ALPINE POLYPHASE METAMORPHISM OF THE OPHIOLITIC SZARVASKŐ COMPLEX, BÜKK MOUNTAINS, HUNGARY

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ABSTRACT

The Szarvaskő ophiolite-like Jurassic complex (southwestern Bükk Mountains, NE Hungary) shows Alpine metamorphic crystallizations related to two distinct events. The younger (Cretaceous) regional event was formerly evidenced by ÁRKAI (1983). The new result presented here prove an earlier hydrothermal (ocean-floor) metamorphic event resulted from the interaction of seawater-derived fluids with the still hot oceanic crust. Metamorphic grade increases downwards from the prehnite-pumpellyite or zeolite facies (metabasalts), through the greenschist facies (metadiabases), to the greenschist/amphibolite transitional and amphibolite facies (metagabbros). In metagabbros, the scarcity of secondary (metamorphic) clinopyroxene and Mg-Fe amphibole, and the compositional changes of the green metamorphic amphibole from hornblende to actinolitic hornblende suggest that the metamorphic crystallization occurred at temperatures decreasing from about 700 to 400°C. The metamorphic clinopyroxene is always poor in its Al content reflecting subsolidus re-equilibration of magmatic clinopyroxene during ocean-floor hydrothermal metamorphism. However, the metamorphic clinopyroxene reported from other ophiolitic gabbro occurrences is much lower in Al and Ti contents (MÉVEL et al., 1978; GIRARDEAU and MÉVEL, 1982; MÉVEL, 1988; GILLIS, 1995) than that of the Szarvaskő complex. The metamorphic amphibole formed in metadiabases is rather actinolitic in composition, and coexists with albite, chlorite and titanite, indicating temperature below 400°C. The phases formed in metabasalts during hydrothermal metamorphism were modified by subsequent Alpine regional prehnite-pumpellyite facies metamorphism. The younger, low temperature regional (dynamothermal) metamorphism produced prehnite-pumpellyite facies assemblages in meta-igneous rocks, and late diagenetic to low temperature anchizonal alteration in the surrounding sedimentary rocks, as stated earlier by ÁRKAI (1983). The compositional variation of Ca-Al silicate minerals (i.e. prehnite, pumpellyite and epidote) could not be taken as indicative of metamorphic conditions because of their high variance occurrences. However, a large variation in Fe content of these minerals was commonly recorded, and appears to be depend largely on the fluid composition, mode of occurrence and/or on the extent of rock alteration, but is weakly related to bulk rock chemistry. In the low variance assemblages, the average X_{Fe}^{3+} values of coexisting prehnite, pumpellyite and epidote are 0.003, 0.137 and 0.008, respectively. This contrasts with the results of CHO et al. (1986), who found that the X_{Fe}^{3+} values most probable increase in order of prehnite, through pumpellyite to epidote.

INTRODUCTION

The oceanic lithosphere is generally subjected to ocean-floor, hydrothermal metamorphism beneath mid-oceanic ridges. Considering a typical vertical profile of an ophiolitic suite, zeolite facies assemblages commonly occur in the pillow lava followed

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downwards by greenschist facies within the sheeted dike complex. The amphibolite facies is restricted to the upper part of the gabbroic horizon (COLEMAN 1977). As an uncommon exception, granulite facies assemblages were also reported from gabbroic rocks (ISHIZUKA, 1985).

The Mesozoic dismembered ophiolite-like magmatic sequence of the Szarvaskő complex in the Bükk Mountains belongs to the innermost tectonic unit of Western Carpathians, forming the southern part of Gemer-Bükk unit (*Fig. 1*). This unit derived from the Inner Dinarides, and their present position has been related to Tertiary large-scale horizontal displacements (KOVÁCS 1982). The most important geological, tectonic and metamorphic features of the Bükk Mountains and the Jurassic Szarvaskő complex and its surroundings are shown in *Figs. 2* and *3*, respectively.

Previous works have shown that the Szarvaskő complex consists mostly of a series of ultrabasic and basic rocks that had been intruded locally (in the Tóbérc quarry) by cross-cutting acidic dikes and veins, showing sharp contact with the enclosing gabbro (SZENTPÉTERY, 1953; BALLA et al. 1983; DOWNES et al. 1990; SADEK GHABRIAL et al. 1994). The sedimentary environment of the ophiolite complex is represented mainly by terrigenous deep-water clastics. The geochemical data obtained by BALLA et al. (1983), DOBOSI (1986) and DOWNES et al. (1990) indicate oceanic tholeiitic nature of the complex. As regarded by DOWNES et al. (1990), the basic magma of Szarvaskő was contaminated by terrigenous sediments during its evolution in a shallow magma chamber. Mineralogically, the gabbros and related intrusive rocks are very similar, but they differ mainly in the proportions of rock forming minerals (SZENTPÉTERY 1938, 1940, 1953). Ilmenite, titanomagnetite and olivine formed first. These were followed by pyroxenes and plagioclase, including diallage, hypersthene, bronzite and basic to intermediate plagioclase with minor amounts of augite and diopside. In some samples olivine also occur. The appearance of hydrous minerals (brown hornblende and biotite), forming large crystals in the latest stage of magmatic crystallization, suggests that the crystallization of the magma within the intrusive bodies occurred at fairly high P_{H2O} which strongly increased with advancing crystallization (BALLA et al. 1983). Plagiogranite consists chiefly of plagioclase, biotite and quartz with garnet and accessories including apatite, zircon and sphene (EMBEY-ISZTIN et al. 1985; DOWNES et al. 1990). Biotite is totally replaced by chlorite, epidote and prehnite (SADEK GHABRIAL et al. 1994). Diabase is the dominant igneous rock type in the Szarvaskő area. Two main kinds of diabases were recognized, namely holocrystalline-grained and aphanitic-glassy ones (SZENTPÉTERY, 1953). Phenocrysts are rare, represented by intermediate plagioclase. Augite is the common mafic silicate, while hornblende and biotite are usually missing. As to DOWNES et al. (1990), the common minerals in basic rocks of the Szarvaskő complex are plagioclase and clinopyroxene with ilmenite, magnetite and pyrite being present as accessories.

As to ÁRKAI (1983), the Szarvaskő metabasites show prehnite-pumpellyite facies alteration. The adjoining sedimentary formations were subjected to regional (dynamothermal) alteration ranging in grade from late diagenesis up to the low-temperature part of the anchizone. Based on the fairly good coincidence of metamorphic grades established in the metabasic rocks and in their sedimentary environment, the prehnite-pumpellyite facies alteration of metabasic rocks was rendered to the regional metamorphic event, which should have obliterated the products of an eventual (but not proved) earlier metamorphic episode (ÁRKAI 1983).

This paper presents the results of a detailed metamorphic petrological study carried out on the whole profile of the ophiolite-like pile of the Szarvaskő complex, paying special
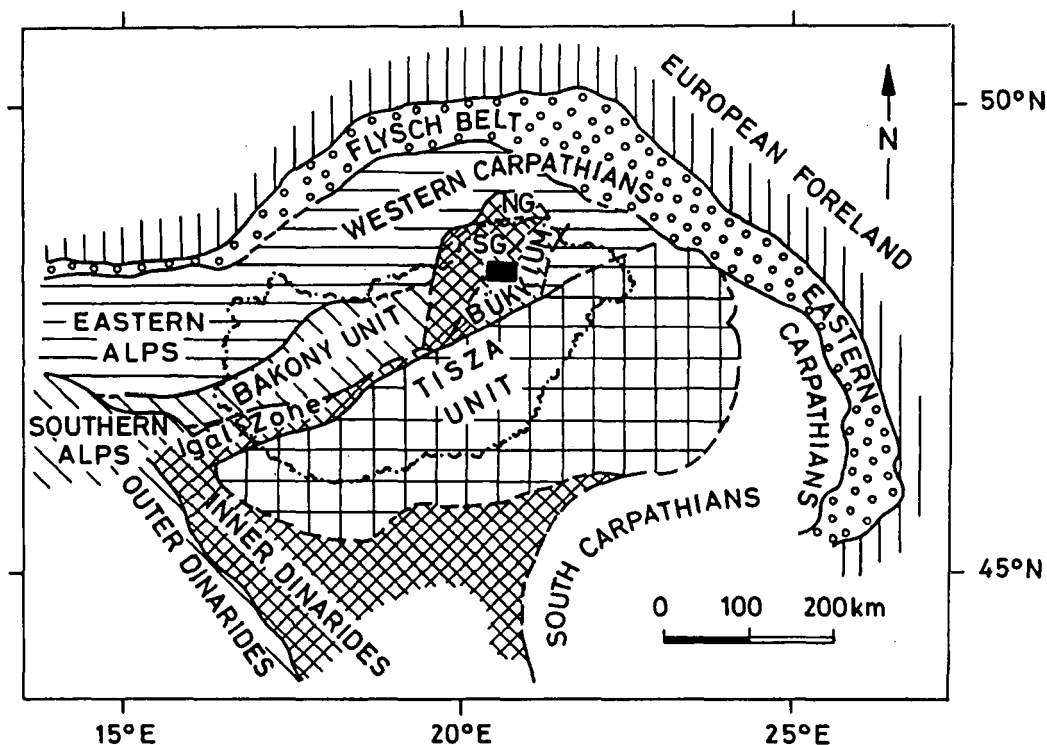
Fig. 1. Position of the Bükkium within the Alp-Carpathian-Dinaric frame.

attention to the distinction between the products of the regional (dynamothermal) and the eventually preceding hydrothermal (ocean-floor) metamorphic events.

MATERIALS AND METHODS

In the Szarvaskő area (*Fig. 3*), samples studied were collected from the Tóbérc quarry (metagranites and metagabbros), the Tardos quarry (metagabbros) and from a series of road cuts, small quarries and valley exposures along a profile, crossing in N-S direction the Szarvaskő complex (metagabbros, metadiabases, metabasalts). Macroscopic and microscopic investigations were done to define rock types, the primary and secondary mineral assemblages, and microstructural features.

