor micacee de la partea superioară a grupului Lotru, în pînza getică. Compoziția mineralogică, realizată în

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trei stadii, cuprinde albit, cuarț, spodumen, muscovit, trifilin-litiofilin, ambligonit, montebrasit, microclin, purpurit, casiterit, magnocolumbit uranifer, blendă, topaz, apatit, alluaudit (?), huréaulit, tavorit, blendă, granați. Sînt menționate pentru prima dată pe teritoriul țării: ambligonitul, purpuritul, magnocolumbitul, huréaulitul și tavoritul.

Absența asocierii cu corpuri intrusive și relațiile cu rocile gazdă indică o geneză metamorfică. Relațiile dintre minerale și compoziția lor argumentează o cristalizare dintr-un fluid, fără o topire observabilă. Condițiile de formare sînt ale ultimului eveniment metamorfic precambrian al grupului Lotru (610 — 625°C, cca 5,5 Kb).

EXPLANATION OF PLATES

Plate I

Fig. 1 — Garnet corroded by cleavelandite-muscovite-amblygonite association. N+, 20 ×.

Fig. 2 — Triphyline relics with tavorite rim, substituted by cleavelandite. N+, 30 ×.

Fig. 3 — Polysynthetically twinned triphyline with huréaulite rim, marginally altered into huréaulite and alluaudite (?). N+, 30 ×.

Fig. 4 — Potassic feldspar marginally substituted by albite, with substitution perthites in contact with purpurite. Purpurite forms secondary phosphates on the margin. N+, 5 ×.

Fig. 5 — Twinned amblygonite included in cleavelandite, N+, 50 ×.

Plate II

Fig. 1 — Graphical synkinematic aggregates of cleavelandite and muscovite corrode quartz. N+, 10 ×.

Fig. 2 — Cleavelandite-topaz vein corroding quartz. N+, 10 ×.

Fig. 3 — Oriented cleavelandite with two generations of triphyline. N+, 10 ×.

Fig. 4 — Triphyline crystal corroded by cleavelandite. Crystal of cassiterite at the cleavelandite-triphyline contact. N+, 10 ×.

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PETROLOGIA ROCILOR METAMORFICE

OBSERVATIONS SUR LES ROCHES CRISTALLOPHYLLIENNES GRANITOÏDES ET FILONIENNES DES MONTS DE CODRU (MONTS APUSENI)[1]

par

NICOLAE STAN[2], MIRCEA ARION[3], ELENA COLIOS[3], ANA ȘERBĂNESCU[2]

Metamorphic rocks. Cambrian. Granite. Sedimentary rocks. Permian. Carboniferous. Volcanoclastics. Sarmatian. Prograde metamorphism. Retrograde metamorphism. Major elements. Minor elements. Apuseni Mountains — North Apuseni Mountains — Codru-Moma Mountains.

Résumé

Le soubassement de la région des monts de Codru comporte des schistes cristallophylliens (série de Botfei) probablement d'âge cambrien, métamorphosés dans le faciès des schistes verts. Ils sont recouverts transgressivement par des dépôts sédimentaires permocarbonifères, ceux permiens étant intercalés par des écoulements rhyolitiques ignimbritiques. Les granitoïdes de Codru (granites, granodiorites, diorites quartzifères, tonalites) sont d'âge permien. Ils forment avec les rhyolites ignimbritiques un complexe magmatique intrusif-effusif acide. Les roches filoniennes représentées par les rhyolites, lamprophyres, spilites et pegmatites sont très rares. Le rétromorphisme de la série de Botfei a eu lieu à la fin du Paléozoïque ou, plus probable, par le glissement de la nappe de Finiș, pendant le Mésozoïque.

Abstract

Observations on the Crystalline Granitoid and Vein Rocks in the Codru Mountains (Apuseni Mountains). The basement of the region consists of crystalline schists (Botfei Series) of Cambrian age metamorphosed in the greenschist facies. They are transgressively overlain by

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Permo-Carboniferous sedimentary deposits, the Permian ones being intercalated with ignimbritic rhyolitic flows. Codru granitoids (granites, granodiorites, quartz diorites, tonalites) are of Permian age. Beside the ignimbritic rhyolites they form an acid intrusive-effusive magmatic complex. Vein rocks represented by rhyolites, lamprophyres, spilites and pegmatites are rarely found. The retromorphism of the Botfei Series took place at the end of the Paleozoic or, more likely, simultaneously with the sliding of the Finiș Nappe during the Mesozoic.

1. Introduction

Les granitoïdes de Codru et les formations cristallophylliennes appartiennent aux unités kimmériennes d'Armanu et Dumbrăvița (Bleahu et al., 1984).

Les granitoïdes apparaissent sous forme d'une bande longue de 15 km, orientée NNW—SSE, tout près des localités de Mărăuș, Botfei, Hășmaș, Susani. La largeur maximum est de 3 km. Ces granitoïdes sont en contact direct avec les schistes cristallophylliens appartenant à la série de Botfei et avec la formation sédimentaire permo-carbonifère. Les schistes cristallins sont recouverts en discordance par les dépôts permo-carbonifères. Ces dépôts comportent des conglomérats laminés, grés vermiculaires, tufs et grès feldspathiques. Au niveau des grès feldspathiques se trouvent intercalés des écoulements de rhyolites ignimbritiques. Chronologiquement suit la formation volcano-sédimentaire sarmatienne et les dépôts quaternaires.

Les roches granitoïdes notamment sont intersectées par des filons de rhyolites, granodiorites—porphyres, spilites et rarement lamprophyres ou pegmatites. Certains filons rhyolitiques traversent même les conglomérats permo-carbonifères.

Dans le cadre des roches granitoïdes, une position particulière est occupée par deux bandes de roches basiques orthoamphibolitiques localisées une dans la vallée de Hășmaș et l'autre dans la vallée d'Albu.

Les roches cristallophylliennes et granitiques des monts de Codru ont été investiguées par Palfy et Rozlozsnik (1939), Tabacu (Bleahu et al., 1958), Cîmpeanu (1970), Georgescu et al. (1972), Barus et Rișa (1973), Nica et al. (1974), Arion (Georgescu et al., 1977) et Stan (Bleahu et al., 1984).

Les schistes cristallins et les granitoïdes ont été mentionnés aussi dans quelques travaux d'anvergure régionale, travaux appartenant à Paucă (1941), Dimitrescu (1964) et Ianovici et al. (1976).

2. Schistes cristallins (série de Botfei)

Cette série se développe surtout dans la partie nord et nord-est du massif granitoïde. Les schistes cristallins affleurent aussi sous forme de bandes plus ou moins développées à l'intérieur du massif granitoïde, dans la plupart des cas ayant des épaisseurs de quelques mètres, qui ne peuvent pas être levées à l'échelle 1 :25.000. C'est pour cette raison qu'on trouve sur la carte, assez souvent, un groupe de telles bandes comme une seule bande de roches cristallophylliennes de 25 à 50 m d'épaisseur.

La série de Botfei, à caractère séricito-chloriteux surtout, comporte des quartzites plus ou moins feldspathiques séricitisés, des schistes quar-

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tzitiques micacés et micaschistes séricito-chloriteux ou schistes séricito-chloriteux. Le passage d'un type de roche à un autre se réalise graduellement. Les structures sont lépidoblastiques, parfois microplissées ou granolépidoblastiques et les textures fréquemment schisteuses, rarement orientées.

La série de Botfei renferme relativement un nombre réduit de minéraux : quartz, séricite, chlorite, muscovite, biotite, albite (plus rare oligoclase acide), accompagnés parfois de grenat et accidentellement d'épidote. Le zircon, l'apatite et la tourmaline ont une apparition sporadique. Y est caractéristique la présence d'une poussière fine de minéraux opaques, diffusée irrégulièrement dans la masse des schistes cristallins.

Dans les bandes de schistes cristallins des roches granitoïdes, tout comme dans les schistes cristallins situés tout près des granitoïdes de Codru, on rencontre fréquemment le grenat, la biotite et la muscovite en état fraîche ou faiblement altérées par rapport à la matrice sériciteuse ou/et chloriteuse, qui les renferme. Parfois on peut observer aussi le développement des minéraux micacés de néoformation orientés obliquement ou perpendiculairement par rapport à la schistosité. A mesure qu'on s'éloigne des roches granitoïdes les porphyroblastes de grenat, biotite et muscovite deviennent plus rares jusqu'à leur disparition totale.

Dans la vallée de Secaciu, prés de la biotite et le grenat de néoformation on a identifié, dans une seule coupe, la sillimanite aussi. Dans la même zone, on a observé, tout prés du granitoïde, dans les schistes cristallins, quelques porphyroblastes entièrement séricitisés qui suggèrent l'existence antérieure de l'andalusite. Ce minéral est mentionné aussi par Palfy et Rozlozsnik (1939) dans la region, au contact des schistes cristallins avec les roches granitoïdes. Les porphyroblastes d'andalusite, séricitisés, ont été observés aussi dans les schistes cristallins de la vallée de Hăşmaş ; ils y ont une position discordante par rapport à la schistosité.

3. Orthoamphibolites (hornblendites)

Ces roches comportent 90 à 95 % hornblende verte, parfois faiblement chloritisée et subsidiairement biotite, quartz, plagioclase kaolinisé séricitisé et minéraux accéssoires : sphène, apatite et rarement grains de minéraux opaques. En des certains cas on passe à des roches plus claires, où la hornblende participe en proportion plus réduite (65 à 75 %) ; parallèlement augmente la teneur en minéraux saliques, plagioclase fortement altéré et quartz.

La composition chimique d'une hornblendite de la vallée de Hăşmaş, exprimée en pourcentages pour les éléments majeurs est la suivante : $SiO_2 = 50{,}93$, $TiO_2 = 0{,}55$, $Al_2O_3 = 7{,}84$, $Fe_2O_3 = 2{,}88$, $FeO = 5{,}10$, $MnO = 0{,}17$, $MgO = 18{,}22$, $CaO = 9{,}57$, $K_2O = 0{,}94$, $Na_2O = 1{,}00$, $P_2O_5 = 0{,}10$, $H_2O^+ = 1{,}23$, $CO_2 = 1{,}01$ (analyste : E. Colios). Pour le même échantillon ont été analysés les éléments mineurs aussi (ppm) : Pb = 12, Cu = 70, Ga = 9,5, Ni = 300, Co = 65, Cr = 1000, V = 130, Sc = 34, Zr = 85, Ba = 160, Sr = 130, Y = 14, Yb = 1,0 (analyste : A. Şerbănescu).

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4. Granitoïdes de Codru

Les intrusions des granitoïdes de Codru occupent une surface de plus de 30 km². Elles sont représentées par des granites et granodiorites et subordonnément diorites quartzifères et tonalites. Les passages entre ces roches se fait graduellement. La texture est massive, la structure holocristalline à grains moyens. Près des failles les structures cataclastiques sont très fréquentes.

Les principaux minéraux compris dans les granitoïdes sont : plagioclase (albite, oligoclase ± andésine), quartz ± feldspath potassique (microcline) ± biotite, rarement hornblende. Beaucoups des granitoïdes situés dans la zone de nord — est (Botfei—Secaciu—Mărăuș) ont des paillettes de muscovite amplement développées, dont la longuer peut atteindre 1 cm. Quelques granitoïdes sont pauvres en minéraux fémiques, réalisant le passage vers les leucogranites ou trondhjémites.

4.1. Chimisme des roches granitoïdes

Pour l'étude chimique ont été prélevés 14 échantillons des principales vallées. Sur le même échantillon on a déterminé tant les éléments majeurs que ceux mineurs aussi.

4.1.1. *Eléments majeurs*. La composition chimique des roches, exprimée en pourcentages, est présentée dans le tableau 1.

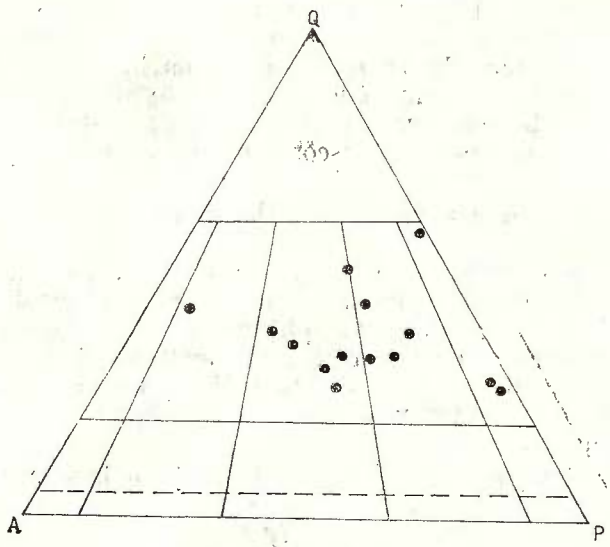

Fig. 1 — Diagramme *Q-A-P*(norme Rittmann).

Dans le diagramme $Q-A-P$ (composition normative—Rittmann, fig. 1) les analyses se situent dans les champs de : granites (syénogranites, monzogranites), granodiorites et tonalites. Dans le champ des tonalites se situent les roches à composition modale correspondant aux tonalites et diorites quartzifères.

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TABLEAU 1

Composition chimique

No	No échant.	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	K_2O	Na_2O	P_2O_5	H_2O+	CO_2	S	Fe(s)	Total	Localisation
1	154	57,89	0,76	17,19	2,43	3,10	0,12	4,05	4,58	2,42	3,98	0,65	1,84	0,71	0,07	0,06	99,85	Vallée Clit
2	179	57,92	0,90	17,36	2,35	3,86	0,12	3,61	5,39	2,50	3,20	0,31	1,95	0,19	0,12	0,10	99,88	Vallée Hăşmaş
3	219	64,42	0,57	18,11	1,23	2,47	0,07	1,49	2,51	3,31	3,54	0,28	1,84	0,00	0,06	0,05	99,95	Vallée Secaciu
4	177	65,65	0,52	16,56	1,43	2,59	0,13	1,66	1,98	3,28	3,62	0,29	1,83	0,11	0,10	0,09	99,84	Vallée Hăşmaş
5	150	65,50	1,73	15,61	1,16	2,25	0,08	1,42	3,19	3,00	3,35	0,29	1,58	0,00	0,07	0,06	100,29	Vallée Clit
6	136	67,27	0,60	16,60	1,18	3,11	0,09	1,10	3,16	1,68	3,35	0,22	0,90	0,23	0,08	0,07	99,70	Vallée Urviş
7	172	67,49	0,68	15,05	1,46	2,38	0,10	1,75	1,74	3,00	3,87	0,30	2,08	0,00	0,10	0,09	100,09	Vallée Hăşmaş
8	137	69,59	0,45	15,96	0,60	1,98	0,09	0,90	2,43	2,66	4,10	0,21	1,31	0,00	0,06	0,05	99,79	Vallée Urviş
9	133	69,93	0,33	16,10	0,77	1,51	0,07	1,05	1,26	3,28	3,75	0,19	0,88	0,44	0,09	0,08	99,75	Vallée Urviş
10	135	70,70	0,45	15,60	0,98	1,63	0,07	0,93	1,96	2,37	3,17	0,29	1,07	0,15	0,21	0,18	99,82	Vallée Urviş
11	161	70,86	0,59	15,32	1,10	1,24	0,06	0,85	1,67	3,28	3,43	0,17	1,22	0,00	0,08	0,07	99,75	Vallée Botfei
12	158	72,04	0,40	15,01	0,92	1,03	0,05	0,75	1,69	2,80	2,90	0,04	1,05	0,35	0,08	0,07	99,98	Vallée Botfei
13	180	73,13	0,30	14,71	0,80	0,83	0,05	0,58	0,91	4,50	2,90	0,21	0,67	0,23	0,06	0,05	99,93	Vallée Hăşmaş
14	226	76,35	0,17	14,19	0,40	0,05	0,03	0,22	0,61	2,40	3,62	0,06	1,11	0,36	0,06	0,05	99,68	Vallée Albului

Analyste: E. Colios.

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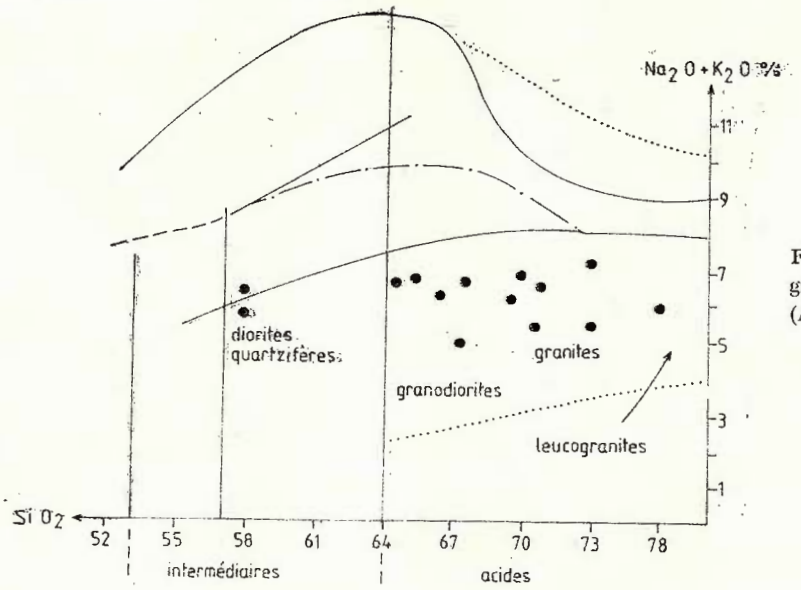

Fig. 2 — Diagramme TAS (Andreeva et al., 1981).

TABLEAU 2

Paramètres Niggli et types de magmas

No.	No. échant.	si	al	fm	c	alk	k	mg	Types de magmas
1	154	186,4	32,6	34,2	15,8	17,4	0,29	0,57	quartz-dioritique
2	179	185,1	32,7	33,8	18,5	15	0,34	0,51	quartz-dioritique
3	219	265,9	44,1	22	11,1	22,9	0,38	0,42	granitique-granodioritique
4	177	280,5	41,7	25,3	9,1	23,9	0,37	0,42	granodioritique
5	150	292	40,4	21,9	15	22,7	0,37	0,42	granodioritique
6	136	295	43,1	23,1	14,8	18,9	0,25	0,31	granodioritique
7	172	303	39,8	26,4	8,4	25,4	0,34	0,44	granitique
8	134	333,4	45,1	17,1	12,5	25,3	0,25	0,38	trondhjémitique
9	133	347,5	47,1	17,7	6,6	28,5	0,37	0,44	leucogranitique
10	135	359,6	46,9	19,1	10,7	23,3	0,33	0,37	leucogranitique-granitique
11	161	363	46,3	16,7	9,2	27,8	0,39	0,39	leucogranitique
12	158	407	49,4	15,6	10,1	25	0,37	0,40	trondhjémitique
13	180	416	49,3	12,8	5,5	32,3	0,51	0,38	trondhjémitiques
14	226	515,7	56,5	5,1	4,4	34	0,30	0,44	leucogranitique

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Dans le diagramme TAS (Andreeva et al., 1981, fig. 2) les granodiorites se classifient en leucogranites, granites, granodiorites et diorites quartzifères.

Les valeurs des paramètres Niggli (tabl. 2) et le diagramme SiO_2-IA (indice d'alcalinité, fig. 3) attestent le caractère calcoalcalin des gra-

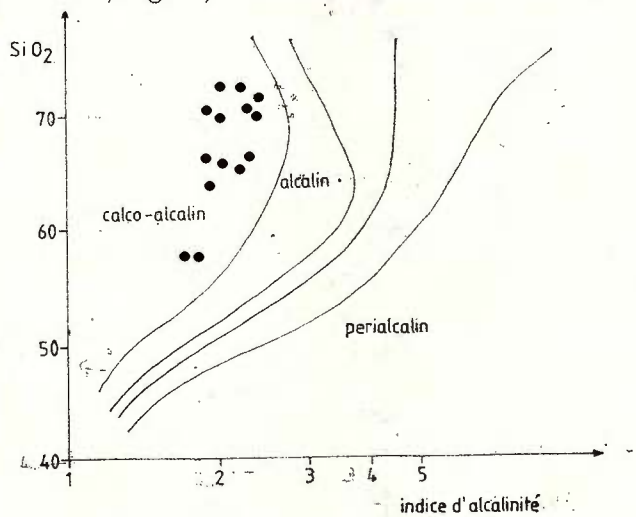

Fig. 3 — Diagramme SiO_2—I.A.

nitoïdes de Codru. Les types de magmas sont les suivants : quartz-dioritiques, granite-granodioritiques, granodioritiques, granitiques leucogranitiques et trondhjémitiques. Ab > An dans les roches granitiques ou leucogranitiques et Ab < An dans les diorites quartzifères. Généralement Ab > Or.

