

Magmatic fluid segregation and overprinting hydrothermal processes in gabbro pegmatites of the Neotethyan ophiolitic Szarvaskő Complex (Bükk Mountains, NE Hungary)

ATTILA PÉNTEK^{1*}, FERENC MOLNÁR¹ and DAVID H. WATKINSON²

¹Department of Mineralogy, Eötvös Loránd University, Pázmány P. sétány 1/C, 1117 Budapest, Hungary; *✉ attilapentek@yahoo.com

²Department of Earth Sciences, Carleton University, 1125 Colonel By Drive, Ottawa K1S5B6, Canada

(Manuscript received September 22, 2005; accepted in revised form June 22, 2006)

Abstract: Pegmatites of the Szarvaskő Ophiolite Complex, Bükk Mts, NE Hungary were classified according to their shape (pockets, dykes) and texture (zoned, homogeneous), representing different stages of fluid enrichment during crystallization of the host gabbro. Local assimilation of adjacent sedimentary rocks increased the volatile-, and incompatible-element content of the melt. Most pegmatites crystallized from a locally segregated hydrous silicate melt, but some were intruded later. Pegmatites have mineral contents similar to their host gabbro, but are enriched in amphibole, biotite, Fe-Ti-oxides, quartz, and apatite. During formation of the pegmatites a fluid phase separated and caused deuteric alteration under magmatic-submagmatic conditions. Post-magmatic sea-floor hydrothermal activity is recognized by intense alteration and formation of a greenschist facies mineral assemblage at temperatures of 250–400 °C. Fluid inclusion studies revealed two aqueous fluid types related to this polyphase hydrothermal process. Alpine regional metamorphism caused intense deformation of the rocks, accompanied by veining of a low-grade metamorphic mineral assemblage. Primary fluid inclusions in vein-filling minerals and chlorite thermometry were used to obtain proposed conditions of 270–285 °C and 150–200 MPa for this process.

Key words: Hungary, Bükk Mts, Alpine metamorphism, sea-floor hydrothermal alteration, deuteric alteration, magmatic fluid, gabbro pegmatite.

Introduction

The Szarvaskő Ophiolite Complex (SOC) is located in the southwestern part of the Bükk Mts, NE Hungary and occupies a territory of about 25 km² (Fig. 1). Gabbroic rocks form the dominant intrusions of SOC and they crop out around the village of Szarvaskő (Fig. 1). They are characterized by very variable texture and mineral composition even within a single outcrop. In some quarries and outcrops, pegmatitic pockets and dykes can also be observed.

Gabbro pegmatites are poorly studied features in ophiolitic gabbros. Manning et al. (1996) and Beard et al. (2002) studied such features in recent oceanic units, and Hoeck et al. (2002) provided data from ophiolitic rocks of the Neotethyan realm. Hence, one of the major aspects of our work was to support the knowledge on gabbro pegmatites in ophiolite-related gabbro intrusions by analysing their relationships to the host rock, textural variations and mineral composition. These observations were used to characterize volatile enrichment, segregation of magmatic fluids and their interaction with silicates during crystallization of the pegmatite bodies.

Previous investigations (Árkai 1983; Árkai et al. 1995; Sadek et al. 1996) have shown that magmatic rocks of the SOC underwent alteration during sea-floor hydrothermal and low-grade regional metamorphic events. Our study is also aimed at outlining characteristics of these processes by combining mineralogical and fluid inclusion data.

Geology of the Szarvaskő Ophiolite Complex

The SOC is situated in the Bükk Mts, a tectonic unit of the ALCAPA (Alpine-Carpathian-Pannonian) terrane, and is in allochthonous position, overlying the Bükk Parautochthonous Unit (Balla et al. 1983; Balla 1984a; Csontos 2000). It has a synformal structure made up of three nappes. The Mónosbél Nappe consists of sedimentary rocks, whereas these are accompanied by magmatic intrusions and pillowed basalts in the Szarvaskő I and Szarvaskő II nappes (Csontos 2000). The magmatic intrusions are hosted by a series of turbiditic shales and sandstones.

The ophiolitic sequence is quite uncommon and different units are missing in comparison to other well-known ophiolites (e.g. mantle section with ultramafic units is completely absent). Cumulate gabbro is abundant and at some places is accompanied by rare ultramafic cumulates (hornblende, Fe-Ti-rich wehrlite). The upper part of the plutonic section with various gabbroic and related plagiogranitic rocks is the best preserved part of the unit (Szentpétery 1953; Sadek & Árkai 1994), and crops out in the Tardos and Tóberc quarries (Fig. 1). The volcanic section is well developed; pillowed basalts are exposed in many outcrops of the SOC (Szentpétery 1953; Balla 1984b).

Petrochemical analyses of mafic-ultramafic rocks of SOC revealed some differences from MORB and therefore

a back arc basin, or a marginal sea environment has been suggested to explain their formation (Kubovics et al. 1990; Downes et al. 1990; Harangi et al. 1996; Aigner-Torres 1996). A similar origin was suggested for some Dinaridic Ophiolite Zone (DOZ) ophiolites (Pamić 1997). Recent geotectonic models favour SOC formation in the Meliata-Vardar Ocean, where sea-floor spreading and magmatism lasted from the Middle Triassic until Late Jurassic/Early Cretaceous (Pamić et al. 2002; Csontos & Vörös 2004). The radiometric age of the SOC magmatic rocks is 166 ± 8 Ma (Árváné Sós et al. 1987), which is comparable to the ages from 189 ± 6.7 Ma to 136 ± 15 Ma obtained from ophiolitic rocks of the DOZ and also partly of the Vardar Zone (VZ) (Pamić et al. 1998; Pamić et al. 2002). According to Csontos & Vörös (2004), the Meliata-Vardar Ocean consisted of a Triassic-Jurassic plate and a Middle-Late Jurassic back arc basin, which could have

been the original site of the SOC. After crystallization of the igneous units of the SOC, the primary rock-forming mineral assemblage underwent a sea-floor hydrothermal alteration (ocean-floor metamorphism), which can be traced throughout the SOC (Sadek et al. 1996).

With the onset of Late Jurassic/Early Cretaceous intra-oceanic subduction of the Meliata-Vardar Ocean (Pamić et al. 2002; Csontos & Vörös 2004), the SOC became a part of the accretionary prism (Csontos 2000). Huge portions of the oceanic lithosphere were obducted onto the Apulian marginal units and are now preserved in the DOZ (Tari & Pamić 1998; Pamić et al. 1998; Csontos & Vörös 2004). The ophiolitic *mélange* and ophiolites of the Dinarides can be traced to the southwest-northeast Periadriatic-Balaton Line (Fig. 1). Smaller units of DOZ within the Mid-Hungarian Zone occur in the Samoborska Gora, Kalnik, Medvednica, and Ivanščica Mountains (Pamić 1997; Slovenec & Pamić

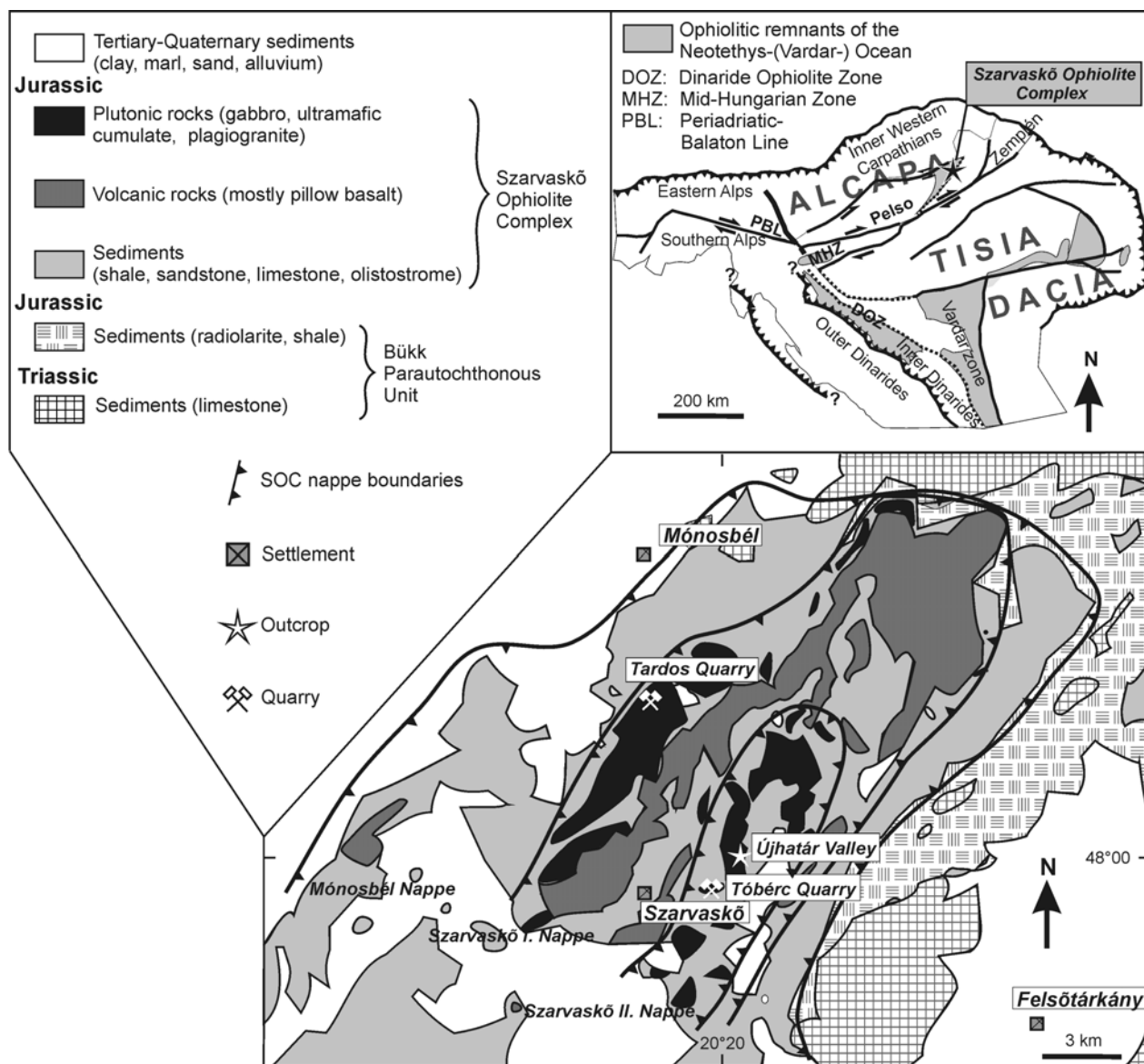


Fig. 1. Geological sketch map of the Szarvaskő Ophiolite Complex (modified after Less et al. 2002 and Kovács et al. 2004).

