MAGMATIC FEATURES AND METAMORPHISM OF PLAGIOGRANITE ASSOCIATED WITH A JURASSIC MORB-LIKE BASIC-ULTRABASIC COMPLEX, BÜKK MOUNTAINS, HUNGARY

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ABSTRACT

Plagiogranite forms cross-cutting dikes, veins with sharp boundaries within the MORB-like Jurassic basic-ultrabasic complex (Szarvaskő complex) of the southwestern Bükk Mountains, NE-Hungary. Earlier petrological and geochemical interpretation suggests that the basic magmas became contaminated by terrigenous sediments in a shallow magma chamber (DOWNES et al. 1990), thus producing atypical, incomplete ophiolitic sequence. In order to characterise the formation of the plagiogranite in question, literature data on oceanic granitic rocks are critically reviewed. The Szarvaskő plagiogranite consists mainly of quartz, plagioclase, biotite ± garnet. Its major element bulk chemistry is similar to that of the oceanic plagiogranites (high Si, moderate Al, low total Fe+Mg and extremely low K contents), crystallised from late differentiated basic melt. The garnet being present in strongly varying proportions in granite is usually accumulated near or at the contact with the metaclastic enclaves. The garnet is almandine-rich, and displays slight chemical zoning of changing character. Most frequently, the Fe content increases and the Ca content decreases from the centres towards the rims of the grains. Its chemistry corresponds to that of the garnet found in metapelites or is plotted near to the field of the igneous silicic rocks in the diagrams used for discrimination. As the investigated metaclastic enclaves are represented by biotite-plagioclase hornfels, and garnet never forms contact metamorphic rim around the enclaves, but it is always dispersed in the granitic melt, the formation of the garnet is most probably a consequence of local assimilation of the metaclastic rocks at a deeper level. The magmatic minerals are extensively altered: the plagioclase is completely albitised, while the biotite is totally altered to chlorite, epidote, prehnite and titanite. The secondary mineral assemblage consists of quartz, albite, chlorite, epidote, prehnite, calcite, titanite, white mica (sericite) and smectite, while pumpellyite is lacking. Illite and chlorite "crystallinity" indices indicate anchizonal metamorphic circumstances, agreeing fairly well with the temperature data (270–350°C, in average, 298°C) obtained by the chlorite-Al^{IV} geothermometer of CATHELINEAU (1988). These data refer to metamorphic conditions corresponding either to the pumpellyite-prehnite-quartz or to the prehnite-actinolite facies. Earlier petrological data of ÁRKAI (1983) prove Cretaceous regional (dynamothermal) metamorphism in the surrounding formations. In order to get evidence on eventual preceding (ocean-floor hydrothermal or burial) metamorphic effects, detailed study of the whole ophiolite-like sequence is required.

INTRODUCTION

Complete ophiolitic sequences consist mainly of ultrabasic and basic rocks. Within these ophiolites, leucocratic rocks ranging in compositions from tonalite to albite granite are usually regarded as the final products in the crystallization of basic magma and they have generally been called keratophyre, quartz-keratophyre, tonalite, trondhjemite, quartz-diorite, albite granite, albitite or plagiogranite (WILSON, 1959, THAYER
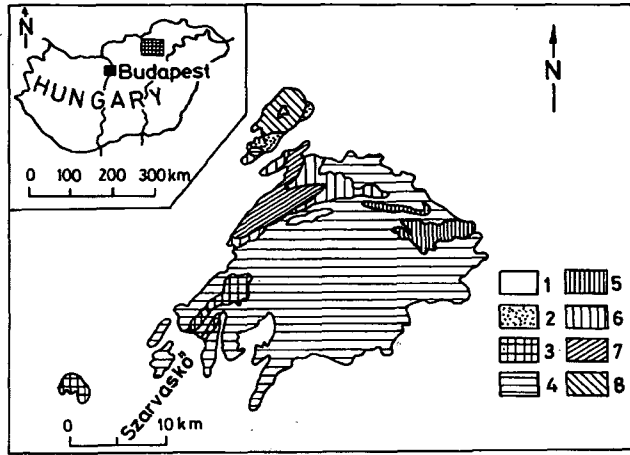
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Fig. 1. Simplified geological sketch map of the Bükk Mountains (after BALLA 1982)
Legend: 1=Cenozoic, 2=Upper Cretaceous, 3=Jurassic basic and ultrabasic rocks (Szarvaskő complex), 4=Triassic and Jurassic sedimentary, 5=Triassic metavolcanic, 6=Permian sedimentary, 7=Middle-Upper Carboniferous sedimentary, 8=Devonian-Lower Carboniferous sedimentary formations metamorphosed in low temperature conditions

1963, COLEMAN and PETERMAN 1975, COLEMAN 1977). These acidic rocks frequently form a small portion in the ophiolitic sequences.

The innermost tectonic units of the Western Carpathians (South Gemeric Unit, Bükk Unit) contain dismembered fragments of Mesozoic ophiolitic suites. In one of their largest occurrences, i. e. in the Szarvaskő basic-ultrabasic complex of **MORB**-type (*Fig. 1*), a great variety of rock types (gabbrodiorite, diorite, plagioclasite, quartz-diorite, granite and their aplitic, pegmatitic and porphyritic variants) were described by SZENTPÉTERY (1953). The rocks with intermediate composition occur mainly as local differentiates ("schliers") within tha basic rock mass with continuous transitions towards the gabbro, while the acidic types form cross-cutting dikes, veins with sharp boundaries. SZENTPÉTERY (1953) also gave the modal and main element compositions of these rocks, the formation of which was explained by crystallisation differentiation.

In spite of the detailed petrographic descriptions of the various types mentioned above, the petrogenetic interpretation of these rocks are lacking. Publishing new rock textural, mineral paragenetic, mineral chemical and bulk chemical data, this paper presents contributions to the petrogenesis of the plagiogranite, with special references to the distinction of the magmatic and metamorphic processes.

NATURE OF GRANITIC ROCKS IN OCEANIC ENVIRONMENT

Mineralogically, the oceanic plagiogranites consist mainly of quartz and plagioclase. Ferromagnesian minerals occur in small amounts and are commonly represented by hornblende, pyroxene or biotite. K-feldspar is completely absent but present only in rare cases. Generally, the plagiogranite rocks can be distinguished from the similar continental leucocratic ones which derived from differentiation of mafic rocks by their

chemical composition. The former ones are characterized by high silica, moderate alumina, low iron-magnesium and extremely low potassium content (COLEMAN and PETERMAN 1975, COLEMAN 1977).

It is obvious that the oceanic leucocratic rocks usually show close relation with associated mafic rocks. Two types of boundaries between the granophyric rocks of Troodos Igneous Complex and the underlying gabbro were distinguished by WILSON (1959): the gabbro may be interdigitated with the more acidic rocks or it may be capped by massive granophyres which are rarely penetrated by gabbro apophyses, whereas the boundary with overlying diabase is generally gradual. Sharp boundary could be recognized by THAYER (1963) between the gabbro and the quartz dioritic rocks; the latters forming dikes, small pipes and irregular masses. Moreover, dikelets of Na-rich trondhjemite and quartz monzonitic composition cutting some granophyric diabase intrusions are described by ENGEL and FISHER (1975). BATLEY *et al.* (1970) referred to the occurrence of light-coloured pillow keratophyric lavas that are underlain by a volcanic succession including basalt and diabase. Furthermore, BATLEY and BLAKE (1974) noticed silica-rich extrusive and intrusive rocks: the finer keratophyre rocks occur as pillow lavas and as massive flows above the gabbro, while the coarser varieties occur as dikes or sills. In very rare cases, these silicic rocks show relationship with the ultrabasic rocks of ophiolites, however, as described by IWAO (1953) in COLEMAN and PETERMAN (1975) intrusions of silicic rocks are commonly present within serpentinized ultrabasic rocks.

Many attempts have been made to define the origin of oceanic leucocratic rocks. The early opinion had been hold by GILLULY (1933) who attributed the origin of albite granite to late-magmatic and post-magmatic metasomatism of an earlier quartz diorite. In fact, these leucocratic oceanic rocks have been shown by most investigators as a product of differentiation processes rather than metasomatic origin.

In the ophiolites of Troodos, the $^{87}Sr/^{86}Sr$ ratios of mafic varieties are similar to those of silicic ones. This similarity is consistent with the idea of low pressure differentiation of a parental subalkaline basaltic magma to produce the K-poor leucocratic rocks (COLEMAN and PETERMAN 1975).

Fractional crystallization experiments were performed on a primitive oceanic tholeiite at dry and 1 atmosphere total pressure by DIXON and RUTHERFORD (1979). In these experiments, the produced residual liquid is of basaltic composition, and with increasing crystallization (after 95% crystallization) the residual liquid became Fe-enriched basalt. This Fe-enriched basalt would be immiscible at a temperature of about 1010°C. A more Fe-enriched basaltic liquid and a granitic liquid which resemble in composition to the oceanic plagiogranites present in certain MOR regions were the immiscible phases produced. Therefore. silicate liquid immiscibility could be responsible for producing plagiogranites in some ophiolites. Furthermore, hydrothermal experiments were made by SPULBER and RUTHERFORD (1983) on a primitive oceanic thoeliite. After 90% crystallization the residual liquid is characterized by high silica and low K_2O contents, similar to low K_2O plagiogranites present in ophiolitic sequence. In these hydrothermal experiments, however, silicate liquid immiscibility did not take place.