For X-ray diffractometric study, the samples were crushed using a jaw crusher, and ground in a mortar mill (Type Pulversette 2, Fritsch) for 3 min. The modal compositions and the illite and chlorite crystallinity indices were evaluated on unoriented powder mounts of the whole rock samples, and on highly oriented (sedimented) <2µm grain size fractions. The latters were separated by differential settling of the rock powder in distilled
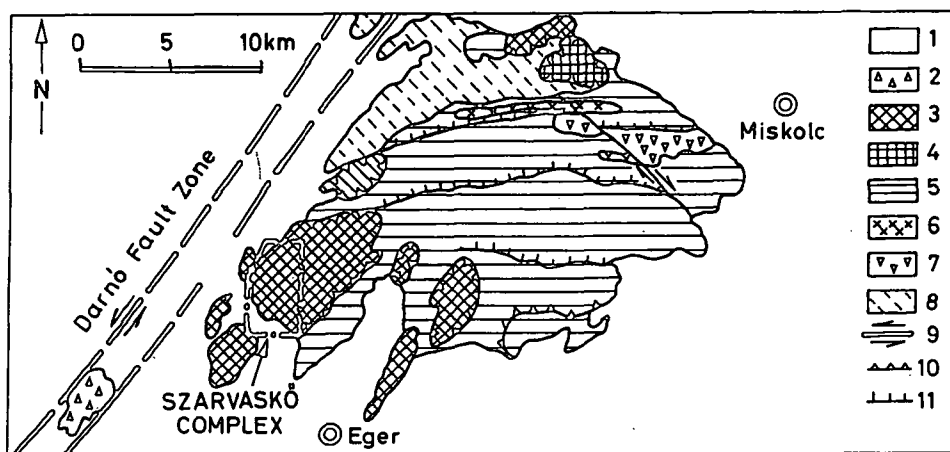
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Fig. 2. Simplified geological and metamorphic map of the Bükk Mountains after CSONTOS (1988), KOVÁCS (1989) and ÁRKAI (1983, 1991). Legend: 1=Tertiary and Quaternary (diagenetic zone), 2=Upper Permian-Mesozoic South Gemer unit (diagenetic zone), 3=Jurassic Szarvaskő-Mónosbél (Little Plateau) Nappe (diagenetic, low-T anchizonal), 4=Triassic Kisfennsík Nape (diagenetic zone), 5-8=Fennsík (Bükk Plateau) Parautochthon, 5=Triassic and Jurassic sedimentary Fms. (anchi-epizonal), 6=Carnic metabasalts (pumpellyite-actinolite facies), 7=Ladinian metaandesites (greenschist facies), 8=Middle Carboniferous-Permian sedimentary Fms. (anchi-epizonal), 9=strike-slip fault, 10=nappe boundary, 11=major imbrications within the nappes.

water, pipetting the suspension onto glass slides and drying at room temperature following the procedure suggested by KÜBLER (1968). In order to determine the expandable clay minerals/components, some of the $<2\mu m$ mounts were solvated by exposing the samples to ethylene glycol vapour in an oven at 80°C for 4 h. The XRD measurements were carried out using a Philips PW-1730 X-ray diffractometer with CuK_α radiation, graphite monochromator, proportional counter, divergence and receiving slits of 1°, at 45 kV, 35 mA, goniometer speeds of 2°/min and "°/min, and time constant of 2 s. The illite and chlorite crystallinity indices, i.e., the full width values measured at half maximum (FWHM) of the 10Å basal reflection of illite-muscovite and the 14 and 7Ĺ basal reflections of chlorite abbreviated as IC(002), ChC(001) and ChC(002) were calibrated against KÜBLER's illite crystallinity scale. The ranges of KÜBLER's anchizone (KÜBLER, 1968, 1990) as measured in the present work are 0.284-0.435 $\Delta°2\Theta$ for IC(002), 0.310-0.427 for ChC(001) and 0.262-0.331 for ChC(002), using $<2\mu m$ grain-size fraction, sedimented, air-dried mounts with a material amount of ca. 2mgcm^{-1} (for details see ÁRKAI, 1991 and ÁRKAI et al. 1995).

The chemical compositions of minerals were determined by a JEOL JCXA-733 electron microprobe at 15 kV and 30 nA, with an electron beam diameter of approximately 5-10 μm. The wave-dispersive spectrometric quantitative analyses were made using the correction methods of BENCE and ALBEE (1968). The standards used for most minerals were orthoclase (Si, Al and K), artificial glass (Fe, Mg and Ca), albite (Na), spessartite (Mn) and rutile (Ti), while those used for garnet were spessartite (Si, Al and Mn), olivine (Mg), hematite (Fe) and Wollastonite (Ca). Relative to the measured values, the standard deviation of the measurements of the given elements or oxides was approximately 2.5%.

The major element chemical analyses were done by a Perkin Elmer 5000 atomic absorption spectrophotometer, using lithium metaborate digestion. Other methods such as gravimetric for SiO_2 and H_2O, permanganometric for FeO and volumetric for CO_2, were also applied.
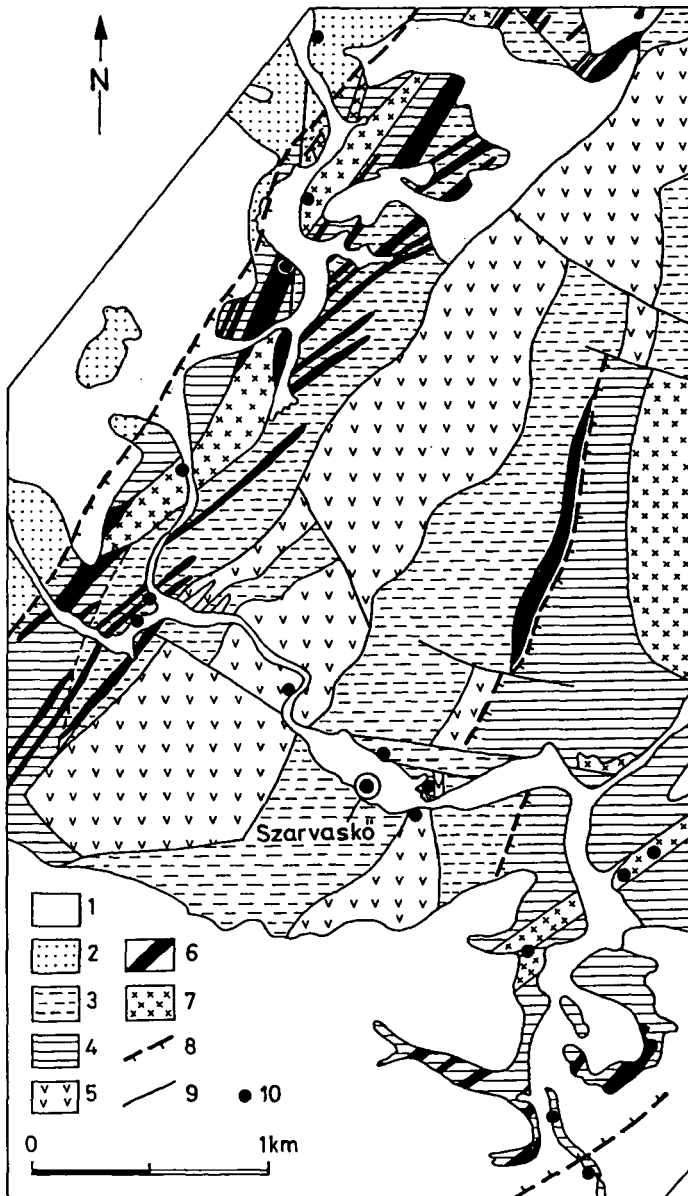
Fig. 3. Geological map of the Jurassic Szarvaskő complex of the Bükk Mountains after BALLA (1982), strongly simplified. Legend: 1=Cenozoic and recent sediments, 2-4=Jurassic sedimentary Fms., 2=Mónosbél Fm.: slate with metachert and sandstone intercalations and limestone olistoliths (Lias-Dogger?), 3=Szarvaskő Fm.: upper, cherty slate (Lias), 4=Szarvaskő Fm.: lower, flysch-like slate (pelitic-silty) with sandstone (Lias), 5-7=igneous rocks, 5=basalt, locally with pillow structure, 6=diabase dykes and sills, 7=cumulate gabbroic intrusives with subordinate ultramafic and acidic differentiates, 8=overthrusting (nappe boundary), 9=fault (within nappe), 10=localities of the investigated sample groups.

The *clastic rocks* associated with the ophiolite complex are characterized by common development of rough fracture cleavage, pencil cleavage, an expression of incipient slaty cleavage and by rare appearance of crenulation cleavage. In the pelitic matrix of some samples, cryptocrystalline elliptic granules possible representing altered microfossils are commonly found. The clastic rocks are characterized by mineral assemblages containing considerable amounts of quartz, plagioclase, white mica and chlorite besides rare amounts of calcite, pyrite, rutile, hematite, chlorite/smectite mixed-layer mineral, smectite and vermiculite. Illite-muscovite (white mica) and chlorite are the dominant phyllosilicates. The quantities of the others phyllosilicates are subordinate. While mixed-layer chlorite/smectite often occurs, smectite and vermiculite were found only sporadically. Cracks are widespread, mostly filled with quartz, calcite and hematite.

The *pillow lavas* are made essentially of fine-grained crystals of plagioclase and clinopyroxene with few sporadic phenocrysts. In some specimens, the matrix shows a notable variation in texture, being glassy in nature. Plagioclase is the dominant mineral occurring usually as acicular or lath-shaped crystals. It is completely albitized: An <5%. The clinopyroxene is of augitic composition, occurring mostly as fine, unaltered crystals. Besides the fine-grained clinopyroxene, few coarse crystals were also recorded. Sometimes, chlorite with traces of titanite replaced partially the clinopyroxene. Ilmenite, hematite, pyrite and calcopyrite are also present as accessories. Other secondary products such as white mica, pumpellyite, Mg-epidote (epidote with MgO content between 0.2 and 1.2%, see CHO. and LIOU 1987), chlorite and rarely chlorite/smectite mixed-layer mineral appear mostly along the fractures cutting across the matrix and the plagioclase crystals, particularly the coarser ones. Veins consist commonly of quartz, albite, calcite, prehnite, pumpellyite, epidote, Mg-epidote and chlorite.

The original porphyritic texture of the fine to medium-grained *metadiabases* is still preserved. The phenocrysts are commonly represented by plagioclase with subordinate clinopyroxene. The groundmass is composed of the same minerals. The plagioclase phenocrysts occur as corroded, fractured crystals. The plagioclase of the groundmass is mostly found as euhedral lath-shaped crystals. The plagioclase is totally recrystallized to albite, showing average anorthite content ranging from 1 to 5%. Traces of pumpellyite are also present as alteration products of plagioclase. The chemistry of clinopyroxene phenocrysts is similar to that of the groundmass clinopyroxene, being represented by augitic composition. Few augite crystals having relatively higher Mg content are enclosed in large, poikilitic plagioclase crystal. The clinopyroxene phenocrysts altered partly or completely to chlorite and actinolite (Plate II/3). Titanite and quartz were occasionally observed in association with chlorite, they may be formed when Ti and Si was released during alteration of clinopyroxene. The clinopyroxene of the groundmass is represented by rounded crystals, that are generally fresh, or rarely, are slightly altered to chlorite. The accessory minerals commonly include apatite, ilmenite, rutile, hematite and zircon. The fissure fillings of these rocks are made of chlorite, quartz, albite, calcite prehnite, pumpellyite and Mg-epidote. Laumontite was found only in one sample. In addition to chlorite, chlorite/smectite mixed-layer mineral was also determined in subordinate quantities in several samples.