2002). Collision of the Bihor-Getic-Serbo-Macedonian upper plate and the Dinaridic High Karst lower plate may be the reason for a 120 Myr (Belák et al. 1995) metamorphic event in the DOZ (Csontos & Vörös 2004). The early Alpine, low-grade metamorphism was also observed in the nappes of Southern Slovakia and Northern Hungary (Árkai 1983; Árkai et al. 1995; Csontos 1999). According to Árkai et al. (1995) and Sadek et al. (1996), this metamorphism was a low temperature anchizonal event (max. 250–300 °C) corresponding to prehnite-pumpellyite facies, with an illite-muscovite K/Ar age of about 120 Ma, which is in good agreement with the data of Belák et al. (1995) for the age of metamorphism in the DOZ.

According to apatite fission-track dating (Árkai et al. 1995), the Szarvaskő Ophiolite Complex reached closure temperatures of 125–70 °C during Late Paleocene–Early Eocene (53 ± 8 Ma). The recent position of the SOC is due to the northeastward escape of the ALCAPA terrane from the Alpine collision zone along the Periadriatic-Balaton Line during the Late Eocene–Oligocene (Kázmér & Kovács 1985; Csontos & Vörös 2004; Fig. 1).

Analytical methods

Detailed field observations of pegmatites and host gabbroic rocks were supplemented by textural analyses of polished pegmatite slabs. Petrography was conducted on polished thin sections of pegmatites and host gabbro from the exposures of the Tóberc quarry and Újhatár Valley (Fig. 1). The chemical composition of minerals was determined using a Cameca MBX electron microprobe by wavelength dispersive method at Carleton University, Ottawa. The analytical standards were a well-characterized suite of synthetic and natural compounds. Analyses were carried out at 15 kV and 15 nA with counting times of 15–20 s (except F with 40 s) or until reaching 40,000 counts.

Microthermometric analyses of fluid inclusions in quartz from pegmatite and late veins were carried out on a Chaixmeca-type microthermometric apparatus. The equipment was standardized for approaching 0.1 °C reproducibility near melting temperatures of pure carbon dioxide (–56.6 °C) and pure water (0.0 °C) and 1 °C reproducibility at 374 °C using synthetic fluid inclusions. Fluid inclusion petrography and microthermometric measurements were carried out using 150–200 µm thick, doubly polished sections.

Occurrences and textural varieties of pegmatites

Tóberc quarry

Various gabbroic and related plagiogranitic rocks are exposed in Tóberc quarry situated 1 km east of Szarvaskő settlement (Fig. 1). Detailed petrographic description of the magmatic rocks was done by Szentpétery (1953), the plagiogranite was also studied by Sadek & Árkai (1994).

The host gabbro of the pegmatites is very heterogeneous, with medium to coarse texture. Coarsening of gab-

bro texture in irregular patches up to a few centimeters in size is a common feature of the rock. Clinopyroxene, plagioclase and amphibole are the major rock-forming minerals accompanied by Fe-Ti-oxides, biotite, apatite and rare orthopyroxene, olivine and sulphides. Xenoliths of the host sediments are abundant. These xenoliths were partly assimilated and recrystallized to coarse-grained massive quartz bodies with a plagioclase-rich granodioritic rim. Sulphide enrichments containing pyrrhotite, pentlandite and chalcopyrite in association with quartz and biotite can also be found along the contact of gabbro with shale. Gabbro is tectonically fragmented and fractures are filled with prehnite, calcite, chlorite, and quartz.

One type of pegmatite forms mostly isometric, round to elliptical pockets with diameter up to 20–30 cm (Fig. 2B and D) and they often have elongated tail-like attachments. These bent tails branch off the pocket and can be traced for a few tens of centimeters in the gabbro. The size of plagioclase, amphibole and biotite crystals is up to 3–4 cm toward the center of the pockets and the boundaries of the pegmatitic bodies are not sharp.

Two sub-types of pocket pegmatites can be distinguished according to their textures and compositions. In pegmatites with homogeneous texture, grain size and mineral composition do not show changes within a pocket (Fig. 2F) and sub- to euhedral hydrous silicate minerals (amphibole, biotite) are more common than in the host gabbro. Some anhedral quartz grains interstitial to the coarse-grained silicates are also present. Pegmatitic pockets with heterogeneous texture often show regular zoning of minerals (Fig. 2E), with a pyroxene-rich rim followed by an inner amphibole- and plagioclase-rich zone, and the latest crystallizing core made up of quartz and plagioclase. Quartz forms interstitial crystals or granophyric intergrowth with plagioclase.

Pegmatite also occurs as dykes that can be followed for several meters in the gabbro. Their terminations are wedge-shaped or abrupt due to tectonism. Pegmatite dykes can also be classified into two subgroups. One of them has mineral composition similar to the host gabbro, with gradual increase of grain size from the margin towards the center (Fig. 2A). Texture is homogeneous and the grain size of rock-forming minerals is up to 3–4 centimeters. The thickness of these dykes does not exceed 5–8 centimeters. The second type of pegmatitic dyke is 15–20 cm thick with sharp contacts toward the gabbro (Fig. 2C). Mineral composition shows a much more differentiated character than that for the host gabbro, because it is enriched in plagioclase and also contains interstitial quartz or granophyric quartz-plagioclase intergrowths. The grain size reaches only 1–2 centimeters.

Újhatár Valley

The homogeneous fine-grained gabbro has a cumulus texture with dominant amphibole and plagioclase. Pyroxene is absent, and Fe-Ti-oxides are less common than in the Tóberc quarry. Late differentiated pegmatitic dykes, xenoliths, and plagiogranitic rocks are absent.

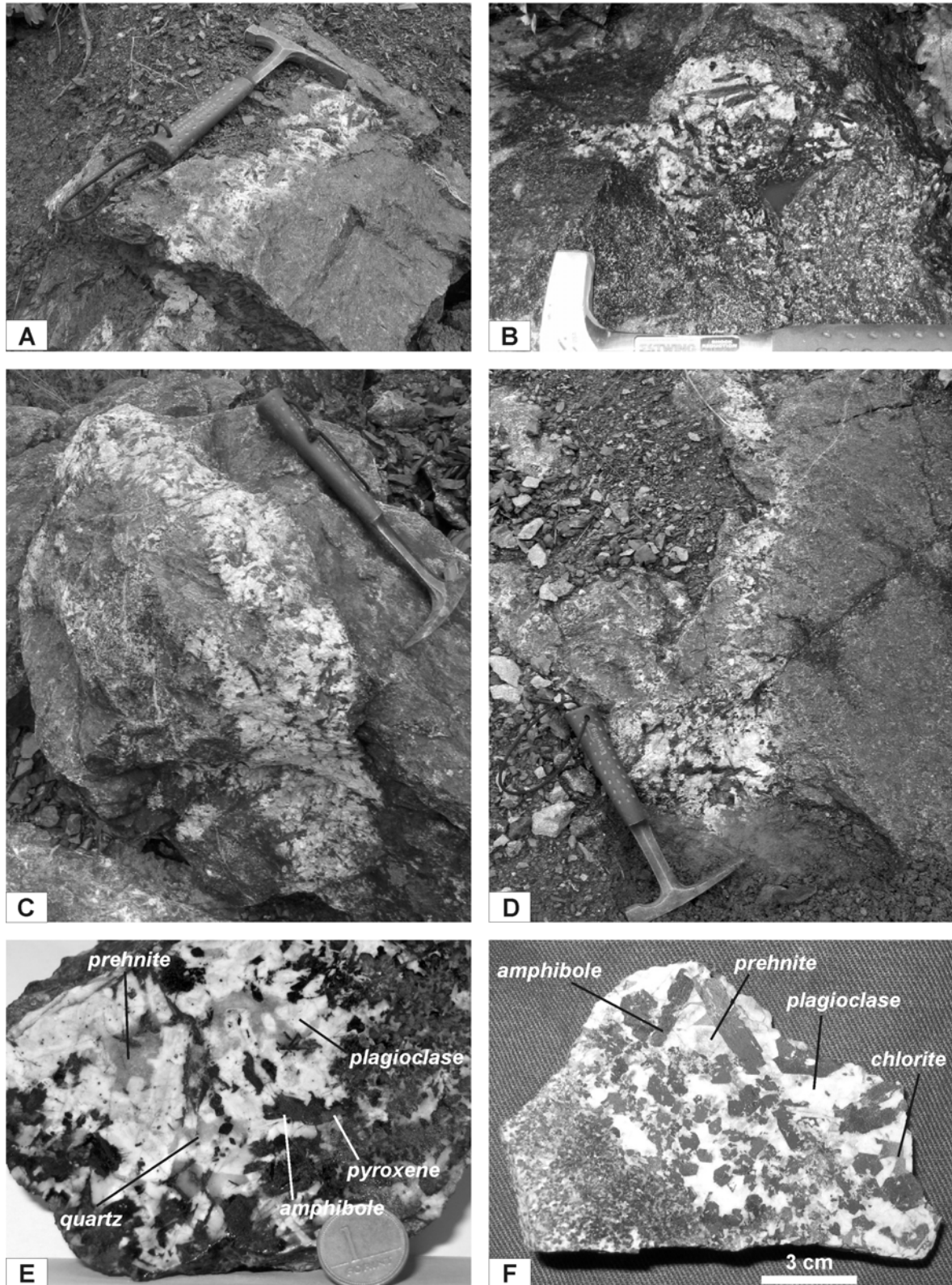


Fig. 2. Photographs of typical pegmatitic structures and textures. **A** — Narrow pegmatitic dyke from the Tóbérc quarry. Note faint boundary and mafic composition. **B** — Pegmatitic pod from the Újhatár Valley outcrop. Note the “tail” branching off the pod. **C** — Thick pegmatitic dyke from the Tóbérc quarry with sharp boundary and strongly evolved composition. **D** — Pegmatitic pod with a bent “tail” from the Tóbérc quarry. **E** — Polished slab of a zoned pegmatitic pod from the Tóbérc quarry. Note mafic rim and the plagioclase- and quartz-rich core of the pod. **F** — Polished slab of a homogeneous pegmatitic pod from the Újhatár Valley outcrop. Note the lack of pyroxene and the dominance of euhedral amphibole.