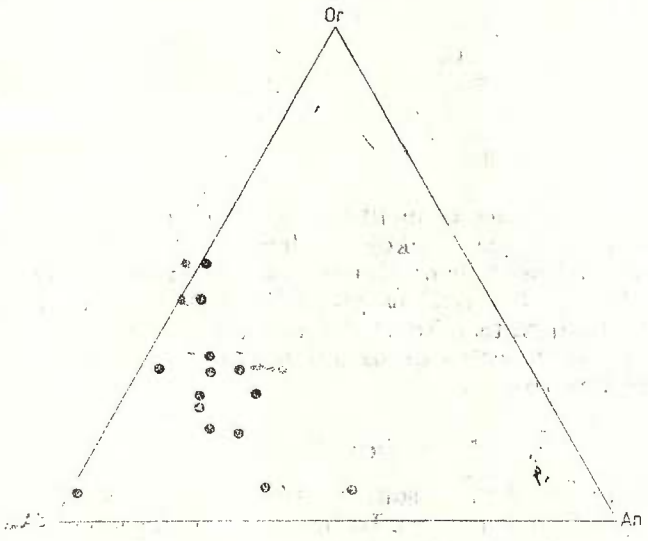

Fig. 4 — Diagramme Or—Ab—An.

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4.1.2. *Eléments mineurs*. La teneur en éléments mineurs des roches granitoïdes a été déterminée par la méthode de la spectrographie d'émission. Les résultats sont présentés dans le tableau 3.

TABLEAU 3

Eléments mineurs (ppm)

No.	No. échant.	Pb	Cu	Sn	Ga	Ni	Co	Cr	V	Sc	La	Zr	Be	Ba	Sr	Y	Yb
1	154	27	9	5	22	37	11	20	130	18	58	122	4,2	980	1030	30	1,5
2	179	17	15	3,5	20	4	13	17	156	20	<30	140	2,3	760	840	26	2,1
3	219	12	8	<	18	15	10	20	65	10	44	150	3,7	850	350	11	<1
4	177	100	35	<2	20	11	7	13	62	8,5	44	150	4,8	790	295	13	<1
5	150	22	25	<2	19	11	6	11	58	8,5	46	168	3,9	860	340	16	<1
6	136	15	8	<2	20	6,5	5	<10	40	10	44	220	2,3	600	300	22	<1
7	172	10	13	<2	17	8	8,5	32	65	15	<30	280	2,6	950	320	35	2,1
8	137	23	5	4,5	20	8,5	4	<10	28	7,5	40	130	5,0	430	310	15	<1
9	133	19	30	7,5	20	9	4	<10	34	7	46	150	4,6	600	280	14	1
10	135	22	31	<2	19	7	4	<10	28	6,5	46	130	4,7	875	320	13	<1
11	161	17	15	6,5	19	6,5	3	<10	23	6	38	135	4,8	710	260	11	<1
12	158	16	10	<2	15	6,5	3	<10	23	5,5	34	130	1,6	750	280	<10	<1
13	180	29	7	<2	14	<2	<2	<10	17	<2	<30	105	1,7	720	170	14	<1
14	226	9	3	<2	17	3	<2	<10	5	4	<30	25	3,2	160	95	<10	<1

Analyste: A. Şerbănescu.

Pour les principaux éléments on a rédigé des diagrammes de variation fonction de l'indice Nockolds-Allen (fig. 5, 6). Les teneurs en Sc, Co, Mg, Ni, Fe, V, Ca et Sr diminuent parallèlement à l'augmentation de l'acidité des diorites quartzifères, granodiorites, vers monzogranites, leucogranites et trondhjémites. Na et Ba ont des valeurs relativement constantes pendant le processus de différentiation. La teneur en K augmente avec l'acidité des roches.

5. Roches filoniennes

Les roches filoniennes sont représentées par des rhyolites, spilites, lamprophyres et pegmatites. La longueur ne dépasse pas 700 ou 800 m, leur épaisseur mesurant quelques mètres.

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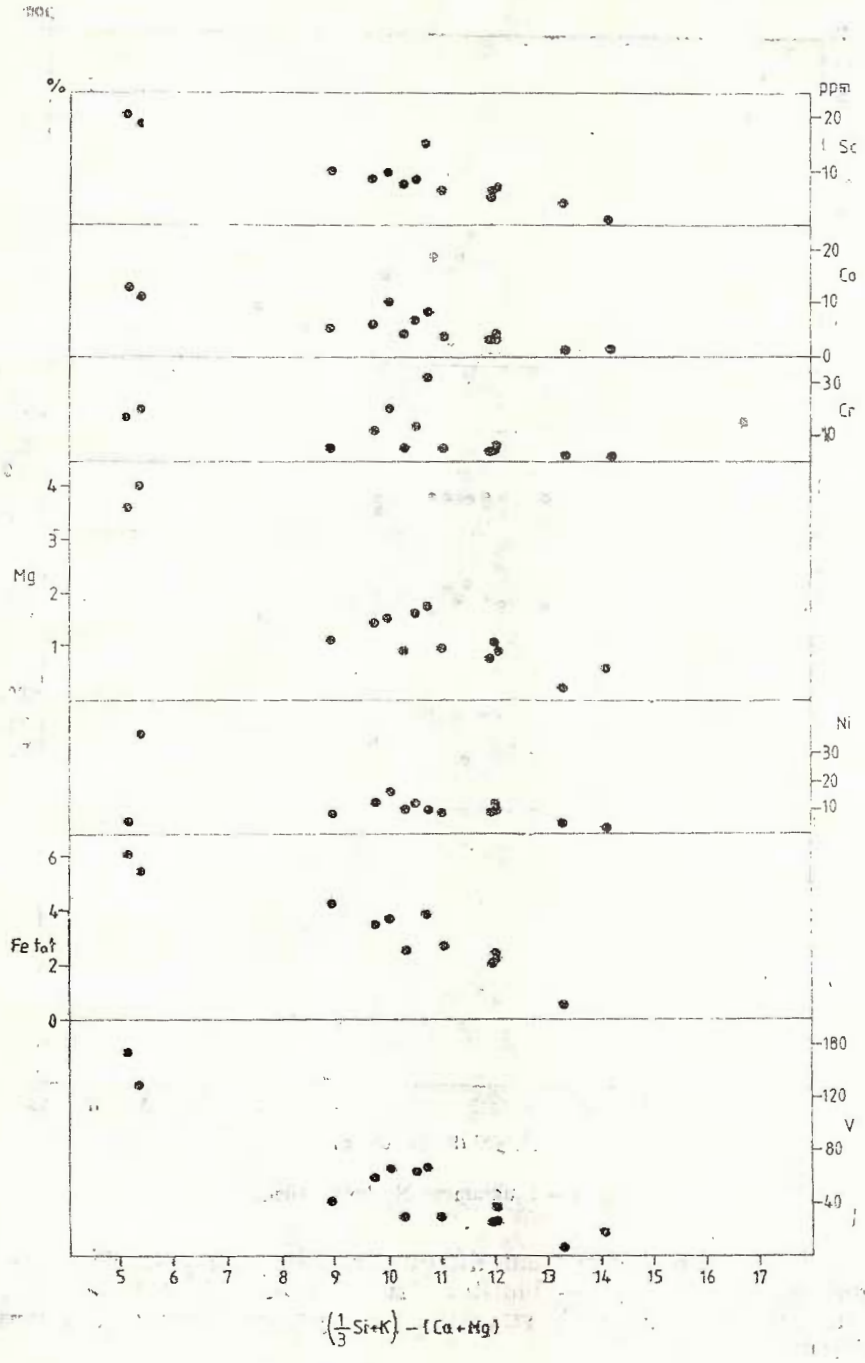

Fig. 5 — Diagramme Nockolds-Allen.

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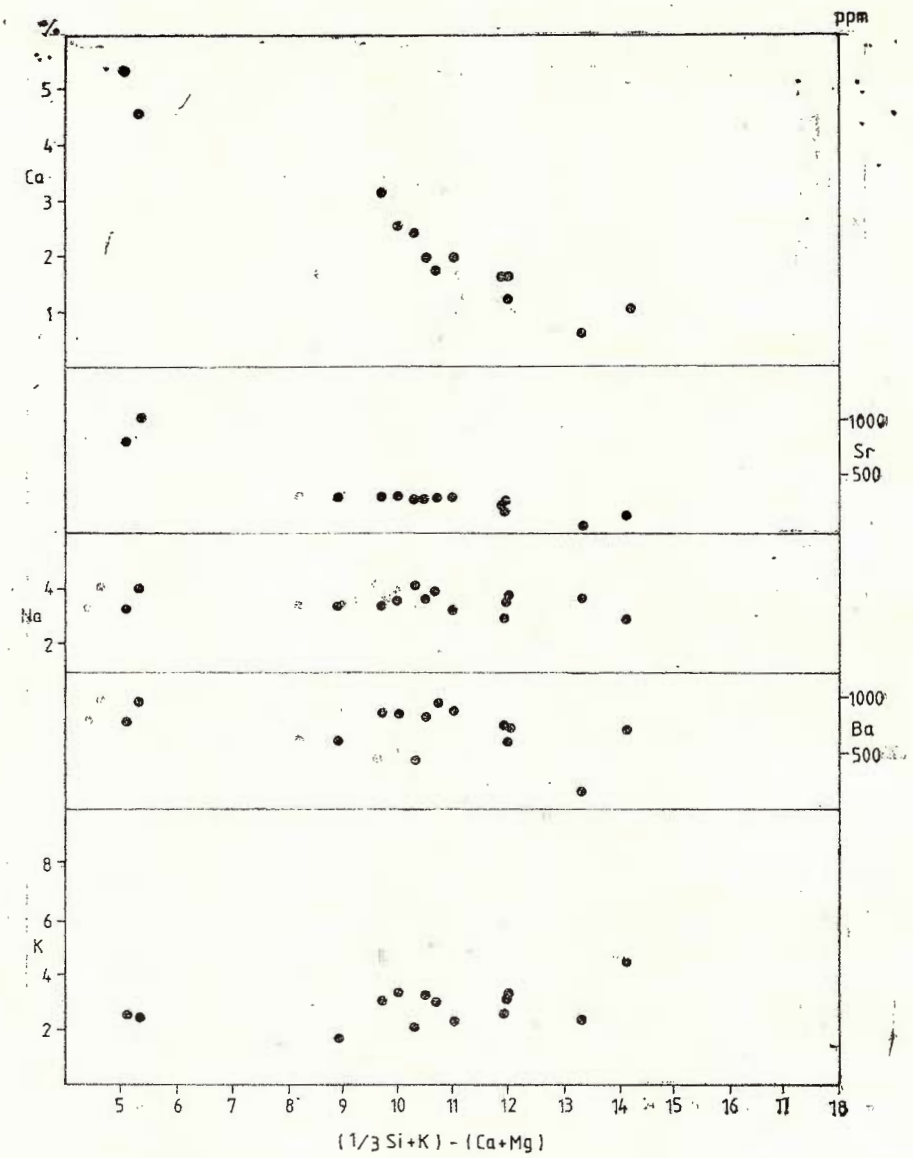

Fig. 6 — Diagramme Nockolds-Allen.

Les rhyolites comportent du quartz, feldspath potassique, albite-oligoclase ± muscovite ± biotite. La masse fondamentale est fortement séricitisée, ayant la structure porphyrique et la texture souvent schisteuse.

Les granodiorite-porphyres présentent une structure microgranulaire porphyrique et comportent des phénocristaux de quartz, albite et orthose.

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Les spilites renferment albite, chlorite, calcite, minéraux opaques. Elles ont une structure basaltique subophitique, la texture massive ou vaguement schisteuse.

Les lamprophyres comportent pyroxènes, hornblende, plagioclase altéré ± biotite incipient chloritisée et minéraux opaques. La structure est panidiomorphe, la texture massive.

Les pegmatites présentent une composition minéralogique simple : feldspath potassique (microclin) ± albite. La structure est holocristalline, la texture massive. La composition chimique de la pegmatite de Valea Mare, exprimée en pourcentages, est la suivante : $SiO_2 = 71,80$, $TiO_2 =$, 0,06, $Al_2O_3 = 14,89$, $Fe_2O_3 = 0,39$, $MnO = 0,03$, $MgO = 0,24$, $CaO =$ 0,56, $K_2O = 8,64$, $Na_2O = 2,45$, $P_2O_5 = 0,16$, $H_2O = 0,23$, $CO_2 = 0,23$ (analyste : E. Colios). On donne les teneurs en éléments mineurs (ppm) : $Pb = 34$, $Cu = 4,5$, $Sn < 2$, $Ga = 8$, $Ni = 4$, $Co < 2$, $Cr < 10$, $V = 4,5$, $Sc < 2$, $La < 30$, $Zr = 12$, $Be = 1,8$, $Ba = 750$, $Sr = 130$, $Y < 10$, $Yb < 1$ (analyste : A. Şerbănescu).

6. Interprétation des données

Tenant compte de la constitution pétrographique de la série de Botfei, on suppose qu'elle a été engendrée par la métamorphose de certains sédiments détritiques principalement psamitiques quartzeux ou quartzo-feldspathiques et rarement pélitiques, argileux.

Il est difficile à déterminer le métamorphisme initial prograde de la série de Botfei, d'une part à cause de la composition chimique de début inadéquat aux sédiments, d'autre part à cause du rétromorphisme manifesté à l'échelle régionale, qui a transformé les éventuels minéraux index en minéraux secondaires. On peut affirmer avec certitude que le faciès à schistes verts, notamment l'isograde à biotite a été atteint. Mais le rétromorphisme de la série de Botfei est évidente, même pour ce faciès : la muscovite, le plagioclase et la biotite sont transformés partiellement ou en totalité en séricite et respectivement en chlorite sur des vastes surfaces.

Un rétromorphisme rétrograde a eu lieu à la fin du Permien ou au commencement du Mésozoïque parce que les conglomérats permo-carbonifères transgressifs sur le soubassement cristallin sont laminés. Ces mouvements qui ont affecté les conglomérats permo-carbonifères ont eu des influences, sans doute, sur le soubassement cristallin de la série de Botfei aussi. Il n'est pas exclus que des phénomènes de rétromorphisme aient eu lieu avant les mouvements herciniens ou alpins aussi, mais ce fait ne peut pas être démontré.

Les mouvements subséquents au métamorphisme régional prograde, qui ont déterminé le rétromorphisme, sont visibles dans les clivages de ciseillement des schistes cristallins, tout comme dans les bandes de cristallin des granitoïdes, bandes à une orientation NNW—SSE, qui pourraient correspondre à un moment de mouvement S_1, en rapport avec les schistes cristallins situés en déhors du granitoïde, ayant des orientations approximativement E—W et qui pourraient correspondre à un autre moment de mouvement S_2.

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Les minéraux de néoformation, la biotite, la muscovite, le grenat, la sillimanite et l'andalusite, inaltérés ou faiblement altérés par rapport à la matrice qui les renferme, disposés obliquement ou parallèlement à la linéation, sont considérés comme minéraux de contact thermique. Les paragenèses minéralogiques du métamorphisme de contact thermique des monts de Codru correspondent, dans la conception de Winkler (1967), au faciès des cornéennes à pyroxènes.

Sous le nom d'intrusions de Codru, Dimitrescu (1964) a inclus tant les roches granitoïdes que celles basiques associées aussi (orthoamphiboles, gabbros, méladiorites). Plus récent, Dimitrescu (1986) décrit les intrusions basiques de Codru, leur attribuant les caractères d'une série appinitique intimement associée aux intrusions trondhjémitique-quartzdioritiques. D'autre part les orthoamphibolites à structures reliques magmatiques peuvent être aussi interprétées soit comme faisant partie de la série de Botfei, soit, plus probable, comme appartenant à une série cristallophyllienne plus ancienne. Dans la dernière variante, les roches basiques de la vallée de Hăşmaş et la vallée d'Albu pourraient être corrélées avec les orthoamphibolites de la série cristallophyllienne de Codru, figurée par Balintoni sur la carte à 1 : 50 000, feuille de Valea Ierii (Hârtopanu et al., 1982). Dans cette conception, les roches les plus vieilles de la région seraient même les orthoamphibolites, probablement d'âge précambrien-cambrien. La série de Botfei, plus faiblement métamorphosée, est probablement cambrienne, tout de même plus ancienne que le Carbonifère.

L'intrusion des granitoïdes de Codru a eu lieu pendant le Permien, peu avant l'action du volcanisme rhyolitique-ignimbritique (Stan, 1989). Ce fait suggère une source magmatique commune qui a alimenté tant les volcanites rhyolitiques que les granitoïdes aussi. Exceptant la ressemblance chimique et minéralogique entre les roches intrusives et celles effusives, tout comme le laps de temps court qui les sépare, on doit tenir compte aussi du fait que certains filons acides qui traversent le granitoïde ou les conglomérats laminés permiens ont une structure hypabyssale. Ils font le passage des granitoïdes vers les rhyolites. On mentionne les filons granodiorite-porphyriques identifiés à Ciuntești et dans la vallée de Secaci, signalés déjà par Istocescu (1971). Ceux-ci représentent les racines des éruptions volcaniques permiennes qui ont engendré les volcanites rhyolitiques (Dimitrescu et al., 1965). Il existe donc, dans les monts de Codru, un édifice magmatique intrusif-effusif d'âge permien.

Les rhyolites ignimbritiques reposent directement sur ces intrusions partialement érodées et affleurées, fait qui suggère que les granitoïdes se sont consolidés dans l'écorce à des profondeurs réduites.

L'absence des éléments de granitoïdes (Istocescu, 1971) dans les conglomérats laminés permo-carbonifères s'explique maintenant par le fait que les intrusions sont plus récentes que les conglomérats.

Les granitoïdes des monts de Codru, d'âge hercinien sont post-cinématiques par rapport aux schistes cristallins de la serie de Botfei. Ces granitoïdes peuvent être parallèlisés avec les granitoïdes de Highiş (Balintoni, 1986) ou le complexe magmatique rhyolitique-granitique permien de Highiş, tel qu'il a été considéré par Giuşcă (1979).

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OBSERVAŢII ASUPRA ROCILOR CRISTALOFILIENE, GRANITOIDE ŞI FILONIENE DIN MUNŢII CODRU (MUNŢII APUSENI)

(Rezumat)

Granitoidele de Codru şi seria cristalofiliană de Botfei fac parte din unităţile chimerice de Armanu şi Dumbrăviţa (Bleahu et al., 1984).

Şisturile cristaline se dezvoltă cu precădere în partea de N şi NE a masivului granitoid. Ele apar de asemenea ca benzi, în majoritatea cazurilor

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cu grosimi metrice. Seria de Botfei are un caracter sericito-cloritos (cuarţite uneori feldspatice, sericitizate, şisturi cuarţitice micacee şi micaşisturi sericito-cloritoase sau şisturi clorito-sericitoase).

La contactul dintre şisturile cristaline şi granitoide apar minerale caracteristice de neoformaţie : frecvent granatul, biotitul şi muscovitul, foarte rar andaluzitul şi cu totul excepţional sillimanitul. Mineralele de contact termic sînt mai proaspete şi dispuse deseori discordant în raport cu mineralele mai vechi care formează matricea sericito-cloritoasă.

Seria de Botfei este acoperită discordant de depozite sedimentare permo-carbonifere alcătuite din conglomerate laminate, gresii vermiculare, tufuri şi gresii feldspatice. La nivelul gresiilor feldspatice se află intercalate curgeri de riolite ignimbritice.

Granitoidele de Codru sînt reprezentate în principal prin granite (monzogranite) şi granodiorite, subordonat prin diorite cuarţifere şi tonalite. Unele granitoide au paiete de muscovit larg dezvoltate ce pot atinge 1 cm lungime. Trecerea către leucogranite se face prin dispariţia treptată a mineralelor femice. Granitoidele de pe Valea Mare (Mărăuş) sînt autometamorfozate (caolinizate, sericitizate).

Granitoidele de Codru străbat depozitele sedimentare permiene la nivelul gresiilor (feldspatice) vermiculare. La rîndul lor, granitoidele sînt străbătute de filoane riolitice şi granodiorit-porfirice care, după Dimitrescu et al. (1965), materializează canalele de alimentare ale vulcanitelor riolit-ignimbritice permiene. Aceste efuziuni, care îşi fac apariţia la nivelul gresiilor feldspatice permiene (Bleahu, 1963), se aşază uneori nemijlocit peste rocile granitoide. Granitoidele şi vulcanitele riolitice formează aşadar un complex magmatic intrusiv-efuziv de vîrstă permiană, alimentat din aceeaşi sursă magmatică.