GEOLOGICAL SETTING

In the southwestern part of the Bükk Mountains, NE Hungary, Mesozoic ophiolitic sequences include plagiogranites which occur in Tóbérc quarry near the village Szar-

vaskő (*Fig. 1*). These sequences are situated in the innermost part of Western Carpathians, and show close relation to the Dinarids and Southern Alps, too. At present these ophiolitic sequences and associated sediments are regarded as ancient parts of the Inner Dinaric belt that moved to their present position as a result of horizontal displacements of the Bükkium towards the east during the Tertiary (KOVÁCS 1982, BALLA 1984).

According to BALLA (1983) and BALLA et al. (1983), these ophiolites form nappe system. The internal structure of these nappes is very complicated with overthrusting and strong folding. Clearly the reconstructed original stratigraphic sequence of Szarvaskő begins with sandy terrigenous flysch sediments which are gradually replaced upwards by clayey sequences. Within this terrigenous sequence, thick intrusive sills ranging in composition from ultrabasic to acidic (including plagiogranite rocks which occur as nests in the intrusive gabbro) are located in the lower part. As regarded by BALLA (1984), these intrusive bodies can be compared with the sheeted dykes of a complete ophiolitic succession. The effusive rocks are situated in the upper part of the sequence and are represented by pillow lavas formed under deep-water conditions. Based on main and trace element (including REE) geochemical and Sr and Nb isotope data of DOWNES et al. (1990) suggested that the basic magmas of the complex were probably originally **MORB**-like, but during evolution in a shallow magma chamber, they became contaminated by terrigenous sediments.

As regarded by ÁRKAI (1983), the basic rocks of Szarvaskő were affected by pumpellyite-prehnite-quartz facies regional metamorphism, while the illite "crystallinity" and vitrinite reflectance data of the surrounding sedimentary rocks refer to low-temperature (anchizonal) metamorphic and late-diagenetic transformations. According to ÁRAKAI (1983) the age of the regional metamorphism is Alpine (Cretaceous, pre-Senonian).

The presence of acidic rocks within ophiolites of Szarvaskő has been defined by some investigators. As shown by SZENTPÉTERY (1953), these rocks exist as nests within gabbro. The plagiogranite consists of plagioclase, biotite and quartz with garnet (EMBEY-ISZTIN et al. 1985). Also DOWNES et al. (1990) reported that these rocks consist mainly of albite + quartz + biotite, with apatite, zircon, sphene and xenotime being present as accessories. Secondary minerals include chlorite, prehnite, actinolite and sericite. Garnets were observed in some quartz diorite, having low Ca- and high Fe-content indicating low pressure metasomatic origin (EMBEY-ISZTIN et al. 1985). The chemistry of these rocks show strong LREE. LILE and Th enrichment and they are strongly depleted in Ti, P and Sr relative to the incompatible elements Zr, Nd and Ce (DOWNES et al. 1990).

Contradiction can be found in discussion regarding the origin of plagiogranites within ophiolites of Szarvaskő. As stated by SZENTPÉTERY (1932, 1938a, 1938b, 1939a, 1939b, 1940, 1953) the Szarvaskő rocks were generated by differentiation which went in two directions, one giving way to ultramafics and the other to intermediate acidic rocks. This opinion has been modified by BALLA et al. (1983) who related their origin not only to differentiation process but also at least to progressive assimilation of the country rocks. Later BALLA (1984), using the analytical data of DOBRETSOV, reported that the garnetiferous varieties were formed by melting of sedimentary or metamorphic rocks while the garnet-free ones by assimilation of this molten material in the gabbro magma. EMBEY-ISZTIN et al. (1985) attributed to the garnet present in these plagiogranites neither xenocrystal nor cognate origin. As chemistry and zoning pattern of garnet strongly vary from grain, garnet might form by metasomatism under different conditions (EMBEY-ISZTIN et al. 1985).

The plagiogranite samples were collected from the Tóbérc quarry and its surroundings located south to the village Szarvaskő (southwestern part of the Bükk Mountains).

The rock samples were examined macroscopically in order to define the mineral composition, and the micro-textural features, with special regards to the distinction of the primary and secondary minerals.

The XRD analysis was used to determine te modal composition, illite "crystallinity" (abbreviated as IC(002) which expresses the half-height width of the illite first 10 Å basal reflection; see KÜBLER 1968, 1990) and chlorite crystallinity, ChC(001) and ChC(002) which are the half-height width values of the chlorite 14 Å and 7 Å reflections, respectively; see ÁRKAI 1991). To determine such parameters, the rock samples were crushed first in a jaw crusher, and then in a mortar mill (Type Pulversette 2, Fritsch) for about 3 min as mentioned by ÁRKAI (1982) to avoid the effects on the quantities of small (~20mm) grain size fractions produced or on the values of IC. These parameters were measured on desoriented powder of the whole rock, highly oriented fraction less than 2μm grain size which were separated from aqueous (clay-water) suspension onto a glass slide, using the method of KÜBLER (1975), and was dried at room temperature. Solvation of $<2\mu$m fraction samples was done holding the preparations in an atmosphere saturated with ethylene glycol at 80°C for 4 hours.

The samples were analyzed by a Philips PW–1730 X-ray diffractometer at 45 KV accelerating voltage and 35 mA amperage. The XRD instrument was used CuK_α radiation, proportional counter, divergence and receiving slits of 1°, graphite monochromator, goniometer speeds of 2°/min and 1/2°/min, and time constant of 2 sec.

In the laboratory of Geochemical Research, the IC measurements were calibrated with those of KÜBLER's laboratory where the anchizone boundaries are 0.25 and 0.42, by using standard rock slab series kindly provided by KÜBLER (see ÁRKAI 1983, 1991).

Microprobe analysis of rock-forming minerals was performed on carbon coated thin sections, using a JEOL Superprobe–733 electron microprobe instrument. The analyses were done at 15 accelerating voltage and 30 nA beam current using a beam size of approximately 5–10 μm. The wave-dispersive spectrometric quantitative analyses were done using Bence-Albee correction methods (BENCE and ALBEE 1968) and spessartite (Si, Al, Mn), olivine (Mg), hematite (Fe), wollastonite (Ca), albite (Na), spessartite (Mn), rutile (Ti) standards for the other minerals.

Whole rock chemical analysis of major elements has been made in the Geochemical Research Laboratory by Zoltán Wieszt, employing AAS Perkin Elmer 5000, using lithium metaborate digestion. Other methods, such as gravimetric for SiO_2 and loss-on ignition, permanganometric for FeO, spectrophotometric for SiO_2, TiO_2 and P_2O_5, and volumetric for CO_2 determinations, have been also used.

PETROGRAPHY

The plagiogranites of Bükk Mountains are fine to medium grained, light coloured rocks displaying typical granitic microstructure and composed mainly of quartz and plagioclase (*Plate I/1*), while K-feldspar has not been recognized in these rocks. The quartz occurs as xenomorphic crystals characterized average grain size ranging from 0.15 to 0.88 mm. Undulating extinction is found in some quartz crystals and notable cracks are also recognized, however, these cracks are sometimes filled with secondary

minerals such as sericite. Feldspar is commonly represented by unzoned, albitized plagioclase. According to the electron microprobe analyses, its composition varies between An_7 and An_1, with an average of An_2. The size of plagioclase may reach 2.34 mm in length and 1.08 mm in width, generally the average size ranges from 0.41 to 0.90 mm in length and from 0.17 to 0.38 mm in breadth. Certainly plagioclase crystals are highly corroded and cracked, and largely altered along cracks and cleavage planes or completely replaced by prehnite laths and sericite (*Plates I/2 and I/3*). Apatite and chlorite are found in plagioclase as inclusions. Biotite is the only primary ferromagnesian mineral present in minor amounts (*Plate I/4*). Biotite was replaced totally by secondary minerals (*Plates II/1, II/2 and II/3*) including chlorite, epidote, prehnite (Fe-rich and Fe-poor) and titanite which in some crystals lies parallel to the (001) planes of chlorite. Inclusions of ilmenite, zircon, rutile, apatite and sphene are observed within altered biotite (*Plate II/4*). Other minerals, mostly ilmenite, pyrite, rutile, zircon and apatite are present as accessories. Garnet is frequently present in different proportions or my be completely absent. Depending on the abundance of garnet, the plagiogranites can be divided into two types, namely, garnet-poor or free and garnet-rich rocks (*Fig. 2*). The former type is usually found in Tóbérc quarry, containing gabbro rocks. It contains rounded, cracked garnet crystals (*Plate III/1*) ranging in size from 0.70 to 1.40 mm. These crystals contain inclusions of quartz, ilmenite, titanite and rutile. In the latter type coarse to fine grained garnets (average size reaches 0.35 mm) are commonly concentrated near to contacts with metasediment enclaves (*Plate III/2*). However, these garnet grains are embedded in the granite (*Plate III/3*), as plagioclase and quartz around garnet crystals commonly resemble those of the granitic rocks. These garnets contain mostly inclusions of quartz and ilmenite and their cracks are filled with titanite and prehnite. In general, colour zoning is absent in both garnet types as they exhibit little changes in their compositions from core to rim. The metasediments are usually associated with garnet-rich variety, and do not show preferred orientation. They seem to be xenoliths within acidic rocks, and consist essentially of fine grained, untwinned plagioclase and of chlorite that appears as a pseudomorph after biotite (*Plate III/4*). Monacite is present as accessory mineral. No quartz could be distinguished in these xenoliths. The plagioclase in these metasediments show prograde zoning: the Ca-content increases towards the edges. The contact between granitic rock and metasediment is sharp.

The fissure filling of these rocks (*Plate IV/1*) are represented by aggregates of chlorite, calcite, prehnite, quartz, albite, apatite and titanite. The iron content of prehnite in the aggregates is varying, the Fe-richer part is more or less located near fractures, and mostly forming exsolution laths (*Plate IV/2*).