The plutonic series comprises mainly *gabbroic rocks* within which nests of leucocratic rocks form a small portion, being regarded as the end products of the crystallization differentiation of the mafic magma. Gabbros exhibit a coarse-grained texture and consist
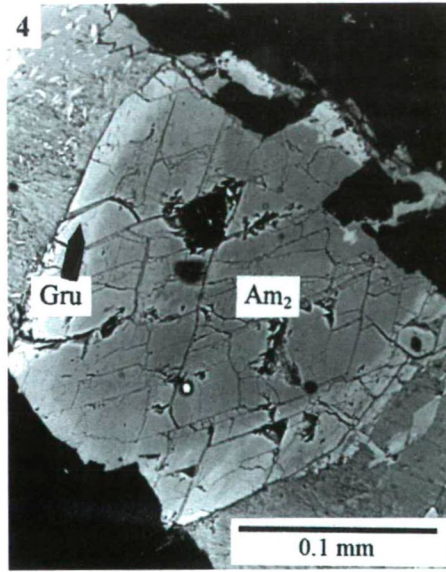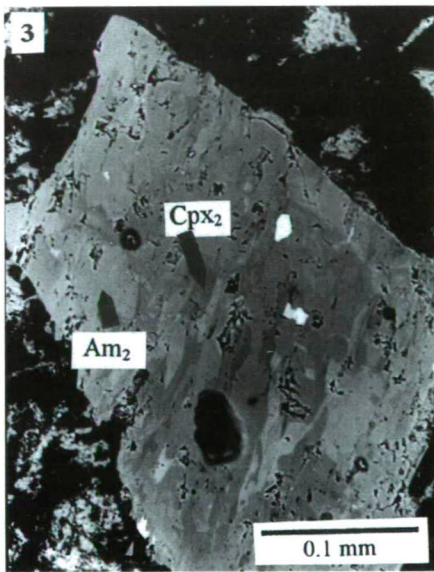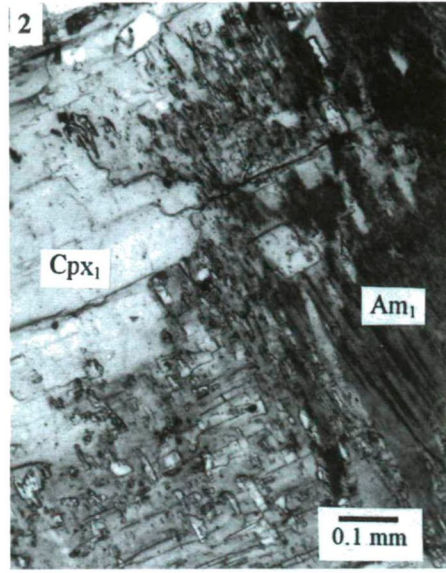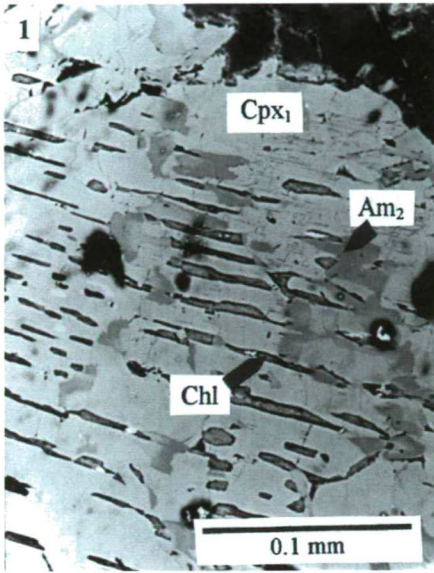
Plate I.

I/1. Alteration stages of clinopyroxene (Cpx₁) from metagabbro, showing patches of green amphibole (Am₂) and chlorite (Chl) along the cleavage planes of Cpx₁. Sample No. DG-6. Back scattered electron (BSE) image.

I/2. Magmatic clinopyroxene (Cpx₁) rimmed by magmatic brown amphibole (Am₁) from the metagabbro sample No. DG-22. Blebs of relic clinopyroxene (Cpx₁) are found in the amphibole rim. Plane polarized light.

I/3. Secondary (metamorphic) clinopyroxene (Cpx₂) forms intergrowths with secondary (metamorphic) amphibole (Am₂) in the metagabbro sample No. DG-27. BSE image.

I/4. Zoned metamorphic amphibole within a chlorite vein with actinolitic hornblende (Am₂) core and grunerite (Gru) rim from the metagabbro sample No. DG-22. BSE image.
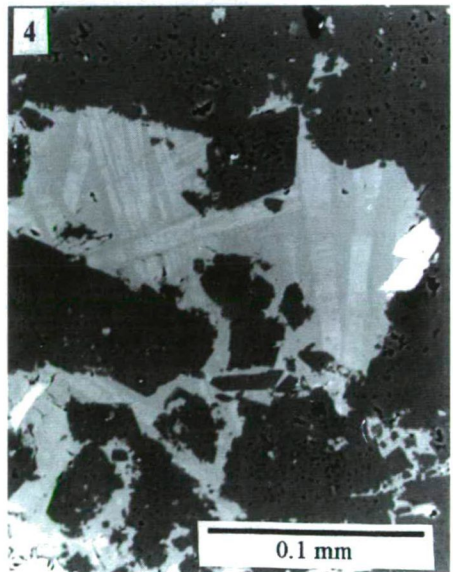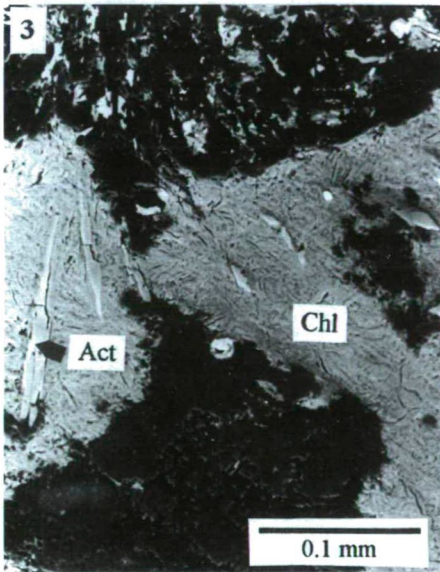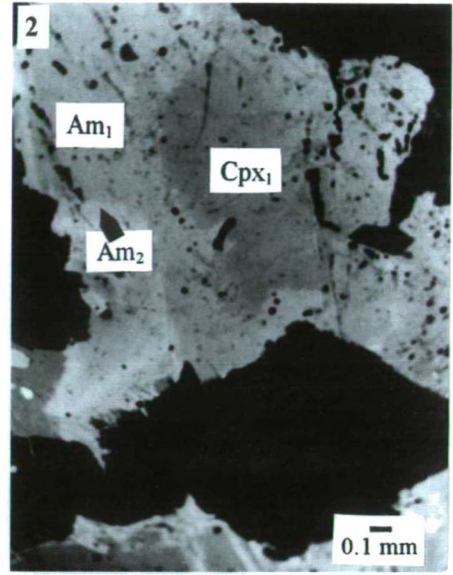
Plate II.

II/1. Fibres of green, metamorphic amphibole in association with chlorite at the edge of a green amphibole grain from the metagabbro sample No. DG-6. Plane polarized light.

II/2. Brown amphibole (Am_1), forming rim around magmatic clinopyroxene (Cpx_1), grades into green amphibole (Am_2) and actinolite and chlorite fibres in metagabbro sample No. DG-6. BSE image

II/3. Clinopyroxene crystals altered to chlorite (Chl) and actinolite (Act) in metadiabase sample No. DG-13. BSE image.

II/4. Exsolution of Fe-rich prehnite in host Fe-poor prehnite in metagranite sample No. DG-5b. BSE image.

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principally of plagioclase, clinopyroxene and amphibole. Chlorite pseudomorphs after olivine (but without relics) were scarcely recognized. Plagioclase occurs mostly as euhedral prismatic crystals, being commonly of albitic composition (An_1 to An_6) with traces of oligoclase (An_{19} to An_{29}). The alteration products of plagioclase include pumpellyite, prehnite, Mg-epidote and calcite. Clinopyroxene is the dominant ferromagnesian mineral that frequently occurs as subhedral or anhedral crystals. The clinopyroxene is augite and salite, sometimes showing simple zoning that displays an enrichment in Fe content and depletion in Mg and Ca contents towards the edges. Clinopyroxene was partially or wholly replaced by amphibole and chlorite. In the early stages of alteration, amphibole forms patches in the mineral, while chlorite is found along the cleavage planes of clinopyroxene (Plate I/1). Metamorphic clinopyroxene occurs scarcely as minute grains at the edge of primary clinopyroxene or forms intergrowth with metamorphic amphibole (Plate I/3). This metamorphic clinopyroxene was distinguished by its higher Si and Ca and lower Al, Ti and Na contents from the magmatic one. Such chemical differences between the two generations of clinopyroxenes are consistent with previous observations in other ophiolitic gabbros (MÉVEL 1987, 1988). The brown primary amphiboles formed in the late (hydrous) stage of magmatic crystallization as large crystals or as rims around clinopyroxene. Like other magmatic amphiboles from oceanic gabbros, they usually exhibit hornblende composition and are enriched in Ti, Al and Na, as compared to the metamorphic amphibole. In the brown rimmed amphibole, blebs of relic clinopyroxene were recognized (Plate I/2). The chemical composition of the rimmed brown amphibole is identical with that of the large individual crystals of brown amphibole. Its origin is, therefore, preferably considered as magmatic or high-temperature deuteric as reported by GIRARDEAU and MÉVEL (1982). This brown amphibole is frequently surrounded by green secondary amphibole that sometimes shows fibrous texture and is characterized by hornblende and actinolitic hornblende composition (Plates II/1 and II/2). This structural relationship was commonly described also in other ophiolitic rocks (MÉVEL 1987, 1988; BORTOLOTTI et al. 1990). In rare cases, zoned amphibole with actinolitic hornblende core and grunerite or cummingtonite rim was observed (Plate I/4). The boundary between the two amphiboles is sharp reflecting a miscibility gap between them during their metamorphic crystallization (see SHIDO 1958). Opaque minerals and apatite are the main accessory minerals, while zircon was rarely recorded. The opaque minerals are represented mainly by homogeneous ilmenite or ilmenite-rutile intergrowth. They are usually mantled by intimately mixed fine grains rich in iron and silica. In few samples, titanomagnetite was also recognized.

The fissure fillings of metagabbros contain prehnite, pumpellyite, chlorite, quartz and calcite. The Fe content of the veined prehnite varies widely from 0.012 to 0.726, and sometimes Fe-rich prehnite forms rim around Fe-poor one. Talc and smectite are rarely found, representing metamorphic and weathering products, respectively.

The petrography of *metagranites* is fully described by SADEK GHABRIAL et al. (1994). Generally, they composed mainly of quartz and plagioclase with minor amounts of biotite that is completely altered to chlorite, prehnite, epidote and titanite. K-feldspar could not be distinguished in these rocks like other oceanic plagiogranites (COLEMAN 1977). Plagioclase crystals are totally albitized (An_1-An_7) and sometimes contain prehnite and white mica along their cracks and cleavage planes. On the basis of the proportion of garnet, these rocks were classified into garnet-poor or -free and garnet-rich granites. The garnet was formed produced by local assimilation of the country rocks (BALLA et al., 1983; SADEK GHABRIAL et al. 1994). The metasediment xenoliths are commonly

associated with garnet-rich rocks, and are composed chiefly of plagioclase with average An_{23} and chlorite that seems to be formed after biotite. No preferred orientation could be observed in these xenoliths. The fissure fillings in metagranites consist of fine aggregates of chlorite, calcite, prehnite, quartz, albite and titanite. Prehnite located near to the fractures usually contains an appreciable amount of iron. In few cases, Fe-rich prehnite forms exsolution laths in host Fe-poor prehnite (Plate II/4).

MINERAL CHEMISTRY

Plagioclase

The magmatic plagioclase, presumably with basic-intermediate composition has been extensively altered. Thus, the plagioclase from the studied meta-igneous rocks is represented mainly by albite (Table 1). In metagabbros, oligoclase (An_{19-29}) was rarely observed in association with albite. There are two possible models for explaining the presence of oligoclase, namely: (i) incomplete albitization of plagioclase and (ii) the oligoclase formed during ocean-floor metamorphism (see ISHIZUKA 1985) and persisted as instable phase during the regional (dynamothermal) metamorphism. Significant amounts of K averaging 0.025 per formula unit (cation number on the basis of 8 oxygen) is found in the oligoclase structure, while the K content in the associated albite is very low, averaging only 0.005. Oligoclase was also recorded from the metasedimentary rocks that form enclaves in the metagranites with an average of An_{23}.
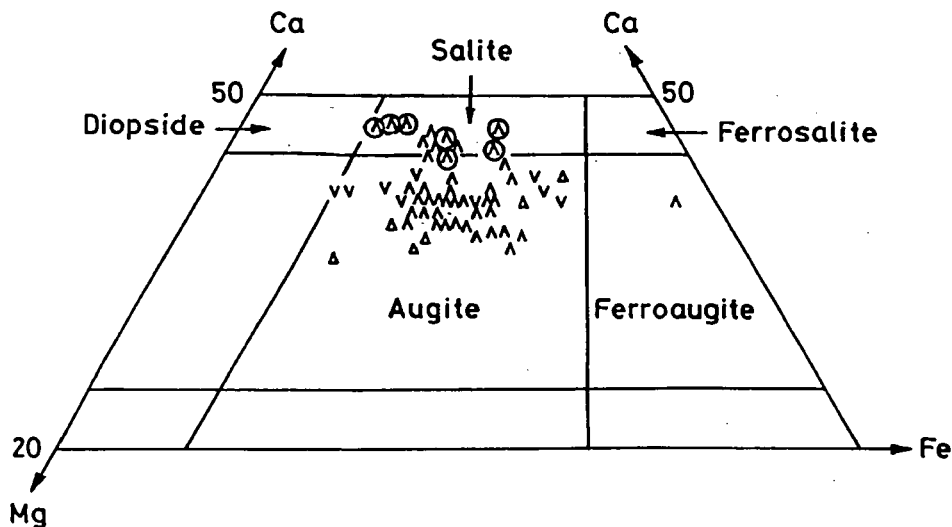
Fig. 4. Compositions of the studied clinopyroxene plotted on the pyroxene quadrilateral. Symbols: ∧ = metabasalt; = metadiabase; metagabbro. The symbols of secondary (metamorphic) clinopyroxene are circled.