În cadrul rocilor granitoide se află cuprinse două lentile de ortoamfibolite ce pot fi paralelizate cu ortoamfibolitele precambrian-cambriene menţionate de Balintoni în bazinul văii Iara (Hîrtopanu et al., 1982). Seria de Botfei, metamorfozată în faciesul de şisturi verzi şi retromorfozată probabil în timpul Mezozoicului datorită alunecării pînzei de Finiş, este mai nouă, de vîrstă posibil cambriană.

Granitoidul de Codru, post-cinematic în raport cu seria de Botfei de vîrstă hercinică, poate fi paralelizat cu granitoidul de Highiş, aşa cum acesta a fost interpretat de Balintoni (1986).

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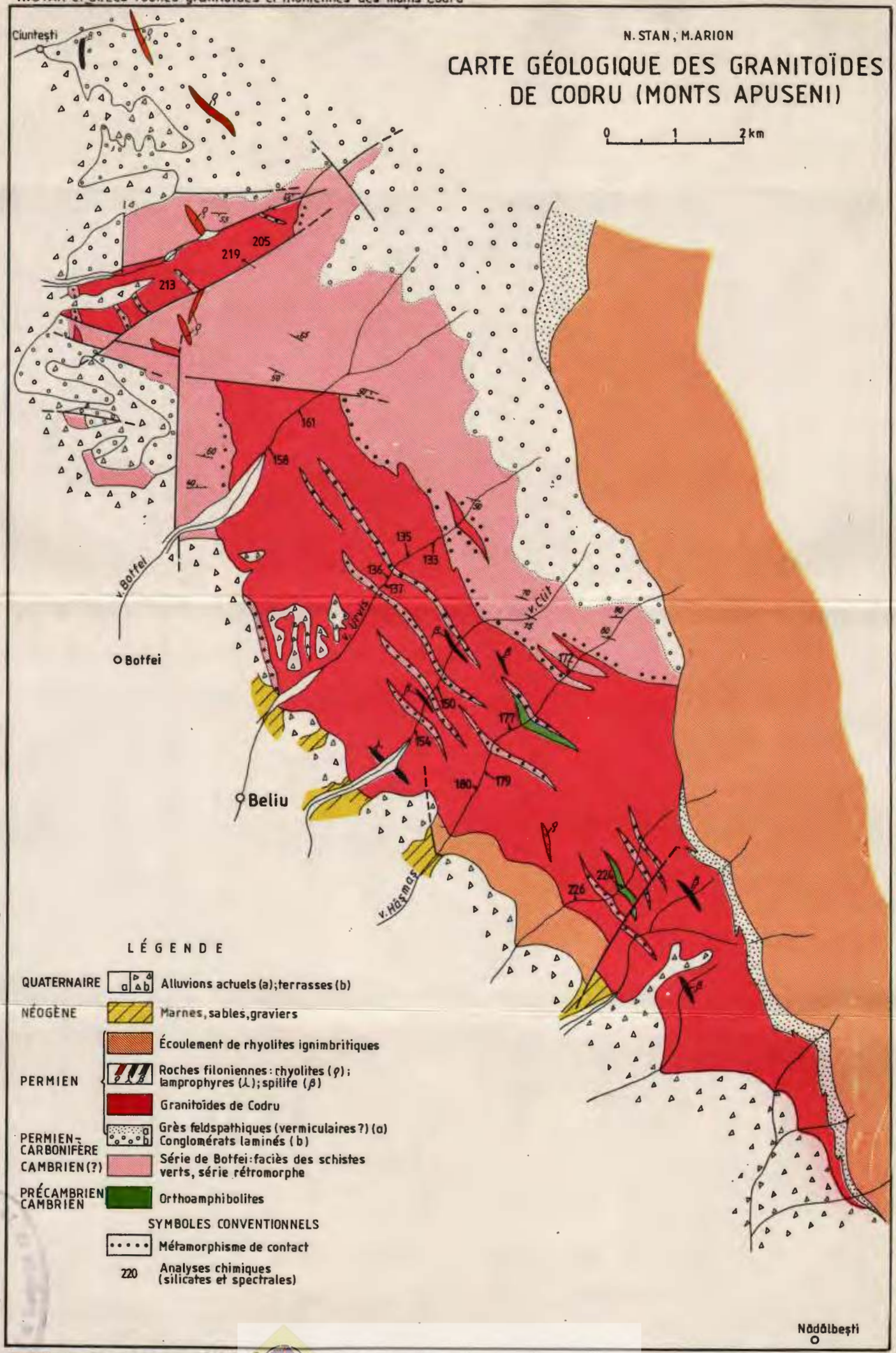
N. STAN, M. ARION

CARTE GÉOLOGIQUE DES GRANITOÏDES DE CODRU (MONTS APUSENI)

0 1 2 km



LÉGENDE

QUATERNAIRE	a b	Alluvions actuels (a); terrasses (b)
NÉOGÈNE		Marnes, sables, graviers
PERMIEN		Écoulement de rhyolites ignimbritiques
		Roches filoniennes: rhyolites (ρ); lamprophyres (λ); spilite (β)
		Granitoïdes de Codru
PERMIEN-CARBONIFÈRE	a b	Grès feldspathiques (vermiculaires?) (a) Conglomérats laminés (b)
CAMBRIEN (?)		Série de Botfei: faciès des schistes verts, série rétromorphe
PRÉCAMBRIEN CAMBRIEN		Orthoamphibolites

SYMBOLES CONVENTIONNELS

- Métamorphisme de contact
- 220 Analyses chimiques (silicates et spectrales)

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POLYMETAMORPHIC BLASTOMYLONITES[1]

by

MIHAI TATU[2]

Blastomylonite. Granite. Prograde metamorphism. Retrograde metamorphism. Caledonian Orogeny. Mobile belts. Polyphase processes. Paragenesis. Recrystallization. South Carpathians — Crystalline Getic — Leaota Mountains.

Abstract

On the basis of the mineralogical-textural analysis four metamorphic events were determined in the evolution of the blastomylonites delimiting the pre-Alpine plane on which the emplacement of the Albești granite took place (M_1 — pre-Grenvillian, M_2 — Caledonian, M_3 — post-Caledonian, and M_4 — Hercynian). Taking into account the outcropping area of the granite and blastomylonite plane, the ampleness of the deformation and of the mineralogical transformations that occurred during M_2, we inferred that at the Caledonian level there was a mobile zone delimiting two lithotectonic entities : Iezer Group and Leaota Group.

Résumé

Blastomylonites polymétamorphiques. L'analyse minéralogique-texturale a mis en évidence quatre événements métamorphiques dans l'évolution des blastomylonites qui délimitent le plan pré-alpin dans lequel a eu lieu la mise en place du granite d'Albești (M_1 — Grenvillien, M_2 — Calédonien, M_3 — post-Calédonien, M_4 — Hercinien). Etant donné l'aire d'affleurement du plan à granites et blastomylonites, l'ampleur de la défformation et des transformations minéralogiques développées en M_2, on considère qu'au niveau du Calédonien a fonctionné une zone mobile qui a séparé deux entités lithotectoniques : le groupe de Iezer et le groupe de Leaota.

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The presence of mylonites and blastomylonites in the polymetamorphic Precambrian fields represents the dynamic image of the relationships between different lithological entities or between the structural units. The place assigned to these relationships depends on the thickness and extension along the strike of the intensely deformed formations, the amount of movement that generated them, the origin and age of the geological entities separated by them. In this sense, blastomylonites occupy a special position due to the blastesis-deformation synchronism, on the one hand, and their outcropping on large areas certifies the existence of mobile zones in the above-mentioned fields, on the other hand.

Considerations on blastomylonites and their significance are found in synthesis papers (Higgins, 1971 ; Antonioz, 1971 ; Bellière, 1971 ; Gibbson, 1977 ; Zwart, 1974 ; Den Tex, 1974) or in papers dealing with special situations from some outcropping zones (Vogel, 1967 ; Antonioz, 1968 ; Engels, 1972 ; Hârtopanu, 1982 ; Hârtopanu et al., 1986 ; Iancu in Năstăseanu et al., 1981 ; Iancu, 1983, 1985 ; Iancu, Hârtopanu, 1982 ; Hann, 1983 ; Mărunțiu, Seghedi, 1983 ; Gherasi et al., 1986 ; Tatu, Săbău, 1987 ; Tatu, Robu, 1987).

In 1974 Den Tex resumed the discussion on the mylonitized mylonites, a term introduced by Christie (1960) and Johnson (1960). Bellière (1971) and Antonioz (1971) were also concerned with it and all of them agreed that the remylonitized mylonites show proof of the polycyclic evolution for the fields where they are identified. The analysis of the recrystallized blastomylonites pointed to the same conclusion. As both the remylonitized blastomylonites and those within which recrystallizations and/or blastesis of the second generation are observed, attest the polycyclicity of the yielding processes, we agree to define them as *polymetamorphic blastomylonites*.

For instance, we shall deal with blastomylonites marking the Albești granite, that crops out in the Leaota Mts on about 600 km^2. It occurs as a tabular body with a remarkable constancy of the chemism and of its "position in the lithostratigraphic column" (Gheuca, Dinică, 1983). Tatu and Săbău (1987) brought arguments in favour of the intrusive character and the magmatic evolution of the Albești granite ; they consider that its emplacement took place during the Cadomian along a tectonic plane attested by the presence of a discontinuous level of almandine-bearing blastomylonites. It is to be mentioned that these blastomylonites occur both in the zone adjacent to the granite and as enclaves within it ; all this made the above-mentioned authors to consider the Albești granite subsequent to the blastomylonite plane. The authors state that the old plane (Săndulescu, 1984) was reactivated during several stages and transposed after a penetrative foliation (S_2) of Caledonian age.

Mineralogical and Textural Aspects

Blastomylonites are rocks of a grey-brownish colour, with a mylonitic texture, made up of garnet, biotite, muscovite, potash and sodium feldspar, quartz, iron oxides and sericite. The microscopic study emphasized the following parageneses and mineralogical assemblages :

1. biotite$_1$ — muscovite$_1$ — garnet — potash feldspar — quartz$_1$ (syn S_1) ;

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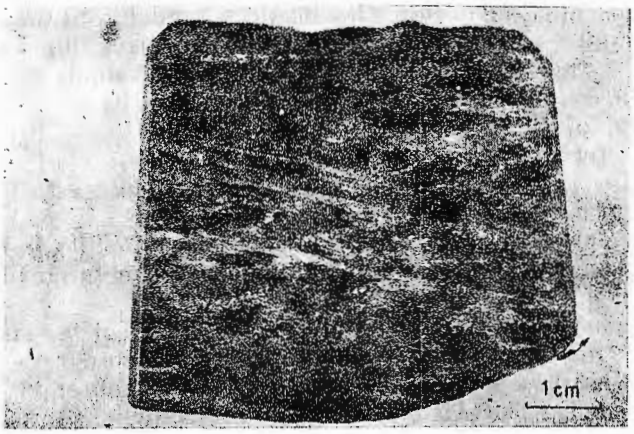
2. biotite$_2$ — albite — quartz$_2$ — iron oxides (syn S_2);
3. muscovite$_2$ (post S_2);
4. sericite-quartz$_3$ (syn S_3).

The first assemblage is strongly marked by the deformational moment D_2, so that its elements occur as relics. Garnet appears as poikiloblasts, with inclusions of biotite$_1$, muscovite$_1$ and quartz$_1$ in fusiform masses of iron oxides (Fig.). Potash feldspar (microcline) also occurs as



Blastomylonite, Podul Runcului, Leaota Mountains.
1, fusiform aggregates of garnet + biotite$_2$ + iron oxides;
2, quartz$_3$ and sericite.

poikiloblasts with inclusions of biotite$_1$, muscovite$_1$ and quartz$_1$, in a high schistose quartz-micaceous mass. In places, in the micaceous environment and lenthened on S_2 occurs a biotite with partly distroyed garnet rims, an aspect similar with that observed in the Albești granite (Reinhard, 1912; Gherasi et al., 1966; Gheuca, Dinică, 1983; Tatu, Săbău, 1987). It is of note that both garnets belong to moment M_1 (both of them being intensely affected by the deformational moment D_2) but the one which borders the biotite was formed later on as compared with the poikiloblasts. All this determined us to regard garnet and microcline as products of a prograde reaction specific to the almandine amphibolite facies (Miyashiro, 1973):

$$\underset{\text{muscovite}}{KAl_3SiO_3O_{10}(OH)_2} + \underset{\text{biotite}}{K(Mg,Fe)_3AlSi_3O_{10}(OH)_2} + 3SiO_2 = \underset{\text{almandine}}{(Mg,Fe)_3Al_2Si_3O} + \underset{\text{potash feldspar}}{2KAlSi_3O_8} + \underset{\text{water}}{2H_2O}$$

This time there is a convergence, the garnet rims of the biotite from granites being the result of a reaction between biotite, plagioclase, quartz and liquid probably during the ascent of the magma (Tatu, Săbău, 1987 cf. Hoschek, 1976).

The deformational event D_2 is accompanied by recrystallization and mineral neoformation yielding biotite$_2$ and iron oxides at the expense of

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garnet, albite by the microcline destabilization, $quartz_2$ by the recrystallization of $quartz_1$ with a flamboyant aspect. The rock displays an intensely schistous aspect due to the phylosilicate bands alternating with quartz-albitic ones including fusiform aggregates of oxides with garnet xenoclasts (Fig.). Potash feldspar porphyroclasts belonging to the first mineralogical assemblage are locally preserved in more micaceous zones. The garnet xenoclasts are corroded by a newly formed biotite.

After this episode occurs the blastesis under static conditions of $muscovite_2$ found as megablasts with a different trending as against S_2.

The next stage is marked by a new S, at a small angle versus S_2 (Fig.), and later on occurs the reorientation in the new stress field of $muscovite_2$ and of the aggregates of garnet + $biotite_2$ + iron oxides. D_3 event is characterized by the syndeformational blastesis of sericite at the expense of muscovite and albite as well as by the recrystallization of $quartz_2$ with a fusiform aspect.

As in case of other zones of the Getic Nappe (Porţile de Fier, Bahna, Godeanu—Iancu, 1983, 1985) the evolution of the metamorphites is marked by four events :

— M_1 is characterized by the presence of the $biotite_1$—$muscovite_1$—garnet—potash feldspar—$quartz_1$ assemblage with a prograde character at the almandine-amphibolite facies (Miyashiro, 1973) synchronous to the deformational moment D_1. The existence of the equilibrium phase $quartz_1$—$biotite_1$—$muscovite_1$ preserved in the garnet and $microcline_1$ porphyroblasts points to a PT field whose maximum reaches values ranging between 600 and 650°C and pressures of 6—8 Kb (Dobreţov et al., 1970). These data define a medium-pressure metamorphism (Barrovian) noticed within the Getic Crystalline (Savu, 1970; Bercia, 1975; Hârtopanu, 1982), probably pre-Grenvillian in age (Iancu, 1983, 1985). Towards the end of this cycle takes place the formation of deep-seated fracture zones in which the movement gradient shows higher values as compared with those of the regional metamorphism. These major dislocations had probably dips of 45—50° which made possible their resuming in the following deformational events.

— M_2 (Caledonian), during which the movement gradient shows high values, is characterized by the appearance of a paragenesis with a retrograde character as compared with the first moment and of a schistosity S_2, highly penetrative, after which the old structure was transposed and even obliterated. At the beginning of the event M_2 in the mentioned zone deformation involved and accompanied the blastesis of the $biotite_2$—$quartz_2$—albite—iron oxides paragenesis situated at the lower limit of the almandine amphibolite facies, biotite zone (500°—525°C and 3—5 Kb, Dobreţov et al., 1970), whereas in the rest of area metamorphism also displayed a retrograde character but at the level of the chlorite zone. Towards the end of the Cadomian the blastomylonite plane entered into a distension regime which allowed the emplacement of the Albeşti granite. Further on, the plane and the granite were resumed during several stages of the following deformational events.

— M_3 is marked, at the blastomylonite level, by the $muscovite_2$ blastesis under static regime, post—S_2;

— M_4 is characterized by the syndeformational occurrence of a retromorphic paragenesis (sericite-$quartz_3$). The deformational moment D_3

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generated a new schistosity S_3, oblique versus S_2 (Fig.), followed by the reorientation of porphyroclasts in the new stress field. This moment is referred, on the basis of the K/Ar data (Pavelescu et al., 1975) to the Hercynian orogenesis. It is widely spread in metamorphites of the Leaota Group, especially in the Călușu Formation.

On this plane, attested by the presence of polymetamorphic blastomylonites, that conditioned tectonically the emplacement of the Albești Granite (Tatu, Săbău, 1987), was released most of the energy during the Caledonian, fact also proved by the ampleness of the mineralogical-textural transformations. The above-mentioned aspects and its extension at the Caledonian level represented a mobile zone that separated (acc. to Balintoni in Gheuca et al., 1987) two lithotectonic entities : Iezer Group, at the lower part (Balintoni, oral communication), and Leaota Group, at the upper part.

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BLASTOMILONITE POLIMETAMORFE

(Rezumat)

Blastomilonitele, ce apar în zona adiacentă granitului de Albeşti şi ca enclave în masa acestuia, atestă existenţa unui plan prealpin pe care s-a realizat punerea în loc a granitului sus-menţionat. Analiza mineralogo-texturală a permis decelarea a patru evenimente metamorfice în evoluţia blastomilonitelor (M_1, prograd, pre-Grenvillian, sin—S_1; M_2, retrograd, Caledonian, sin—S_2; M_3, static, post-Caledonian; M_4, retrograd, Hercinic, sin—S_3). Dată fiind aria de aflorare a planului cu granite şi blastomilonite, amploarea deformării şi a transformărilor mineralogice desfăşurate în M_2, considerăm că la nivelul Caledonianului, în acest areal a funcţionat o zonă mobilă ce a separat două entităţi litotectonice: grupul Iezer şi grupul Leaota.

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BOOK REVIEW

J. V. SMITH, W. L. BROWN : *"Feldspar Minerals", vol. 1, Crystal Structures, Physical, Chemical and Microtextural Properties.* Second edition. Springer-Verlag, 1988, 639 p.

Volume 1 and 2 of "Feldspar Minerals", edited by J. V. Smith, were published in 1974 and volume 3 was to be published later on. Due to the enormous advances in knowledge of feldspars, a new version of "Feldspar Minerals", revised and extended, was needed.

Volume 1 (1988 edition) edited by J. V. Smith and W. L. Brown includes materials from both volumes of 1974 edition (Smith) but reorganized as regards the structure, new chapters being included and others revised because of the great advances obtained by TEM, spectrometry, minute determination of the trace elements, and by experimental studies.

The book is subdivided into four parts and twenty chapters, a rich list of references split for ease of consultation into two parts (pre-1973 and post-1973), name index, geographical index and subject index.

Part 1 — Crystal Structures — (113 p.) comprises 5 chapters dealing with general aspects concerning the internal structure, nomenclature, order-disorder, geometry of crystalline framework and complex crystal structures.

Part 2 — Physical Properties and Experimental Techniques (178 p.) — consists of 7 chapters dealing with the knowledge of the cell geometry and atomic structure of feldspars using X-ray and neutron diffraction techniques as well as feldspars determination by common microscopic techniques or by transmission electron microscopy — IR, NMR, NQR, EPR, NGR (Mössbauer), XAS, EXAFS, etc. A special chapter is dedicated to the physical properties of feldspars.

Part 3 — Chemical Properties and Experimental Techniques (95 p.) — consists of two chapters ; the former refers to the specific techniques used for the study of the feldspars and the latter describes in detail their chemical properties.

Part 4 — Diffusion, Growth, Twins and Intergrowths (253 p.) — is the largest one including 6 chapters dealing with : diffusion phenomena in feldspars and measurement methods of them ; growth of crystals which implies nucleation and transport and subsequent atoms array ; twins ; intimate intergrowths of two or several feldspars (perthites, peristerites, Bøggild intergrowths) ; exolution and miscibility problems. Intergrowths of feldspars with other minerals — myrmekites, symplectites, spherulites, granophyric intergrowths, graphic granites, etc. — are also presented.

This revised and extended edition of "Feldspar Minerals" is of a great interest for specialists in this domain as it includes the latest results in feldspars study obtained by most advanced methods as well as a help in determining some of feldspars properties.