Subparallel bent dykes are the most common pegmatite bodies. Due to tectonism, the extent of dykes is unknown and it is not possible to conclude if these subparallel structures were horizontal sills, or vertical dykes. Dykes are just up to 5 cm thick and do not exhibit sharp contacts with the gabbro. Pocket-like pegmatites are also present, but less frequent. Similar to the host gabbro, pegmatite bodies exhibit a homogeneous, shape-independent texture. They are dominated by euhedral amphibole crystals up to 12 cm long (Fig. 2B and F). Plagioclase also forms euhedral to subhedral, but smaller crystals. Quartz is absent in these pegmatites.

Petrography and mineral chemistry

Primary minerals

Clinopyroxene occurs only in heterogeneous pegmatites of the Tóberc quarry. Euhedral crystals, up to 1 cm in size, are intergrown with anhedral ilmenite (Fig. 3B). Being the first crystallizing phase, it often grows along the boundary with the host gabbro and it is absent in the inner pegmatite zones. Pyroxene is strongly altered: only large crystals preserved fresh cores. Rims are always altered to amphibole, whereas chloritization occurs along cracks and cleavage planes. Fresh pegmatitic clinopyroxene corresponds to augite (Morimoto 1989) with a composition range of En_{37-39} Fs_{20-23} Wo_{39-41} (Table 1). The *Mg*-number decreases from 80–85 in the host gabbro (Aigner-Torres 1996) to 62–66 in the pegmatites at Tóberc quarry. Ti is also enriched in the pegmatitic augite. The Al and Na contents are almost the same as they are in the host gabbro, but the Ca-content is lower in the pegmatitic pyroxene (Table 1). The differentiated character of the parent melt is supported by negligible chromium content even in the rock forming pyroxene of the gabbro.

Plagioclase occurs as twinned anhedral grains up to 3 cm in diameter in pegmatites of the Tóberc quarry, and they always show intense alteration (sericitization and saussuritization). The original composition corresponds to andesine-labradorite (An_{35-60}) in the gabbro (Aigner-Torres 1996) and An_{44-50} in pegmatites (Table 2). Potassium contents are always low (less than 1 mol % Or component), and about 0.5 wt. % FeO is also present. The plagioclase in pegmatites of the Újhatár Valley is subhedral to anhedral, up to 1 cm in size. It often forms inclusions in amphibole, and is relatively fresh. The composition is almost pure albite (Ab_{98-99}) with less than 1 mol % Or component and with about 0.2–0.3 wt. % FeO. The fresh and homogeneous appearance of plagioclase rules out its origin by alteration of a more calcic feldspar.

Amphibole replaces rims of pyroxene, but it also occurs interstitially between plagioclases as well as in the form of euhedral grains (Fig. 3A). Up to 12 cm long prisms occur in pegmatites of the Újhatár Valley. Amphibole is always rich in Al, Ti and Na; the *Mg*-numbers are mostly less than 50 (Table 3). At the Újhatár Valley locality, tschermakite,

Table 1: Chemical composition of clinopyroxene from pegmatites and host gabbro of the Tóberc quarry (TOB). Data for host gabbro clinopyroxene from Aigner-Torres (1996). $\text{Mg\#} = \text{Mg}/(\text{Mg} + \text{Fe}_{\text{total}})$, En = enstatite, Fs = ferrosillite, Wo = wollastonite.

Sample no.	Pegmatite			Host Gabbro	
	TOB 10-3	TOB 64-5	TOB 64-6	TOB 2D core	TOB 2D rim
SiO ₂	51.56	52.02	50.91	52.32	52.16
TiO ₂	0.80	1.05	1.13	0.57	0.58
Al ₂ O ₃	2.06	2.15	3.34	2.90	2.58
Cr ₂ O ₃	0.01	0.00	0.01	0.50	0.56
FeO	13.73	12.50	11.64	6.23	6.00
MnO	0.47	0.33	0.36	0.15	0.09
MgO	12.83	13.37	13.02	16.83	16.32
CaO	18.79	18.77	19.15	20.84	20.95
Na ₂ O	0.32	0.38	0.44	0.31	0.28
K ₂ O	0.00	0.00	0.00	0.00	0.00
Total	100.57	100.57	100.00	100.65	99.52
Mg#	62.50	65.58	66.61	86.25	83.93
En	37.40	39.27	38.83	48.80	47.30
Fs	23.22	21.14	20.11	7.78	9.06
Wo	39.38	39.59	41.07	43.43	43.64

Table 2: Chemical composition of primary and secondary plagioclase from Újhatár Valley (UH) and Tóberc quarry (TOB) pegmatites. Ab = albite, An = anortite, Or = orthoclase.

Sample no.	Primary				Secondary		
	TOB 6-1	TOB 6-5	UH 11-2	UH 11-4	TOB 6-3	TOB 6-4	TOB 6-6
SiO ₂	55.73	57.76	69.65	69.45	65.10	68.05	66.31
Al ₂ O ₃	27.57	27.17	19.68	19.74	21.67	20.60	21.78
FeO	0.84	0.30	0.19	0.18	0.14	0.14	0.34
CaO	9.81	9.12	0.12	0.20	2.87	1.20	1.61
Na ₂ O	5.78	6.27	11.71	11.43	9.51	10.87	9.78
K ₂ O	0.09	0.16	0.05	0.14	0.18	0.07	0.70
SrO	0.00	0.00	0.00	0.00	0.00	0.00	0.08
Total	99.83	100.77	101.40	101.13	99.46	100.93	100.60
Ab	51.4	55.0	99.2	98.3	84.8	93.9	87.9
An	48.1	44.1	0.6	1.0	14.1	5.7	8.0
Or	0.5	0.9	0.3	0.8	1.0	0.4	4.2

Fe-tschermakite, hastingsite, Fe-hastingsite and pargasite are present, while edenite and Fe-edenite are the dominant compositions in the Tóberc quarry according to classification of Leake (1997). Halogen contents are relatively high for both localities, but they are mostly higher at the Tóberc quarry (Table 3), with 0.1–0.2 wt. % Cl and 0.1–0.15 wt. % F (Fig. 4). In some cases, alteration of amphibole to chlorite occurs along fractures and cleavages.

Apatite is rare in the gabbro but it is a common accessory phase in pegmatites. Euhedral grains (with up to 2 mm length in pegmatites) occur as inclusions in amphibole, plagioclase and ilmenite. Compositional zoning has not been observed. The apatite from pegmatites is always fluorine-rich (1.1–2.3 wt. %) and contains 0.4–0.8 wt. % chlorine (Table 4 and Fig. 4). The total REE concentrations are between 3400 and 9200 ppm. Ce and Y are the most abundant elements followed by Nd and La. There is no significant difference in terms of halogen content between the localities but the total amount of REE is much