Clinopyroxene

Most of the analysed clinopyroxenes are augite (*Fig. 4*). Their Mg/Fe ratios depend strongly on the bulk rock chemistry. As to MÉVEL (1987, 1988), the most significant differences between the magmatic and metamorphic clinopyroxenes are related to the abundances of Ti, Al, Si, Ca and Na. The metamorphic clinopyroxene is always richer in
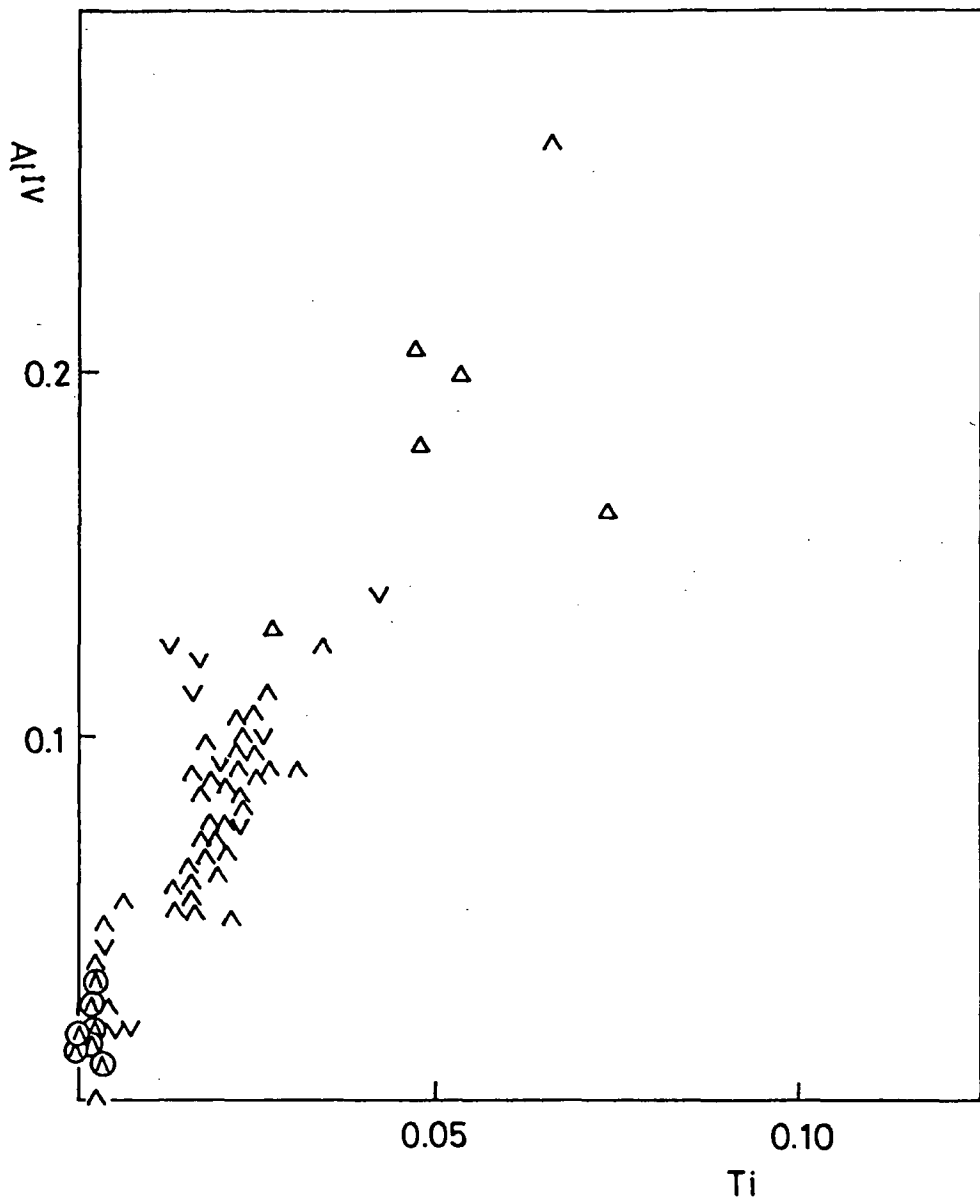
Fig. 5. The plot of Ti against Al^{IV} of clinopyroxene from the Szarvaskő complex. Symbols: = metabasalt; = metadiabase; = metagabbro. The symbols of secondary (metamorphic) clinopyroxene are circled.

Si and Ca, and poorer in Al, Ti and Na than the magmatic one (see also GILLIS 1995). The present data show that the clinopyroxene of metabasalts contains more Al and Ti than that of the intrusive rocks (Table 2). This may be attributed to subsolidus re-equilibration of clinopyroxene during ocean-floor metamorphism which might also cause the appearance

of metamorphic clinopyroxene in metagabbros. The continuous transition between the formation of magmatic and metamorphic clinopyroxenes may be related to the initial high temperature phase of hydrothermal metamorphism which might cause incomplete recrystallization of primary magmatic clinopyroxene resulting in quasy continuous changes in clinopyroxene chemistry (*Fig. 5*). Thus, the continuous transition between primary and secondary clinopyroxenes may be the result of non-equilibrium character of hydrothermal metamorphism that is common in such metamorphic settings. The notable depletion of Al and Ti in some clinopyroxenes from metadiabases and metagabbros seems also to depend on the concentration of iron: the Fe-rich clinopyroxene is always characterized by low values of Al and Ti (DEER et al. 1964).

Amphibole

Magmatic brown amphibole is widespread in metagabbros as large, individual crystals or as rims around clinopyroxene grains. This type of amphibole is of hornblende and tschermakitic hornblende composition (*Fig. 6*), being enriched in Ti, Al and Na and depleted in Si as compared to metamorphic amphiboles (Table 3). Nonetheless, the TiO_2 and K_2O contents of these magmatic amphiboles (reaching maximum of 2.84 and 0.33 weight-%, respectively) are not comparable to those reported by MÉVEL (1988) from oceanic gabbros (4% TiO_2 and 0.6-0.8% K_2O), but they resemble to those given by GIRARDEAU and MÉVEL (1982); EVARTS and SCHIFFMAN (1983).

The metamorphic amphiboles are widespread in metagabbros, forming rims around magmatic amphibole and clinopyroxene. They are almost characterized by actinolitic hornblende and hornblende compositions (*Fig. 6*), and vary in colour from pale brown to green with decreasing Ti content. The pale brown amphibole is a relatively Ti-rich hornblende that is followed by green actinolitic hornblende, reflecting variations of Si, Al^{IV} and Ti with decreasing crystallization temperature of amphiboles. The dark green amphibole relates to the increase in Fe content. In rare cases, sharply zoned amphibole with actinolitic hornblende core and Mg-Fe amphibole (grunerite and cummingtonite) rim was recognized, suggesting a miscibility gap between the two amphiboles during their metamorphic crystallization. In metadiabases, metamorphic amphiboles are uncommonly developed at the expense of magmatic clinopyroxene. The former ones are rather actinolitic in composition, and are plotted in the actinolite field or at the limit between the actinolite and actinolitic hornblende fields *(Fig. 6)*. The Fe/Mg ratios of amphiboles seem to be largely controlled by bulk rock composition.

Chlorite

In FOSTER's (1962) diagram used for classification, the chlorites of the studied rocks fall mostly in the brunsvigite field (Table 4). Not only the Fe/Mg ratios, but also the Al^{IV} contents of chlorite are strongly governed by bulk rock compositions, although the Al^{IV} content of chlorite has been used as an empiric geothermometer by several authors (CATHELINEAU 1988; JOWETT 1991 and others). For example, there is a significant difference in Al^{IV} contents between the chlorites of metagranite and metagabbro (Table 4), although the temperature of metamorphism should have been the same for both rock types. The petrogenetic evaluation of chemistry and structural characteristics of chlorites and their precursor phases from the Szarvaskő complex is outlined by ÁRKAI and SADEK GHABRIAL (1996).
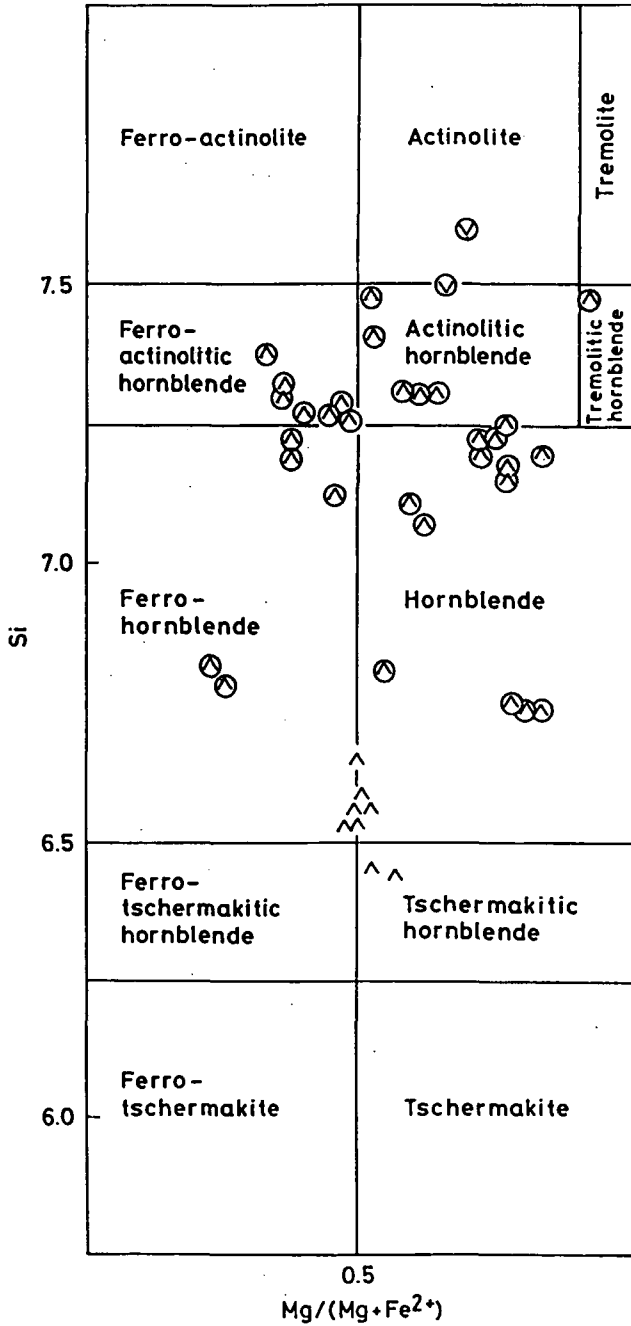
Fig. 6. A plot of Si versus Mg/(Mg+Fe^{2+}) of amphiboles from the Szarvaskő complex (after LEAKE 1978). Symbols: = metadiabase; = metagabbro. The symbols of secondary (metamorphic) amphibole are circled.