D. Russo-Săndulescu

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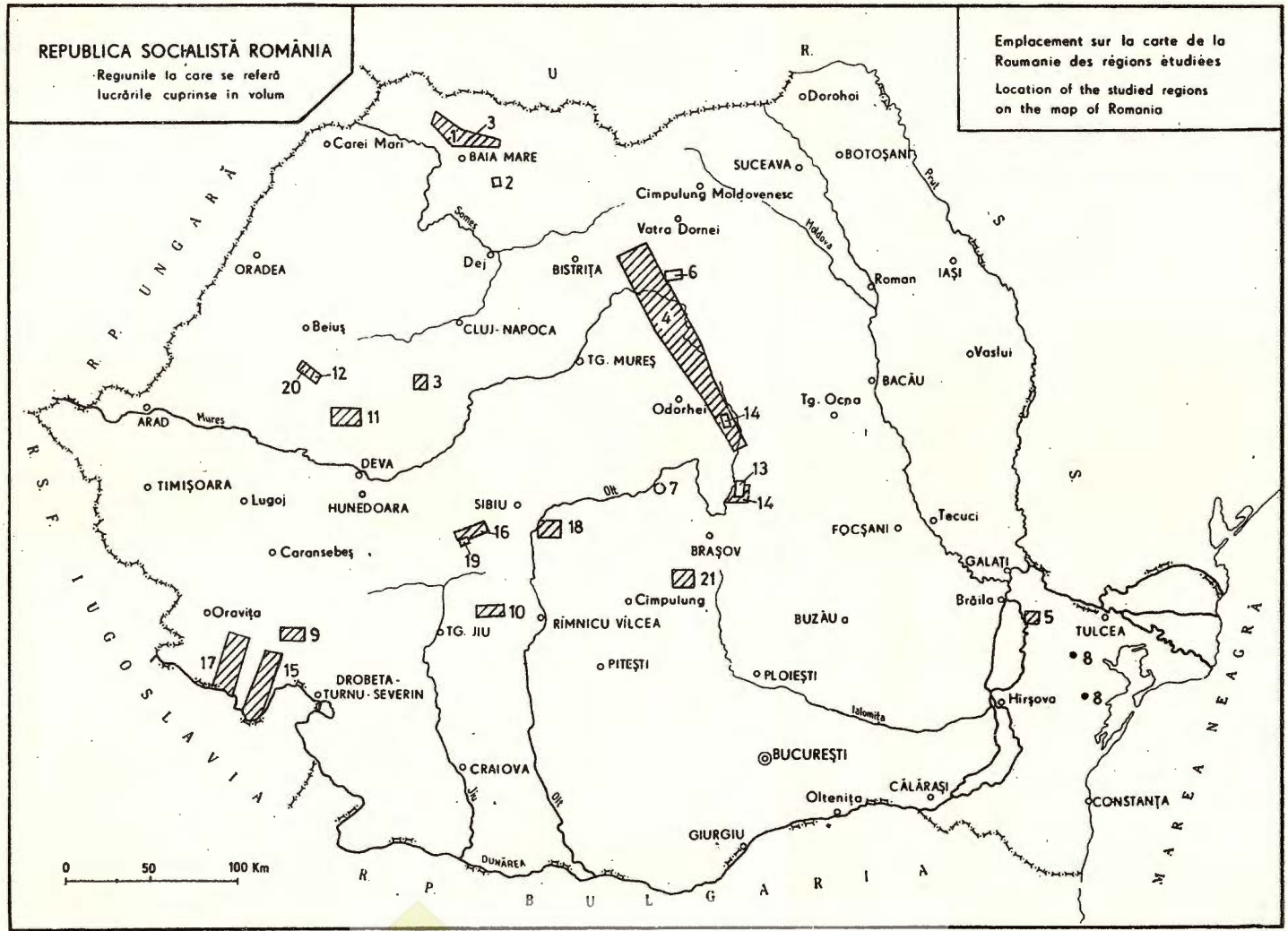

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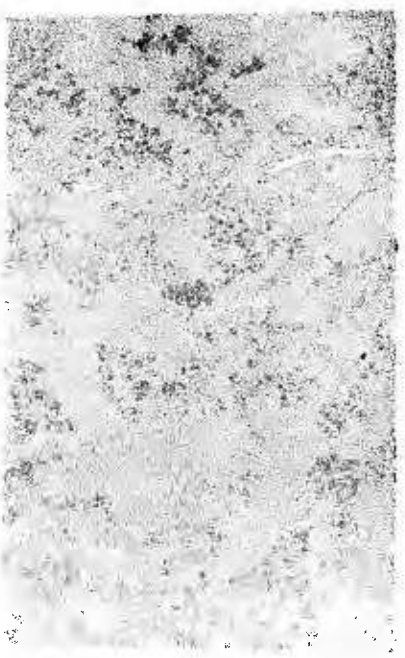

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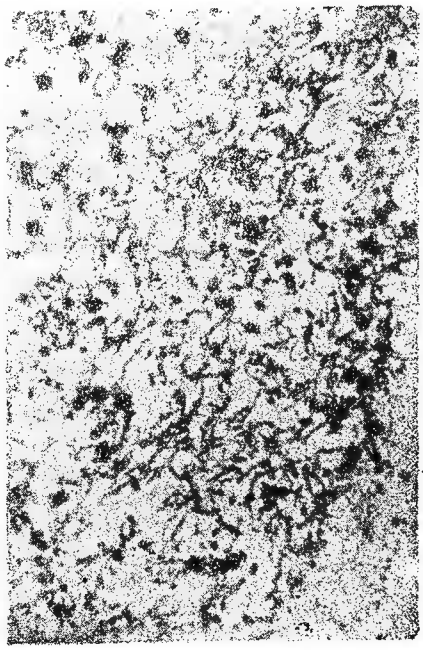

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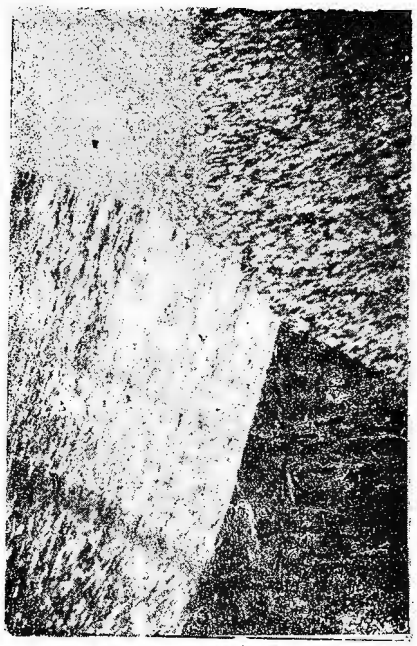

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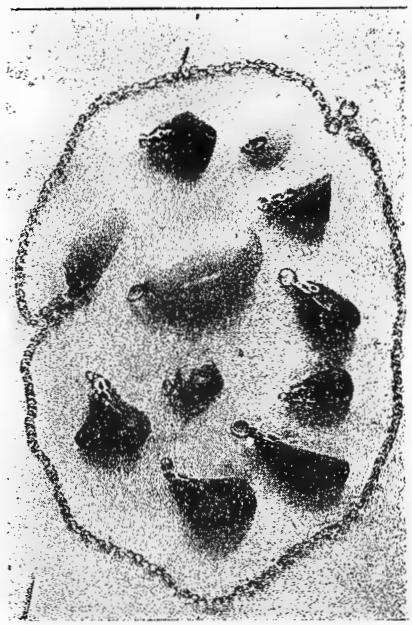

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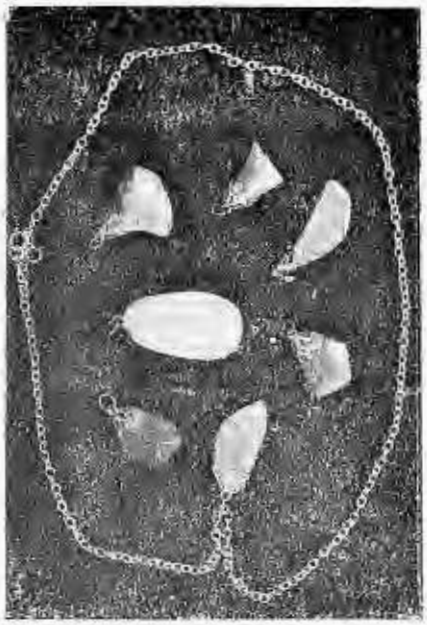

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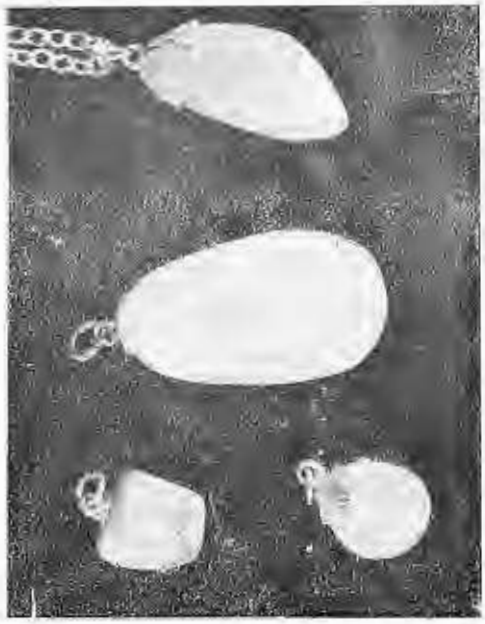

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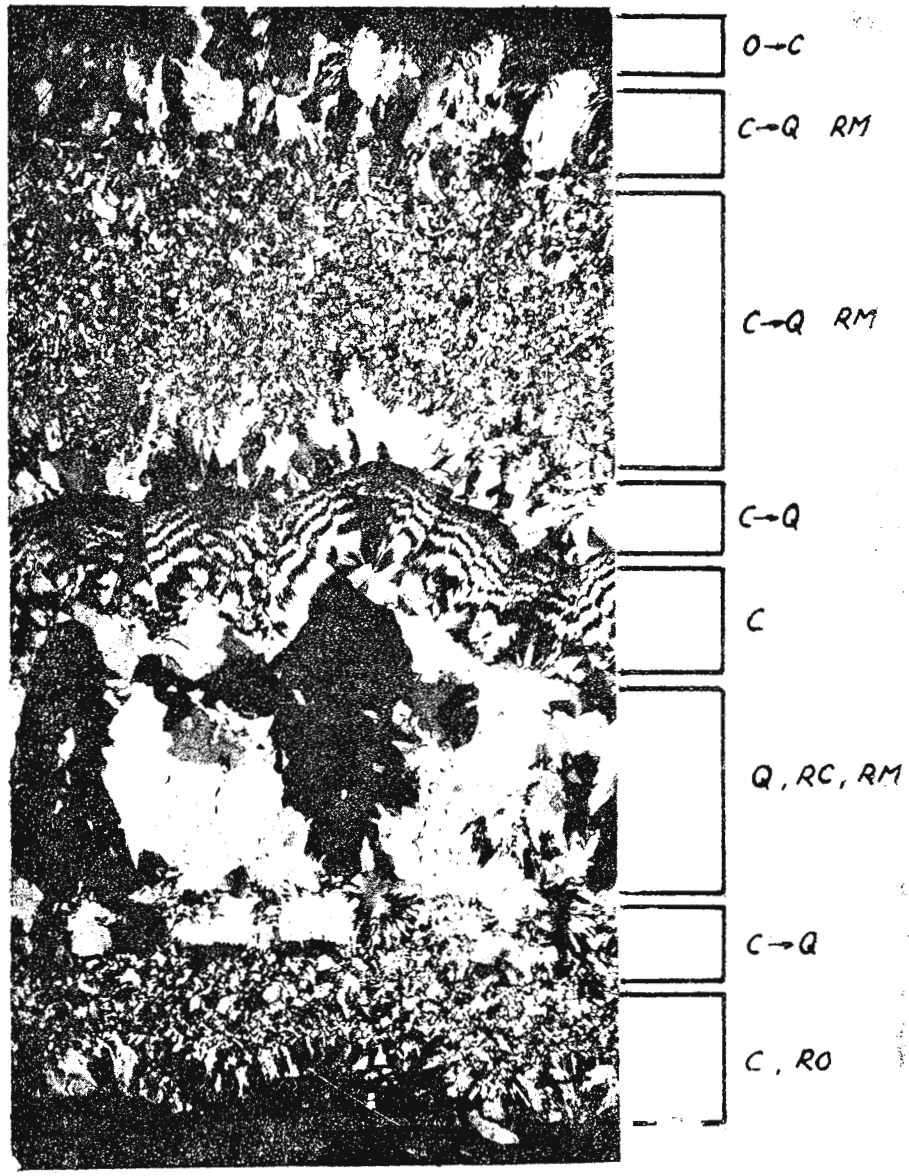

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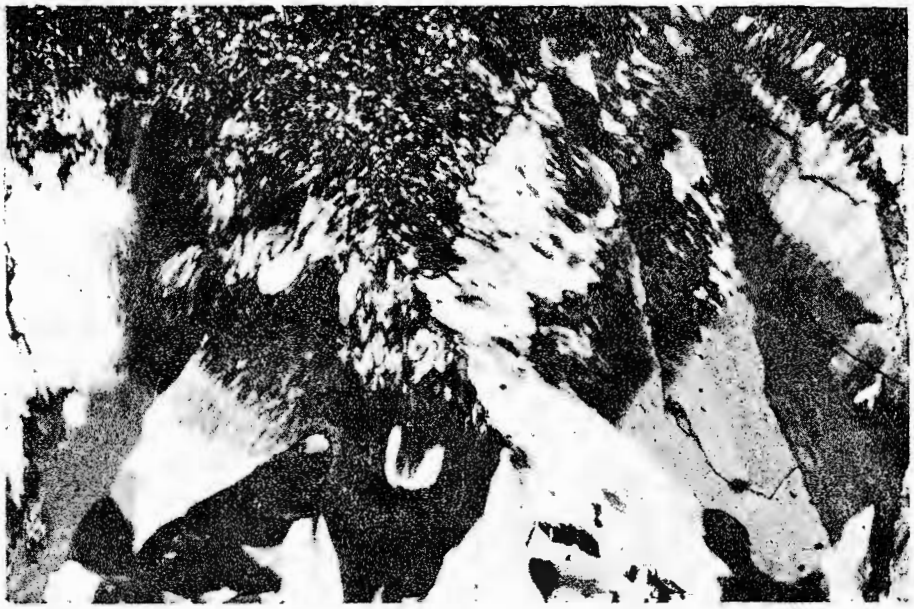

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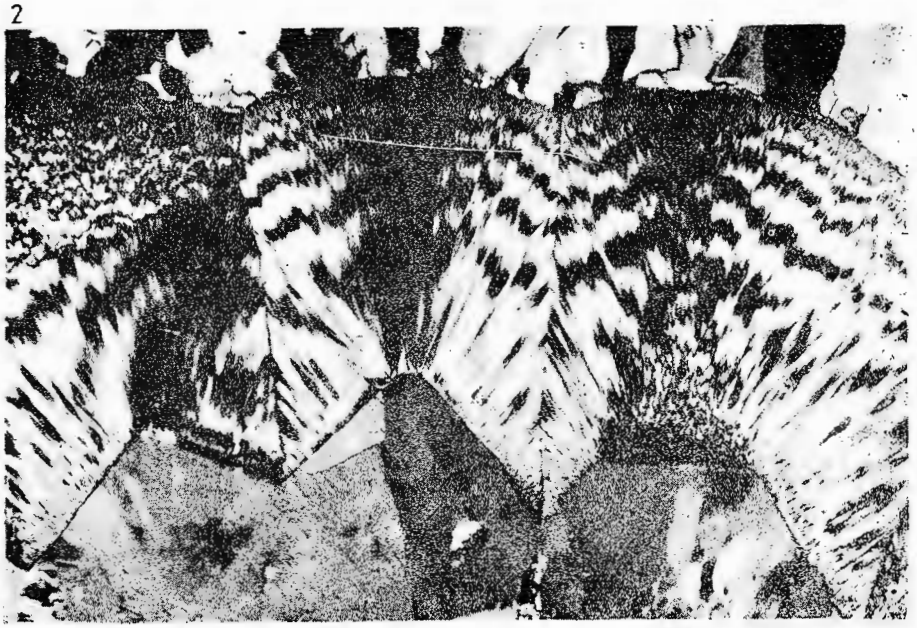

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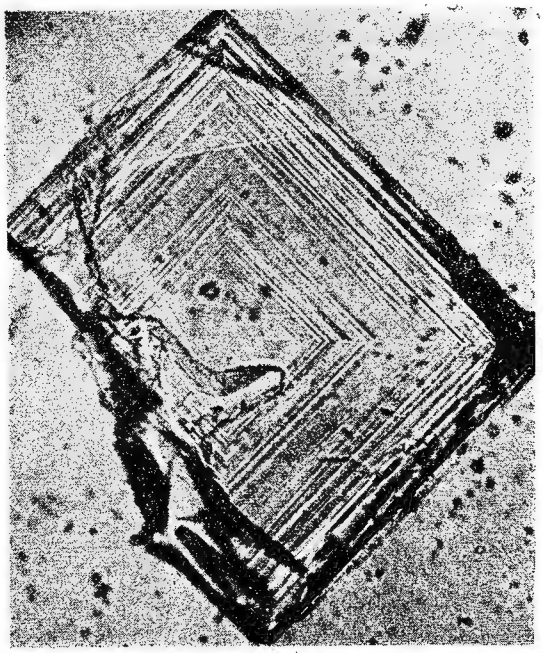

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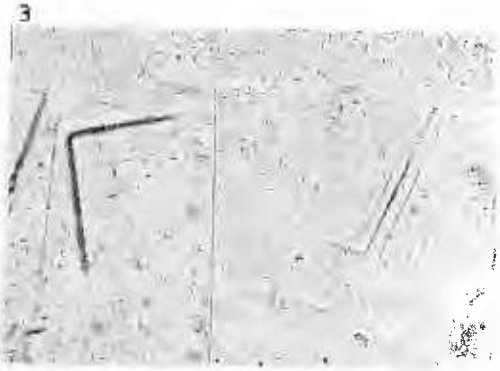

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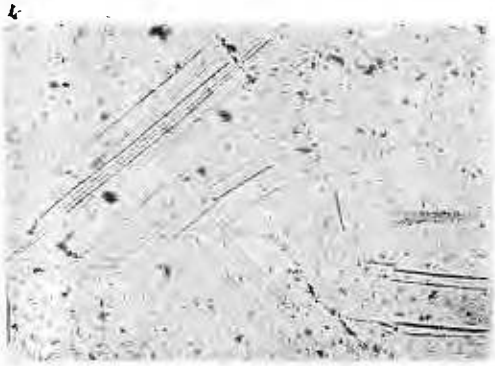

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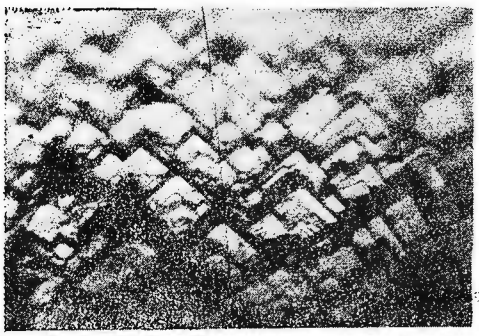

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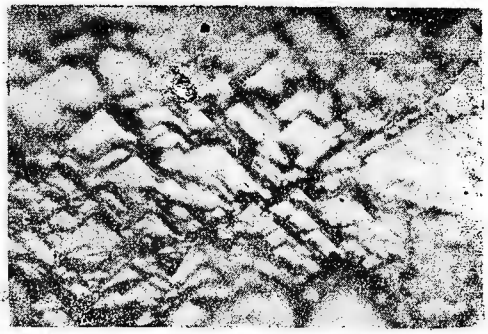

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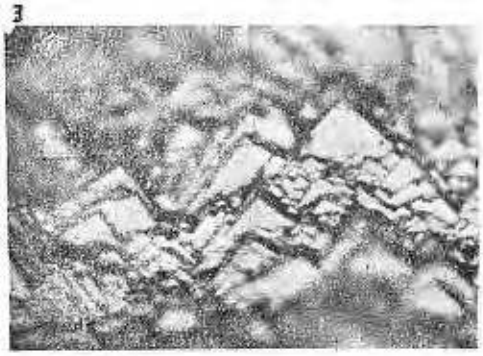