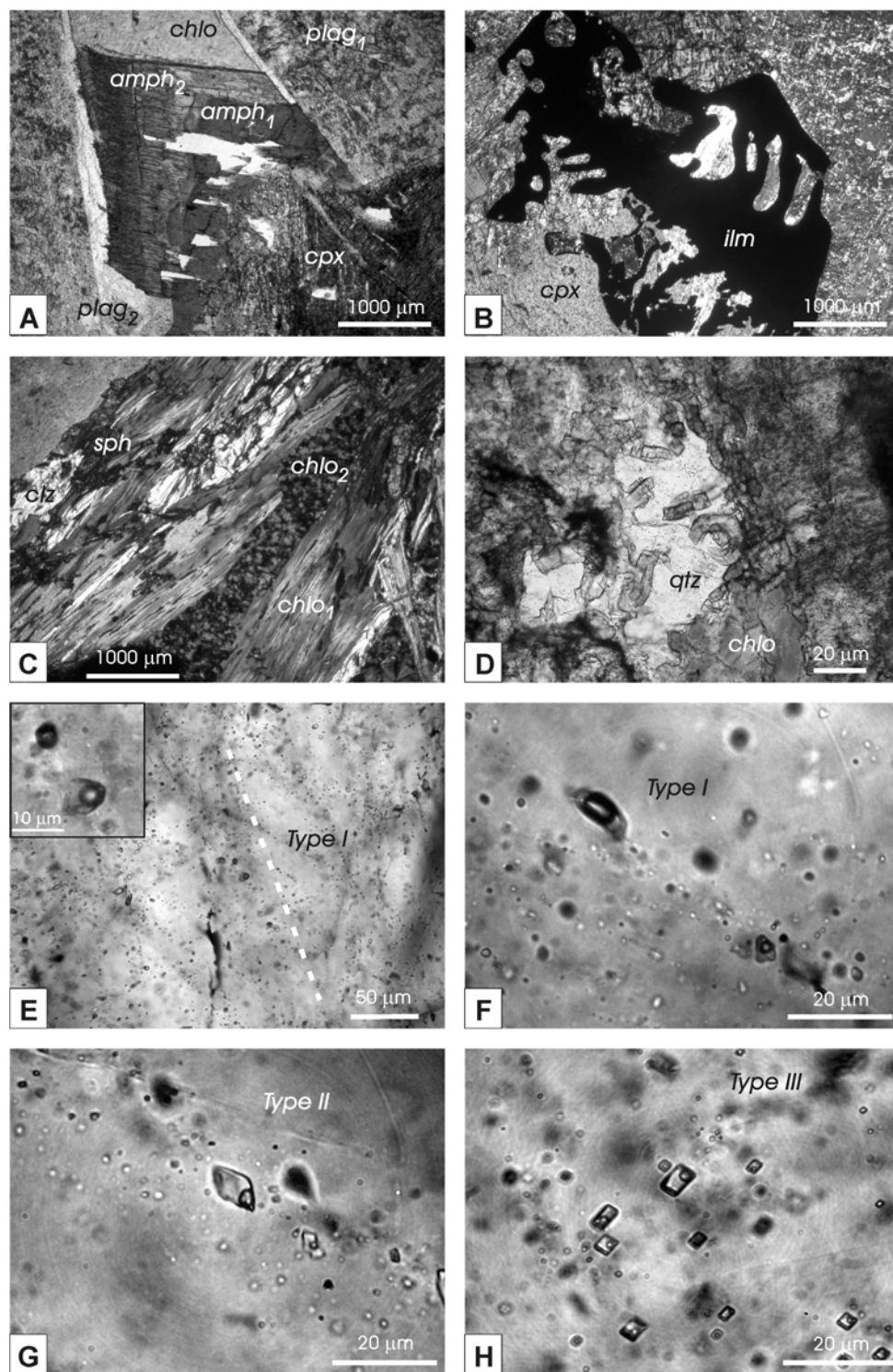


Fig. 3. Photomicrographs of gabbro pegmatites, showing mineral phases and fluid inclusion types. All photos in transmitted light, plane polars, except photo (B) and (C) in crossed polars. **A** — Typical mineral assemblage with primary and alteration phases. Clinopyroxene (cpx) is replaced by zoned deuterite (amph₁) and zoned sea-floor hydrothermal amphibole (amph₂). Primary plagioclase (plag₁) is altered and overgrown by a fresh rim of albite (plag₂). The vug is filled by fine-grained chlorite (chlo). Fibrous Fe-actinolite (last amphibole generation) has suffered brittle deformation. **B** — Clinopyroxene (cpx) intergrown with primary anhedral ilmenite (ilm). **C** — Chlorite (chlo₁), sphene (sph) and clinozoisite (clz) replacing biotite. Fine-grained chlorite (chlo₂) is also visible. **D** — Alpine metamorphic quartz (qtz) intergrown with fine-grained chlorite (chlo). **E** — Magmatic quartz containing solid- and fluid-inclusions. Broken line shows strike of secondary fluid inclusion plane hosting fluid type I. **F** — Secondary fluid inclusion plane of type I fluid in magmatic quartz. Note different vapour ratio at room temperature. **G** — Secondary fluid inclusion plane of type II fluid in magmatic quartz. Note the 15 vol. % of vapour at room temperature. **H** — Primary assemblage of type III fluid inclusions in metamorphic calcite.

Table 3: Chemical composition of magmatic and hydrothermal amphibole from Újhatár Valley (UH) and Tóberc quarry (TOB) pegmatites. Act = actinolite, Ed = edenite, Grun = grunerite, Hast = hastingsite, Hb = hornblende, Parg = pargasite, Tsch = tschermakite; (Fe-) means always the ferro-endmember, (Mg-) means the magnesio-endmember; Mg# = $Mg/(Mg + Fe^{2+})$ (Leake 1997). Number of cations based on 23 (O).

Sample no.	Magmatic				Deuteric			Sea-floor hydrothermal			
	UH 3-3	UH 3-7	UH 11-1	TOB 10-1	UH 3-6core	TOB 10-2core	TOB 19-7rim	TOB 64-1	TOB 64-2	TOB 19-6	TOB 19-4
	Hast	Parg	Fe-Tsch	Fe-Ed	Mg-Tsch	Hast	Fe-Hb	Grun	Grun	Fe-Act	Fe-Act
SiO ₂	41.55	41.27	43.04	43.44	44.30	42.73	46.23	49.84	47.37	49.44	50.28
TiO ₂	3.03	3.43	2.73	2.51	0.34	2.71	1.66	0.46	0.45	0.61	0.08
Al ₂ O ₃	10.65	11.83	10.32	8.47	11.83	9.31	6.48	2.22	2.25	3.64	2.43
Cr ₂ O ₃	0.00	0.00	0.00	0.00	0.00	0.00	0.03	0.02	0.00	0.00	0.02
FeO	20.38	18.04	21.86	23.22	18.33	21.75	22.80	31.06	39.01	24.43	25.93
MnO	0.30	0.25	0.41	0.38	0.28	0.41	0.36	0.71	1.05	0.33	0.36
MgO	8.33	9.16	7.96	7.64	10.74	8.03	8.29	9.49	4.75	7.47	6.77
CaO	10.27	10.86	10.33	9.96	10.00	9.92	10.27	2.53	1.98	10.88	11.14
Na ₂ O	2.39	2.73	2.10	2.39	2.24	2.61	1.86	0.58	0.51	0.76	0.29
K ₂ O	0.21	0.25	0.29	0.28	0.30	0.42	0.27	0.12	0.02	0.32	0.05
Cl	0.06	0.10	n.a.	0.22	0.05	0.16	0.19	0.18	0.02	0.09	0.03
F	0.05	0.04	n.a.	0.16	0.07	0.14	0.04	0.00	0.01	0.00	0.00
Total	97.22	97.96	99.04	98.67	98.48	98.19	98.25	97.21	97.42	97.88	97.38
Si	6.330	6.215	6.439	6.599	6.491	6.498	6.977	7.702	7.605	7.513	7.728
Al (IV)	1.670	1.785	1.561	1.401	1.509	1.502	1.023	0.298	0.395	0.487	0.272
Al (VI)	0.242	0.315	0.260	0.115	0.533	0.167	0.129	0.106	0.031	0.165	0.168
Ti	0.347	0.388	0.307	0.287	0.037	0.310	0.188	0.053	0.054	0.070	0.009
Cr	0.000	0.000	0.000	0.000	0.000	0.000	0.004	0.002	0.000	0.000	0.002
Fe ³⁺	0.400	0.220	0.451	0.450	0.680	0.400	0.390	0.150	0.210	0.210	0.100
Fe ²⁺	2.190	2.050	2.284	2.500	1.570	2.370	2.490	3.860	5.030	2.900	3.230
Mg	1.891	2.056	1.775	1.730	2.345	1.820	1.865	2.186	1.137	1.692	1.551
Mn	0.039	0.032	0.052	0.049	0.035	0.053	0.046	0.093	0.143	0.042	0.047
Ca	1.676	1.752	1.657	1.621	1.570	1.616	1.660	0.419	0.341	1.771	1.834
NaB	0.214	0.187	0.214	0.248	0.229	0.264	0.228	0.131	0.055	0.150	0.058
NaA	0.492	0.610	0.395	0.456	0.407	0.505	0.316	0.043	0.104	0.074	0.029
K	0.041	0.048	0.055	0.054	0.056	0.081	0.052	0.024	0.004	0.062	0.010
Total	15.532	15.658	15.449	15.510	15.463	15.587	15.368	15.067	15.108	15.136	15.039
Cl	0.015	0.026	n.a.	0.056	0.013	0.041	0.048	0.046	0.005	0.024	0.008
F	0.024	0.019	n.a.	0.080	0.031	0.066	0.018	0.000	0.004	0.000	0.000
Mg#	46.3	50.1	43.7	40.9	59.9	43.4	42.8	36.2	18.4	36.8	32.4

higher and local Ba and Sm enrichments also occur in apatite from the Tóberc quarry.

Biotite forms flakes up to 1 cm around amphibole in pegmatites, while it is rare in the host gabbro. The biotite flakes are mostly replaced by chlorite.

Quartz is the last crystallizing phase in the central parts of pegmatites in the Tóberc quarry; however, it is completely absent in the pegmatites of the Újhatár Valley and in the host gabbro. This quartz contains many inclusions of apatite and fluids (Fig. 3E).

Ilmenite is often intergrown with pyroxene (Fig. 3B) and forms rims around pyroxene. Ilmenite also occurs as exsolution lamellae or anhedral inclusions in magnetite both in the gabbro and its pegmatites.

Magnetite is fresh in the gabbro, but altered to brown amphibole and biotite in pegmatites. In this latter case only the ilmenite exsolution lamellae remained from the original titanomagnetite grains.

Alteration of primary minerals

Petrography revealed three alteration stages of primary minerals developed in unequal extent at various localities.

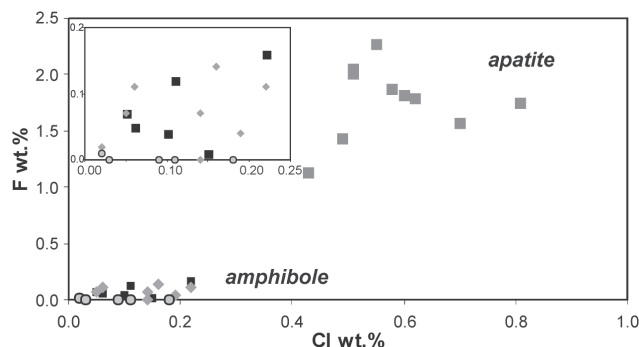


Fig. 4. Cl versus F content of amphibole and apatite from the Újhatár Valley and Tóberc quarry. The inset shows halogene content of amphibole in detail. Note the variable F and Cl content in magmatic and deuteric amphibole, while hydrothermal amphibole contain almost no fluorine. ■ — magmatic, ◆ — deuteric, ● — hydro-thermal amphibole compositions, ■ — apatite.

The first alteration was probably caused by fluids coexisting with the parent melt of pegmatites and is thus defined as a deuteric alteration. The deuteric alteration is absent in the gabbro and occurs only in pegmatites. This stage is

Table 4: Chemical composition of apatite from the Újhatár Valley (UH) and Tóbérc quarry (TOB) pegmatites. ¹ phase hosting the apatite inclusion (amph = amphibole, plag = plagioclase) ² OH and H₂O calculated on the basis of (F+Cl+OH) = 1 per formula unit; X_F, X_{Cl}, X_{OH} = mole fraction of F, Cl, OH, respectively.