Based on the XRD studies, the minerals present in the garnet-free or -poor plagiogranite of Tóbérc quarry are plagioclase, quartz and chlorite with minor prehnite and calcite. Traces of ilmenite, zircon, rutile, apatite, titanite, pyrite, epidote, kaolinite, smectite and garnet are also found. On th other hand, the minerals of the garnet-rich granites are mainly plagioclase, quartz, chlorite and garnet with minor calcite. Apatite, zircon, ilmenite, titanite, hematite, sphene, epidote, and prehnite are present as traces (Table I.) The measured XRD parameters are shown in Tables II and III.

The contact between granite and gabbro is more or less sharp. The average grains sizes of plagioclase and quartz decrease towards this contact (chilled margin).

1

2

3

4

Plate I
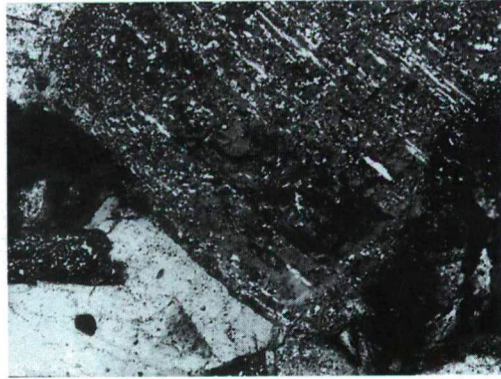
1. Microtexture of plagiogranite (sample No. 1). Mineral composition includes quartz, plagioclase and chlorite. Fine cracks are filled with prehnite. Plane polarized light, approximate width of the photo is 6 mm.

2. Coarse plagioclase crystal altered along cleavage planes to sericite, while its cracks are filled with chlorite (sample No. 1). Crossed nicols, approximate width of the photo is 0.75 mm.

3. Laths of prehnite formed after plagioclase. Sample No. 5a. Crossed nicols, approximate width of the photo is 0.75 mm.
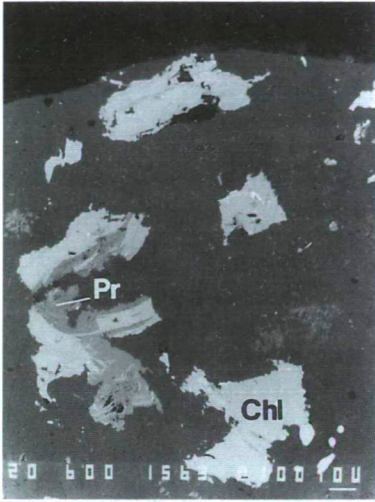
4. Chloritized biotite. Sample No 5c. Plane polarized light, approximate width of the photo is 0.75 mm.
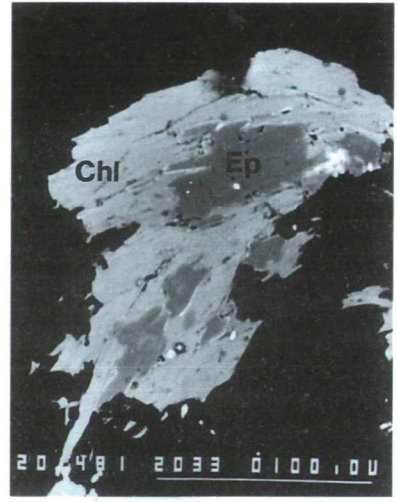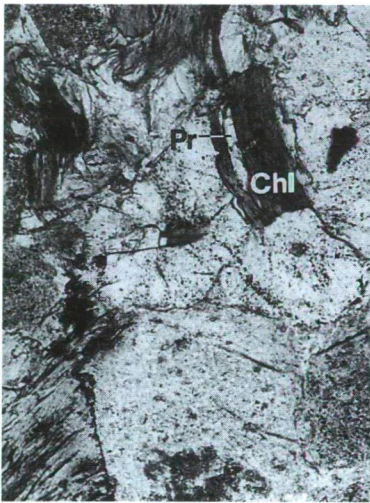
1

2

3


4

Plate II

1. Biotite altered to chlorite (Chl) and prehnite (Pr) and plagiclase to prehnite (Pr). Sample No 1. Back scattered electron (BSE) image, the length of the white bar in the lower right part of the photo is 0.1 mm.

2. Biotite crystal totally replaced by chlorite (chl) and epidote (Ep). Sample No. 34b. BSE image, the length of the white bar in the lower right part of the photo is 0.1 mm.

3. Biotite crystal altered to chlorite (Chl) and prehnite (Pr.). Sample No. 1. Plane polarized light, approximate width of the photo is 0.75 mm.

4. Inclusions of zircon and apatite within chloritized biotite. Sample No. 1. BSE image, the lenght of the white bar in the lower right part of the photo is 0.1 mm.

1

2

Plate III
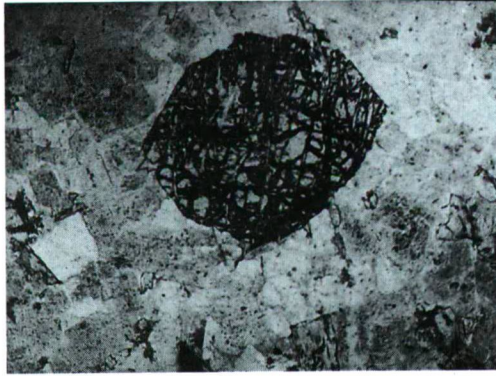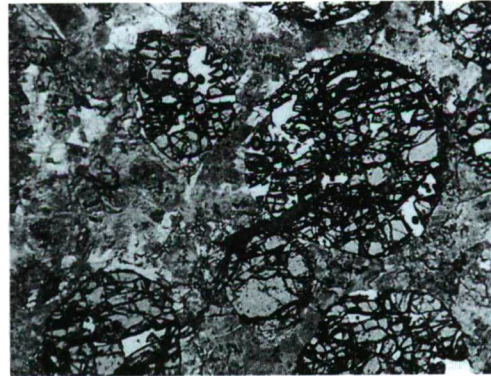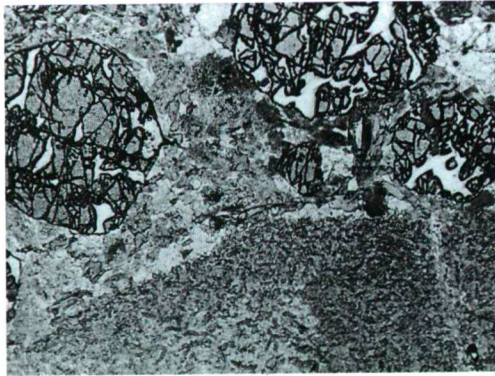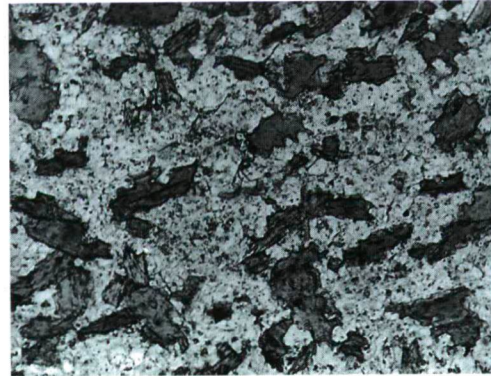
1. Garnet from garnet-poor granite. Sample No. 5b. Plane-polarized light, approximate width of the photo is 3 mm.

2. Abundant garnets from garnet-rich plagio-granite. Sample No. 34c. Plane-polarized light, approximate width of the photo is 6 mm.

3. Sharp contact between garnet-rich plagio-granite and metasediment enclave. Sample No. 34b. Plane-polarized light, approximate width of the photo is 6 mm.

4. Metasediment xenolith consists mainly of plagioclase and chloritized biotite. Preferred orientation is absent. Sample No. 34b. Plane-polarized light, approximate width of the photo is 0.75 mm.

3

4

Plate IV

1. Fracture filling with prehnite and calcite in garnet-poor plagiogranite. Sample No. 5c. Crossed nicols, approximate width of the photo is 3 mm.

2. Exsolution of Fe-rich prehnite in host Fe-poor prehnite. Sample No. 5b. BSE image, the legth of the white bar in the lower right part of the photo is 0.1 mm.

Fig. 2(a–b).

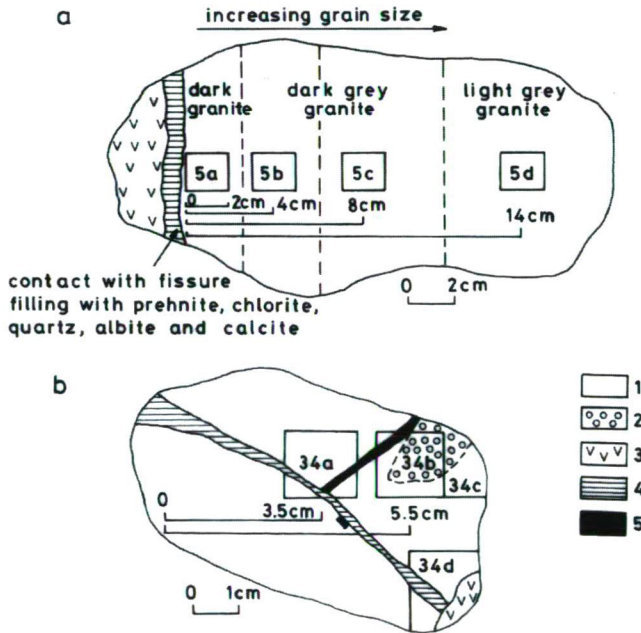Sketch drawing of plagiogranite types of the Tóbérc quarry and the surroundings. 2a is a part of a garnet-poor granite dike showing the contact with gabbro. 2b is a garnet-rich granite also contacting gabbro but containing metasedimentary enclave, too.