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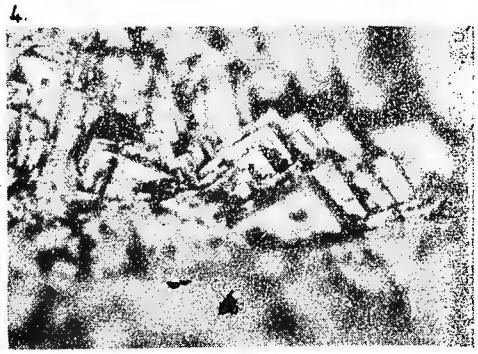

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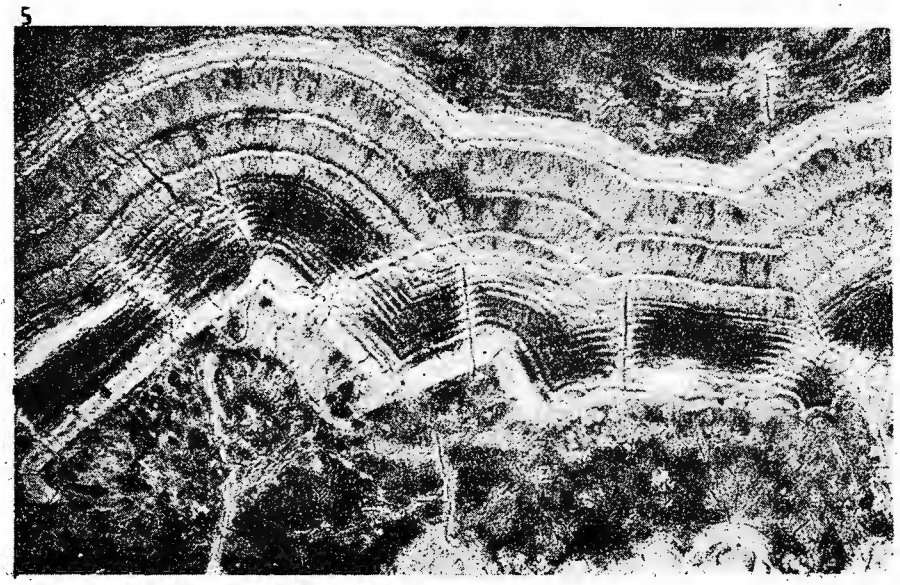

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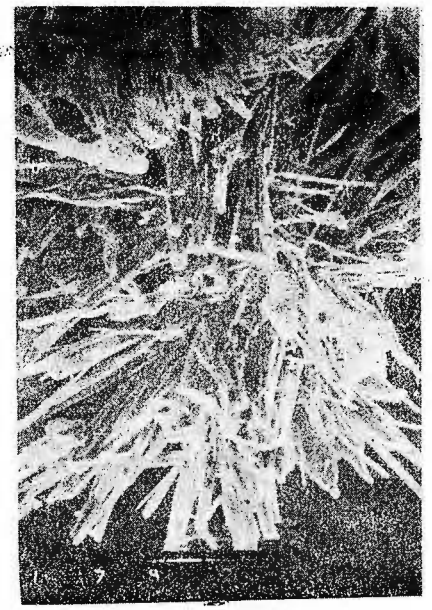

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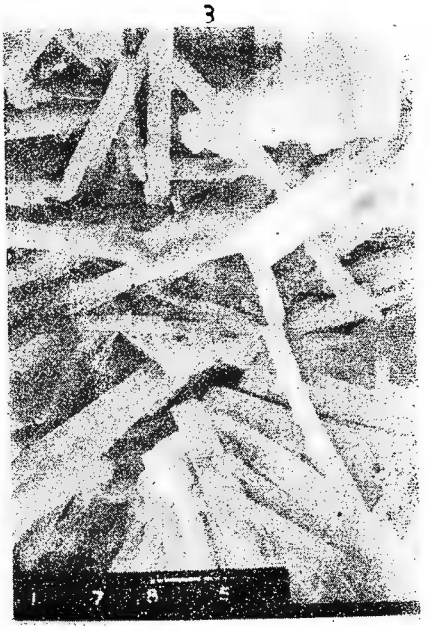

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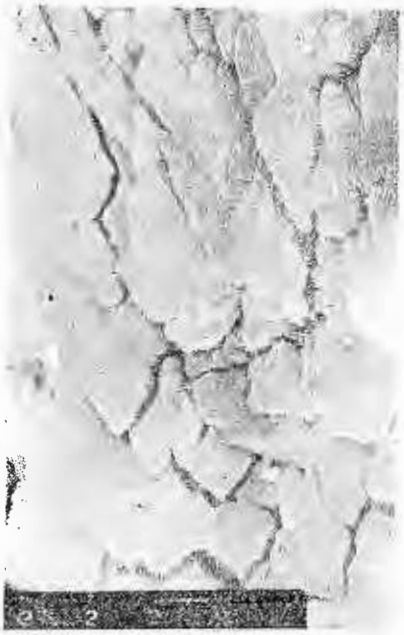

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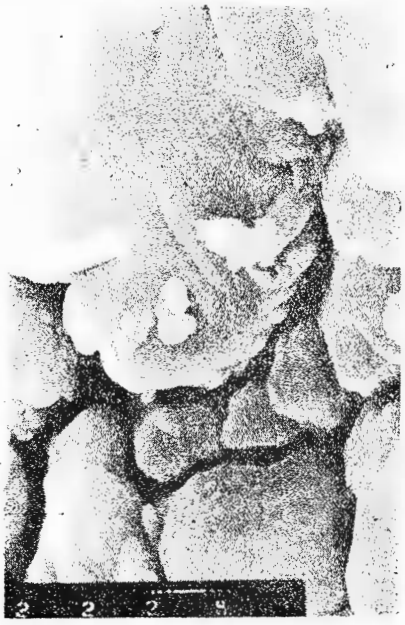

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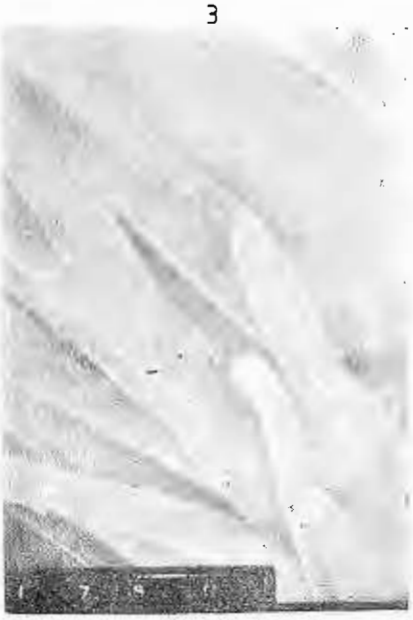

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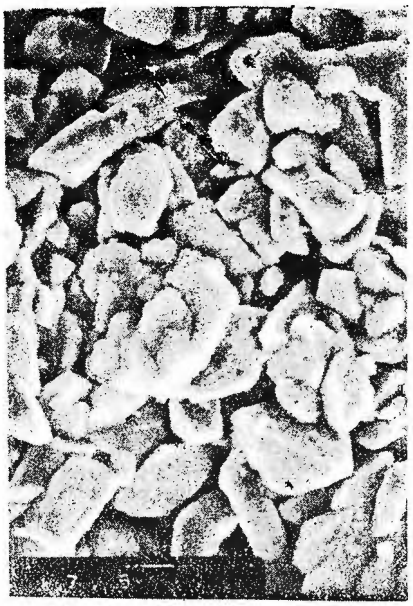

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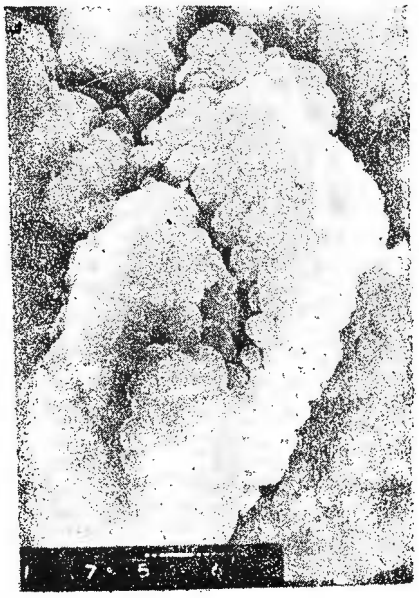

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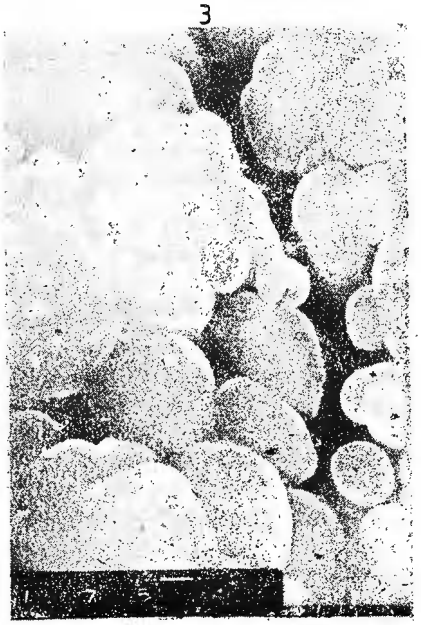

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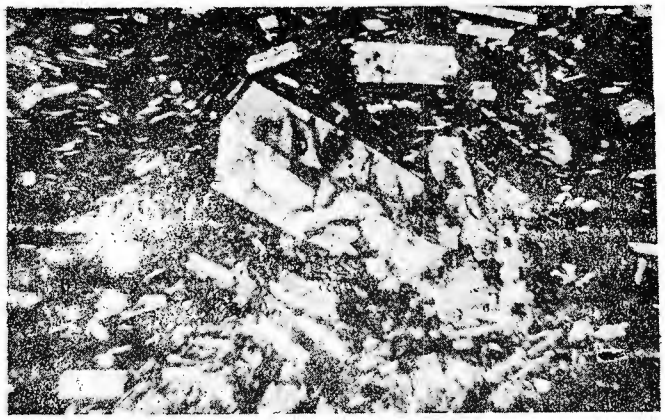

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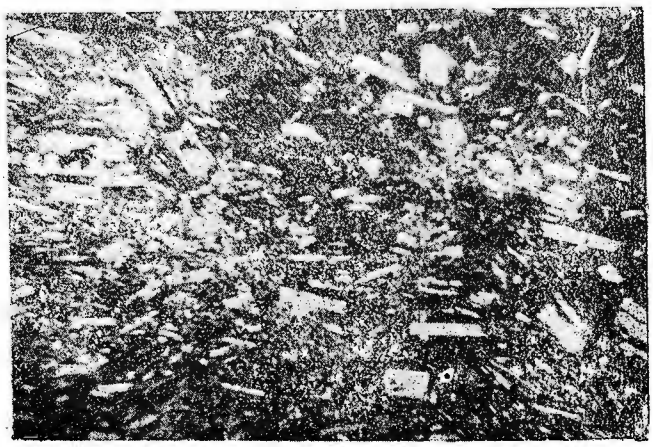

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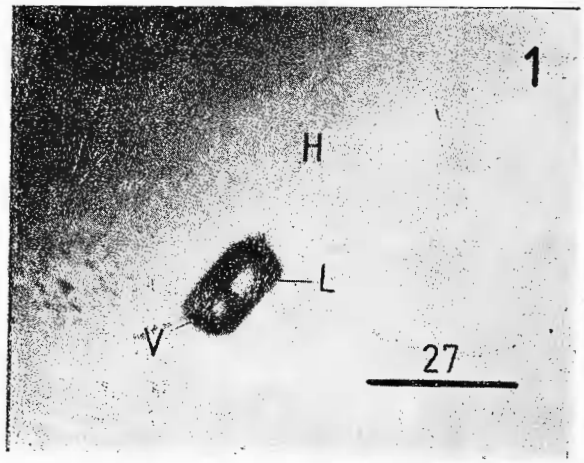



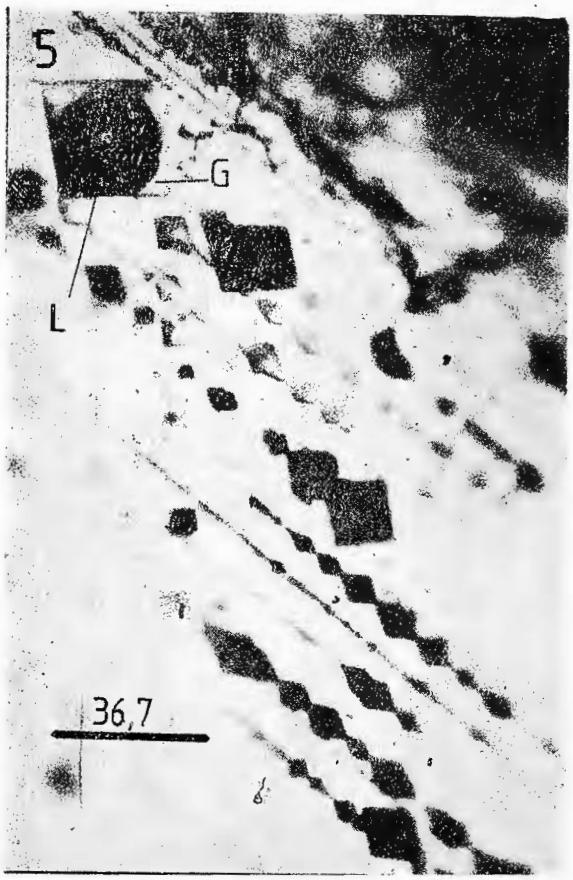

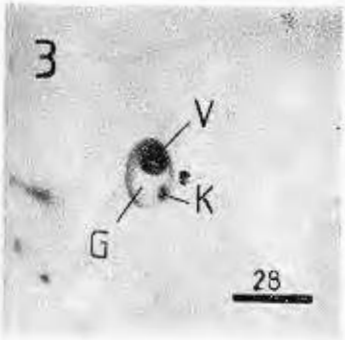

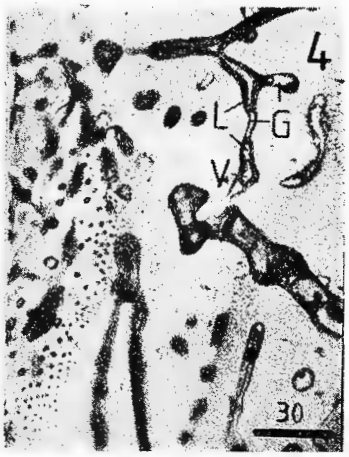

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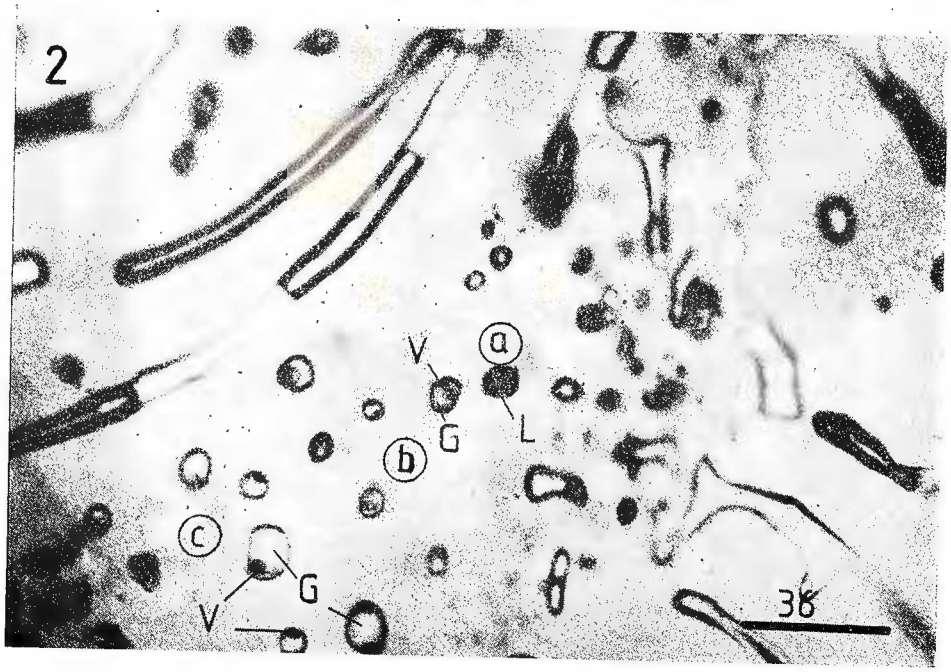

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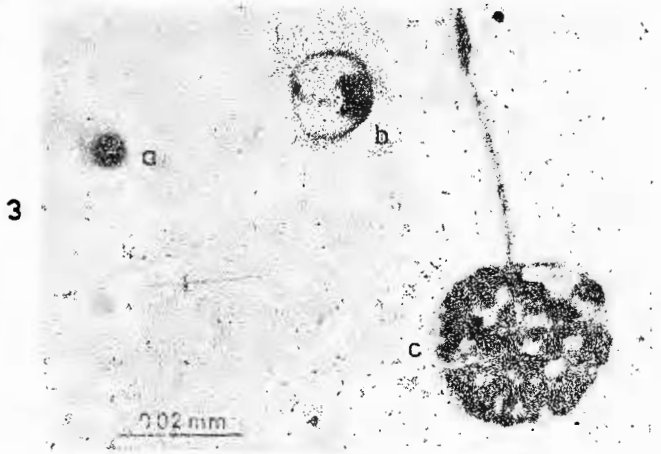

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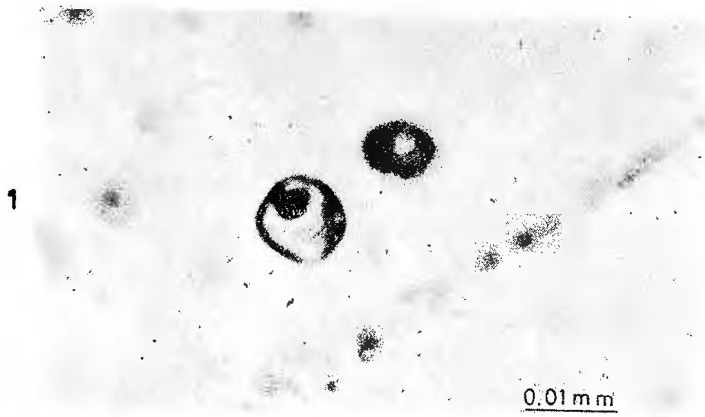



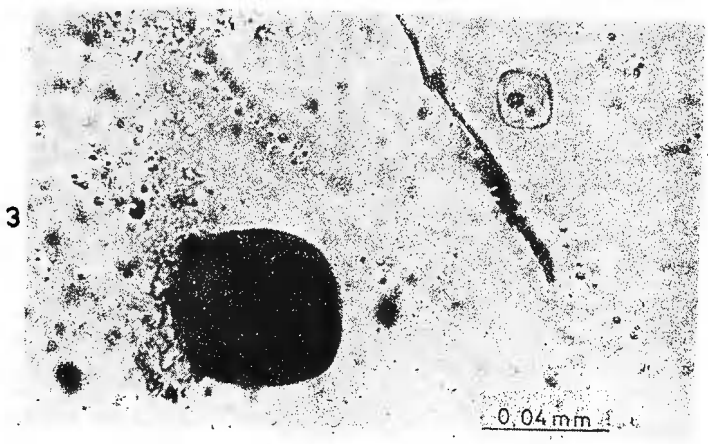

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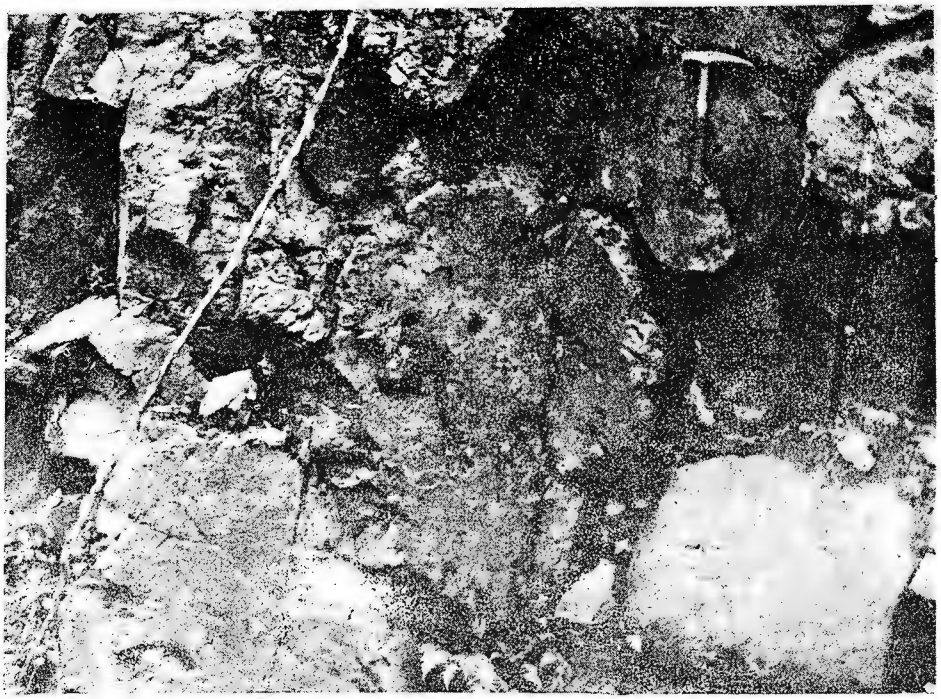