Host ¹	UH3-1-centre	UH3-1-rim	UH3-2	UH3-3	TOB64-1	TOB64-2	TOB64-3	TOB10-1	TOB10-2	TOB10-3
	amph	amph	amph-plag	plag	amph	amph	amph	plag	plag	plag
P ₂ O ₅	42.31	41.79	41.83	41.18	42.08	41.21	42.20	42.56	42.28	41.99
SiO ₂	0.20	0.26	0.17	0.31	0.28	0.19	0.24	0.16	0.21	0.16
As ₂ O ₅	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
SO ₃	0.00	0.00	0.00	0.01	0.02	0.00	0.00	0.04	0.01	0.01
CaO	54.74	54.64	55.87	55.68	54.87	55.07	55.32	54.67	54.30	54.75
Na ₂ O	0.05	0.07	0.05	0.03	0.02	0.07	0.10	0.05	0.08	0.07
FeO	0.46	0.53	0.34	0.25	0.51	0.50	0.50	0.49	0.58	0.49
MnO	0.11	0.22	0.10	0.13	0.13	0.09	0.06	0.06	0.10	0.07
SrO	0.01	0.02	0.01	0.00	0.01	0.04	0.02	0.05	0.02	0.02
BaO	0.02	0.00	0.00	0.00	0.00	0.07	0.09	0.14	0.17	0.18
Y ₂ O ₃	0.13	0.07	0.08	0.10	0.23	0.16	0.22	0.15	0.24	0.16
La ₂ O ₃	0.04	0.16	0.10	0.09	0.05	0.01	0.06	0.04	0.15	0.05
Ce ₂ O ₃	0.12	0.20	0.12	0.12	0.11	0.15	0.18	0.17	0.31	0.20
Pr ₂ O ₃	0.03	0.02	0.04	0.08	0.00	0.06	0.05	0.00	0.08	0.00
Nd ₂ O ₃	0.04	0.14	0.00	0.00	0.15	0.16	0.22	0.08	0.14	0.09
Sm ₂ O ₃	0.04	0.00	0.00	0.08	0.11	0.10	0.14	0.00	0.00	0.00
H ₂ O ²	1.13	0.74	0.70	0.97	0.77	0.75	0.58	0.86	0.68	0.74
F	1.12	1.74	2.00	1.43	1.79	1.81	2.27	1.56	2.04	1.87
Cl	0.43	0.81	0.51	0.49	0.62	0.60	0.55	0.70	0.51	0.58
Total	99.85	100.68	101.22	99.98	100.98	100.29	102.22	100.92	101.22	100.69
O=F,Cl	0.57	0.91	0.96	0.71	0.89	0.90	1.08	0.81	0.97	0.92
Total	99.28	99.77	100.26	99.27	100.09	99.39	101.14	100.11	100.25	99.77
X _F	0.30	0.46	0.53	0.38	0.47	0.49	0.60	0.41	0.54	0.50
X _{Cl}	0.06	0.12	0.07	0.07	0.09	0.09	0.08	0.10	0.07	0.08
X _{OH} ²	0.64	0.42	0.40	0.55	0.44	0.43	0.32	0.49	0.39	0.42

mainly characterized by the formation of zoned amphibole. Clinopyroxene has been partly replaced by brown amphibole, which has a composition similar to interstitial and euhedral brown amphibole. However, replacing amphibole has a gradual change of composition toward the rim, with decrease in Na, Al, Ti and F contents and increase in Si content (Table 3, Fig. 5). This is also manifested by change of their colour from brown to green (Fig. 3A). The amphibole zoning is typical of pegmatites in the Tóbérc quarry, and it is less developed in the pegmatites of the Újhatár Valley. The green zones of the amphibole correspond to Fe-hornblende (Leake 1997). Alteration of magnetite in pegmatites appears to be synchronous with the amphibole alteration of pyroxene. Magnetite crystals are totally decomposed, but their ilmenite exsolution lamellae have remained intact. During this process biotite flakes crystallized among the lamellae. Alteration of plagioclase to sericite probably started at this stage but was overprinted during later events.

The early stage (deuteric) alteration was overprinted by sea-floor hydrothermal alteration in pegmatites and their host gabbro of the Tóbérc quarry. This type of alteration is very weakly developed in the Újhatár Valley. During this stage, plagioclase underwent intense sericitization, albitization and saussuritization. Biotite was replaced by fine-grained chlorite and clinozoisite (Fig. 3C), and titanite crystallized along cleavage planes. The outer rims of amphibole and pyroxene show chloritization, which can be accompanied by pyritization. According to the classification of Hey (1954),

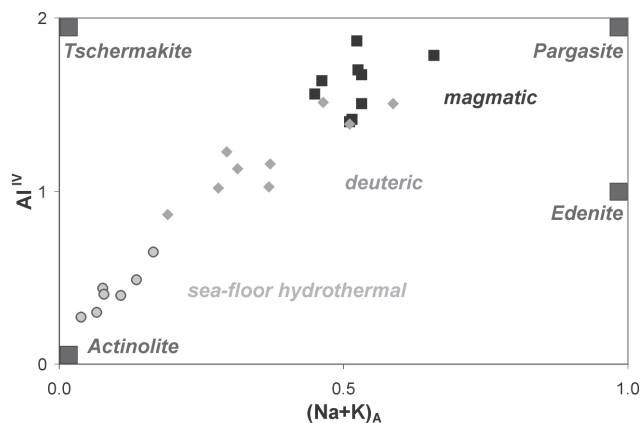


Fig. 5. Amphibole compositions plotted in an Al^{IV} versus (Na+K)_A diagram. ■ — magmatic, ◆ — deuteric, ● — hydrothermal amphibole compositions.

chlorite corresponds to ripidolite and brunsvigite (Table 5). Around the zoned brownish-greenish deuteric amphibole, greenish-colourless amphibole also formed during the hydrothermal alteration. The early and late amphibole grains exhibit sharp boundaries (Fig. 3A). Hydrothermal amphibole also occurs as euhedral, wedge-shaped crystals on the surface of earlier amphibole grains. The chemical compositions of these hydrothermal amphiboles correspond to ferroactinolite and grunerite (Table 3). They are also commonly intergrown with chlorite and overgrown by more ferro-

Table 5: Chemical composition and calculated formation temperature (T) of chlorite from the Újhatár Valley (UH) and Tóbérc quarry (TOB). (Chlorite in vug; replacing amphibole, biotite or plagioclase) (Temperature in °C, K+ML: according to Kranidiotis & MacLean 1987; Z+F: according to Zang & Fyfe 1995).

Sample no.	TOB19-3	TOB6-1	TOB6-3	UH11-3	UH3-2
	brunsvigite	ripidolite	brunsvigite	ripidolite	brunsvigite
Position	repl. amph	in vug	repl. bio	repl. plag	repl. bio
SiO ₂	26.60	25.68	26.25	25.12	25.72
Al ₂ O ₃	18.92	20.11	17.54	19.72	19.81
TiO ₂	0.04	0.08	0.13	0.05	0.04
Cr ₂ O ₃	0.01	0.00	0.00	0.00	0.02
FeO	30.45	30.70	37.49	33.38	34.41
MnO	0.28	0.26	0.27	0.29	0.31
MgO	11.70	11.83	7.52	9.82	8.04
K ₂ O	0.01	0.05	0.09	0.00	0.15
CaO	0.14	0.00	0.00	0.00	0.17
Cl	0.01	n.d.	n.d.	n.d.	0.05
F	0.02	n.d.	n.d.	n.d.	0.00
	88.18	88.71	89.28	88.39	88.72
number of cations based on 28 (O)					
Si	5.728	5.505	5.809	5.497	5.633
Al (IV)	2.272	2.495	2.191	2.503	2.367
Sum	8.000	8.000	8.000	8.000	8.000
Al (VI)	2.530	2.587	2.382	2.582	2.746
Ti	0.006	0.012	0.022	0.009	0.006
Cr	0.002	0.000	0.000	0.000	0.003
Fe ²⁺	5.485	5.505	6.936	6.107	6.304
Mn	0.052	0.047	0.050	0.054	0.057
Mg	3.755	3.782	2.481	3.201	2.626
K	0.003	0.013	0.025	0.000	0.041
Ca	0.032	0.000	0.000	0.000	0.040
Sum	11.865	11.947	11.896	11.952	11.823
T (K+ML)	302	326	304	332	321
T (Z+F)	236	259	213	254	235

actinolitic amphibole (TOB19-4 in Table 3). This dark green (even more Fe-rich) fibrous ferro-actinolite is intergrown with newly formed albite (Fig. 3A), which also forms fresh rims around the altered primary plagioclase. The hydrothermal amphibole is devoid of fluorine, and is characterized by variable chlorine content up to 0.18 wt. % (Table 3 and Fig. 4). The Al^{IV} vs. (Na+K)_A diagram of magmatic, deuteric and hydrothermal amphibole generations (Fig. 5) shows the “pargasitic trend” which has been described from recent and ancient ocean-floor gabbros (Prichard & Cann 1982; Ito & Anderson 1983; Mével 1988; Talbi et al. 1999). The albite has a minor anorthite component (An₅₋₈) and its Or content is up to 4.2 mol % (Table 2). Titanite formed around ilmenite lamellae at this stage of alteration (Sadek et al. 1996).

In the Újhatár Valley, the sea-floor hydrothermal alteration is manifested by sericitic alteration of plagioclase. Newly formed albite and amphibole generations are sporadic. Alteration of biotite is the same as described above.

The third stage of alteration can be related to the Alpine low-grade metamorphic event and can thus be observed in the whole SOC. The prehnite-pumpellyite facies mineralization (Sadek et al. 1996) appears mostly in veinlets. The

main phases in the veinlets are prehnite, calcite and quartz. These minerals are intergrown with fine-grained chlorite (Fig. 3D), corresponding to ripidolite (Table 5). The metamorphic assemblage locally replaces plagioclase. Deformation related to the low-grade metamorphism is widespread, and locally resulted in formation of cataclastic rock fabrics. Ductile deformation can also be observed as kink bands (Fig. 3A) in amphibole and folding of the chlorite pseudomorphs after biotite.

Mineral thermometry

Formation conditions of the pegmatites are difficult to determine because of the intense alteration of most of the primary phases. Remnants of fresh coexisting amphibole and plagioclase from the Újhatár Valley yielded formation temperatures of about 800–850 °C (Holland & Blundy 1994). This temperature range appears to be geologically acceptable; however, the use of this geothermometer is not recommended for such acid plagioclase. The semiquantitative, Al- and Ti-in-amphibole thermometer of Ernst & Liu (1998) yielded temperatures of 900–950 °C and 880 °C for samples from the Újhatár Valley and the Tóbérc quarry, respectively. Aigner-Torres (1996) established formation temperatures of up to 930 °C for the host gabbro using the amphibole-plagioclase thermometer of Holland & Blundy (1994). In general, the available data indicate formation temperatures of the SOC pegmatites between 800 and 900 °C.