Epidote

Epidote is widespread in metagranites as replacing magmatic biotite, and as filling vugs, in which it is usually associated with prehnite, chlorite, calcite, albite and quartz. The $X_{Fe}3+$ values (where $X_{Fe}3+ = Fe^{3+}/[Fe^{3+}+Al]$) of both textural types of epidote are apparently identical, ranging from 0.234 to 0.248. Epidote of metabasites mostly shows high Mg contents approaching pumpellyite composition (Table 5). Similar epidotes were named Mg-epidote by CHO and LIOU (1987), who mentioned that this mineral is predominant in somewhat higher grade rocks (transitional zone between the prehnite-pumpellyite and greenschist facies, and greenschist facies) than the Mg-poor epidote, the latter being common in prehnite-pumpellyite facies rocks. In the Szarvaskő ophiolite complex, Mg-epidote was more commonly observed in the prehnite-pumpellyite facies metabasic rocks than in prehnite-actinolite facies metagranites.

In metabasalts, epidote from low variance assemblages contains very low Fe content with $X_{Fe}3+$ values averaging 0.008. The Fe contents of epidote don't show a trend towards lower values with increasing metamorphic grade, as suggested by NYSTRÖM (1983).

Prehnite

Prehnite was observed in all types of the investigated meta-igneous rocks, being more common in the plutonic than in the extrusive ones. It usually occurs as vein fillings (or vugs) in association with other minerals, and as replacing pre-existing minerals such as plagioclase and biotite.

The compositional variation of prehnite in term of $X_{Fe}3+$ (where $X_{Fe}3+ = Fe^{3+}/[Fe^{3+}+Al]$) largely depends on its mode of occurrence (Table 6). The lowest Fe contents, with $X_{Fe}3+$ values averaging 0.025, are found in prehnite formed within plagioclase. Prehnite formed after ferromagnesian silicate minerals is characterized by intermediate $X_{Fe}3+$ values ranging from 0.057 to 0.129. The Fe contents of prehnite found in veins and vugs are highly varying even in two neighbouring prehnite crystals, with $X_{Fe}3+$ values attaining a maximum of 0.358. Prehnites formed at the expense of magmatic minerals constitute high variance assemblages, whereas the veined prehnites constitute variable, sometimes low variance assemblages. The veined prehnite characteristic of high variance assemblages coexisting only with chlorite and quartz ± calcite ± sphene, forms aggregates in which the Fe contents of the individual prehnite crystals strongly vary: the Fe-richer prehnite lies always close to the border. Sometimes, Fe-poor prehnite is rimmed by Fe-rich prehnite, or encloses exsolution laths of Fe-rich prehnite. Further research is needed to explain the possible reasons of these phenomena which may be related to non-equilibrium crystallization of prehnite as well as to temperature dependent chemistry of prehnite, including Fe^{3+} immiscibility at low temperatures. Zoned prehnite with Fe-rich core and Fe-poor rim was also recognized. In contrast, prehnite associated with other Ca-Al silicate minerals + chlorite + quartz from low variance assemblages shows consistently very low $X_{Fe}3+$ values in metabasalts ranging from 0.002 to 0.004, and higher $X_{Fe}3+$ values in metagranites ranging from 0.036 to 0.292.

As a whole, the chemistry of prehnite cannot be assumed as a reliable indicator of metamorphic temperature, confirming the earlier statements of CHO et al. (1986), BEVINS and MERRIMAN (1988). More likely, the chemistry of the fluid phase and the mode of occurrence may control the composition of prehnite.
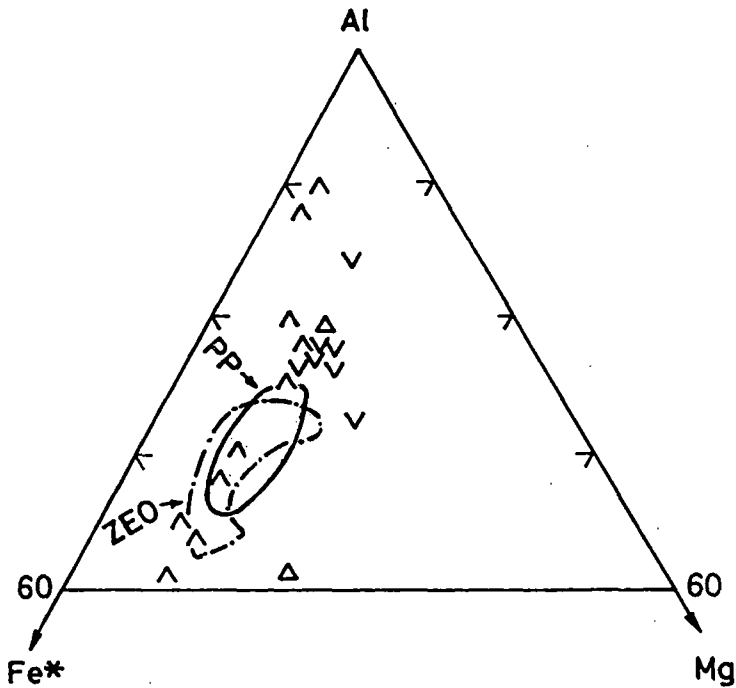
Fig. 7. Al-Fe*-Mg diagram displaying pumpellyite compositions of the studied rocks. Solid and dash-dotted lines border the compositional fields of low-variance pumpellyite from prehnite-pumpellyite and zeolite facies, respectively, given by CHO et al. (1986). Symbols: = metabasalt; = metadiabase; = metagabbro.

Pumpellyite

Pumpellyite has been recorded in all types of metabasic rocks, but is lacking in metagranites. It forms granular aggregates in veins with other minerals, and replaces plagioclase and clinopyroxene in the matrix. Pumpellyite is abundant in high variance assemblages, where it is characterized by extensive Fe-Al substitution (Table 7). Therefore, the pumpellyite should not be used as an indicator of metamorphic grade as regarded by CHO et al. (1986). The $X_{Fe}3+$ values of pumpellyite from low variance assemblages are relatively lower than those from high variance assemblages. Due to the smaller number of data obtained from pumpellyite of low variance assemblages, these $X_{Fe}3+$ values cannot be applied for determining the metamorphic grade. Displaying high Al contents, the analysed low variance pumpellyites from the Szarvaskö area are plotted outside the compositional fields of low variance pumpellyites given by CHO et al. (1986) for zeolite and prehnite-pumpellyite facies (*Fig. 7* and Table 7). It seems to be very likely that compositional variations in pumpellyite are governed mostly by the extent of host rock alteration (see also NYSTRÖM 1983) and by the effective fluid chemistry, while the bulk composition shows little control on pumpellyite chemistry.

The average $X_{Fe}3+$ values of coexisting prehnite, pumpellyite and epidote are 0.003, 0.137 and 0.008, respectively. These values contrast with the results of CHO et al. (1986), who found that the $X_{Fe}3+$ values most probably increase in order of prehnite, through pumpellyite to epidote.

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SZARVASKŐ COMPLEX

| gabbro plagiogranite | diabase | pillow basalt |

- Laumontite
- Prehnite
- Pumpellyite
- Mg-epidote
- Epidote
- Actinolite
- Actinolitic hornblende
- Hornblende
- Fe-Mg amphibole
- Clinopyroxene
- Chlorite
- Albite
- Calcic plagioclase
- Quartz
- Sphene
- Calcite

Fig. 8. Distribution of metamorphic minerals in the simplified vertical profile of the Szarvaskő ophiolite complex.

Other minerals

Laumontite is the only zeolite mineral that occurs only in one metadiabase sample. Its composition is generally homogeneous, having an ideal formula. Laumontite might form at lower temperatures than the other metamorphic minerals, perhaps in a late stage of cooling (uplift). Garnets from garnet-rich and garnet-poor metagranites are characterized by similar composition, being almandine-rich (Alm_{75-76}, Pyr_{7-11}, $Spess_{9-14}$, $Gross_{4-5}$). The chemistry of garnet has been fully described and interpreted by SADEK GHABRIAL et al. (1994).

METAMORPHISM

The ophiolitic rocks of the Szarvaskő complex display significant change in their mineral parageneses, reflecting the effects of different metamorphic events, including also local metasomatic processes. The combined effect and superposition of these different metamorphic processes which may be attributed to ocean-floor hydrothermal, burial and regional metamorphic events, are likely to produce various secondary minerals, almost showing a wide variation in their composition. Therefore, the evolution path of these rocks is rather difficult to reconstruct.
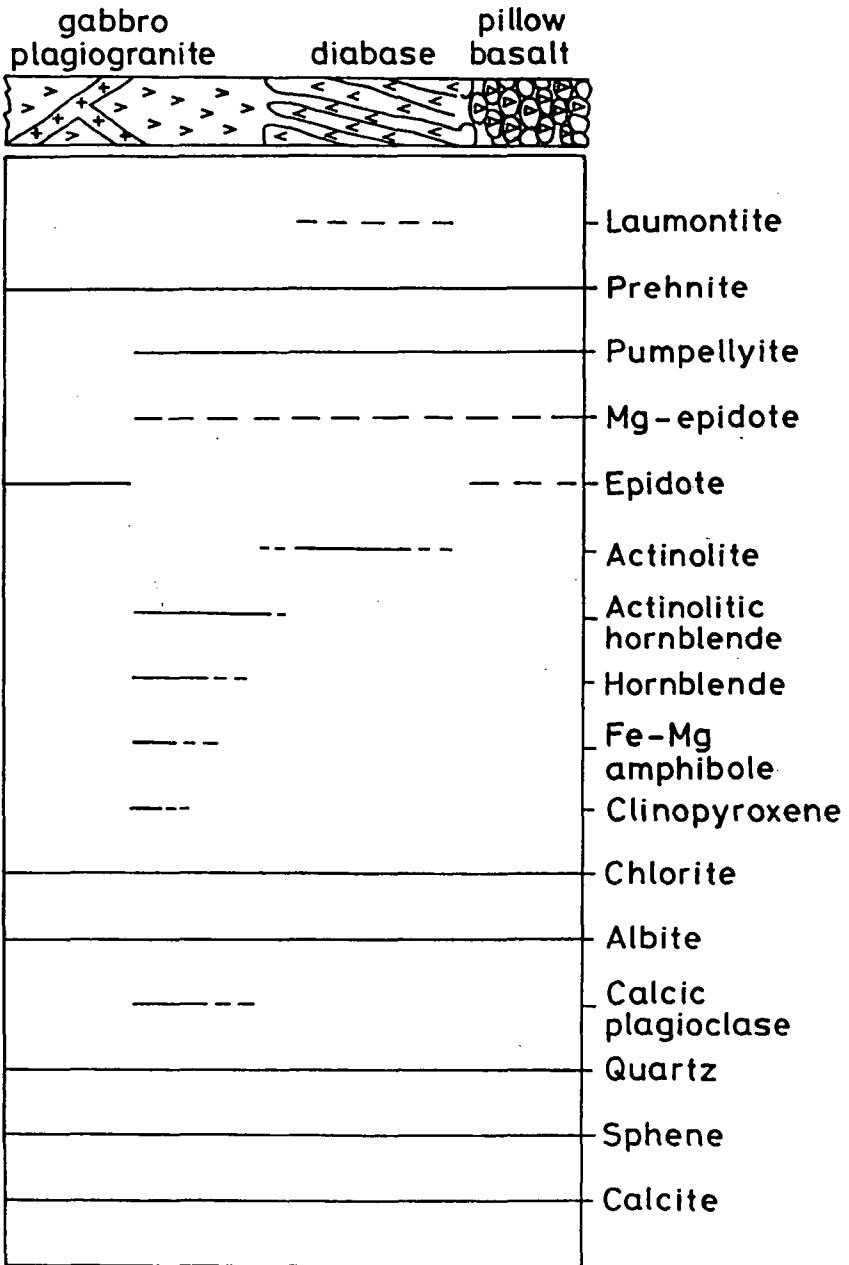
Hydrothermal (ocean-floor) metamorphism