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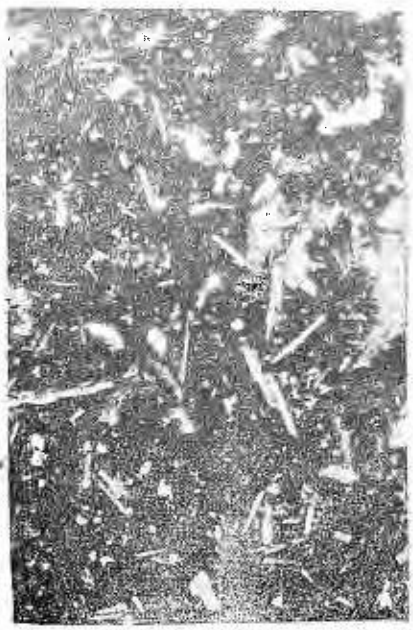

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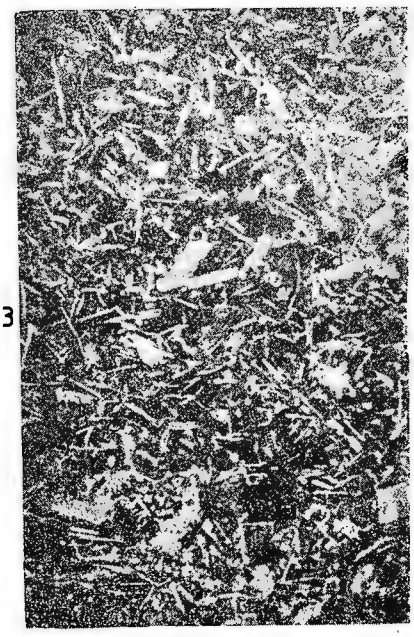

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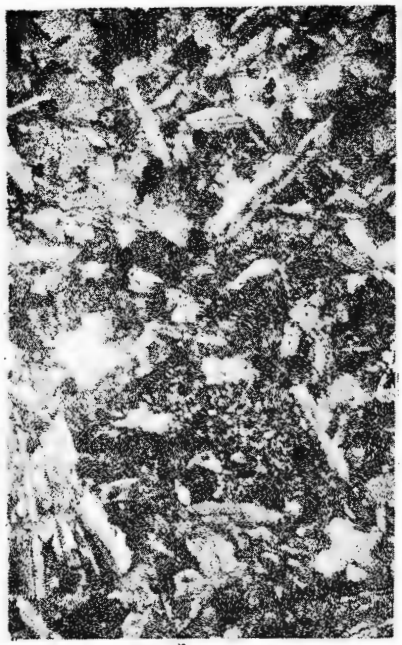

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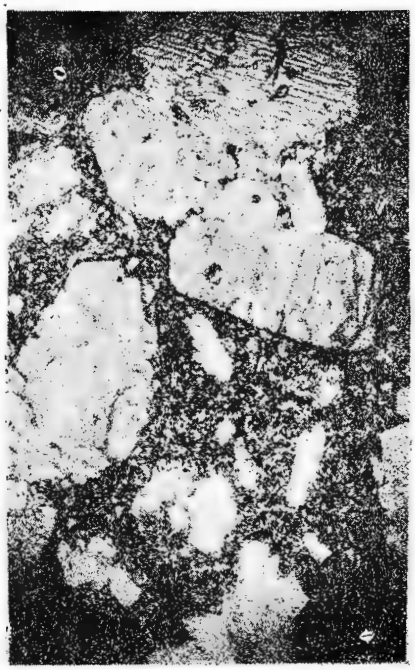

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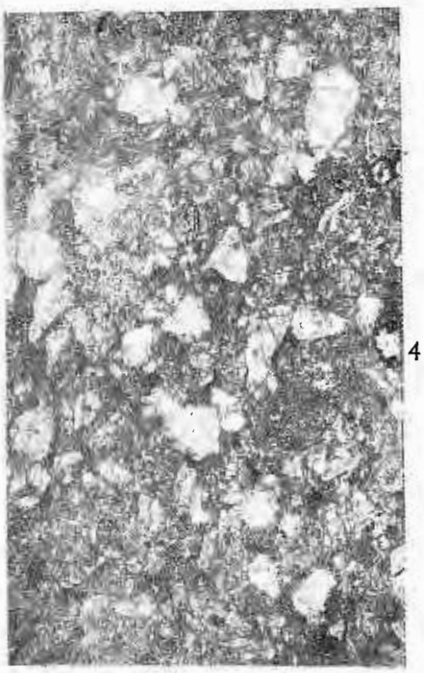

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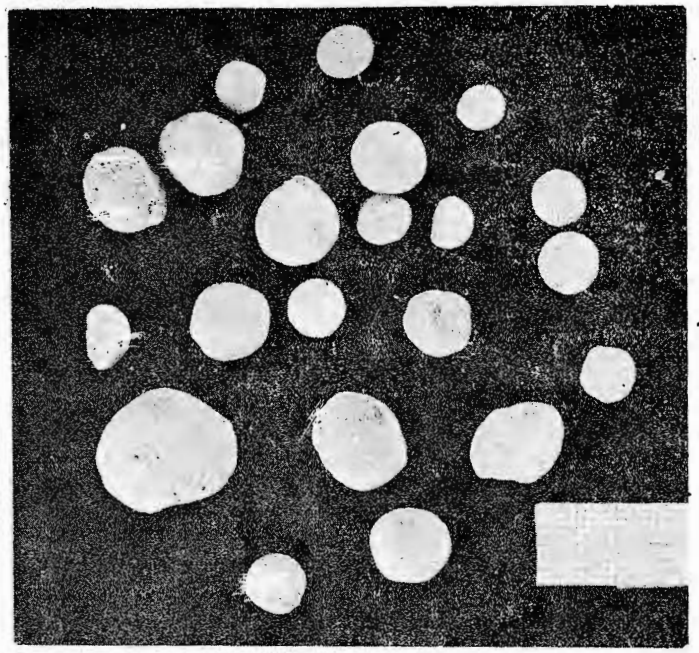

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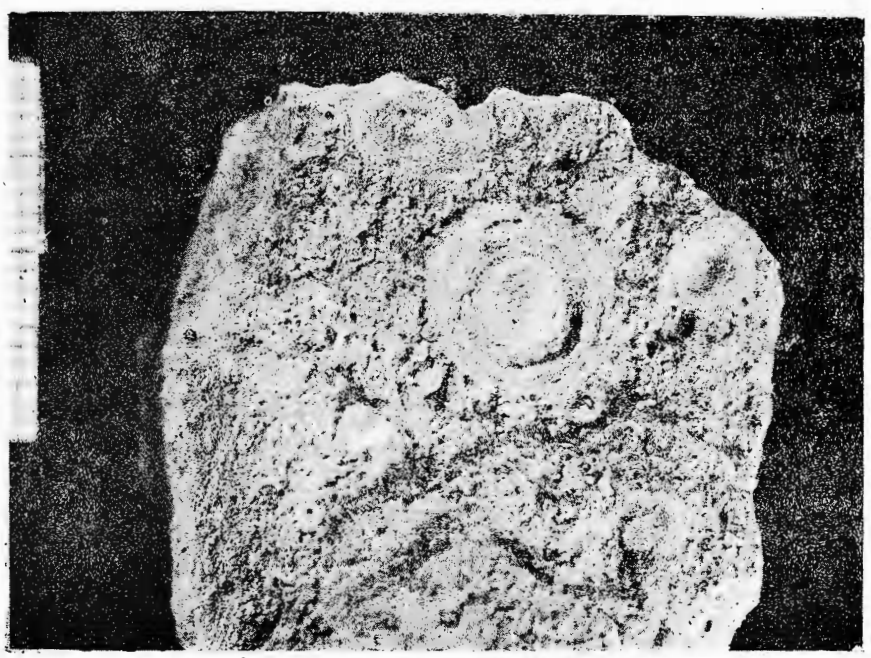

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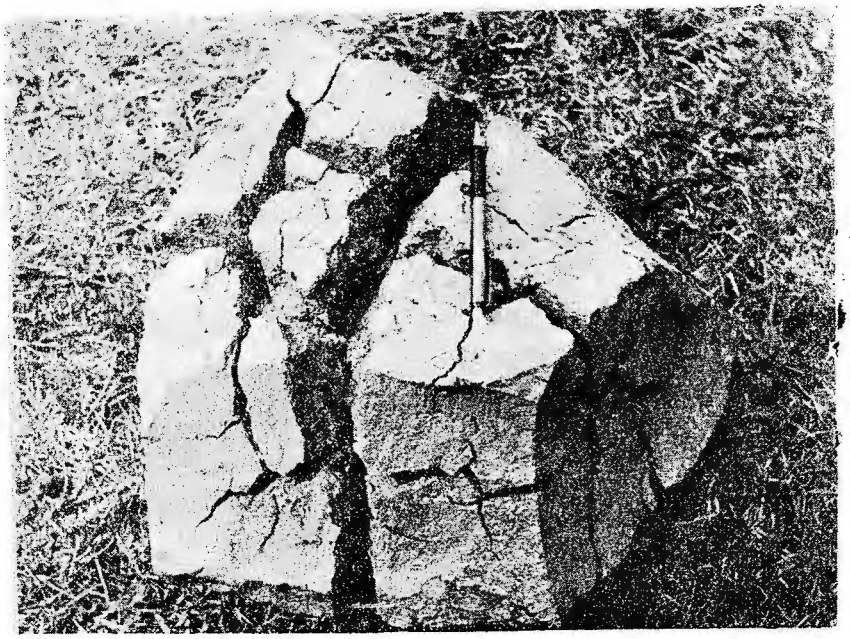

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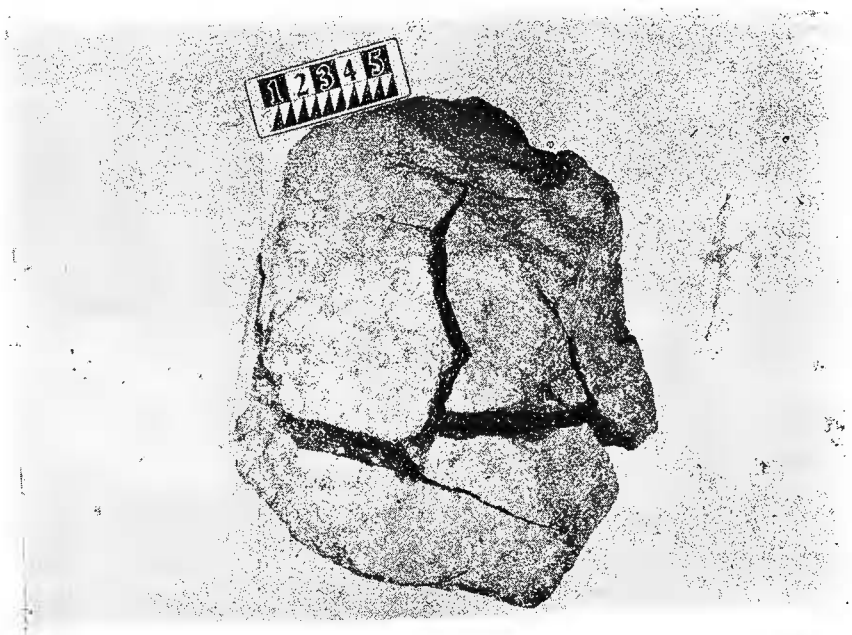

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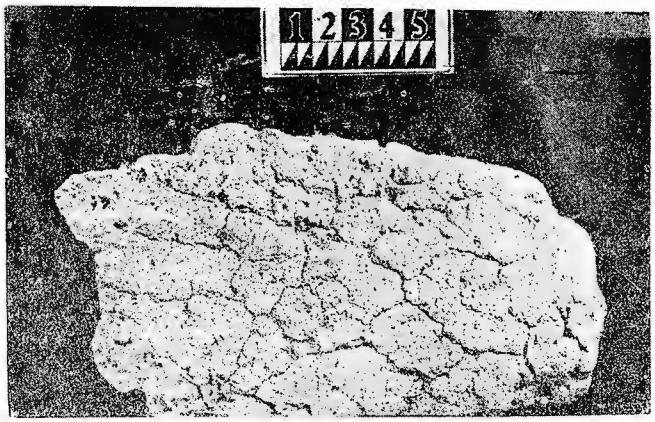

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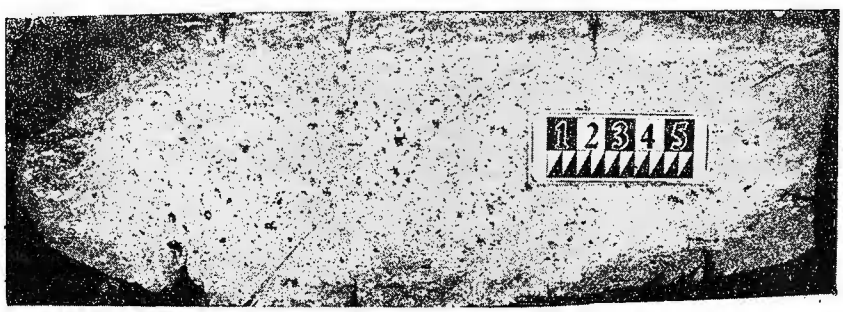

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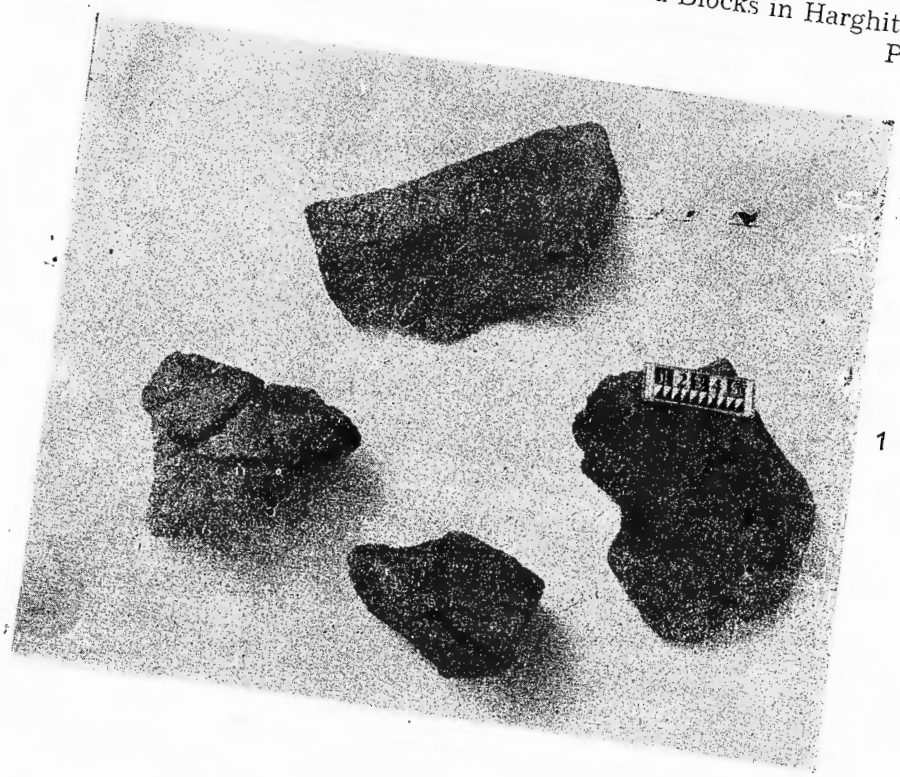

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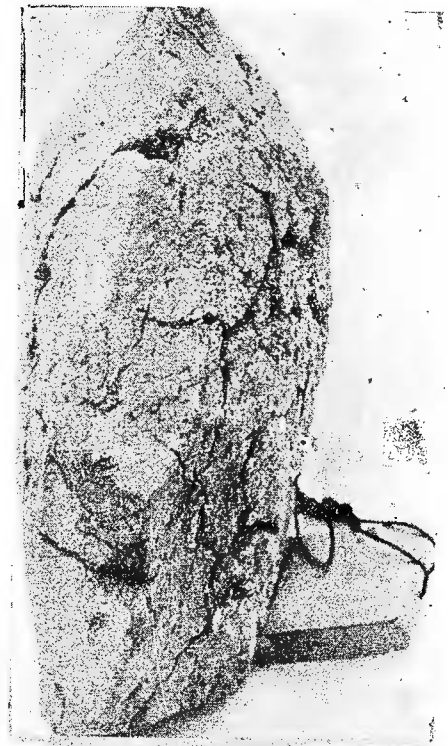

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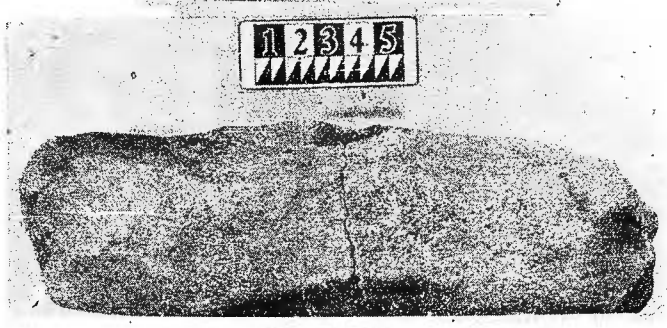

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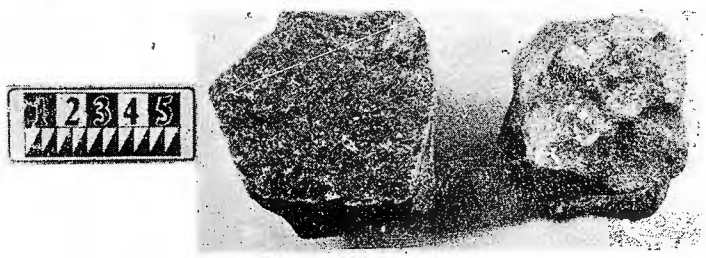

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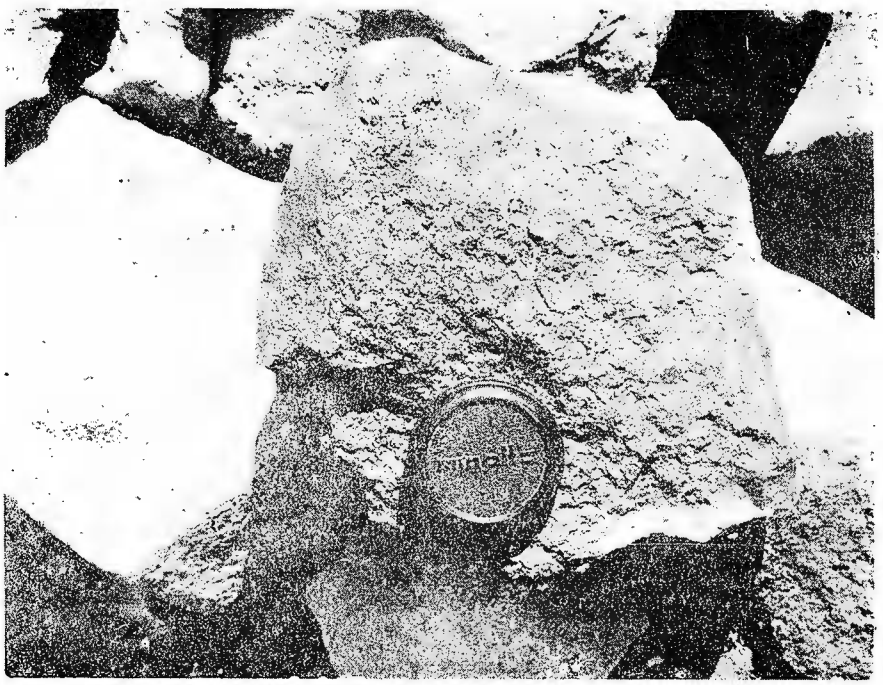