Deuteric alteration took place immediately after formation of primary phases and the temperature was continuously decreasing as indicated by the zoned amphibole. The mineral assemblage, textural features, and the amphibole thermometer of Ernst & Liu (1998) indicate magmatic-submagmatic conditions and a temperature of 700–900 °C.

The sea-floor hydrothermal assemblage corresponds to greenschist facies conditions as was also pointed out by Sadek et al. (1996). Similar phases were reported from ocean-floor gabbros, which suffered hydrothermal alteration of this grade (Ito & Anderson 1983; Talbi et al. 1999). The presence of actinolite instead of hornblende confirms the temperatures below 400 °C (Gillis 1995; Talbi et al. 1999). The Al^{IV}-content of chlorite replacing biotite and amphibole can also be used as a thermometer. Chlorite thermometers of Cathelineau (1988) and Kranidiotis & MacLean (1987) yielded almost similar temperatures of 290–330 °C, while the equation of Zang & Fyfe (1995) always provided lower temperatures of about 210–260 °C. As the thermometer of Cathelineau (1988) is not recommended for such iron-rich chlorite, and according to Sadek et al. (1996) the composition of chlorite in SOC is dependent on bulk-rock compositions, care must be exercised when interpreting the results of the chlorite geothermometry.

The Alpine regional metamorphism is characterized by prehnite-pumpellyite facies conditions (Árkai 1983; Árkai et al. 1995; Sadek et al. 1996), and a maximum tempera-

ture of 250–300 °C (Árkai et al. 1995). Chlorite thermometers of Cathelineau (1988) and Kranidiotis & MacLean (1987) yielded 325–340 °C, while that of Zang & Fyfe (1995) 250–260 °C.

Fluid inclusion studies

Three types of fluid inclusions were distinguished on the basis of their phase compositions and origin in pegmatitic and hydrothermal quartz, as well as calcite.

Type I inclusions are secondary and occur in healed cracks of pegmatitic quartz. These inclusions have variable liquid to vapour ratios at room temperature (Fig. 3E and F) from about 20–30 vol. % vapour phase to essentially vapour-dominated inclusions. The variable volumetric phase ratios indicate boiling of their parent fluids. The aqueous liquid-dominated inclusions homogenized between 297 and 334 °C (Fig. 6). The average salinity of the aqueous phase corresponds to 3 NaCl equiv. wt. % (Fig. 7).

Type II inclusions also occur along secondary fractures in pegmatitic quartz (Fig. 3G), and exhibit constant, 15–20 vol. % vapour phase at room temperature. Homogenization to liquid occurred between 218 and 257 °C (Fig. 6). The average salinity of these inclusions is 3.4 NaCl equiv. wt. %, but the scatter is larger than that for the Type I fluid inclusions (Fig. 7).

Type III inclusions are primary in quartz and calcite from prehnite-bearing veins, thus being indicative of the low grade metamorphic overprint in the SOC. Type III inclusions are isolated in quartz or form populations along growth zones of calcite (Fig. 3H): accidentally trapped chlorite flakes confirm a primary origin for these inclusions, because the chlorite is deposited together with quartz and calcite in late veinlets. The vapour phase occupies about 10 vol. % at room temperature. Homogenization of type III inclusions took place between 159 and 187 °C (Fig. 6), and salinities ranged from 1.4 to 6.45 NaCl equiv. wt. % with an average of 3.4 wt. % (Fig. 7).

Discussion

Our observations reveal that orthopyroxene and olivine occur in subordinate amounts, and Fe-Ti-oxides are common in gabbro of SOC. The differentiated character of the SOC gabbro is also indicated by the mineral chemistry. This study has shown many differences between differentiation and the late crystallization products of the two studied localities. At the Tóbérc quarry, strongly contaminated ferrogabbro occurs, which most probably formed in the roof zone of a magma chamber. The interaction of the melt with the enclosing sedimentary rocks is indicated by assimilation of sedimentary xenoliths and occurrences of contact rocks. These features provide evidence of reaction between gabbro and the surrounding sedimentary rocks, and suggest mobilization of volatiles and other elements during emplacement of gabbro intrusions.

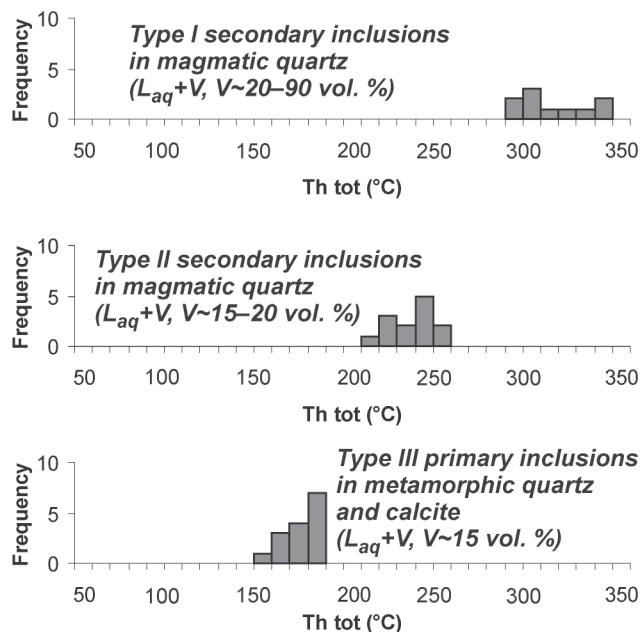


Fig. 6. Frequency distribution diagram of homogenization temperatures (Th tot) for fluid inclusion types from pegmatitic quartz of Tóbérc quarry.

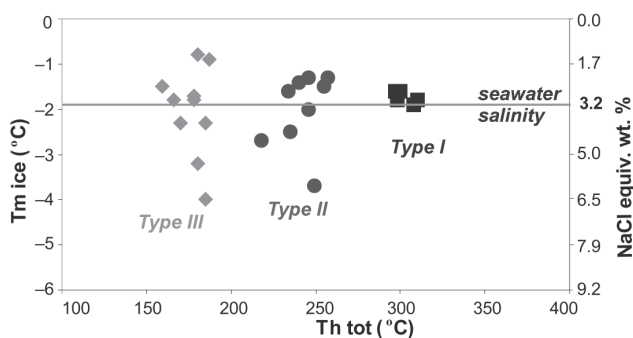


Fig. 7. Plot of homogenization temperature (Th tot) versus final melting temperature of ice (Tm ice) for fluid inclusion types from pegmatitic quartz of Tóbérc quarry. Black line represents final melting temperature of ice for mean sea water. Salinities (NaCl equiv. wt. %) calculated from final melting temperature of ice.

The original fluid contents of the various melts cannot be established, but it is obvious that the formation of pegmatites was promoted by the assimilated volatiles. During crystallization, the melt became fluid-enriched, and this process resulted in formation of different pegmatitic structures. The gabbro contains small patches where the increased concentration of fluids is visualized by increasing grain size of the rock-forming minerals. The pegmatitic pockets indicate segregation of a hydrous melt, which was enriched in fluids, volatiles (e.g. halogens), and other incompatible elements. This is revealed by the fact that pegmatites contain abundant hydrous silicates, apatite and quartz. The major element composition of the rock-forming phases and the abundance of Fe-Ti-oxides show that iron-content increased in the pegmatites. Zon-

ing of some pegmatitic pockets with a pyroxene-rich rim and a quartz-rich core, accompanied by increasing grain size indicates crystallization from margins to the core. Similar zoned pegmatites have also been described from orogenic gabbroic plutons (Lovering & Durrell 1959; Beard & Day 1986). The homogeneous pegmatite pockets have most probably crystallized simultaneously, and the melt could not differentiate. Cooling and synchronous tectonism of the crystallizing gabbro resulted in fracturing of the rock, and the segregated hydrous melt could also move along the fractures. This is shown by the "tails" of the pegmatite pockets and occurrences of narrow pegmatitic dykes. The ductile deformation of the "tails" reveals that formation of pockets occurred between solidus and liquidus of the host gabbro. The straight strike, but diffuse contacts of the narrow dykes show that the host gabbro was brittle, but still hot enough to react with the hydrous melt. The thick dykes with sharp contacts correspond to a later differentiation stage, which is also supported by their strongly evolved composition. In contrast with the former pegmatites, their melt seems not to have been derived by a local segregation. It is more probable that the parent melts segregated in a different part of the crystallizing magma chamber, and were intruded later into this rock unit.

Gabbroic rocks in the Újhatár Valley are highly evolved, homogeneous fine-grained amphibole gabbros. The parent melt may have originally been more differentiated, but was not affected by assimilation. The evolved character is confirmed by mineral composition and chemistry (amphibole instead of ortho- and clinopyroxene, almost pure primary albite). The absence of sedimentary xenoliths and contact rocks indicate a greater distance from the contact and/or a deeper position in the magma chamber. The latter is also favoured by the absence of late pegmatitic dykes and plagiogranitic intrusions. The greater distance from the contact resulted in less extensive assimilation and contamination of the melt, which could explain the absence of quartz, and the lower halogen, and rare earth contents, in comparison to Tóberc. The origin of fluids responsible for the formation of pegmatitic segregations was dominantly magmatic.

Exsolution of magmatic fluids is revealed by deuteric alteration in many gabbroic systems containing pegmatites (Ballhaus & Stumpfl 1986; Beard & Day 1986; Watkinson & Ohnenstetter 1992; Li et al. 2004). Separation of a fluid phase during the formation of pegmatites at Szarvaskő is evident by quartz-feldspar granophyric texture and the composition of apatite with F/(F+Cl) ratios of about 0.8–0.9. During fractional crystallization of the melt, the ratio of F versus Cl does not change; therefore this process cannot be responsible for the formation of the fluorine-rich apatite (Boudreau & McCallum 1989). In contrast, fluid separation leads to enrichment of chlorine in the fluid phase, while fluorine remains in the residual melt. Thus, apatite precipitating from the melt is fluorine rich (Boudreau & McCallum 1989; Meurer & Boudreau 1996; Meurer & Natland 2001). The segregated fluid could not be identified in our fluid inclusion studies, which is most probably due to decrepitation of inclusions during later hydrothermal processes.