Legend: 1=garnet-poor granite, 2=garnet-rich granite, 3=gabbro, 4=white fissure filling (prehnite, calcite, quartz), 5=metasedimentary xenolith

TABLE 1

Modal composition of the granites studied

Sample	Pl	Qtz	Chl	Ms	Alm	Smc	Kln	Prh	Pmp	Cal	Ep	Hem	Aln	Py	Ttn	Ilm	Zrn	Rt	Ap
1	+	+	x	tr	tr	tr	tr	o	–	–	tr	tr	tr	tr	tr	–	–	–	tr
5a	+	+	x	tr	tr	tr	–	o	–	o	tr	–	–	–	–	tr	–	–	tr
5b	+	+	x	tr	tr	tr	–	o	–	o	tr	–	–	–	–	tr	–	–	tr
5c	+	+	x	tr	tr	tr	tr	o	–	o	tr	tr	–	–	tr	tr	–	tr	tr
5d	+	+	x	tr	–	tr	tr	o	–	o	tr	tr	tr	–	tr	–	tr	–	tr
34a	+	+	x	tr	tr	tr	–	tr	–	tr	tr	–	–	–	–	–	–	–	–
34b	+	+	x	tr	+	tr	–	tr	–	tr	tr	tr	–	–	tr	–	tr	–	tr
34c	+	+	x	tr	+	tr	–	tr	–	o	tr	–	–	–	–	tr	–	–	tr
34d	+	+	x	tr	–	tr	–	o	–	o	tr	tr	–	–	tr	tr	–	–	tr
34e	+	+	x	tr	–	tr	–	tr	–	tr	tr	–	–	–	–	–	–	–	tr
35	+	+	x	tr	+	tr	tr	tr	–	–	tr	o	–	–	–	–	–	–	tr
36	+	+	x	x	–	–	–	–	–	–	tr	–	–	–	–	–	–	–	tr

Pl = plagioclase, **Qtz** = quartz, **Chl** = chlorite, **Ms** = muscovite, **Alm** = almandine, **Smc** = smectite, **Kln** = kaolinite, **Prh** = prehnite, **Pmp** = pumpellyite, **Cal** = calcite, **Ep** = epidote, **Hem** = hematite, **Aln** = allanite, **Py** = pyrite, **Ttn** = titanite, **Ilm** = ilemenite, **Zrn** = zircon, **Rt** = rutile, **Ap** = apatite.

+ = common   x = considerable   o = minor   **tr** = traces   – = missing

TABLE 2
Mean values and standard deviations of phyllosilicate "crystallinity" indices ($\Delta 2°\Theta$) of the granite samples

Goniometer speed	XRD parameters	Whole rock	2μm fractions
2°/min	IC(002)	0.220 (n=1)	0.369±0.066 (n=6)
	ChC(001)	0.307 (n=1)	0.332±0.045 (n=7)
	ChC(002)	0.276±0.021 (n=11)	0.282±0.026 (n=12)
1/2°/min	IC(002)	–	0.300±0.049 (n=5)
	ChC(001)	–	0.226±0.025 (n=9)
	ChC(002)	0.229±0.012 (n=12)	0.246±0.024 (n=11)

ROCK CHEMISTRY

Chemical analyses of nine samples from the studied area are given in Table IV. It is clear that the acidic garnet-free rocks are chemically different from the garnet-rich ones. These differences are largely attributed to the abundance of the garnet and the presence of metasediment xenoliths in the latter ones. In general, these plagiogranites are characterized by high silica and very low K_2O percent. However, the silica content in samples No. 34_g, 34_{g+e}, and 35 are relatively very low resulting from an enrichment in garnet, furthermore, sample No. 36 has high K_2O percent, showing normative Or (about 20) more than normative Ab (about 18) resulting from the effect of metasediments.

In the An-Ab-Or diagram (*Fig. 3*) the analyses plot on the low pressure of feldpsar field. In the same diagram, most of samples are situated in the tonalite or trondhjemite field, except sample No. 36 which is located in the granite field, showing high K_2O percent.

The $(F0+0.9Fe_2O_3)$-(K_2O+Na_2O)-MgO diagram (AFM diagram) is used to show the differentiation trends leading to the oceanic plagiogranites. The oceanic leucocratic rocks usually follow the tholeiite trend (COLEMAN and PETERMAN 1975). The analyzed rocks show more or less the same trend, whereas sample Nos. 34_g, 34_{g+e} and 35 situate nearer to the $(FeO+0.9Fe_2O_3)$ corner showing lower total alkali and higher total iron contents (*Fig. 4*).

A plot of $SiO_2\%$ against $K_2O\%$ (*Fig. 5*) displays that most of garnet-poor granite samples fall within the oceanic plagiogranite field, except sample No. 5a which plots outside the field having lower SiO_2 and K_2O contents, that may be due to contamination with basic rocks. On the other hand, samples Nos. 34_g, 34_{g+e} and 35 plot near to the field of subalkaline oceanic basalts and gabbros due to the presence of garnet,

TABLE 3

Some phyllosilicate "crystallinity" indices measured on airdried (AD) and ethylene glycol-solvated mounts of the <2μm fractions of granite samples

goniometer speed		2°/min						1/2°/min					
sample		1	5b	5d	34b	35	averages	1	5b	5d	34b	35	averages
IC(002)	AD	0.385	0.437	–	–	–	0.411 (n=2)	0.291	–	–	0.375	0.272	0.313 (n=3)
	EG	0.386	0.323	0.353	0.353	0.350	0.353±0.022	0.315	–	–	0.303	0.269	0.296 (n=3)
ChC(001)	AD	0.295	–	0.350	0.351	0.394	0.348±0.041	0.230	0.257	0.239	0.266	0.262	0.251±0.016 (n=5)
	EG	0.284	0.376	0.299	0.311	0.350	0.324±0.038 (n=5)	–	–	–	–	0.250	0.250 (n=1)
ChC(002)	AD	0.308	0.300	0.313	0.266	0.320	0.301±0.021 (n=5)	0.248	0.251	0.295	0.248	0.236	0.256±0.023 (n=5)
	EG	0.281	0.289	0.281	0.282	0.282	0.283±0.003 (n=5)	0.248	0.248	0.257	0.225	0.249	0.245±0.012 (n=5)

Table 4

Chemical analyses of the studied rocks (weight per cent).

	1	5a	5b	5d	34g	34g+e	34d	35	36
SiO_2	69.99	58.72	66.42	68.30	49.18	51.36	61.43	49.97	66.53
TiO_2	0.29	0.52	0.43	0.28	0.56	0.54	0.33	1.18	0.52
Al_2O_3	13.86	15.63	14.46	13.65	16.60	16.46	18.10	16.22	16.19
Fe_2O_3	0.49	1.08	0.61	0.60	3.67	3.23	1.69	2.25	1.52
FeO	2.63	5.50	4.29	2.90	17.80	14.02	3.37	15.68	3.96
MnO	0.05	0.19	0.11	0.05	2.25	1.65	0.71	1.48	0.22
MgO	0.72	1.44	0.89	0.80	2.35	2.21	0.08	2.94	1.63
CaO	2.22	7.32	3.31	3.78	2.52	2.74	3.36	2.56	0.31
K_2O	0.48	0.05	0.17	0.14	0.24	0.32	0.53	0.32	3.35
Na_2O	6.28	5.37	6.47	6.45	2.64	3.70	7.89	3.06	2.15
$-H_2O$	0.15	0.09	0.09	0.07	0.08	0.10	0.07	0.12	0.12
$+H_2O$	1.69	2.76	1.96	1.61	1.96	2.43	1.61	3.18	3.30
CO_2	0.03	0.83	0.39	0.79	0.37	0.37	0.42	0.04	0.00
P_2O_5	0.41	0.23	0.13	0.08	0.15	0.14	0.11	0.50	0.06
Sum	99.29	99.73	99.73	99.50	100.37	99.27	99.70	99.50	99.86
C I P W norm									
q	25.19	10.83	18.71	22.46	10.45	9.35	5.30	9.97	35.75
c	0.02	–	–	–	8.63	6.24	– 7.47	8.61	
or	2.84	0.30	1.00	0.83	1.42	1.89	3.13	1.89	19.80
ab	53.14	45.44	54.75	54.58	22.34	31.31	66.77	25.89	18.19
an	8.15	18.40	9.91	7.88	9.18	10.34	12.41	9.18	1.15
cpx	–	9.52	2.74	4.55	–	–	0.79	–	–
di	–	3.05	0.74	1.54	–	–	0.03	–	–
hd	–	6.46	2.00	3.01	–	–	0.76	–	–
opx	5.83	7.44	7.68	4.14	38.77	30.76	5.35	35.06	9.63
en	1.79	2.17	1.87	1.28	5.85	5.50	0.18	7.32	1.06
fs	4.04	5.27	5.80	2.86	32.92	25.26	5.16	27.74	5.57
mt	0.71	1.57	0.88	0.87	5.32	4.68	2.45	3.26	2.20
il	0.55	0.99	0.82	0.53	1.06	1.03	0.63	2.24	0.99
hm	–	–	–	–	–	–	–	–	–
ru	–	–	–	–	–	–	–	–	–
ap	0.97	0.54	0.31	0.19	0.36	0.33	0.26	1.18	0.14
cc	0.07	1.89	0.89	1.80	0.84	0.84	0.96	0.09	–

g = garnet-rich part,,   g+e= garnet+enclave-rich part

An

• garnet-poor rocks⎱
■ garnet-rich rocks⎰ (this work)

