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NATURE OF THE MAGMA SOURCE OF THE SZARVASKÖ COMPLEX (NE-HUNGARY): PETROLOGICAL AND GEOCHEMICAL CONSTRAINTS

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ABSTRACT

The Szarvaskö complex exposes a rare fragment of Jurassic mafic and ultramafic igneous rocks that have been described as a dismembered portion of a Mesozoic ophiolitic sequence, formed in a setting related to the opening of a narrow basin of the Vardar ocean. These magmatic rocks are mainly extrusive pillow basalts, along with gabbroic sills intruded into terrigenous shales, and minor ultramafic and plagiogranite bodies. The basalts and some of the gabbros have fractionated N-MORB-like trace element patterns, but low ε_{Nd} values, indicating a possible enriched source component as well. Other gabbros as well as the ultramafic rocks have the lowest REE patterns and are regarded as cumulates. The plagiogranites on the other hand, show inverse patterns with overall high trace element contents, and remarkable Eu anomalies. This suite of samples cannot be related solely by fractional crystallization to a common parental magma. Rather, they represent a combination of processes where the originally N-MORB-like magmas fractionated variable amounts of olivine+plagioclase+clinopyroxene±chromite and then assimilated terrigenous sediments abundantly present in the area. There is no reliable evidence for a subduction-related component in these rocks.

INTRODUCTION

Hungary is essentially a huge low-lying basin, encircled by the mountains of the Carpathians to the north and east, the Eastern Alps to the west, and the Dinarides to the south. The tectonic and magmatic evolution of this country reflects its position in the Intra-Carpathian Area (ICA). The ICA, with its three major blocks Alcapa, Tisza, and Dacia, has a very complex tectonic structure and various magmatic associations that have been formed before the Cenozoic and then were juxtaposed afterwards by large lateral displacements (Csontos et al., 1992a; Haas et al., 1990; Harangi et al., 1996). One of the most striking expressions of those Cenozoic tectonic events was certainly the eastward escape of the Alcapa block from the Alpine area under compression (Csontos et al., 1992b; Kázmér and Kovács, 1989). On the other hand, some ophiolites or ophiolitic-like bodies situated within the ICA mark important stages of the Mesozoic magmatic evolution of the northwestern part of the Tethys, as in the case of the rise and demise of the prominent Vardar ocean.

Ophiolites are thought to represent a section through the oceanic lithosphere. The current interpretation from petrologic studies is that ophiolites have formed either in mid-ocean ridge-like environments, or above subduction zones (Elthon, 1991; Gass, 1989; Pearce, 1991; Searle, 1992; Taylor et al., 1992). An increasing number of supra-subduction zone ophiolites have been assigned to various ophiolites worldwide. This assignment is based especially on geochemical grounds, which often show elemental anomalies in lavas and gabbros of ophiolite complexes.

In this work we examine and review the geochemistry of the magmatic rocks of the Szarvaskö complex. Situated in the SW Bükk Mountains (Fig. 1) and occupying an area about 8 km long and 3 km wide, this complex forms one of the largest Jurassic mafic bodies within the Carpathian chain. The main objectives of the study are to constrain the long-standing problem of the petrologic controls on the Szarvaskö magmatism, to delineate the nature of the Szarvaskö crust giving insight into the magma sources, and to compare the complex with other ancient and modern oceanic analogues. Since the various oceanic magmatic events in the ICA were related to the opening, subduction, and reopening of the oceanic branch of NW-Tethys, it is reasonable to suppose multiple mantle sources. Accordingly, the study of these igneous rocks provides an opportunity to evaluate the geochemical characteristics of an ophiolite originating along an ocean ridge behind a subduction zone.

THE SZARVASKÖ COMPLEX

The Szarvaskö complex, located in the southwestern part of the Bükk Mountains (Fig. 1) belongs to the innermost tectonic unit of the Western Carpathians (see review in Harangi et al., 1996). It is composed of several thrust units in a complicate imbricated structure (Balla et al., 1983; Balla, 1987) and it has previously been mapped (Balla et al., 1983) as a series of gabbroic sills intruded into Jurassic deep-water terrigenous shales, with abundant pillow basalts. These rocks were interpreted as fragments of a dismembered Mesozoic ophiolitic sequence (Balla and Dobretsov, 1984; Balla et al., 1983).