Fluid segregation accompanying pegmatite formation resulted in deuteric alteration of minerals. Alteration of magnetite, formation of biotite, replacement of pyroxene by brown amphibole, and the formation of zoned amphibole can be related to this process. Change of chemical composition in amphibole indicates cooling from magmatic to sub-magmatic temperatures, where hornblende was the stable phase. The high fluorine content of amphiboles, up to 0.14 wt. %, may be explained by magmatic fluid segregation rather than by sea-floor hydrothermal origin. Coogan et al. (2001) distinguished magmatic and sea-floor hydrothermal amphibole in the Mid-Atlantic Ridge gabbros in a similar way.

According to studies on recent ocean-floor gabbro, brittle deformation of rock accompanied by seawater influx is significant below a temperature of about 800 °C (Ito & Anderson 1983; Talbi et al. 1999; Coogan et al. 2001). Temperature of the seawater-dominated hydrothermal systems depends on the depth and the distance from the ridge axis. The character of alteration caused by this hydrothermal process is further dependent on spreading rate, cooling rate and the fracturing of rock (Ito & Anderson 1983; Mével 1988; Gillis 1995; Talbi et al. 1999).

Sea-floor hydrothermal alteration occurs in all rocks of the SOC and was first described by Sadek et al. (1996). In the Tóberc quarry, our studies revealed a mineral assemblage corresponding to greenschist facies conditions and temperatures below 400 °C. The hydrothermal alteration was a polyphase process, which is documented by several generations of Fe-actinolite and grunerite. Intense chloritization, sericitization and the formation of albite, clinozoisite and titanite are also typical of this alteration. Sea-floor hydrothermal alteration of the Újhatár Valley rocks is less intense, which can be explained by a greater original depth to the sea floor and/or less intense fracturing of the rock body which hindered circulation of seawater.

In our fluid inclusion studies we could identify two fluid generations (type I and II) responsible for the sea-floor hydrothermal alteration. Both fluids have salinities clustering around that of mean seawater (~3.2 ‰). The slight differences may have been caused by fluid/rock interactions. Similar boiling fluids to type I were found by Vanko et al. (1992) in gabbro of the Mid-Atlantic Ridge. Assuming a pure NaCl-H₂O system, boiling of the fluid should have occurred at around 7–8 MPa of pressure (Zhang & Frantz 1987). If calculated under hydrostatic pressure this would mean a depth of 800 m below sea level. This shallow depth could be plausible for a sea-floor in a back-arc basin environment; however, the presence of a minor amount of CO₂ that is undetectable by petrography and microthermometry cannot be excluded (Hedenquist & Henley 1984). Therefore the estimated paleodepth is the minimum depth for entrapment of boiling fluids in secondary inclusions of pegmatitic quartz (which has certainly formed at a greater depth). Fluids similar to our fluid type II, were also reported from recent oceanic environments (Vanko et al. 1992; Kelley & Früh-Green 2001). The mean homogenization temperature of about 240 °C gives a

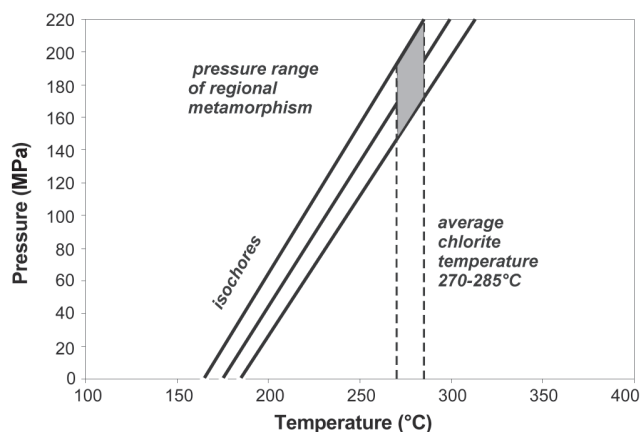


Fig. 8. Modeling of entrapment conditions for type III fluid inclusions. Grey coloured area shows proposed entrapment conditions. Isochores were calculated according to Zhang & Frantz (1987).

minimum entrapment temperature, and a lower limit for the sea-floor hydrothermal alteration.

Being primary in phases of the Alpine low-grade metamorphic veining, type III fluid inclusions are indicative of that process. Considering the fact that the fine-grained chlorite, calcite and quartz were growing together, chlorite thermometry and isochores of fluid inclusions define entrapment conditions of the low grade metamorphic fluids (Fig. 8), ranging between 270–285 °C and 150–200 MPa.

Conclusions

Pegmatitic structures studied in two areas in the SOC may be classified according to their textural and mineralogical properties. Pegmatitic patches, pockets and narrow dykes precipitated from a locally segregated hydrous melt,

whereas thick and felsic dykes were intruded later. Homogeneous pegmatitic pods crystallized simultaneously in contrast to zoned pods which indicate further differentiation of the hydrous melt. Variation of the mineral assemblages and their chemical composition in pegmatite bodies was also influenced by enrichment of volatiles and incompatible elements in the melt due to interaction with the host sedimentary rocks. The formation temperatures of the pegmatites are difficult to determine, but their crystallization most probably occurred between 800 and 900 °C (Table 6).

During crystallization of pegmatites a fluid phase separated, as indicated by granophyric textures and apatite compositions. This magmatic fluid caused deuteritic alteration of the primary pegmatitic assemblage, which is revealed by alteration of magnetite and pyroxene, accompanied by formation of biotite and zoned amphibole. This process took place under continuously cooling magmatic-submagmatic temperatures.

The pegmatites and host gabbro underwent postmagmatic alteration due to sea-floor hydrothermal activity: the responsible fluids were identified in two fluid inclusion generations. Typical of this process is alteration of all primary phases and formation of Fe-actinolite, grunerite, chlorite, clinozoisite and albite. Mineral assemblage, mineral- and fluid-inclusion thermometry indicate a polyphase hydrothermal process between 250 and 400 °C.

The Alpine regional metamorphism caused brittle and ductile deformation of the rocks discernible at outcrop- and microscopic scales. This deformation was accompanied by intense veining of a prehnite-chlorite-quartz-calcite-feldspar assemblage. This low-grade Alpine metamorphic overprint occurred at around 270–285 °C at 150–200 MPa.

Acknowledgments: This work was supported by the NSF (OTKA) No. T 049633 to F. Molnár and by the Cana-

Table 6: Synthetic table summarizing main characteristics of magmatic, hydrothermal and metamorphic processes revealed in gabbro pegmatites from the Tóberc quarry. ¹mean values.

Process	Mineral Assemblage	Temperature	Fluid
Magmatic	clinopyroxene, plagioclase, brown amphibole, apatite, Fe-Ti-oxides, biotite, quartz	~ 800–900 °C (mineral assemblage; Ernst & Liu 1998; Holland & Blundy 1994)	–
Deuteritic	brown amphibole, hornblende, biotite, ilmenite lamellae	~ 700–800 °C (mineral assemblage; Ernst & Liu 1998; Holland & Blundy 1994)	exsolved magmatic fluid (not present as inclusions)
Sea-floor hydrothermal	Fe-actinolite, grunerite, chlorite, clinozoisite, albite	~ 250–400 °C (mineral assemblage; Kranidiotis & MacLean 1987; Zang & Fyfe 1995)	seawater derived fluids Type I: Th = 300 °C 2.96 wt. % NaCl equiv. ¹ Type II: Th = 238 °C 3.36 wt. % NaCl equiv. ¹
Alpine regional metamorphism	prehnite, chlorite, calcite, quartz, feldspar	~ 250–300 °C (mineral assemblage; Árkai et al. 1995) ~ 270–285 °C (Kranidiotis & MacLean 1987; Zang & Fyfe 1995)	metamorphic fluid Type III: Th = 175 °C 3.4 wt. % NaCl equiv. ¹

dian-Hungarian (CAN-2/04) and Croatian-Hungarian (HR-17/2004) Science and Technology Cooperation, both funded by the Technological and Innovation Fund, Hungary. Some funding from NSERC Grant A7874 to D.H. Watkinson is acknowledged. Authors express their thanks to Prof Friedrich Koller, University of Vienna for valuable discussions and suggestions. Significant improvements to the manuscript resulted from critical reviews by Dr. Vratislav Hurai and two anonymous referees.