Tonalite
Adamellite

Granite

Trondhjemite

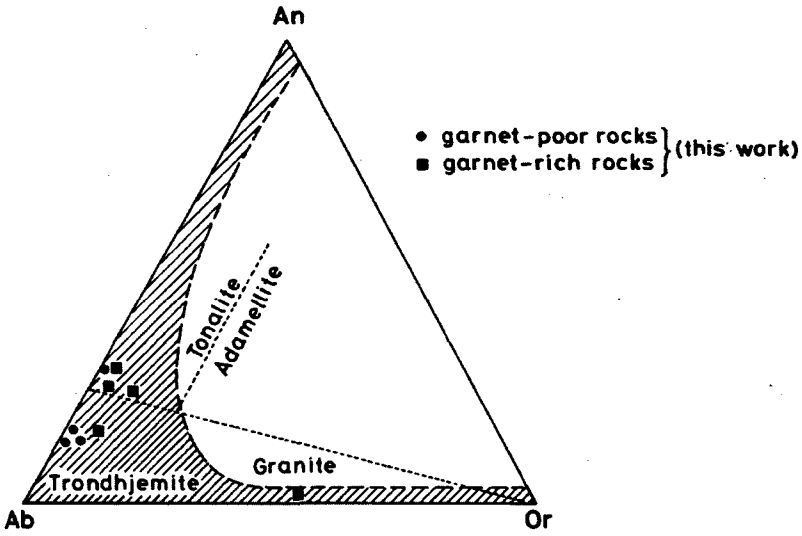Ab   Or

Fig. 3.
Ab-An-Or triangular diagram for plagiogranitic rocks. Shaded area: low pressure feldspar field (<5Kbar); dotted lines separate different rock types on the basis of their feldspar contents (O'CONNER 1965)

$FeO + 0.9 \ Fe_2O_3$

• garnet-poor rocks⎱
■ garnet-rich rocks⎰ (this work)

Parent magma,
Troodos

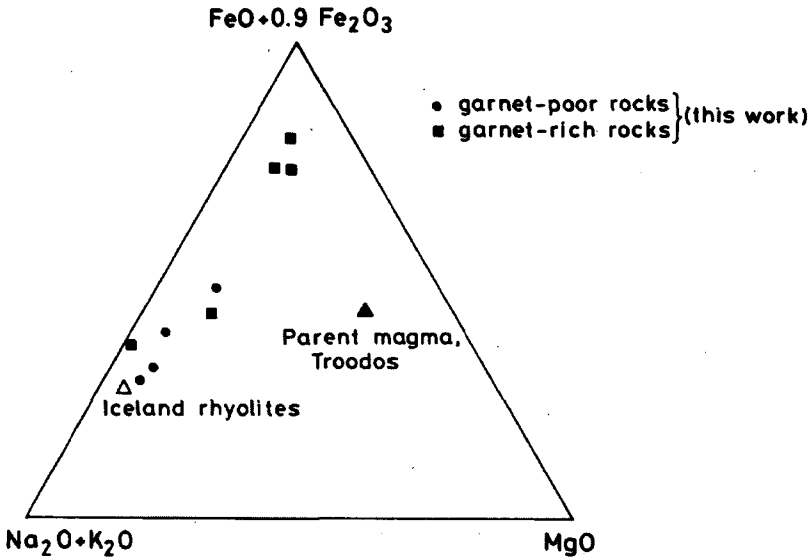Iceland rhyolites

$Na_2O + K_2O$ MgO

Fig. 4.
AFM diagram showing plagiogranites of the studied area comparing with the Troodos plagiogranites. Parent magma of Troodos ophiolite and Iceland rhyolites are also plotted (COLEMAN and PETERMAN 1975)
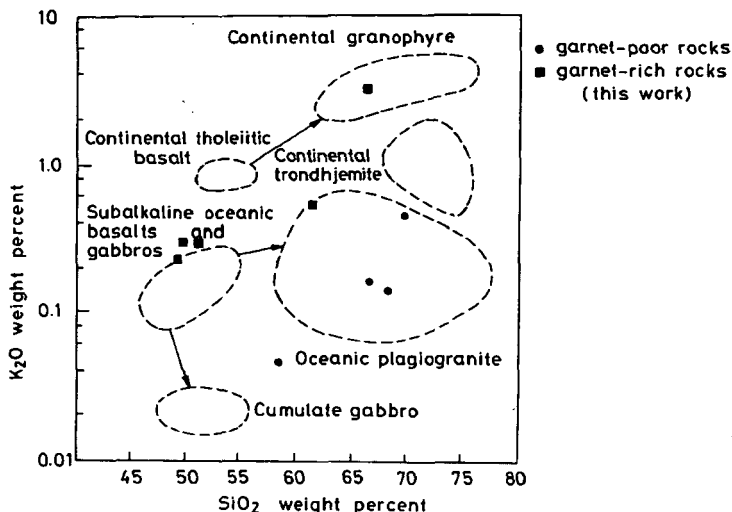
55

Fig. 5.
A plot of SiO₂ againts K₂O for Szarvaskő plagiogranites and common rocks of equivalent composition for comparison (COLEMAN and PETERMAN 1975).

while sample No. 36 lies within continental granophyre field as a result of metasediment contamination.

According to COLEMAN and PETERMAN (1975), the ($K_2O \times 100$)/($K_2O + Na_2O$) ratio is about 49% for granophyres, but that of oceanic leucoratic rocks is about 5%. The calculated ratios for the analyzed samples mostly range from 0.9 to 9.47, so that the higher values are shown by garnet-rich rocks (in sample No. 36 this ratio reaches 60.9%).

In fact, Szarvaskő plagiogranites contain higher content of Rb (using analytical data of BALLA et al. 1983, DOWNES et al. 1990), which may be due to submarine alteration or country rock contamination. In addition, the REE patterns show significant enrichment of light REE, strong negative Eu anomalies and slight heavy REE depletion when compared with the REE patterns of Troodos oceanic plagiogranites (KAY and SENECHAL 1976).

$^{87}Sr/^{86}Sr$ ratios of Szarvaskő plagiogranites (BALLA et al. 1983; DOWNES et al. 1990) range from 0.70669–+0.00001 0.70754–+0.00001 which is obviously higher than those of Troodos (reach 0.7060; COLEMAN and PETERMAN 1975). This enrichment in isotope ratios may be attributed to rock alteration.

MINERAL CHEMISTRY

The chemistry of rock-forming minerals from plagiogranites of Szarvaskő have been determined by electron microprobe analysis (Table V).

The plagioclase of plagiogranite is characterized by high sodium and very low calcium content (average An-content vary from 1 to 4%). On the other hand, the plagioclase in metasediments that form enclaves in the granite, shows relatively higher anorthite content with an average of An_{23}. The latter exhibits notable composional

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zoning with increasing Ca-content towards the grains' rims. Furthermore, the potassium content of plagioclase is very low.

Chlorite usually occurs as replacement of original biotite or as filling in veins and cracks of plagioclases. Chlorite of metasediment xenoliths is very similar to those occurring in acidic rocks. The only detectable difference is an increasing in potassium-content (average K of 0.247) in the former case, comparing with the latter case, where the average K varies from 0.011 to 0.045.

Prehnite generally exists an infilling veins and cracks and also as alteration products of both biotite and plagioclase. Two types of prehnites could be observed, the first is Fe-rich with average Fe_2O_3 content of 4.65, whereas the second type is Fe-poor with average Fe_2O_3 content of 0.51.

Epidote analyses refer to considerable amounts of REE. In fact, both epidotes and allanites were formed after biotite and plagioclase, respectively.

The analyzed garnets from garnet-poor granites and from garnet-rich rocks exhibit the same chemical composition. They are almandine-rich, with composition ranges of Alm_{75-76} Pyr_{7-11} $Spess_{9-14}$ $Gross_{4-5}$. Garnets frequently show slight, irregular chemical zoning without significant change in composition from core to rim. The distribution of Mg, Fe, Mn and Ca in the individual zoned crystals from both rock types are shown in *Figs. 6* and *7*. From *Fig 6*. it is obvious that Fe is the only element displaying an increase towards the rim, while Mg, Mn and Ca vary in irregular patterns. In *Fig. 7a*, both Fe and Mn increase towards the rims, coupled with a decrease in Mg content, while Ca exhibits slight change. In an other grain *(Fig. 7b)* Fe changes against Ca, while both Mg and Mn do not show significant difference between the core and rim. In the Alm-Pyr-(Spess+Gross) triangular diagram, the garnets of this work are located within the field of metapelites *(Fig. 9)*. Moreover, using the $(Mg^{++}+Fe^{++})-Mn^{++}-Ca^{++}$ triangular diagram *(Fig. 8)*, these garnets fall between the fields of igneous silicic rocks and the representative points of the garnet from the metamorphic rocks which surround the Vedrette di Ries pluton (BELLINI *et al.* 1979).
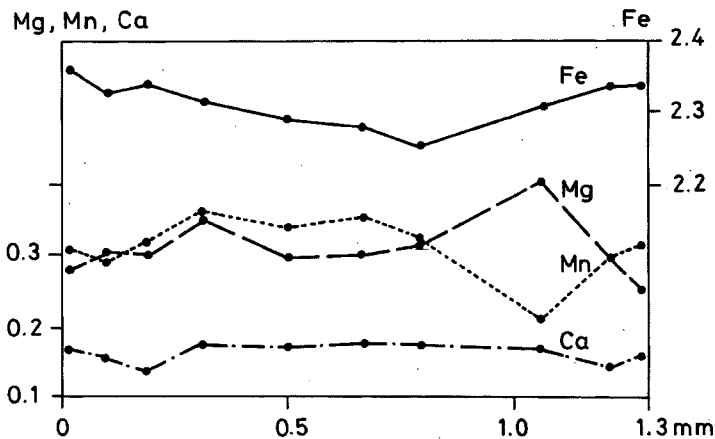