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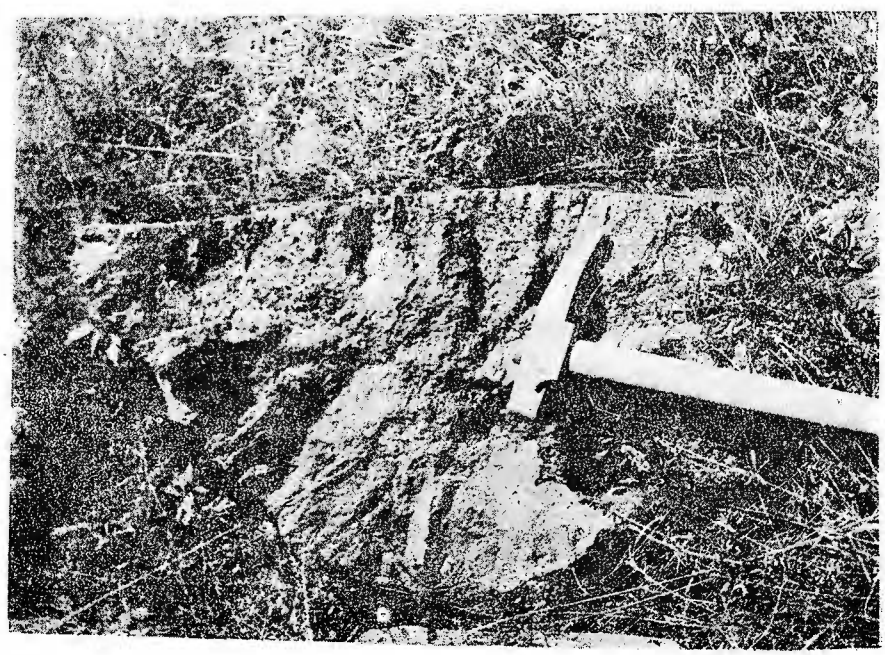

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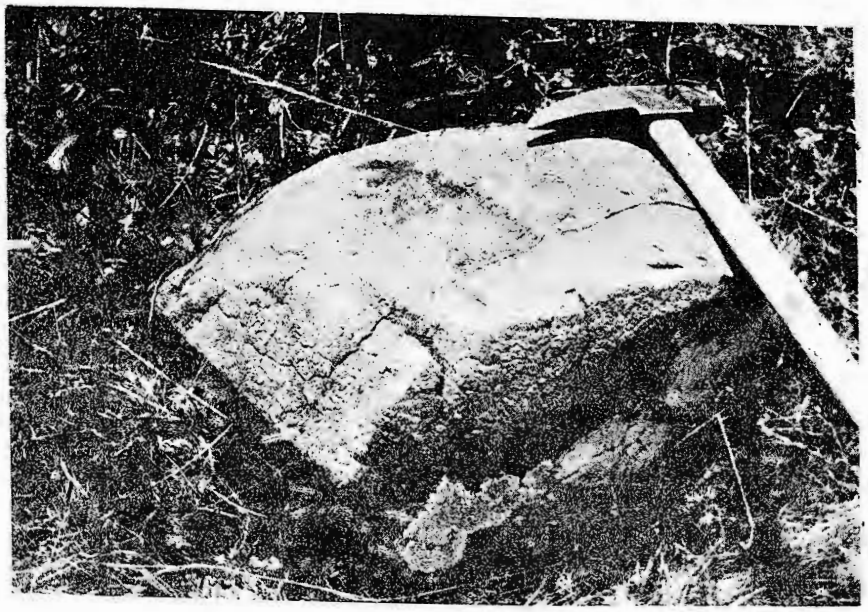

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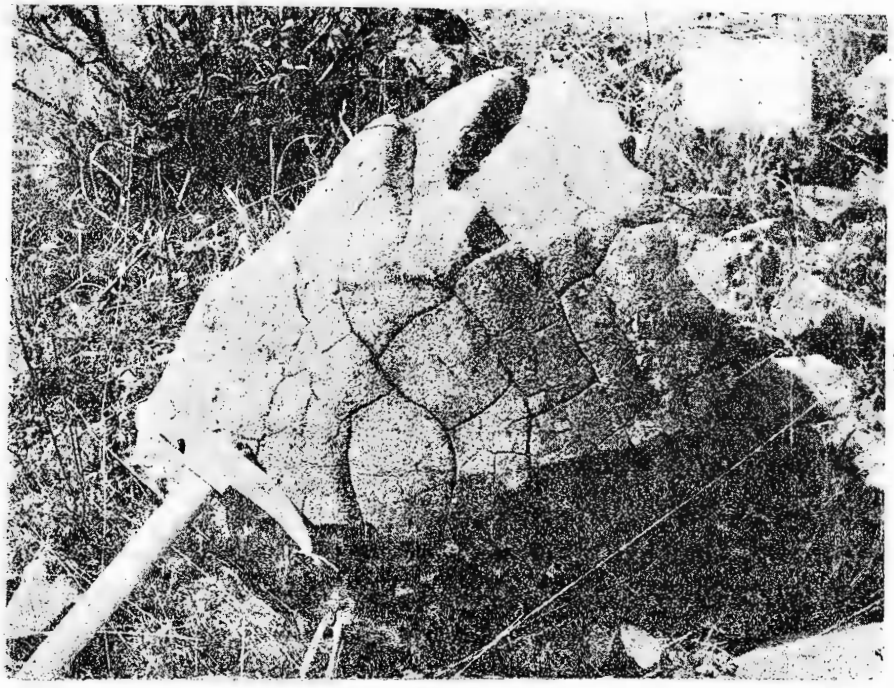

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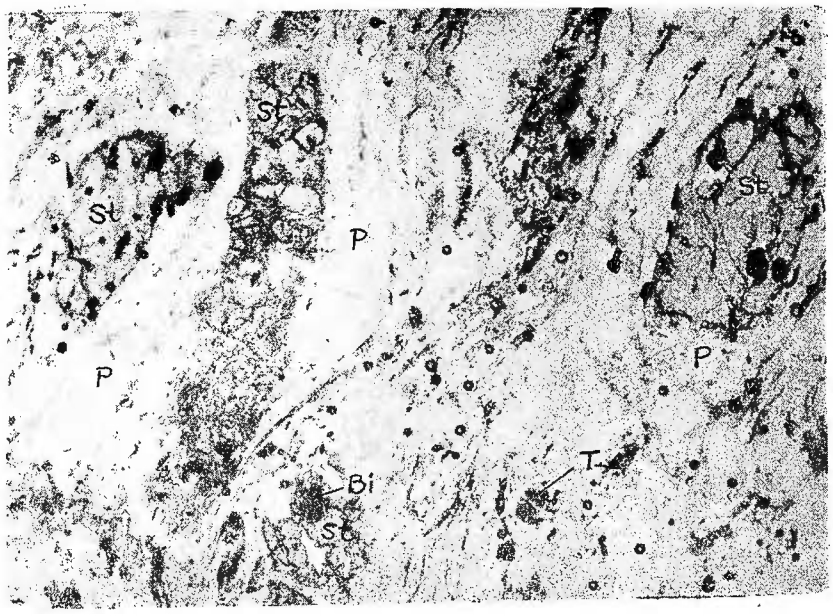

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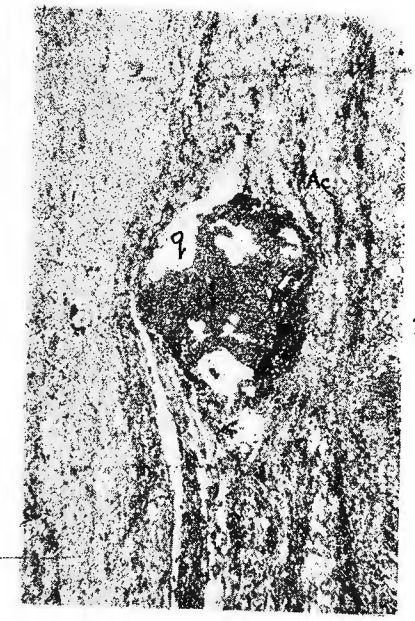

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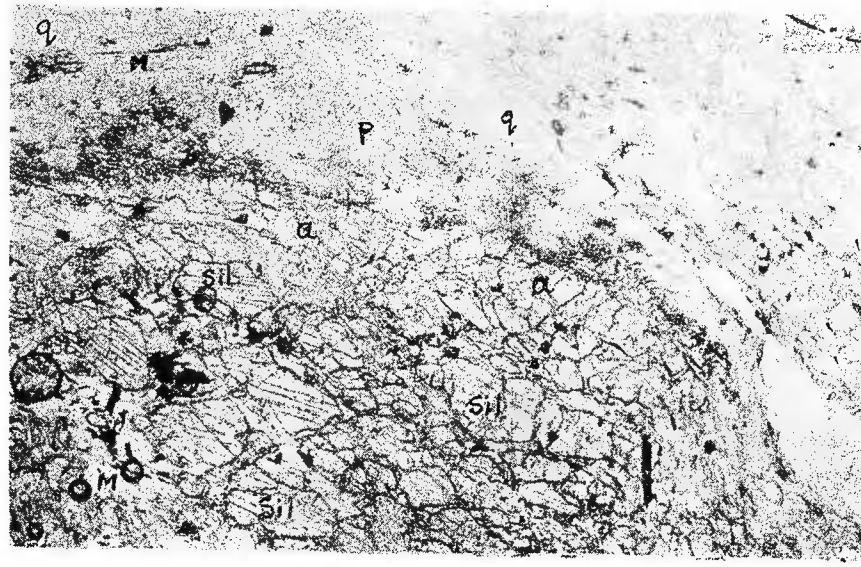

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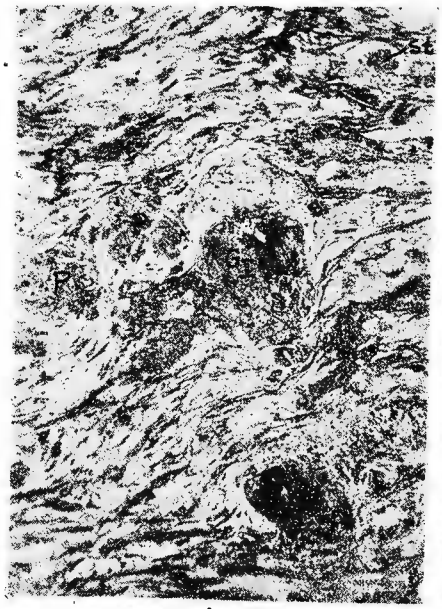

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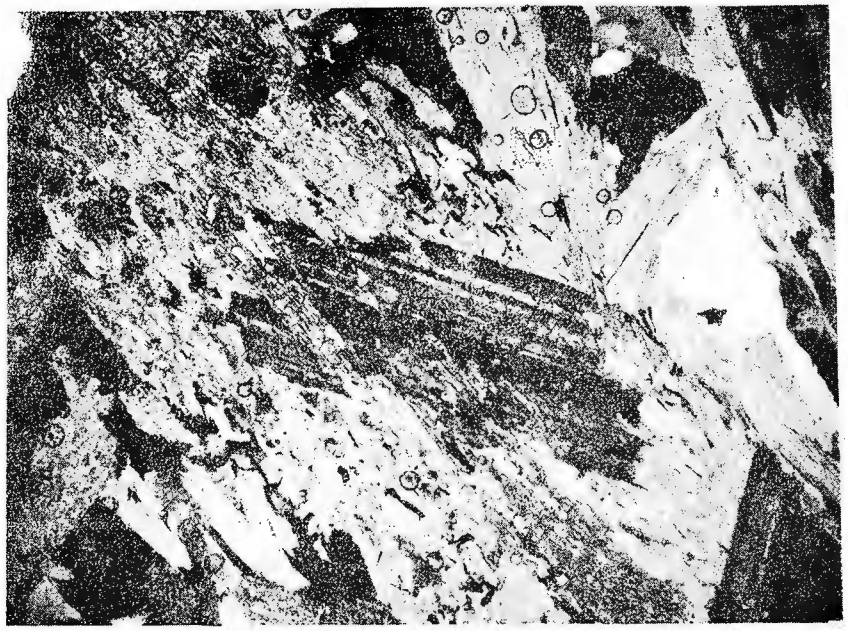

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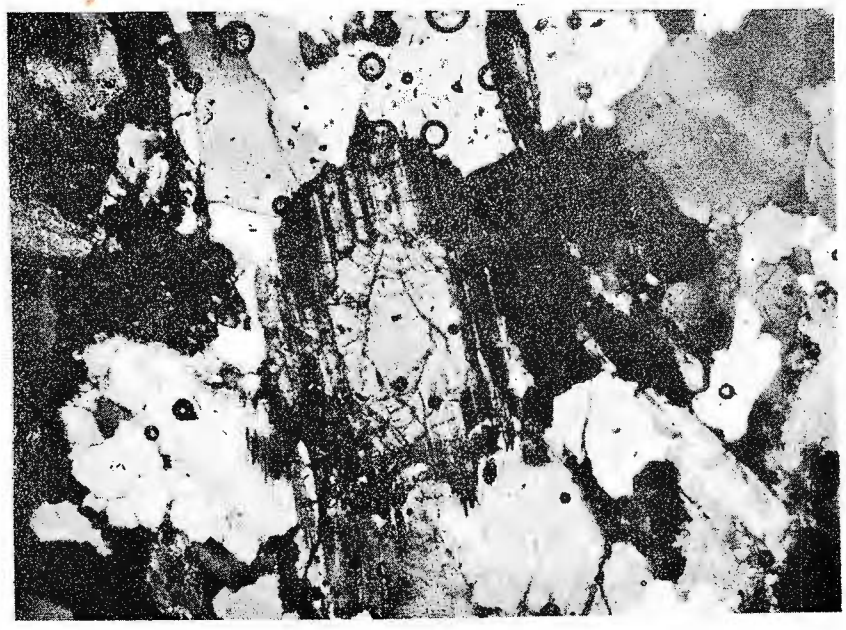

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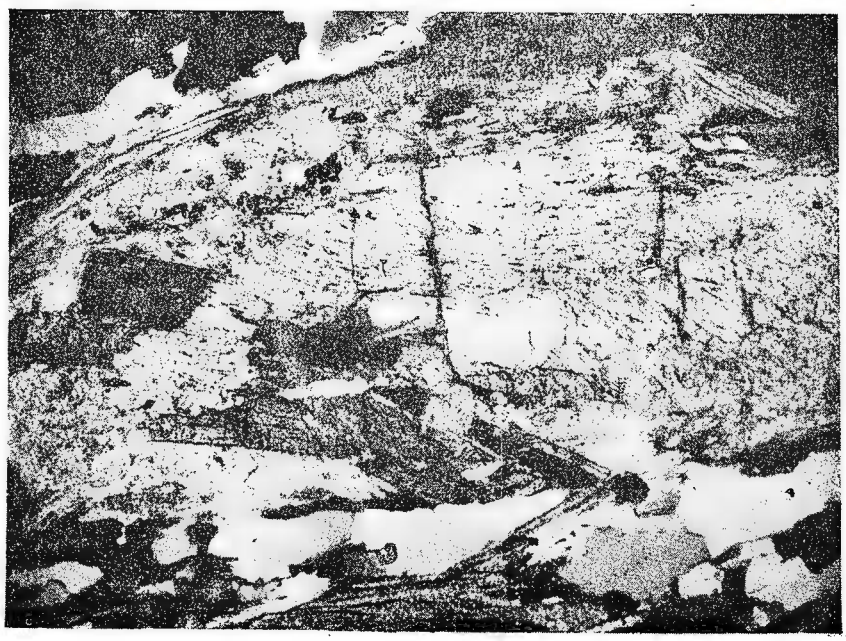

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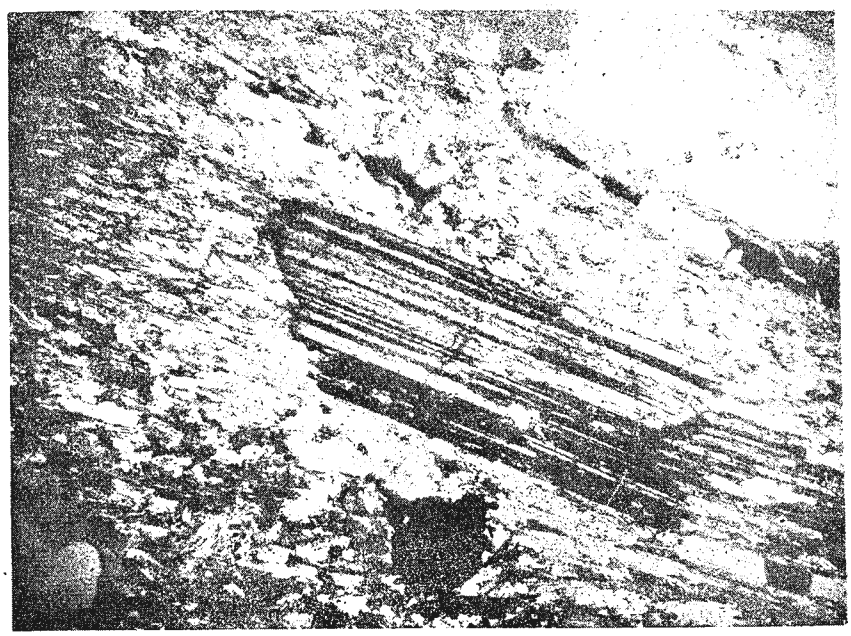

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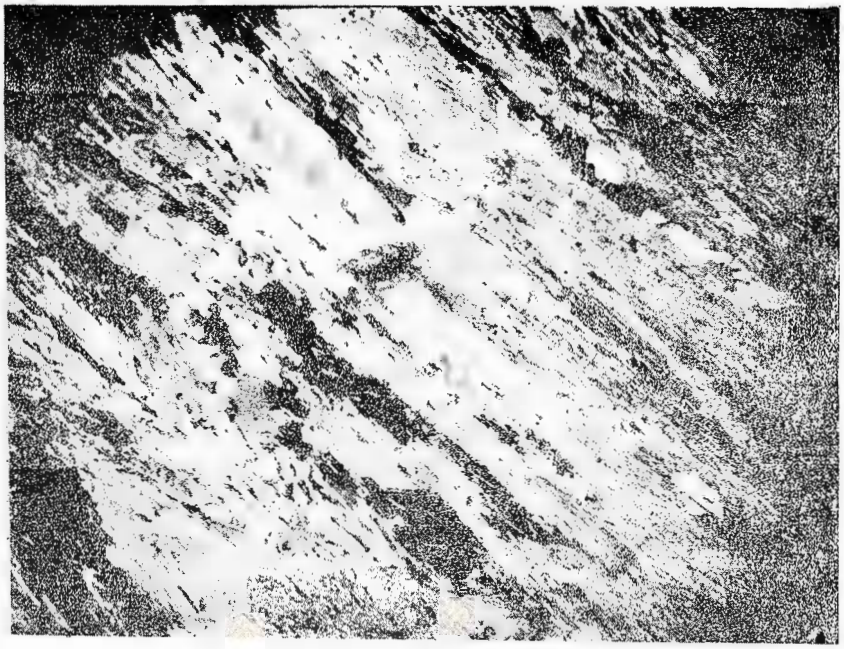

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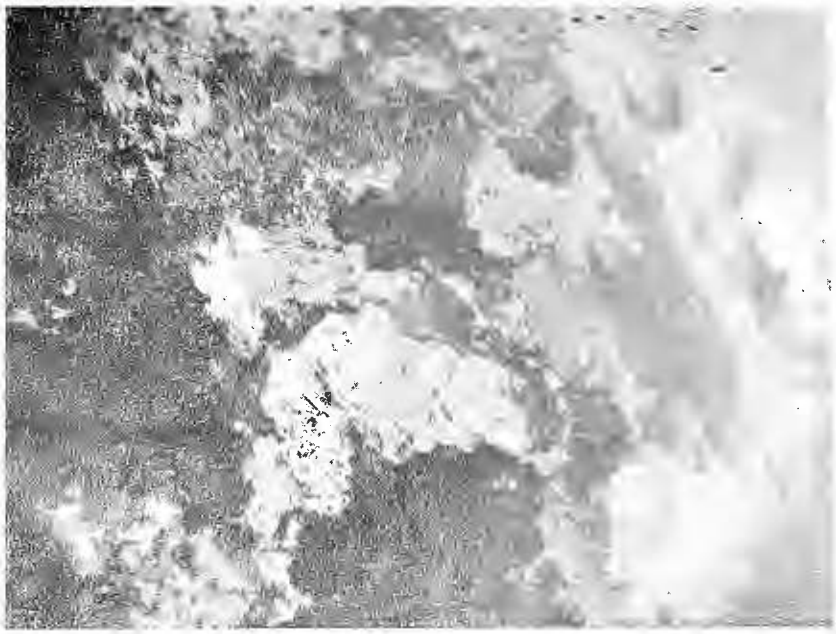