References

- Aigner-Torres M. 1996: Petrology and geochemistry of the Szarvaskő igneous rocks Bükk Mountains — NE Hungary. *M.Sc. Thesis*, Vienna.
- Árkai P. 1983: Very low- and low-grade Alpine regional metamorphism of the Paleozoic and Mesozoic formations of the Bükkium, NE Hungary. *Acta Geol. Hung.* 26, 83–101.
- Árkai P., Balogh K. & Dunkl I. 1995: Timing of low-temperature metamorphism and cooling of the Paleozoic and Mesozoic formations of the Bükkium, innermost Western Carpathians, Hungary. *Geol. Rdsch.* 84, 334–344.
- Árváné Sós E., Balogh K., Ravaszné Baranyai L. & Ravasz Cs. 1987: K/Ar dates of Mesozoic igneous rocks in some areas of Hungary. *Ann. Rep. Hung. Geol. Inst.* 1985, 295–307 (in Hungarian).
- Balla Z. 1984a: The North Hungarian mesozoic mafics and ultramafics. *Acta Geol. Hung.* 27, 341–357.
- Balla Z. 1984b: On the Szarvaskő pillow lavas (Bükk Mts, N. Hungary). *Ann. Rep. Hung. Geol. Inst.* 1982, 72–100 (in Hungarian).
- Balla Z., Hovorka D., Kuzmin M. & Vinogradov V. 1983: Mesozoic ophiolites of the Bükk Mountains (North Hungary). *Ofioliti* 8, 5–46.
- Ballhaus C.G. & Stumpfl E.F. 1986: Sulfide and platinum mineralization in the Merensky Reef: evidence from hydrous silicates and fluid inclusions. *Contr. Mineral. Petrology* 94, 193–204.
- Beard J.S. & Day H.W. 1986: Origin of gabbro pegmatite in the Smartville intrusive complex, northern Sierra Nevada, California. *Amer. Mineralogist* 71, 1085–1099.
- Beard J.S., Fullagar P.D. & Krishna Sinha A. 2002: Gabbroic pegmatite intrusions, Iberia Abyssal Plain, ODP Leg 173, Site 1070: Magmatism during a Transition from Non-volcanic Rifting to Sea-floor Spreading. *J. Petrology* 43, 5, 885–905.
- Belák M., Pamić J., Kolar-Jurkovšek T., Pécskay Z. & Karan D. 1995: Alpine regional metamorphic complex of Medvednica Mt. (northwestern Croatia). *Proc. 1. I. Croatian Geol. Congress, Opatija. Inst. Geol., Croatian Geol. Soc., Zagreb*, 67–70 (in Croatian).
- Boudreau A.E. & McCallum I.S. 1989: Investigations of the Stillwater Complex: Part V. Apatites as indicators of evolving fluid composition. *Contr. Mineral. Petrology* 102, 138–153.
- Cathelineau M. 1988: Cation site occupancy in chlorites and illites as a function of temperature. *Clay Miner.* 23, 471–485.
- Coogan L.A., Wilson R.N., Gillis K.M. & MacLeod C.J. 2001: Near-solidus evolution of oceanic gabbros: Insights from amphibole geochemistry. *Geochim. Cosmochim. Acta* 65, 4339–4357.
- Csontos L. 1999: Structural outline of the Bükk Mts (N Hungary). *Földt. Közl.* 129, 4, 611–651 (in Hungarian).
- Csontos L. 2000: Stratigraphic reevaluation of the Bükk Mts (N Hungary). *Földt. Közl.* 130, 1, 95–131 (in Hungarian).
- Csontos L. & Vörös A. 2004: Mesozoic plate tectonic reconstruction of the Carpathian region. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 210, 1, 1–56.
- Downes H., Pantó G., Árkai P. & Thirlwall M.F. 1990: Petrology and geochemistry of Mesozoic igneous rocks, Bükk Mountains, Hungary. *Lithos* 24, 201–215.
- Ernst W.G. & Liu J. 1998: Experimental phase-equilibrium study of Al- and Ti-contents of calcic amphibole in MORB — A semiquantitative thermobarometer. *Amer. Mineralogist* 83, 952–969.
- Gillis K.M. 1995: Controls on hydrothermal alteration in a section of fast-spreading oceanic crust. *Earth Planet. Sci. Lett.* 134, 473–489.
- Harangi Sz., Szabó Cs., Józsa S., Szoldán Zs., Árvá-Sós E., Balla M. & Kubovics I. 1996: Mesozoic igneous suites in Hungary: Implications for genesis and tectonic setting in the Northwestern Part of Tethys. *Int. Geol. Rev.* 38, 336–360.
- Hedenquist J.W. & Henley R.W. 1985: The importance of CO₂ on freezing point measurements of fluid inclusions: Evidence from active geothermal systems and implications for epithermal ore deposition. *Econ. Geol.* 80, 1379–1406.
- Hey M.H. 1954: A new review of the chlorites. *Mineral. Mag.* 30, 277–292.
- Hoeck V., Koller F., Meisel T., Onuzi K. & Kneringer E. 2002: The Jurassic South Albanian ophiolites: MOR- vs. SSZ-type ophiolites. *Lithos* 65, 143–164.
- Holland T. & Blundy J. 1994: Non-ideal interactions in calcic amphiboles and their bearing on amphibole-plagioclase thermometry. *Contr. Mineral. Petrology* 116, 433–447.
- Ito E. & Anderson A.T. Jr. 1983: Submarine metamorphism of gabbros from the Mid-Cayman Rise: Petrographic and mineralogic constraints on hydrothermal processes at Slow-Spreading Ridges. *Contr. Mineral. Petrology* 82, 371–388.
- Kázmér J. & Kovács S. 1985: Permian-Paleogene paleogeography along the eastern part of the Insubric-Periadriatic Lineament system: Evidence for continental escape of the Bakony-Drauzug Unit. *Acta Geol. Hung.* 28, 71–84.
- Kelley D.S. & Früh-Green G.L. 2001: Volatile lines of descent in submarine plutonic environments: Insights from stable isotope and fluid inclusion analyses. *Geochim. Cosmochim. Acta* 65, 19, 3325–3346.
- Kovács S., Ebner F., Gawlick H.J., Pamić J., Haas J., Gaetani M., Castellaria A., Kräutner H.G., Karamata S., Krstić B., Sudar M., Mello J., Polák M. & Mioc D. 2004: Tectonostratigraphic terrane and paleoenvironment maps of the circum-Pannonian region: Maximum Neotethyan Spreading (Middle Jurassic) Environments. *Hung. Geol. Inst. Release*.
- Kranidiotis P. & MacLean W.H. 1987: Systematics of shlorite alteration at the Phelps Dodge Massive sulfide deposit, Matagami, Quebec. *Econ. Geol.* 82, 1898–1911.
- Kubovics I., Szabó Cs., Harangi Sz. & Józsa S. 1990: Petrology and petrochemistry of Mesozoic magmatic suites in Hungary and adjacent areas: An overview. *Acta Geod. Geoph. Mont. Hung.* 25, 3–4, 345–371.
- Leake B.E. 1997: Nomenclature of amphiboles. *Canad. Mineralogist* 35, 219–246.
- Less Gy., Gulácsi Z., Kovács S., Pelikán P., Pentelényi L., Rezessy A. & Sásdi L. 2002: Geological map of the Bükk Mts. (1:50,000). *Hung. Geol. Inst. Release* (in Hungarian).
- Li C., Ripley E.M., Merino E. & Maier W.D. 2004: Replacement of base metal sulfides by actinolite, epidote, calcite, and magnetite in the UG2 and Merensky Reef of the Bushveld Complex, South Africa. *Econ. Geol.* 99, 173–184.
- Lovering J.K. & Durrell C. 1959: Zoned gabbro pegmatites of Ereka peak, Plumas county, California. *J. Geol.* 67, 3, 253–268.
- Manning C.E., Weston P.E. & Mahon K.I. 1996: Rapid high-temperature metamorphism of East Pacific Rise gabbros from Hess Deep. *Earth Planet. Sci. Lett.* 144, 123–132.
- Mével C. 1988: Metamorphism of oceanic layer 3, Gorrige bank,

- Eastern Atlantic. *Contr. Mineral. Petrology* 100, 496–509.
- Meurer W.P. & Boudreau A.E. 1996: An evaluation of models of apatite compositional variability using apatite from the Middle Banded series of the Stillwater Complex, Montana. *Contr. Mineral. Petrology* 125, 225–236.
- Meurer W.P. & Natland J.H. 2001: Apatite compositions from oceanic cumulates with implications for the evolution of midocean ridge magmatic systems. *J. Volcanol. Geoth. Res.* 110, 281–298.
- Morimoto N. 1989: Nomenclature of pyroxenes. *Canad. Mineralogist* 27, 143–156.
- Pamić J. 1997: The northwesternmost outcrops of the Dinaridic ophiolites: a case study of Mt. Kalnik (North Croatia). *Acta Geol. Hung.* 40, 37–56.
- Pamić J., Gušić I. & Jelaska V. 1998: Geodynamic evolution of the Central Dinarides. *Tectonophysics* 297, 251–268.
- Pamić J., Tomljenović B. & Balen D. 2002: Geodynamic and petrogenetic evolution of Alpine ophiolites from the central and NW Dinarides: an overview. *Lithos* 65, 113–142.
- Prichard H.M. & Cann J.R. 1982: Petrology and mineralogy of Dredged Gabbro from Gettysburg Bank, Eastern Atlantic. *Contr. Mineral. Petrology* 79, 46–55.
- Sadek G.D. & Árkai P. 1994: Magmatic features and metamorphism of plagiogranite associated with a jurassic MORB-like basic-ultrabasic complex, Bükk Mountains, Hungary. *Acta Mineral. Petrology* 35, 41–69.
- Sadek G.D., Árkai P. & Nagy G. 1996: Alpine polyphase metamorphism of the ophiolitic Szarvaskő Complex, Bükk Mountains, Hungary. *Acta Mineral. Petrology* 37, 99–128.
- Slovenec D. & Pamić J. 2002: Geology of the Vardar Zone ophiolites of the Medvednica mountain area located along the Zagreb-Zemlin Line (NW Croatia). *Geol. Carpathica* 53, 1, 53–39.
- Szentpétery Zs. 1953: Diabase and gabbro bodies in Southern Bükk. *Ann. Hung. Geol. Inst.* 41, 1, 3–102 (in Hungarian).
- Talbi E.L.H., Honnorez J., Clauer N., Gauthier-Lafaye F. & Stille P. 1999: Petrology, isotope geochemistry and chemical budgets of oceanic gabbros-seawater interactions in the Equatorial Atlantic. *Contr. Mineral. Petrology* 137, 246–266.
- Tari V. & Pamić J. 1998: Geodynamic evolution of the northern Dinarides and the southern part of the Pannonian Basin. *Tectonophysics* 297, 269–281.
- Vanko D.A., Griffith J.D. & Erickson C.L. 1992: Calcium-rich brines and other hydrothermal fluids in fluid inclusions from plutonic rocks, Oceanographer Transform, Mid-Atlantic Ridge. *Geochim. Cosmochim. Acta* 56, 35–47.
- Watkinson D.H. & Ohnenstetter D. 1992: Hydrothermal origin of platinum-group mineralization in the Two Duck Lake Intrusion, Coldwell Complex, Northwestern Ontario. *Canad. Mineralogist* 30, 121–136.
- Zang W. & Fyfe W.S. 1995: Chloritization of the hydrothermally altered bedrock at the Igarapé Bahia gold deposit, Carajás, Brazil. *Mineral. Deposita* 30, 30–38.
- Zhang Y.G. & Frantz J.D. 1987: Determination of the homogenization temperatures and densities of supercritical fluids in the system NaCl-KCl-CaCl₂-H₂O using synthetic fluid inclusions. *Chem. Geol.* 64, 335–350.