Hydrothermal metamorphism occurring at the oceanic ridge axes, has been commonly resulted in formation of disequilibrium mineral assemblages, while the original igneous fabrics have been preserved. The distribution of secondary minerals in the Szarvaskő complex is depicted in *Fig. 8*. In metagabbros, albite and traces of oligoclase coexist replacing the more calcic, magmatic plagioclase. Magmatic clinopyroxene and brown amphibole have been uralitized producing heterogeneous crystals of green amphibole that are frequently zoned from hornblende to actinolitic hornblende showing decrease in their Al^{IV} and Ti contents towards the rims. This compositional trend can be related to the decrease of temperature during the crystallization of amphiboles. In rare cases, clinopyroxene has probably also recrystallized, forming minute grains of metamorphic clinopyroxene. In addition, Fe-Mg amphibole also appears as rims around metamorphic green amphibole grains. Thus, metagabbros show crystallization in amphibolite and greenschist/amphibolite transitional facies (700-400°C; see also LIOU et al. 1974 and MOODY et al. 1983), displaying a gradual cooling from upper amphibolite facies condition that is responsible for the development of secondary clinopyroxene, hornblende to actinolitic hornblende, Fe-Mg amphibole and calcic plagioclase, to lower amphibolite or greenschist/amphibolite transitional facies that is characterized by actinolitic hornblende + oligoclase + albite ± sphene ± chlorite assemblage. The gradual cooling of gabbro and the reacting fluid phase might be a consequence of changing tectonic position, namely, the spreading away from the ridge axis.

Most of the primary minerals of metagranites have also been altered. The plagioclase is of albitic composition. Other main constituents are chlorite, epidote, prehnite and titanite representing pseudomorphs after pre-existing biotite. The metagranites are characterized by albite + chlorite + prehnite + epidote + titanite assemblage. As suggested by EVARTS and SCHIFFMAN (1983), the presence of epidote-bearing and actinolite-free assemblage indicates greenschist facies metamorphism and is characteristic of low pressure conditions, while prehnite is paragentically late. Moreover, this assemblage may

also be equivalent to the low pressure prehnite-actinolite facies of LIOU et al. (1985,1987), forming between ca. 200 and 350°C.

The secondary minerals of metadiabases include albite, chlorite and amphibole that belongs to the actinolitic group. The two latter minerals were formed after primary clinopyroxene, indicating that the metadiabases were subjected to greenschist facies metamorphism (<400°C; LIOU et al. 1974 and MOODY et al. 1983).

No evidence is available on the eventual ocean-floor hydrothermal metamorphism of the pillow basalt. The basaltic layers might have been partially recrystallized into either zeolite or prehnite-pumpellyite facies associations during hydrothermal alteration as have been recorded from other ophiolitic pillow lavas (COLEMAN 1977; ÁRKAI 1983; DI GIROLAMO et al. 1992). The younger regional (dynamothermal) metamorphic event, corresponding to prehnite-pumpellyite facies, should has overprinted and obliterated the lower temperature associations formed eventually during the ocean-floor alteration.

Regional (dynamothermal) metamorphism

As it was demonstrated in the previous part, there are considerable uncertainties in the genetic interpretation of low-temperature mineral assemblages. These assemblages may be attributed to (i) the late, cooling stages of hydrothermal metamorphism; (ii) the somewhat younger, Cretaceous regional (dynamothermal) metamorphism as proved earlier by ÁRKAI (1983); or (iii) the combined, superimposed effects of the formers two events, both producing mainly prehnite-pumpellyite facies assemblages.

Out of these, the explanation (iii) seems to be the most probable, because it is not possible to distinguish between the alteration products of both metamorphic processes indicated by (i) and (ii), and the compositions of secondary minerals are not homogeneous, referring to non-equilibrium conditions.

During the Alpine regional metamorphic event, the adjoining sedimentary rocks of the Szarvaskő complex suffered late diagenetic – low-temperature anchizonal orogenic transformations, the estimated temperature of which agrees fairly well with that of the prehnite-pumpellyite facies determined in the metadiabase and metagabbro varieties (ÁRKAI 1983).

The secondary mineral assemblages of the metasedimentary rocks (quartz + albite + white mica + chlorite + calcite ± smectite ± vermiculite ± chlorite/smectite mixed-layer mineral ± pyrite ± rutile ± hematite) are not diagnostic for determining exactly the metamorphic grade. The averages of IC(002), ChC(001) and ChC(002) of the <2μm grain-size fraction samples of the metaclastic rocks are 0.388, 0.452 and 0.398, respectively, using a goniometer speed of 2°/min (Table 8). The IC(002) average indicates anchizonal conditions. Based on the relatively high chlorite crystallinity values, these rocks fall in the diagenetic range, using the chlorite crystallinity boundaries of the anchizone proposed by ÁRKAI (1991). This may be due to the broadening of the 7Å and 14Å chlorite peaks in the presence of other minerals having similar spacing such as vermiculite, smectite and chlorite/smectite mixed-layers. As a whole, most of metamorphic indicators show that the sedimentary environment of Szarvaskő complex underwent transitional late diagenetic- – low-temperature anchizonal regional transformations.

The matabasic rocks of the Szarvaskő complex exhibit prehnite-pumpellyite facies assemblages which include quartz, chlorite, prehnite, pumpellyite, epidote and calcite. Excluding pumpellyite, similar assemblages were found in the metagranites (SADEK GHABRIAL et al. 1994). The lack of pumpellyite in metagranites can be properly explained by inappropriate bulk composition. Actinolitic amphibole was also recorded from the

metadiabases, and its origin relates to ocean-floor (hydrothermal) metamorphism (see the preceding part). The refined petrogenetic grid of FREY et al. (1991) showed that the bulk rock composition plays an important role in determining the mineral assemblages in metabasites transitional between sub-greenschist and greenschist facies, because the P-T fields of the four sub-greenschist facies (i.e. ZEO, PrP, PrA and PA) partially to completely overlap. This might explain the juxtaposition of assemblages having different low-grade metamorphic facies and being developed at the same P-T conditions. Two spatially related assemblages (Prh-Pmp-Ep-Chl-Qtz and Prh-Act-Ep-Chl-Qtz) were recognized by POWELL et al. (1993) in a hyaloclastic metabasalt sample, whereas the compositions of both epidote and prehnite are considerably different reflecting the presence of two distinct univariant assemblages rather than one invariant assemblage. BEVINS and ROBINSON (1993) described prehnite-pumpellyite and prehnite-actinolite facies assemblages in metabasites associated with epizonal pelites. Prehnite-pumpellyite facies assemblages occur in metabasites with $MgO/(MgO+FeO)_{chlorite}$ ratios of <0.54, while prehnite-actinolite facies assemblages are present in metabasites with $MgO/(MgO+FeO)_{chlorite}$ ratios of >0.54. These authors also recognized actinolite-bearing prehnite-pumpellyite facies assemblage in metabasites with $MgO/(MgO+FeO)_{chlorite}$ ratio of 0.50 in epizonal circumstances. On the other hand, metabasites with the prehnite-pumpellyite facies assemblages were found in association with diagenetic and anchizonal pelites. In these metabasites, the range of $MgO/(MgO+FeO)_{chlorite}$ ratios is rather broad (0.31-0.65). The analysed chlorite from the Szarvaskő metadiabases, which are associated with diagenetic – low-temperature anchizonal pelitic rocks, has relatively low $MgO/(MgO+FeO)$ ratios ranging from 0.29 to 0.51. Therefore, the actinolite is considered to be of hydrothermal (ocean-floor) metamorphic origin. The IC(002) averages of the <2µm fraction samples measured at 2°/min goniometer speed are 0.369 Δ°2θ for the metagranites and 0.282 Δ°2θ for the metagabbros, indicating anchizonal alterations for both rock types. These conditions based on illite crystallinity data may be considered as roughly equivalent to those of the prehnite-pumpellyite facies. Based on the chlorite crystallinity boundaries of the anchizone given by ÁRKAI (1991) for metasediments, the metagranites show high temperature anchizonal metamorphism, while the metabasic rocks fall mostly in the diagenetic range or near to the diagenesis/anchizone boundary that corresponds to transitional conditions between zeolite and prehnite-pumpellyite facies. The larger chlorite crystallinity values of the metabasic rocks can be explained by the presence of interstratified smectitic and/or vermiculitic components in chlorite. As also given in Table 7, the chlorite crystallinity averages decrease downwards in the ophiolite-like pile from the pillow lava through the diabase down to the cumulate gabbro. This trend may be interpreted as a remnant of the late, cooling stage of ocean-floor metamorphism characterized by steep thermal gradient. In this case, the younger orogenic (dynamothermal) event could not totally obliterate these differences. Furthermore, the differences between the chlorite crystallinity values obtained from metagranites and from metabasic rocks relate most probably to the effects of bulk chemistries on chlorite chemistry (ÁRKAI and SADEK GHABRIAL 1996). The temperatures (metabasalt: 236°C; metadiabase: 224-252°C; metagabbro: 230-280°C; metagranite: 276-350°C) estimated by the chlorite-Al geothermometer of CATHELINEAU (1988) support this difference between metagranites and metabasic rocks.

TABLE 1

Averages of chemical compositions (weight-%) and cation numbers of plagioclase

Rock type	metabasalt	metadiabase		metagabbro			metagranite					enclave metasediment
Sample No.	DG-33	DG-13	DG-16	DG-2a	DG-21b	DG-22	DG-1	DG-5b	DG-5d	DG-34b	DG-34d	DG-34e
n	1	4	4	3	4	6	4	4	4	4	4	3
SiO_2	66.66	68.12	67.03	63.09	66.71	67.76	67.61	67.21	69.16	66.66	68.23	62.31
Al_2O_3	19.59	20.48	20.28	22.56	20.52	19.89	20.46	20.25	19.16	20.57	20.10	24.05
CaO	1.02	0.17	0.92	4.57	0.64	0.83	0.60	0.32	0.15	1.10	0.11	4.77
Na_2O	10.95	11.99	10.70	8.51	11.81	11.04	11.38	11.03	11.33	10.58	12.20	8.79
K_2O	0.08	0.04	0.04	0.43	0.07	0.11	0.11	0.56	0.09	0.23	0.06	0.16
Total	98.30	100.80	98.97	99.16	99.75	99.63	99.68	99.36	100.15	99.13	100.70	100.08
number of the cations on the basis of 8 (O)												
Si	2.959	2.954	2.954	2.810	2.928	2.967	2.948	2.959	3.008	2.196	2.965	2.753
Al	1.025	1.047	1.054	1.186	1.062	1.026	1.05	1.051	0.996	1.069	1.029	1.257
Ca	0.049	0.008	0.043	0.219	0.030	0.039	0.029	0.015	0.007	0.052	0.005	0.226
Na	0.943	1.008	0.914	0.675	1.005	0.937	0.962	0.942	0.956	0.905	1.028	0.752
K	0.005	0.002	0.002	0.025	0.004	0.006	0.006	0.031	0.005	0.013	0.003	0.009
Total	4.981	5.019	4.967	4.915	5.029	4.975	4.996	4.999	4.971	4.939	5.031	4.992
An%	5	1	5	24	3	4	3	2	1	5	1	23