The complex chiefly consists of terrigenous sediments with sandstones dominating the bottom and shales the top of the sequence. Its mafic-ultramafic rocks are mainly represented by pillow lavas and sills of gabbroic rocks. These mafic and ultramafic rocks have been interpreted as representing olistoliths in an olistostrome series (Balla, 1987). The magmatic rocks of the Szarvaskö Complex consist of an extensively differentiated basalt-gabbro crustal association with minor ultramafic rocks and plagiogranites intruded into the shaly sediments. The volcanic sequence, by far the most abundant in the complex, is predominantly made up of

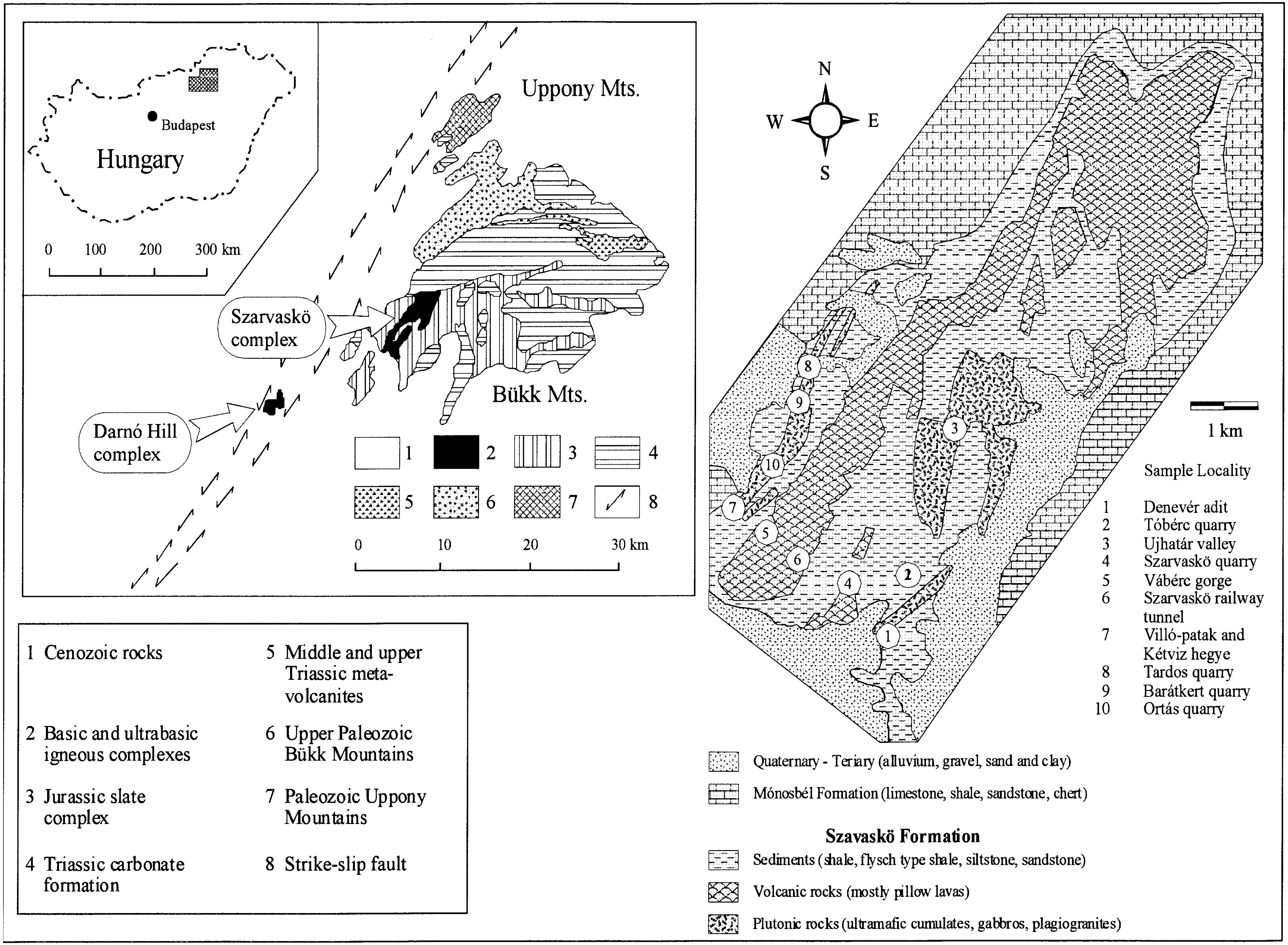
Hungary
Budapest
0 100 200 300 km
Uppony Mts.
Bükk Mts.
Szarvaskö complex
Darnó Hill complex
1 2 3 4
5 6 7 8
0 10 20 30 km
1 Cenozoic rocks
2 Basic and ultrabasic igneous complexes
3 Jurassic slate complex
4 Triassic carbonate formation
5 Middle and upper Triassic meta-volcanites
6 Upper Paleozoic Bükk Mountains
7 Paleozoic Uppony Mountains
8 Strike-slip fault
N
W
E
S
1 km
Sample Locality
1 Denevér adit
2 Tóbérc quarry
3 Ujhatár valley
4 Szarvaskö quarry
5 Vábérc gorge
6 Szarvaskö railway tunnel
7 Villó-patak and Kétviz hegye
8 Tardos quarry
9 Barátkert quarry
10 Ortás quarry
Quaternary - Teriary (alluvium, gravel, sand and clay)
Mónosbél Formation (limestone, shale, sandstone, chert)
Szavaskö Formation
Sediments (shale, flysch type shale, siltstone, sandstone)
Volcanic rocks (mostly pillow lavas)
Plutonic rocks (ultramafic cumulates, gabbros, plagiogranites)

Fig. 1 - Sketch map of the Bükk Mountains, NE-Hungary, with a geological sketch map of the Szarvaskö complex, modified after Balla et al. (1983). Samples localities are also shown.