Fig. 6.
Zoning of a garnet grain from garnet-poor granites of Szarvaskő.
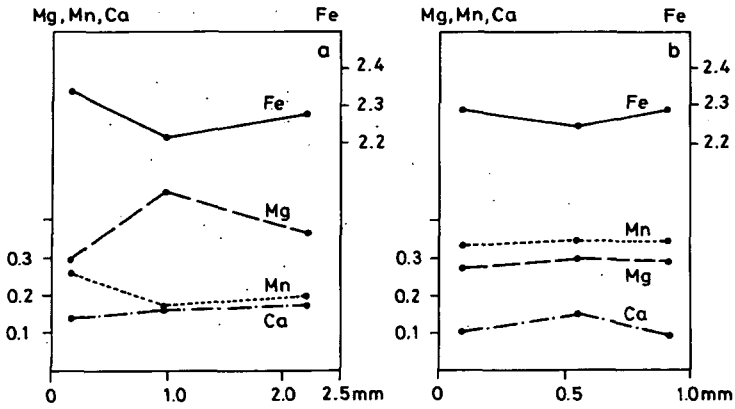
Fig. 7 (a–b)
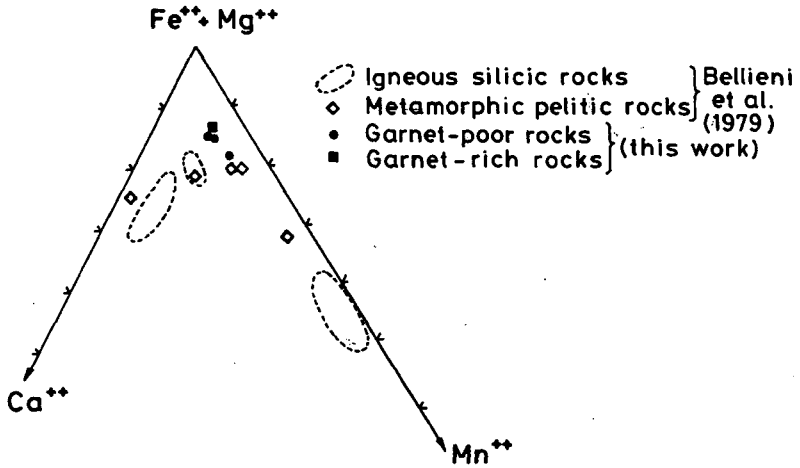Zoning of garnet grains from garnet-rich granites of Szarvaskő

Fig. 8.
A plot of (Fe^{++}+Mg^{++})–Mn^{++}–Ca^{++} (after BELLIENI *et al.* 1979) illustrating the composition of the studied garnet

TABLE 5

Electron microprobe data of rock minerals in weight percent and number of cations per unit cell
plagioclase analyses

sample	1	5b	5d	34b	34e	34d
n	4	4	4	4	3	4
SiO_2	67.61±0.96	67.21±0.26	69.16±2.13	66.66±1.47	62.31–60.48: 65.68**	68.23±0.33
Al_2O_3	20.46±0.85	20.25±0.12	19.16±1.48	20.57±0.49	24.05–21.91: 25.59	20.10±0.21
CaO	00.60±0.64	0.32±0.11	0.15±0.10	1.10±0.30	4.77– 2.04: 6.32	0.11±0.01
Na_2O	11.38±0.49	11.03±0.64	11.33±0.70	10.58±0.67	8.79– 7.92: 10.34	2.20±0.27
K_2O	0.11±0.04	0.56±0.58	0.09±0.06	0.23±0.14	0.16– 0.13: 0.22	0.06±0.02
Total	99.68±0.26	99.36±0.28	100.15±0.24	99.13±1.92	100.08–99.69:100.62	100.69±0.50
numbers of cations on the basis of 8 (0)						
Si	2.948±0.039	2.959±0.004	3.008±0.077	2.196±0.039	2.753–2.884:2.672**	2.965±0.011
Al	1.051±0.041	1.051±0.009	0.996±0.080	1.069±0.014	1.257–1.332:1.121	1.029±0.011
Ca	0.029±0.030	0.015±0.005	0.007±0.005	0.037±0.023	0.226–0.299:0.096	0.005±0.002
Na	0.962±0.043	0.942±0.053	0.956±0.062	0.905±0.058	0.752–0.880:0.682	1.028±0.020
K	0.006±0.002	0.031±0.033	0.005±0.003	0.013±0.008	0.009–0.012:0.008	0.003±0.001
Total	4.996±0.011	4.999±0.015	4.971±0.068	4.939±0.111	4.992–4.980:5.004	5.031±0.016
	An_3	An_2	An_1	An_4	An_{23}	An_1

e = enclave metasediment   n = number of analyses   ** the highest and the lowest values

TABLE 5 contd

Garnet analyses

	1	5a	5b	34b
n	6	5	10	6
SiO_2	36.18±0.29	37.04±0.28	36.04±0.33	37.03±0.17
Al_2O_3	21.04±0.17	20.80±0.13	20.97±0.20	20.80±0.23
FeO	32.50±1.91	33.41±0.34	33.91±0.52	33.65±0.56
MgO	1.75±0.50	2.26±0.09	2.62±0.29	2.72±0.70
MnO	5.95±3.17	4.41±0.38	4.22±0.63	3.97±1.10
CaO	1.46±0.37	1.79±0.05	1.86±0.16	1.56±0.42
Total	98.88±1.85	99.71±0.05	99.61±0.31	99.73±0.38
numbers of cations				
Si	2.975±1.014	3.007±0.008	2.942±0.020	2.995±0.017
Al	2.039±0.010	1.990±0.009	2.018±0.018	1.982±0.015
*Fe^{2+}	2.235±0.121	2.268±0.021	2.315±0.032	2.276±0.041
Mg	0.214±0.016	0.273±0.010	0.318±0.034	0.329±0.080
Mn	0.415±0.221	0.303±0.026	0.292±0.045	0.272±0.075
Ca	0.128±0.033	0.156±0.004	0.163±0.014	0.135±0.035
Total	8.006±0.015	7.998±0.006	8.048±0.013	7.988±0.011
	Pyr_7	Pyr_9	Pyr_{10}	Pyr_{11}
	Alm_{76}	Alm_{76}	Alm_{75}	Alm_{76}
	$Spess_{14}$	$Spess_{10}$	$Spess_9$	$Spess_9$
	$Gross_4$	$Gross_5$	$Gross_5$	$Gross_4$

* total iron calculated as FeO and Fe^{2+}, respectively   n = number of analyses
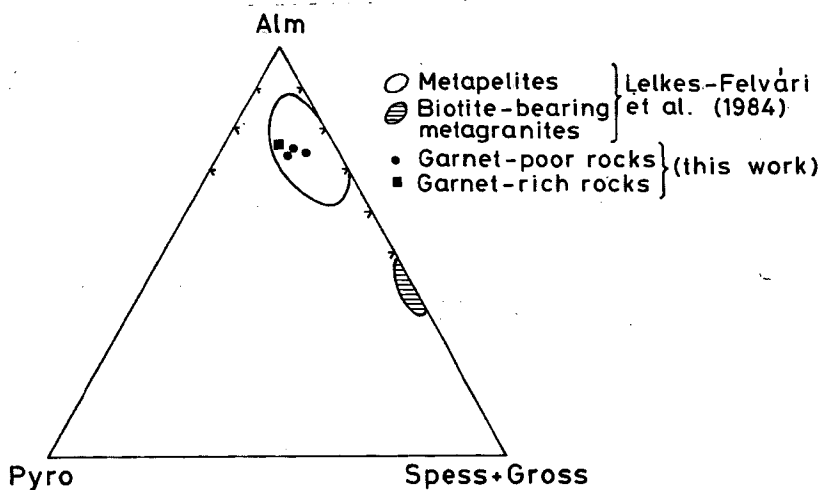
Fig. 9.
Alm-Pyr-(Spess+Gross) diagram of LELKES-FELVÁRI *et al.* (1984) showing the distribution of the studied garnet