n = number of analyses

TABLE 2

Averages of chemical compositions (weight-%) and cation numbers of clinopyroxene

Rock type	metabasalt	metadiabase		metagabbro							
Sample No.	DG-33	DG-13	DG-16	DG-6	DG-21b	DG-22		DG-27		DG-37c	DG-34dd
n	5	2	8	4	7	6	1**	6	6**	9	7
SiO_2	48.32	51.52	50.51	50.85	50.65	51.16	53.03	50.85	52.66	50.76	49.97
TiO_2	1.80	0.53	0.62	0.70	0.71	0.56	0.00	0.75	0.04	0.50	0.93
Al_2O_3	4.13	3.69	1.97	2.74	2.05	1.38	0.31	1.72	0.45	1.70	3.07
*FeO	11.28	6.33	11.88	9.99	13.62	11.96	10.65	11.67	9.33	11.58	10.11
MnO	0.25	0.17	0.26	0.30	0.42	0.55	0.35	0.36	0.36	0.39	0.38
MgO	14.92	17.33	13.33	14.33	13.41	11.64	13.55	14.32	13.71	14.18	13.38
CaO	18.59	20.90	20.34	19.92	19.35	19.85	22.88	19.50	23.27	19.81	21.38
Na_2O	0.29	0.26	0.30	0.38	0.24	0.30	0.07	0.31	0.14	0.34	0.34
K_2O	0.01	0.00	0.00	0.03	0.00	0.00	0.00	0.02	0.00	0.02	0.02
Total	99.59	100.64	99.21	99.23	100.45	100.39	100.85	99.51	99.96	99.28	98.61
number of the cations on the basis of 6 (O)											
Si	1.826	1.882	1.924	1.915	1.915	1.949	1.981	1.925	1.977	1.928	1.890
Al^{IV}	0.171	0.118	0.076	0.085	0.085	0.050	0.014	0.075	0.020	0.070	0.110
Al^{VI}	0.013	0.041	0.013	0.037	0.007	0.012	0.000	0.002	0.000	0.006	0.026
Ti	0.051	0.015	0.018	0.020	0.020	0.016	0.000	0.021	0.001	0.014	0.026
*Fe^{2+}	0.358	0.193	0.380	0.315	0.431	0.478	0.333	0.370	0.294	0.368	0.320
Mn	0.008	0.006	0.008	0.010	0.013	0.018	0.011	0.012	0.012	0.013	0.012
Mg	0.838	0.944	0.755	0.805	0.755	0.660	0.755	0.808	0.767	0.803	0.753
Ca	0.754	0.819	0.830	0.804	0.784	0.811	0.916	0.791	0.937	0.807	0.866
Na	0.021	0.018	0.022	0.027	0.018	0.022	0.005	0.023	0.010	0.025	0.025
K	0.001	0.000	0.000	0.002	0.000	0.000	0.000	0.001	0.000	0.001	0.001
Total	4.042	4.036	4.026	4.019	4.029	4.015	4.015	4.029	4.018	4.035	4.030
Ca%	38.5	41.8	42.0	41.6	39.5	41.2	45.4	39.9	46.6	40.6	44.4
Fe+Mn%	18.7	10.2	19.7	16.8	22.4	25.2	17.1	19.3	15.2	19.1	17.0
Mg%	42.8	48.0	38.3	41.6	38.1	33.6	37.5	40.8	38.2	40.3	38.6

* Total iron calculated as FeO and Fe^{2+}
** secondary clinopyroxene
n = number of analyses

TABLE 3

Averages of chemical compositions (weight-%) and cation numbers of amphibole. The calculations of cation numbers follow the scheme of ROBINSON et al. (1982)

Rock type		metadiabase	metagabbro							
Sample No.		DG-13	DG-2a		DG-6		DG-22	DG-27		DG-37c
n		2**	4***	1**	7***	4**	9**	3***	3**	8**
SiO_2		52.28	43.25	48.07	43.61	49.02	48.27	46.77	49.63	49.48
TiO_2		0.16	2.75	0.61	1.85	1.52	0.74	1.62	0.72	0.54
Al_2O_3		1.73	9.26	5.72	8.82	4.37	3.01	8.19	4.26	2.48
*FeO		19.66	20.12	21.03	21.54	19.80	27.14	11.87	17.12	22.74
MnO		0.22	0.30	0.43	0.29	0.26	0.36	0.17	0.32	0.39
MgO		12.78	8.40	10.13	7.83	10.47	7.42	15.05	13.07	10.98
CaO		11.22	10.10	9.97	10.67	11.12	8.98	11.70	11.28	8.60
Na_2O		0.20	2.32	1.04	1.94	0.91	1.49	1.67	1.19	1.59
K_2O		0.06	0.29	0.22	0.36	0.20	0.36	0.17	0.17	0.27
Total		98.31	96.79	97.22	96.91	96.67	97.77	97.21	97.76	97.07

		number of the cations on the basis of 23 (O)								
Si		7.543	6.551	7.075	6.647	7.326	7.259	6.745	7.229	7.267
Al	T	0.294	1.449	0.925	1.353	0.642	0.534	1.255	0.731	0.429
Fe^{3+}		0.163	–	–	–	0.032	0.207	–	0.040	0.304
Al		–	0.205	0.067	0.230	0.135	–	0.137	–	–
Ti		0.017	0.313	0.068	0.210	0.058	0.084	0.175	0.078	0.060
Fe^{3+}	M_1	0.888	0.604	1.238	0.575	0.557	1.173	0.654	0.719	1.402
Mg	M_3	2.747	1.896	2.223	1.770	2.303	1.663	3.235	2.838	2.405
Fe^{2+}		1.321	1.946	1.350	2.178	1.913	2.033	0.778	1.326	1.087
Mn		0.027	0.037	0.054	0.037	0.034	0.046	0.020	0.039	0.046
Fe^{2+}		–	–	–	–	–	–	–	–	–
Mn	M_4	–	0.001	–	–	–	0.001	–	–	–
Ca		1.735	1.639	1.573	1.743	1.782	1.448	1.808	1.764	1.355
Na		0.056	0.361	0.297	0.257	0.208	0.401	0.192	0.211	0.433
Na	A	–	0.320	–	0.313	0.056	0.032	0.255	0.124	0.018
K		0.010	0.057	0.041	0.071	0.038	0.070	0.031	0.032	0.051
Total		14.801	15.37	14.911	15.384	15.084	14.951	15.285	15.131	14.857
Al		0.294	1.653	0.992	1.582	0.777	0.534	1.392	0.731	0.429
Na		0.056	0.681	0.297	0.570	0.264	0.434	0.466	0.335	0.452
A		0.010	0.377	0.041	0.384	0.094	0.102	0.285	0.156	0.069
$(Al+Fe^{3+})^{IV}$		0.458	1.449	0.925	1.353	0.674	0.741	1.255	0.771	0.733

*   total iron calculated as FeO
**   green amphibole
***   brown amphibole
n = number of analyses

TABLE 4

Averages of chemical compositions (weight-%) and cation numbers of chlorite

Rock type	metabasalt	metadiabase		metagabbro						metagranite				enclave metased.
Sample	DG-33	DG-13	DG-16	DG-6	DG-21b	DG-22	DG-37c	DG-34dd	DG-1	DG-5b	DG-5d	DG-34b	DG-34d	DG-34e
n	5	10	5	4	8	8	7	7	9	7	4	4	4	2
SiO_2	28.18	29.28	28.82	26.78	26.96	27.27	27.11	27.37	27.21	24.45	26.33	26.20	25.69	26.94
TiO_2	0.00	0.08	0.06	0.07	0.11	0.56	0.67	1.02	1.54	0.27	1.02	0.95	0.49	1.59
Al_2O_3	16.35	18.19	15.67	17.45	15.35	14.60	14.05	16.79	17.64	19.83	17.60	17.91	18.79	17.89
*FeO	30.95	25.88	27.75	27.80	39.04	37.34	37.42	29.09	31.54	36.18	31.51	36.06	34.94	33.13
MnO	0.40	0.39	0.39	0.22	0.30	0.34	0.42	0.36	0.69	0.38	0.55	0.50	0.42	0.49
MgO	10.73	14.74	13.42	12.63	6.41	6.88	7.47	11.20	8.94	7.02	8.56	6.73	6.72	6.48
CaO	0.67	0.65	0.56	0.70	0.49	0.40	0.52	1.01	0.75	0.24	0.67	0.52	0.94	0.72
Na_2O	0.03	0.02	0.02	0.06	0.02	0.02	0.02	0.01	0.04	0.03	0.04	0.05	0.05	0.04
K_2O	0.02	0.01	0.02	0.03	0.03	0.04	0.04	0.08	0.16	0.04	0.12	0.14	0.07	0.88
Total	87.33	89.24	86.71	85.75	88.71	87.45	87.72	86.94	88.52	88.45	86.41	89.05	88.11	88.16

number of the cations on the basis of 28 (O)

Si	6.153	6.051	6.227	5.874	6.075	6.186	6.149	5.964	5.898	5.445	5.864	5.786	5.704	5.938
Al^{IV}	1.847	1.949	1.773	2.126	1.925	1.814	1.851	2.036	2.102	2.555	2.136	2.2148	2.296	2.062
Al^{VI}	2.358	2.481	2.218	2.391	2.151	2.090	1.905	2.280	2.406	2.651	2.481	2.448	2.626	2.586
Ti	0.000	0.013	0.009	0.012	0.019	0.095	0.113	0.167	0.251	0.046	0.171	0.156	0.082	0.264
*Fe^{2+}	5.654	4.484	5.014	5.104	7.365	7.082	7.100	5.312	5.722	6.740	5.869	6.667	6.498	6.106
Mn	0.075	0.068	0.072	0.041	0.058	0.066	0.080	0.067	0.127	0.073	0.103	0.094	0.080	0.090
Mg	3.491	4.528	4.321	4.127	2.149	2.325	2.525	3.633	2.887	2.330	2.844	2.211	2.226	2.129
Ca	0.158	0.144	0.130	0.164	0.118	0.098	0.127	0.235	0.137	0.056	0.161	0.121	0.221	0.170
Na	0.012	0.008	0.009	0.026	0.009	0.010	0.009	0.006	0.018	0.014	0.019	0.021	0.023	0.018
K	0.007	0.003	0.001	0.009	0.008	0.013	0.011	0.023	0.045	0.011	0.035	0.040	0.021	0.247
Total	19.755	19.729	19.774	19.873	19.877	19.779	19.870	19.724	19.621	19.920	19.684	19.758	19.776	19.608