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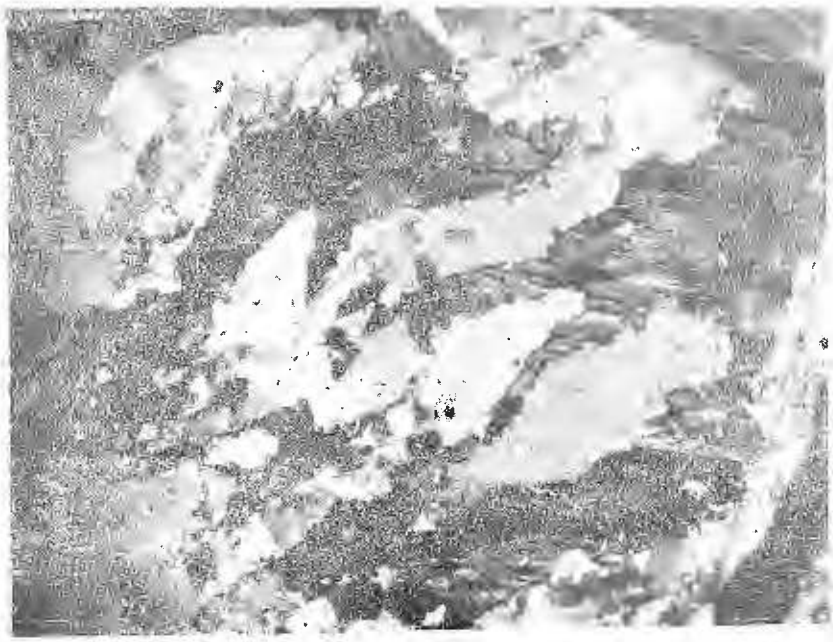

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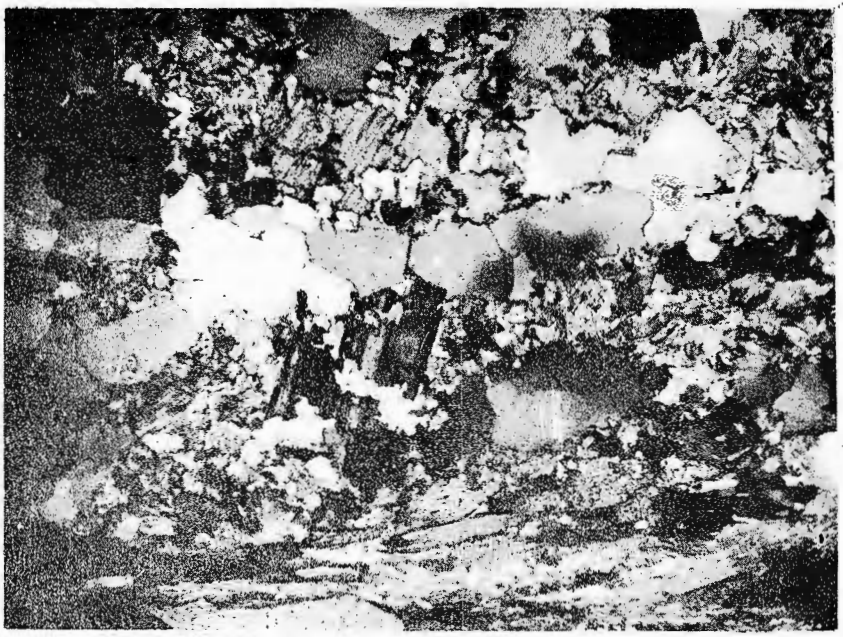

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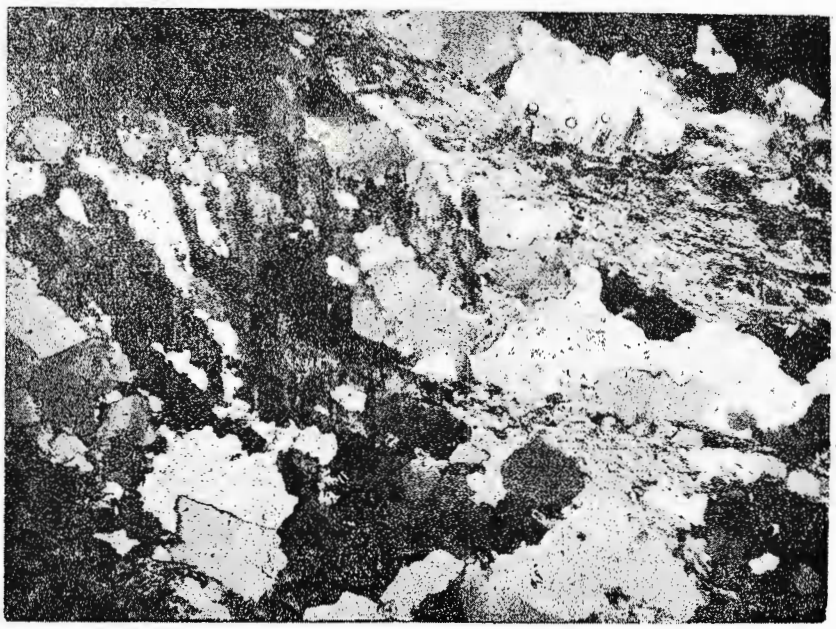

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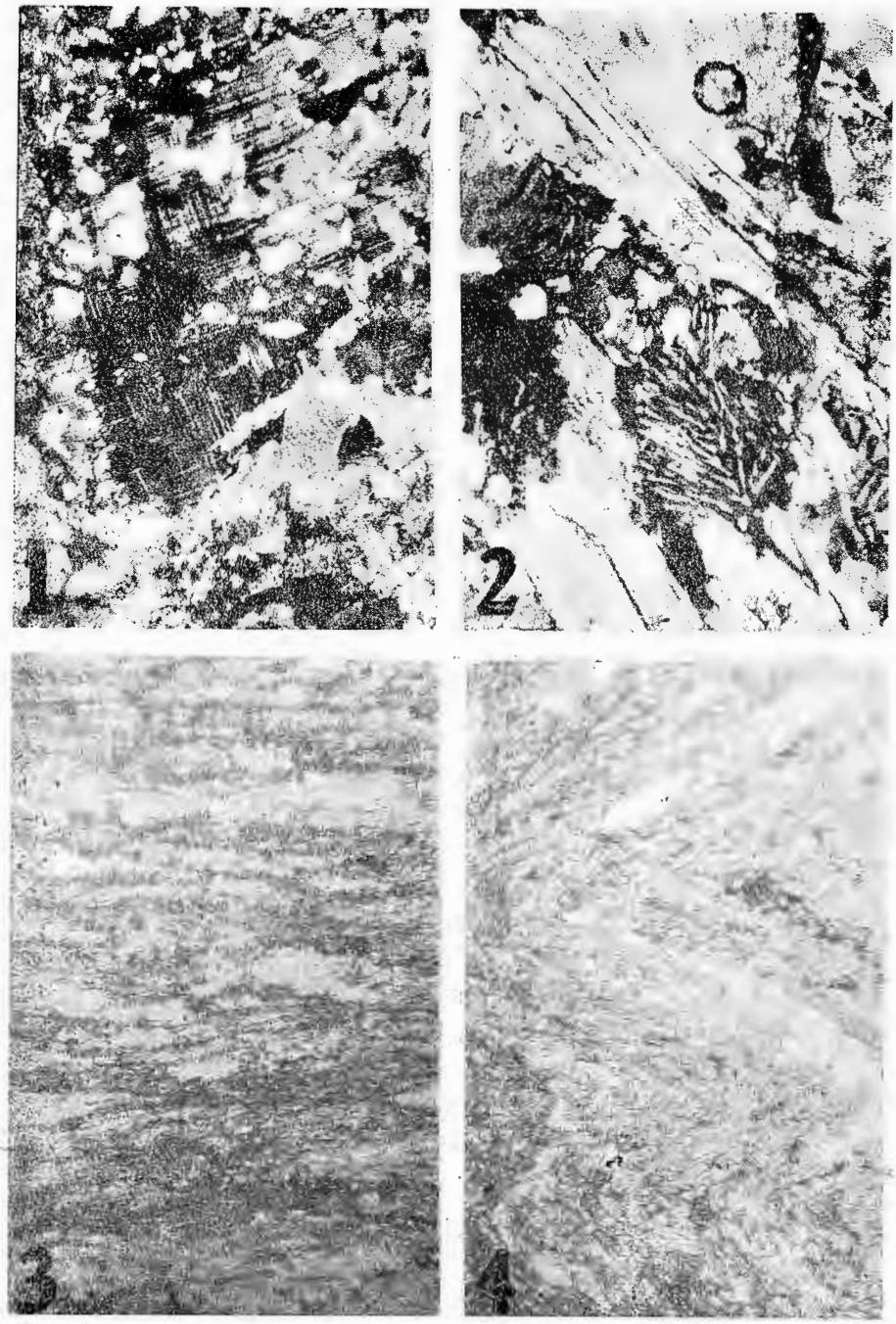

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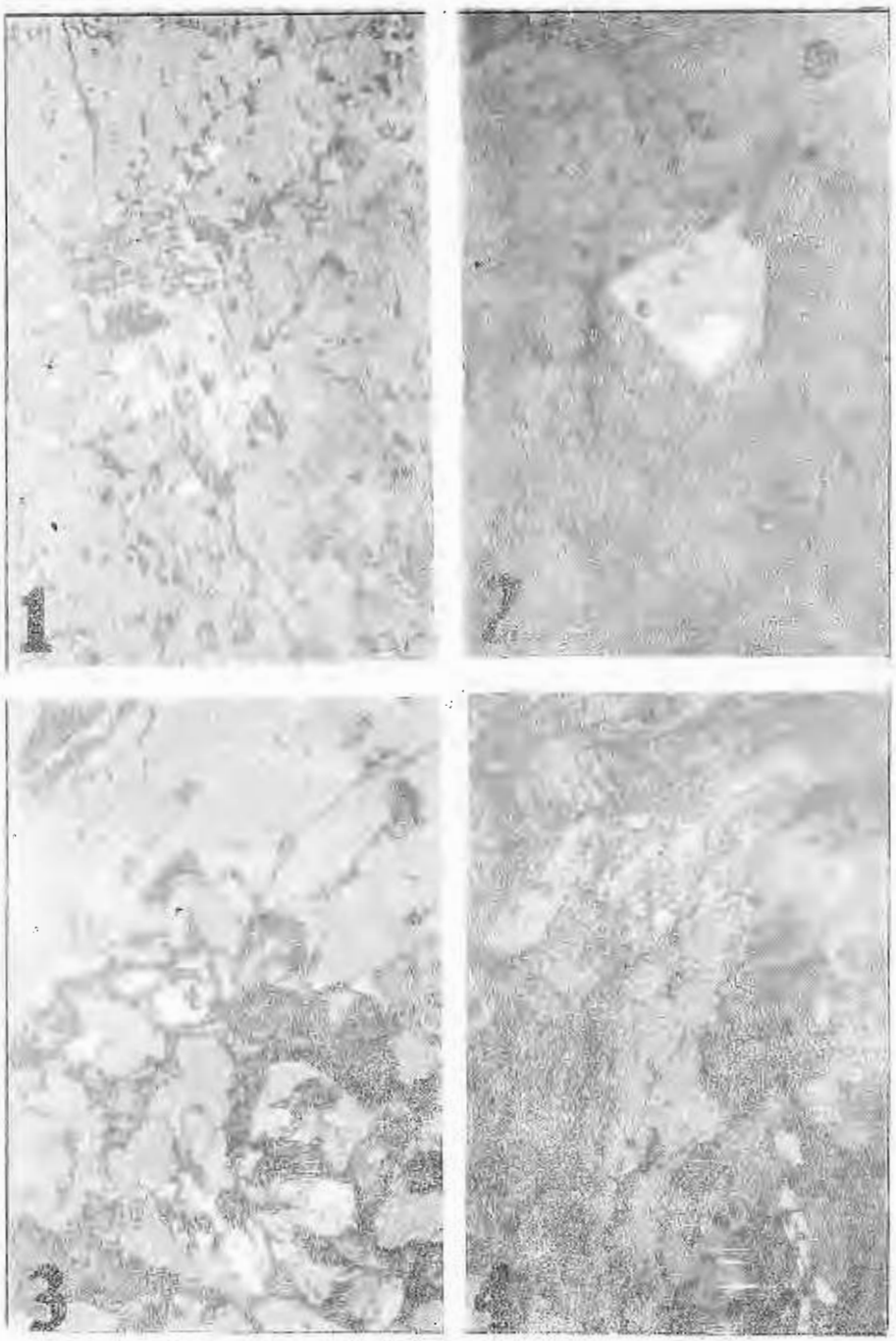

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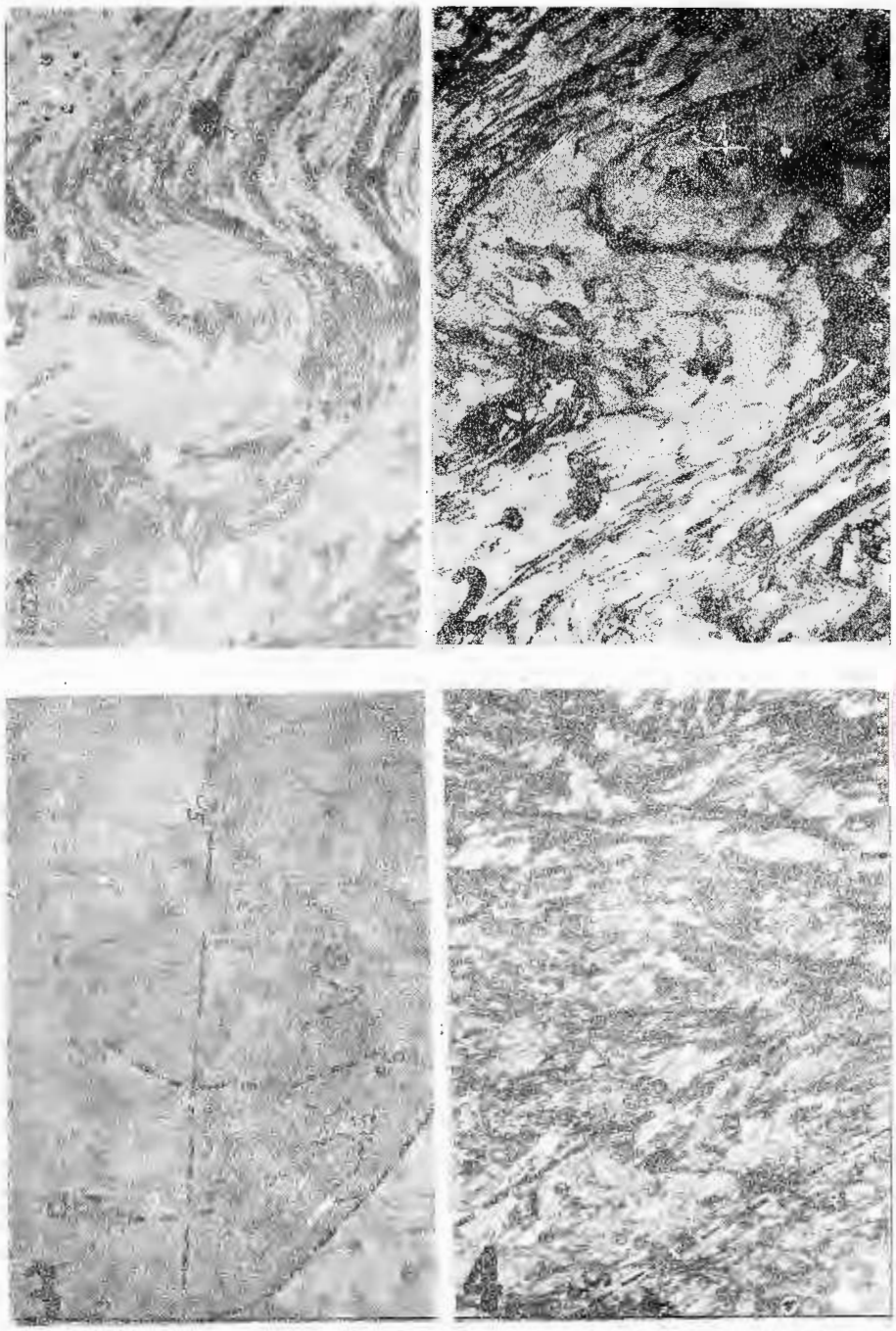

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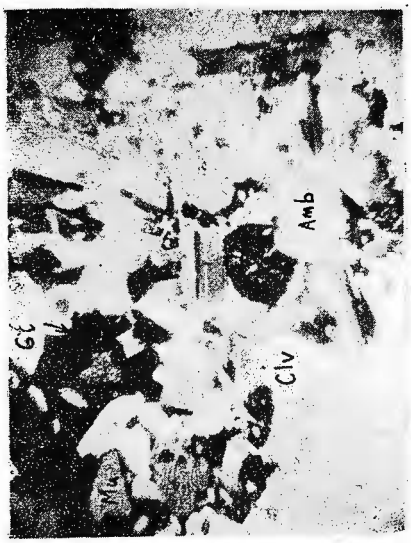

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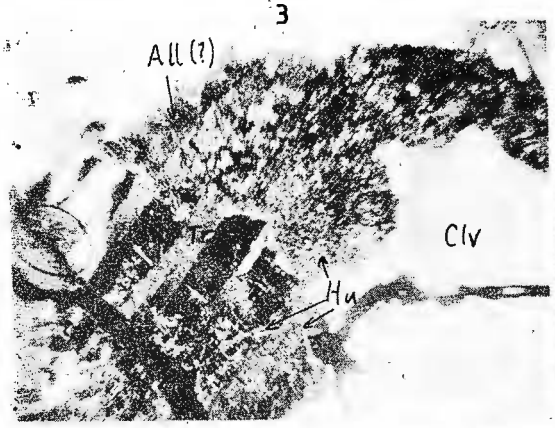

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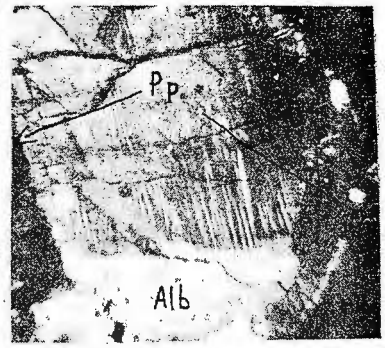

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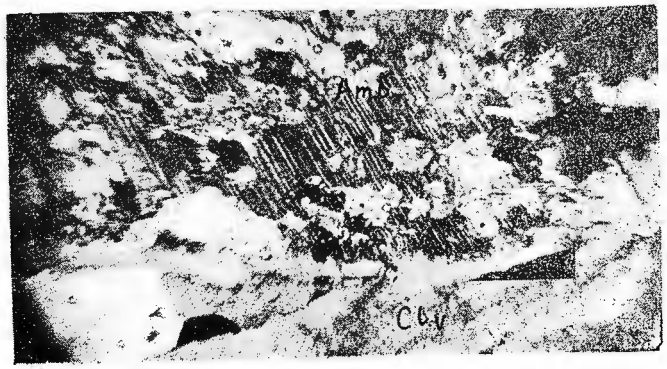

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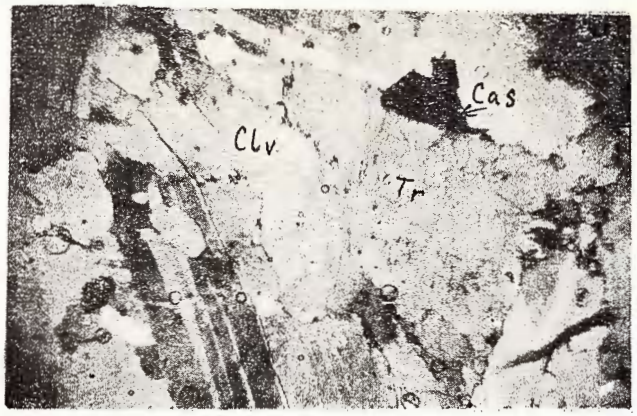

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Comitetul Geologic t. XXXVII–LII/1(1953–1966)
Comitetul de Stat al Geologiei t.LII/2–LV/1(1967–1969)
Institutul Geologic t.LV/2–LX(1970–1974)
Institutul de Geologie și Geofizică – à partir du tome LXI(1975)

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