TABLE 5. contd

Chlorite analyses

sample	1	5b	5d	34b	34e	34d
n	9	7	4	4	2	4
SiO_2	27.21±0.72	24.45±0.88	26.33±0.83	26.20±0.68	26.94–26.81:27.07**	25.69±1.45
Al_2O_3	17.64±0.72	19.83±1.08	17.60±0.60	17.91±0.44	17.89–17.67:18.11	18.79±1.69
*FeO	31.54±2.24	36.18±1.41	31.51±0.27	36.06±2.36	33.13–32.80:33.45	34.94±1.20
MgO	8.94±1.91	7.02±0.67	8.56±0.34	6.73±1.09	6.48– 6.46: 6.50	6.72±1.69
MnO	0.69±0.34	0.38±0.14	0.55±0.12	0.50±0.02	0.49– 0.48: 0.49	0.42±0.07
CaO	0.750.34	0.24±0.19	0.67±0.14	0.52±0.43	0.72– 0.51:0.93	0.94±1.18
Na_2O	0.04±0.01	0.03±0.01	0.04±0.04	0.05±0.04	0.04– 0.03: 0.05	0.05±0.04
K_2O	0.16±0.14	0.04±0.03	0.12±0.07	0.14±0.09	0.88– 0.55: 1.21	0.07±0.03
TiO_2	1.54±0.57	0.27±0.34	1.02±0.26	0.95±0.90	1.59– 1.53: 1.65	0.49±0.62
Total	88.52±0.87	88.45±0.64	86.41±0.71	89.05±0.45	88.16–87.41:88.89	88.11±0.91
numbers of cations on the basis of 28 (0)						
Si	5.898±0.129	5.445±0.178	5.864±0.128	5.786±0.078	5.938– 5.927: 5.948**	5.704±0.235
Al^{IV}	2.102±0.129	2.555±0.178	2.136±0.128	2.214±0.078	2.062– 2.073: 2.052	2.296±0.235
Al^{VI}	2.406±0.092	2.651±0.115	2.481±0.109	2.448±0.086	2.586– 2.569: 2.602	2.626±0.293
*Fe	5.722±0.446	6.740±0.289	5.869±0.096	6.667±0.550	6.106– 6.086: 6.126	6.498±0.487
Mg	2.887±0.590	2.330±0.213	2.844±0.124	2.211±0.355	2.129– 2.122: 2.135	2.226±0.557
Mn	0.127±0.600	0.073±0.026	0.103±0.022	0.094±0.004	0.090– 0.089: 0.091	0.080±0.014
Ti	0.251±0.094	0.046±0.057	0.171±0.044	0.156±0.146	0.264– 0.252: 0.276	0.082±0.104
Ca	0.137±0.078	0.056±0.046	0.161±0.032	0.121±0.100	0.170– 0.119: 0.220	0.221±0.273
Na	0.018±0.005	0.014±0.006	0.019±0.018	0.021±0.015	0.018– 0.014: 0.022	0.023±0.016
K	0.045±0.038	0.011±0.009	0.035±0.019	0.040±0.023	0.247– 0.156: 0.337	0.021±0.009
Total	19.621±0.125	19920±0.095	19.684±0.077	19.758±0.125	19.608–19.555:19.661	19.776±0.157

e = enclave metasediment   n = number of analyses   * total iron calculated as FeO and Fe^{2+}, respectively   ** the highest and the lowest values

TABLE 5. contd

Prehnite analyses

sample	1	5b dark	5d light	5d	34b	34d
n	4	4	4	4	3	5
SiO_2	43.91±0.17	43.47±0.24	42.69±0.32	43.10±0.50	44.49–44.00:44.84**	42.81±0.54
Al_2O_3	23.39±0.96	24.15±0.17	21.21±0.79	23.74±0.11	23.64–23.44:23.77	22.65±1.48
*Fe_2O_3	0.94±0.62	0.51±0.10	4.65±0.90	1.44±0.73	0.25– 0.21: 0.31	1.73±1.74
MgO	0.04±0.08	0.01±0.01	0.01±0.01	0.10±0.05	0.00– 0.00	0.01±0.02
MnO	0.03±0.01	0.02±0.02	0.00±0.00	0.01±0.01	0.02– 0.00: 0.04	0.03±0.02
CaO	26.97±0.16	27.61±0.22	27.11±0.24	27.39±0.24	26.85–26.20:27.26	27.26±0.70
Na_2O	0.13±0.11	0.04±0.10	0.03±0.01	0.02±0.01	0.35– 0.10: 0.84	0.06±0.07
K_2O	0.02±0.01	0.01±0.01	0.01±0.01	0.03±0.01	0.04– 0.02: 0.05	0.03±0.05
TiO_2	0.05±0.00	0.07±0.05	0.04±0.03	0.06±0.02	0.02– 0.00: 0.03	0.02±0.03
Total	95.49±1.11	95.89±0.41	95.75±0.48	95.88±1.09	95.66–95.03:96.20	95.59±1.37
numbers of cations on the basis of 22 (O)						
Si	6.059±0.052	5.977±0.011	5.970±0.006	5.947±0.024	6.107– 6.087: 6.118**	6.000±0.019
Al	3.803±0.117	3.914±0.009	3.498±0.120	3.861±0.044	3.826– 3.815: 3.840	3.740±0.213
*Fe^{3+}	0.098±0.065	0.053±0.011	0.496±0.099	0.150±0.074	0.026– 0.022: 0.032	0.183±0.187
Mg	0.009±0.016	0.002±0.002	0.003±0.003	0.021±0.011	0.000– 0.000	0.002±0.001
Mn	0.004±0.001	0.002±0.002	0.001±0.001	0.001±0.001	0.002– 0.000: 0.005	0.002±0.001
Ca	3.989±0.029	4.70±0.023	4.064±0.008	4.051±0.061	3.951– 3.849: 4.018	4.095±0.065
Na	0.035±0.030	0.011±0.002	0.008±0.002	0.006±0.004	0.092– 0.026: 0.223	0.017±0.018
K	0.004±0.001	0.002±0.001	0.003±0.001	0.006±0.002	0.006– 0.004: 0.009	0.006±0.009
Ti	0.006±0.011	0.007±0.005	0.004±0.003	0.005±0.003	0.002– 0.000: 0.003	0.002±0.003
Total	14.005±0.020	14.038±0.010	14.038±0.005	14.048±0.020	14.013–13.976:14.065	14.049±0.025

* total iron calculted as Fe_2O_3 and Fe^{3+}, respectively   ** the highest and the lowest values   **n** = number of analyses

TABLE 5. contd.

Epidote analyses

sample	1	34b	34d
n	2	2	3
SiO_2	37.41–37.19:37.63**	36.36–38.03:38.69**	37.87–37.33:38.42**
Al_2O_3	24.50–24.35:24.65	24.58–24.45:24.71	23.98–23.46:24.32
*Fe_2O_3	11.72–11.62:11.80	12.19–11.83:12.54	10.04– 5.81:12.22
MgO	0.05– 0.03: 0.07	0.03– 0.00: 0.06	0.80– 2.20: 0.04
MnO	0.10– 0.06: 0.13	0.31– 0.27: 0.34	0.24– 0.05: 0.22
CaO	23.32–23.03:23.59	23.47–23.19:23.75	23.08–22.82:23.36
Na_2O	0.04– 0.04: 0.04	0.06– 0.04: 0.07	0.01– 0.00: 0.02
K_2O	0.00– 0.00	0.03– 0.02: 0.04	0.00– 0.00
TiO_2	0.21– 0.19: 0.22	0.30– 0.23: 0.36	0.61– 0.00: 1.43
Total	97.33–97.13:97.52	99.32–99.09:99.55	96.54–92.64:98.55

	numbers of cations on the basis of 25 (O)		
Si	5.962– 6.938: 8.985**	5.997– 5.967: 6.026**	6.052– 6.140: 5.979**
Al	4.602– 4.565: 4.639	4.529– 4.522: 4.536	4.518– 4.368 4.684
*Fe^{3+}	1.408– 1.395: 1.420	1.436– 1.388: 1.483	1.201– 0.719: 1.446
Mg	0.012– 0.007: 0.017	0.007– 0.000: 0.014	0.194– 0.009: 0.539
Mn	0.013– 0.008: 0.018	0.041– 0.036: 0.045	0.021– 0.007: 0.029
Ca	3.982– 3.942: 4.022	3.933– 3.900: 3.965	3.955– 3.843: 4.066
Na	0.012– 0.012: 0.012	0.017– 0.012: 0.021	0.003– 0.000: 0.006
K	0.000– 0.000	0.006– 0.004: 0.008	0.000– 0.000
Ti	0.025– 0.023: 0.026	0.035– 0.027: 0.042	0.072– 0.000: 0.170
Total	16.015–16.012:16.017	15.998–15.994:16.002	16.017–15.942:16.161

* total iron calculted as Fe_2O_3 and Fe^{3+}, respectively   ** the highest and the lowest values
n = number of analyses

TABLE 5. contd.

Allanite analyses

sample	1	5d
n	3	3
SiO$_2$	33.69– 32.98: 34.36**	34.71–34.50: 34.72**
Al$_2$O$_3$	22.51– 22.17: 22.82	23.01–22.76: 23.83
*Fe$_2$O$_3$	11.05– 10.23: 11.92	11.52–11.31: 11.94
MgO	0.35– 0.08: 11.92	0.11– 0.08: 0.12
CaO	14.53– 13.89: 15.75	15.53–14.80: 16.62
MnO	0.00– 0.00	0.34– 0.27: 0.38
TiO$_2$	0.58– 0.20: 1.47	0.00– 0.00
Y$_2$O$_3$	0.56– 0.31: 1.02	4.17– 4.40: 4.61
La$_2$O$_3$	2.96– 2.15: 3.39	1.31– 0.85: 1.54
Ce$_2$O$_3$	9.10– 6.44: 1.64	4.30– 2.68: 5.18
Pr$_2$O$_3$	1.00– 0.80: 1.25	0.41– 0.34: 0.49
Nd$_2$O$_3$	3.64– 3.22: 3.82	1.94– 1.33: 2.33
Sn$_2$O$_3$	0.78– 0.82: 0.70	0.68– 0.45: 0.92
Eu$_2$O$_3$	0.08– 0.08: 0.09	0.02– 0.00: 0.05
Gd$_2$O$_3$	0.42– 0.41: 0.44	0.50– 0.46: 0.58
Dy$_2$O$_3$	0.19– 0.12: 0.28	0.72– 0.65: 0.80
Ar$_{23}$	0.00– 0.00	0.37– 0.28: 0.43
Yb$_2$O$_3$	0.00– 0.00	0.37– 0.24: 0.51
Total	101.58–100.31:103.08	100.12–98.71:101.11

numbers of cations on the basis of 25 (O)