* Total iron calculated as FeO and Fe^{2+} n = number of analyses

TABLE 5

Averages of chemical compositions (weight-%) and cation numbers of epidote

Rock type	metabasalt		metadiabase	metagabbro		metagranite	
Sample No.	DG-33	DG-33**	DG-13**	DG-37c**	DG-1	DG-34b	DG-34d
n	2	10	2	1	2	2	2
SiO_2	37.59	36.58	38.13	37.38	37.41	38.36	38.14
TiO_2	0.00	0.00	0.01	0.00	0.21	0.30	0.92
Al_2O_3	28.74	22.57	27.34	26.46	24.50	24.58	23.89
$*Fe_2O_3$	0.33	9.58	5.44	3.29	11.72	12.19	12.15
MnO	0.33	0.12	0.24	0.03	0.10	0.31	0.22
MgO	0.00	0.63	0.76	0.73	0.05	0.03	0.10
CaO	24.12	22.65	21.89	23.28	23.31	23.47	23.09
Na_2O	0.05	0.13	0.02	0.04	0.04	0.06	0.01
K_2O	0.01	0.00	0.00	0.00	0.00	0.03	0.00
Total	91.17	92.26	93.83	91.21	97.34	99.33	98.52
number of the cations on the basis of 25 (O)							
Si	6.160	6.123	6.129	6.157	5.962	5.997	6.008
Al^{IV}	–	–	–	–	0.038	0.016	0.012
ΣZ	6.160	6.123	6.129	6.157	6.000	6.013	6.020
Al^{VI}	5.551	4.443	5.180	5.137	4.564	4.513	4.425
$*Fe^{3+}$	0.041	1.219	0.658	0.453	1.408	1.436	1.442
Ti	0.000	0.000	0.001	0.000	0.025	0.035	0.109
ΣY	5.592	5.662	5.839	5.590	5.997	5.984	5.976
Mn	0.046	0.017	0.033	0.004	0.013	0.041	0.029
Mg	0.000	0.157	0.181	0.179	0.012	0.007	0.022
Ca	4.238	4.067	3.769	4.110	3.982	3.933	3.899
Na	0.016	0.040	0.005	0.013	0.012	0.017	0.002
K	0.001	0.001	0.000	0.000	0.000	0.006	0.000
ΣW	4.301	4.282	3.988	4.306	4.019	4.004	3.952
$X_{Fe^{3+}}$	0.007	0.215	0.113	0.081	0.236	0.241	0.246

* total iron calculated as Fe_2O_3 and Fe^{3+}

** Mg-epidote

n = number of analyses

TABLE 6

Averages of chemical compositions (weight-%) and cation numbers of prehnite

Rock type	metabasalt		metagabbro				metagranite			
Sample	DG-33	DG-2a	DG-21b	DG-22	DG-34dd	DG-1	DG-5b	DG-5d	DG-34b	DG-34d
n	3	4	6	11	4	4	8	4	3	7
SiO_2	42.47	44.50	43.33	43.86	42.54	43.91	43.08	43.47	44.49	42.91
TiO_2	0.00	0.02	0.08	0.06	0.03	0.05	0.05	0.07	0.02	0.02
Al_2O_3	23.46	22.83	22.11	21.13	22.94	23.39	22.68	24.15	23.64	22.33
$*Fe_2O_3$	0.06	0.38	3.42	3.01	2.37	0.94	2.58	0.51	0.25	2.08
MnO	0.03	0.04	0.01	0.06	0.03	0.03	0.01	0.02	0.02	0.03
MgO	0.01	0.01	0.04	0.29	0.13	0.04	0.01	0.01	0.00	0.01
CaO	26.89	26.44	26.62	26.14	26.73	26.97	27.36	27.61	26.85	27.14
Na_2O	0.05	0.11	0.00	0.02	0.06	0.13	0.04	0.04	0.35	0.05
K_2O	0.04	0.04	0.03	0.04	0.02	0.02	0.01	0.01	0.04	0.02
Total	93.01	94.37	95.64	94.61	94.85	95.49	95.82	95.89	95.66	94.58
number of the cations on the basis of 22 (O)										
Si	6.011	6.183	6.019	6.141	5.951	6.059	5.974	5.947	6.107	6.018
Al^{IV}	1.989	1.817	1.981	1.859	2.049	1.941	2.026	2.053	1.893	1.982
ΣZ	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al^{VI}	1.920	1.922	1.633	1.627	1.730	1.862	1.679	1.808	1.933	1.707
$*Fe^{3+}$	0.006	0.040	0.366	0.320	0.250	0.098	0.271	0.150	0.026	0.221
Ti	0.000	0.002	0.009	0.006	0.003	0.006	0.005	0.005	0.002	0.002
ΣY	1.926	1.964	2.008	1.953	1.983	1.966	1.955	1.963	1.961	1.930
Mn	0.004	0.005	0.001	0.008	0.004	0.004	0.001	0.001	0.002	0.004
Mg	0.001	0.002	0.009	0.060	0.026	0.009	0.002	0.021	0.000	0.002
Ca	4.083	3.938	3.963	3.923	4.009	3.989	4.067	4.051	3.951	4.080
Na	0.013	0.029	0.000	0.004	0.016	0.035	0.009	0.006	0.092	0.013
K	0.008	0.008	0.005	0.006	0.005	0.004	0.003	0.006	0.006	0.005
ΣX	4.109	3.982	3.978	4.001	4.060	4.041	4.082	4.085	4.051	4.104
$X_{Fe^{3+}}$	0.003	0.020	0.183	0.164	0.126	0.050	0.139	0.077	0.013	0.115

* total iron calculated as Fe_2O_3 and Fe^{3+}

n = number of analyses

TABLE 7

Averages of chemical compositions (weight-%) and cation numbers of pumpellyite

Rock type	metabasalt	metadiabase			metagabbro	
Sample No.	DG-33	DG-13	DG-16	DG-21b	DG-22	DG-37c
n	2	1	6	3	3	4
SiO_2	36.12	37.25	36.64	36.20	37.44	36.75
TiO_2	0.00	0.00	0.03	0.19	0.56	0.00
Al_2O_3	23.58	26.93	23.88	20.76	21.57	25.17
*Fe_2O_3	10.23	4.38	7.02	15.89	10.60	3.59
MnO	0.15	0.22	0.10	0.06	0.07	0.07
MgO	3.07	1.81	2.33	1.56	1.61	1.75
CaO	19.81	23.35	22.16	19.68	21.36	23.00
Na_2O	0.02	0.02	0.02	0.05	0.00	0.06
K_2O	0.02	0.00	0.01	0.03	0.25	0.00
Total	93.01	93.96	92.19	94.42	93.46	90.39
number of the cations on the basis of 16 (O)						
Si	5.918	5.944	6.012	6.002	6.174	6.072
Al^{IV}	0.124	0.056	0.012	0.016	–	–
ΣZ	6.042	6.000	6.024	6.018	6.174	6.072
Al^Y	4.000	4.000	3.996	3.977	3.931	4.000
*Fe^Y	–	–	–	–	–	–
Ti	0.000	0.000	0.004	0.023	0.069	0.000
ΣY	4.000	4.000	4.000	4.000	4.000	4.000
Al^X	0.430	1.009	0.610	0.064	0.261	0.898
*Fe^X	1.270	0.531	0.875	2.002	1.330	0.500
Mg	0.748	0.431	0.568	0.386	0.396	0.429
ΣX	1.448	1.971	2.053	2.452	1.987	1.827
Mn	0.021	0.030	0.014	0.008	0.010	0.010
Ca	3.482	3.994	3.899	3.499	3.775	4.072
Na	0.006	0.006	0.007	0.016	0.000	0.019
K	0.003	0.000	0.002	0.006	0.053	0.001
ΣW	3.512	4.030	3.922	3.529	3.838	4.102
X_{Fe*}	0.223	0.096	0.160	0.331	0.241	0.093

* total iron calculated as Fe_2O_3 and Fe^{3+}

n = number of analyses

TABLE 8

Mean values and standard errors (in parentheses) of phyllosilicate crystallinity indices ($\Delta^o 2\theta$)

XRD parameters	Rock type	Whole rock, AD		<2μm fractions, AD		<2μm fractions, EG	
		2°/min	½°/min	2°/min	½°/min	2°/min	½°/min
ChC(001)	meta-pelite	0.348 (0.020)	0.323 –	0.452 (0.023)	– –	0.394 –	– –
	meta-basalt	– –	– –	0.449 (0.008)	0.399 (0.013)	0.536 –	0.454 –
	meta-diabase	0.404 (0.020)	0.364 –	0.388 (0.014)	0.332 (0.014)	0.482 (0.028)	0.424 (0.022)
	meta-gabbro	0.353 –	– –	0.421 (0.026)	0.372 (0.015)	0.343 (0.007)	0.301 (0.014)
	meta-granite	0.271 (0.016)	– –	0.332 (0.017)	0.226 (0.008)	0.324 (0.017)	0.250 –
ChC(002)	meta-pelite	0.358 (0.011)	0.290 (0.022)	0.398 (0.012)	0.306 (0.022)	0.313 –	0.275 –
	meta-basalt	0.410 (0.012)	0.370 (0.013)	0.372 (0.010)	0.342 (0.009)	0.392 (0.008)	0.357 –
	meta-diabase	0.337 (0.006)	0.287 (0.011)	0.359 (0.011)	0.321 (0.010)	0.380 (0.014)	0.346 (0.012)
	meta-gabbro	0.314 (0.011)	0.256 (0.010)	0.337 (0.009)	0.306 (0.010)	0.310 (0.009)	0.296 (0.014)
	meta-granite	0.276 (0.006)	0.229 (0.003)	0.282 (0.008)	0.246 (0.007)	0.283 (0.001)	0.245 (0.005)
IC(002)	meta-pelite	0.364 (0.014)	0.234 (0.023)	0.388 (0.023)	0.262 (0.053)	0.288 (0.047)	0.253 (0.046)
	meta-basalt	– –	– –	– –	– –	– –	– –
	meta-diabase	– –	– –	0.352 –	– –	– –	– –
	meta-gabbro	– –	– –	0.285 (0.024)	0.245 (0.016)	0.318 –	0.244 –
	meta-granite	0.220 –	– –	0.369 (0.027)	0.300 (0.022)	0.353 (0.010)	0.296 –

AD = Air-dried preparations of the <2μm fractions.
EG = Ethylene glycol-solvated preparations of the <2μm fractions.

CONCLUSION

In the Mesozoic dismembered ophiolite complex of the Szarvaskő area (Bükk Mountains, NE Hungary), imprints of two metamorphic events were distinguished. In the initial stage of hydrothermal (ocean-floor) metamorphism, metamorphic minerals characteristic of high temperature and low pressure (e.g. new generation of clinopyroxene, calcic amphibole and a Fe-Mg amphibole) were formed. The grade of hydrothermal metamorphism increases downwards from the prehnite-pumpellyite or zeolite facies (metabasalt), through the greenschist facies (metadiabase), to the greenschist/amphibolite transitional and amphibolite facies (metagabbro). The existence of second generation of clinopyroxene, namely an Al-poor clinopyroxene in metagabbros indicates re-equilibration of magmatic clinopyroxene during hydrothermal metamorphism that has been rarely recorded so far in the literature. The changes in amphibole chemistry and in mineral associations refer to continuous cooling (retrogression) in the course of hydrothermal metamorphism. The composition of the metamorphic amphibole formed during hydrothermal metamorphism in the metagabbros grades from hornblende to actinolitic hornblende while that of metadiabase is actinolitic in composition, reflecting variations in its Si, Al and Ti contents with decreasing grade upwards in the ophiolite profile. The younger (Alpine, Cretaceous) orogenic (dynamothermal) metamorphism - as it was shown by ÁRKAI (1983) - produced mainly prehnite-pumpellyite facies assemblages in matabasic rocks and late diagenetic to low-temperature anchizonal alterations in the associated metaclastic rocks. This younger, regional metamorphic event could not perfectly obliterate and re-equilibrate the earlier products of hydrothermal metamorphism. As a consequence, typical non- equilibrium, mixed assemblages have persisted.

The large variations in the compositions of Ca-Al silicate minerals suggest that the fluid chemistry, mode of occurrence and the extent of rock alteration rather than temperature are the main controls of their compositions. In the low variance assemblages, the average $X_{Fe}3+$ values of coexisting prehnite, pumpellyite and epidote are 0.003, 0.137 and 0.008, respectively. This contrasts with the suggestion of CHO et al. (1986) who mentioned that the $X_{Fe}3+$ values most probable increase in order of prehnite, through pumpellyite to epidote.

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