	1	5d
Si	5.731– 5.708: 5.774**	5.816– 5.783: 5.834**
Al	4.515– 4.466: 4.553	4.545– 4.483: 4.631
*Fe^{3+}	1.427– 1.296: 1.520	1.455– 1.428: 1.503
Mg	0.088– 0.021: 0.181	0.027– 0.021: 0.030
Ca	2.646– 2.518: 2.835	2.788– 2.665: 2.954
Mn	0.000– 0.000	0.048– 0.038: 0.053
Ti	0.074— 0.009: 0.186	0.000– 0.000
Y	0.050– 0.028: 0.091	0.371– 0.310: 0.411
La	0.186– 0.133: 0.217	0.080– 0.052: 0.095
Ca	0.567– 0.396: 1.674	0.263– 0.164: 0.317
Pr	0.062– 0.049: 0.079	0.025– 0.021: 0.030
Nd	0.221– 0.193: 0.237	0.116– 0.080: 0.139
Sm	0.046– 0.041: 0.50	0.039– 0.026: 0.053
Eu	0.005– 0.005: 0.005	0.001– 0.000: 0.003
Gd	0.023– 0.023: 0.024	0.028– 0.025: 0.032
– Dy	0.011– 0.007: 0.016	0.039– 0.035: 0.043
Yb	0.000– 0.000	0.019– 0.015: 0.022
Total	15.680–15.636:15.751	15.644–15.622:15.685

n = number of analyses
* total iron calculted as Fe$_2$O$_3$ and Fe^{3+}, respectively
** the highest and the lowest values

1. The primary nature of plagiogranites

The plagiogranites of the Szarvaskő area are usually occur in small amounts as nests within the gabbroic rocks of the ophiolitic sequence, showing sharp contact with them.

Mineralogically, the investigated rocks are made of quartz and unzoned albitized plagioclase, with minor biotite which were completely replaced by secondary minerals. In comparison with the oceanic plagiogranites of Troodos (COLEMAN and PETERMAN 1975), the Szarvaskő plagiogranites differ in the presence of unzoned plagioclase with albitic compositions, and in the lack of pyroxene and amphibole (the primary ferromagnesian minerals are represented only by biotite). BAILEY and BLAKE (1974) described zoned plagioclase phynocrysts (An_{50-62}) in a matrix of albite and quartz in the extrusive silicic rocks, while in the intrusive ones referred to the presence of albitized plagioclase phenocrysts. This may be explained by the post magmatic replacement of early formed calcic plagioclase. Moreover, the above authors notes small amounts of biotite in addition to amphibole in coarser varieties (see also GILLULY 1933).

The major element abundances in Szarvaskő plagiogranites is similar to those of oceanic ones showing high silica, moderate alumina, low total Fe+Mg, extremely low K_2O contents. When comparing with the data of COLEMAN and PETERMAN (1975), normative Or is usually less than 4 except sample No. 36 which contains metasediment xenoliths showing high normative Or (~20%), whereas normative An is too small, rarely reaching 18% (An_{21-61} in Troodos plagiogranites), that may be resulted from albitization of early formed plagioclase. In general, Szarvaskő rocks usually occupy the oceanic plagiogranitic field (*Figs. 3* and *5*).

The Rb content in the studied rocks is relatively high (reaches 15 pmm). Comparing with other oceanic silicic rocks the enrichment of Rb may be due to submarine alteration or due to the contamination by country rocks (DOWNES et al. 1990). In the quartz-monzonitic dikelet of the Central Indian ridge, the concentration of Rb is 84 ppm (ENGEL and FISHER 1975). Significant light REE enrichment of Szarvaskő plagiogranite compared with Troodos plagiogranite may be attributed to country rock contamination. Higher values of $^{87}Sr/^{86}Sr$ may be resulted from alteration as shown by COLEMAN and PETERMAN (1975).

From the above discussion it can be suggested that the Szarvaskő plagiogranites are primary oceanic plagiogranitic rocks crystallized from late differentiated basic melt.

2. The origin of the garnet

The composition of the garnet from garnet-poor rocks is very similar to that of garnet-rich varieties: both being almandine-rich. They occur as hypidiomorphic crystals with slightly resorbed outlines and show varying zoning patterns. In general, Fe content increases and Ca content decreases towards the rims. The inclusions are represented by quartz, ilmenite, sphene and rutile.

In garnet-rich varieties, garnets usually accumulated in the granitic rocks near or at the contact with the metasedimentary enclaves. On the other hand, however garnets do not form contact metamorphic rims in metasedimentary enclaves. They are always dispersed in granitic melt.

According to their chemistry, the investigated garnets are located either within the metapelite field or near to the igneous silicic rock field (*Fig. 8* and *9*).

From the above observations it can be concluded that garnets formed by local assimilation of the country rocks. This conclusion supports the former opinion of BALLA et al. (1983), but contrasts with the metasomatic origin suggested by EMBEY-ISZ-TIN et al. (1985).

3. Post-magmatic alteration of plagiogranite

Most of the primary magmatic minerals of the plagiogranites are obviously altered. Thus, the original composition of plagioclase which crystallized from magmatic melt is not known, as it was completely albitized, presumably by post-magmatic processes. In addition, prehnite and white mica (sericite) were formed, mainly along cleavage surfaces and cracks of plagioclase. Biotite totally altered to chlorite, epidote, prehnite and titanite. It has been shown that the Szarvaskő plagiogranites are characterized by secondary mineral associations commonly including chlorite, epidote, prehnite, albite, calcite, titanite and sericite, while pumpellyite is completely lacking.

As proved by ÁRKAI (1983), the metabasic rocks associated with these plagiogranites were affected by pumpellyite-prehnite-quartz facies metamorphism. The averages of IC(002), ChC(001) and ChC(002) measured on $<2\mu$m fractions, using a goniometer speed of 2°/min, are 0.369 ± 0.066, 0.329 ± 0.042 and 0.282 ± 0.026 $\Delta°28$, respectively. IC(002) average value indicates anchizonal regional metamorphism (see the zone boundaries of KÜBLER 1975) or falls into the zone of deep diagenesis (see ÁRKAI, 1987). On the basis of IC, the temperature must be of ca. 230°C. Moreover, using the ChC(001) and ChC(002) anchizonal boundaries given by ÁRKAI (1991), the chlorite crystallinity of the first and second basal reflections of the studied rocks suggest anchizonal regional metamorphism of temperature of about 280°C which corresponds to the conditions of the pumpellyite-prehnite-quartz facies at low geothermal gradient or to the prehnite-actinolite facies at high geothermal metamorphic gradient (very low pressure facies). Using the chlorite geothermometer of CATHELINEAU (1988) elaborated originally for andesitic rocks and proved later on for metabasic rocks and metasandstones, the calculated temperature values of the metamorphism of the Szarvaskő plagiogranites vary between 276 and 350°C (in average, 298°C). The variations in chemical compositions of certain secondary minerals (prehnite, chlorite) may refer to local changes in equilibrium conditions.

Finally, it may be concluded that the secondary mineral assemblage of the plagiogranite might form in three various metamorphic regimes, i. e., during ocean-floor hydrothermal, burial and regional (dynamothermal) processes, from which the last Cretaceous regional metamorphic event was already proved (ÁRKAI 1983). The superposition of these processes seems to be also possible. Local metasomatic effects resulting from the circulating hydrothermal solutions and/or from the surrounding sedimentary sequence and/or from sea water interactions might also contribute to the formation of secondary minerals. The microstructural features and the mineral assemblage of prehnite + epidote + chlorite + albite + quartz of the plagiogranite alone do not give sufficient evidence either for determining exactly the metamorphic facies or the paleotectonic system of the transformation. Chlorite "crystallinity" values together with the prehnite + epidote + chlorite + albite + quartz assemblage may suggest metamorphic condition corresponding to the low pressure type prehnite-actinolite facies of LIOU et al. (1985) who mentioned that other assemblages characteristic of this facies include epidote + prehnite ± hematite ± magnetite could be recognized in active geothermal systems.

CONCLUSIONS

Based on new petrographic, bulk chemical and mineral chemical data the following conclusions can be drawn on the petrogenesis of the granitic rocks associated with basic-ultrabasic sequence of the Jurassic Szarvaskő complex, Bükk Mountains, Hungary.

1. Major element bulk chemical compositions of the investigated granitic rocks is similar to that of the oceanic plagiogranites, and refer to crystallization from late differentiated basic melt.

2. Almandine-rich garnet found in varying proportions in granite is accumulated near to or at the contacts with metasedimentary (biotite-plagioclase hornfels) enclaves. As no direct spatial connection can be observed between the garnets and the enclaves (garnet are always dispersed in the granitic melt), the formation of garnet is explained by local assimilation of the country rocks in a deeper level and not by direct (contact metamorphic or metasomatic) effects caused by the granitic melts on the enclaves.

3. The primary (magmatic) minerals of the plagiogranite was commonly altered: plagioclase was completely albitised (±prehnite, white mica), biotite was totally altered to chlorite, epidote, prehnite and titanite. The secondary mineral assemblage (quartz, albite, chlorite, epidote, prehnite, white mica, calcite and titanite), illite and chlorite "crystallinity" indices as well as the data obtained by the chlorite-Al" geothermometer refer to low temperature (ca. 280°–300°C) metamorphism. Further studies on the whole ophiolite-like sequence and the surrounding sediments are needed to prove eventual ocean-floor (hydrothermal) or burial metamorphic events which might precede the Cretaceous regional (dynamothermal) effect documented by ÁRKAI (1983).

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