Table 1 - Major and trace element analyses of representative Szarvaskö samples

Sample Rock type	WH5 Ultramafic	UJ3 Cum-Gabbro	TOB2D Gabbro	TAR1 Gabbro	TAR3-7 Gabbro	TAR4A Gabbro	TAR4D Gabbro	TOB6B Plagiogranite	TOB3C Plagiogranite	VAR1C Basalt	RW1 Basalt	SZ1B Basalt
SiO_2	31,71	45,88	49,83	48,30	51,83	49,44	50,88	70,17	58,00	48,93	49,67	50,90
TiO_2	10,96	3,90	1,93	1,68	2,49	3,75	2,80	0,28	0,39	1,78	1,66	1,82
Al_2O_3	1,92	17,92	16,63	14,94	13,05	13,71	13,63	15,57	19,35	14,54	14,70	14,61
$Fe_2O_3^{tot}$	33,11	13,16	12,27	11,92	15,74	16,74	14,74	3,99	4,30	11,53	11,42	12,54
MnO	0,52	0,18	0,19	0,18	0,25	0,30	0,23	0,06	0,04	0,16	0,19	0,17
MgO	14,62	4,49	5,16	6,23	3,73	4,14	4,05	0,48	0,82	6,33	6,17	5,59
CaO	5,67	9,45	7,36	10,17	6,27	6,37	6,89	1,51	5,15	10,58	8,40	6,87
Na_2O	0,48	3,31	3,77	3,25	3,87	3,78	4,56	5,64	7,35	2,06	3,56	4,33
K_2O	0,06	0,32	0,67	0,25	0,28	0,32	0,19	0,38	0,14	0,81	0,40	0,05
P_2O_5	0,03	0,06	0,18	0,15	0,36	0,31	0,26	0,05	0,09	0,16	0,16	0,17
LOI	1,71	1,70	2,60	3,01	2,00	1,70	1,70	1,21	3,45	3,46	2,52	2,98
Total	100,78	100,36	100,59	100,09	99,86	100,56	99,94	99,33	99,08	100,33	98,85	100,04
XRF analyses												
Nb[1]	14,8	4,42	4,08	3,74	6,8	6,8	6,63	9,35	10,88	3,23	3,4	4,25
Zr	46	65	131	98	222	188	178	404	494	127	120	137
Y	19	21	39	33	65	53	59	65	63	38	37	40
Sr	17,3	299	220	268	163	68	89	96,2	48	158	193	109
Pb	<5	<5	<5	<5	<5	<5	<5	11	11	<5	<5	<5
Ga	17	24	21	20	26	25	24	18	23	19	18	20
Zn	149	87	94	86	108	93	96	31	34	86	84	95
Cu	42	49	45	58	23	20	14	9	40	73	63	52
Ni	143	33	52	64	22	22	23	13	21	69	65	52
Co	109	45	33	43	30	35	34	13	19	39	33	35
Cr	214	48	62	142	18	22	26	9	19	200	191	100
V	1392	661	342	302	235	376	389	13	27	303	325	342
INAA analyses												
Sc	64,60	32,00	31,10	38,20	33,20	36,50	38,60	7,43	8,85	43,20	39,20	35,90
Rb	2,46	18,00	29,00	16,30	10,00	5,00	6,00	13,00	10,00	27,00	13,40	2,00
Cs	0,69	0,52	2,61	0,49	0,55	0,29	0,25	0,61	0,27	1,27	0,27	0,28
La	1,22	2,70	5,80	4,85	8,97	8,65	7,96	27,20	32,70	5,22	4,94	5,18
Ce	3,21	7,53	15,00	14,50	27,70	27,50	24,90	67,10	78,50	15,60	15,60	16,20
Nd	3,36	5,29	11,30	11,20	21,70	19,30	19,30	38,20	41,80	12,60	12,80	12,90
Sm	1,28	1,93	3,69	3,69	7,46	6,15	6,22	9,23	9,54	4,13	4,15	4,24
Eu	0,34	1,05	1,53	1,42	2,54	2,21	2,19	2,12	3,24	1,49	1,44	1,47
Gd	1,67	2,40	4,91	4,48	8,93	8,78	7,50	12,60	13,90	5,77	4,91	5,33
Tb	0,29	0,41	0,97	0,88	1,69	1,53	1,38	2,20	2,12	1,03	0,90	0,96
Yb	1,37	1,92	4,04	3,28	6,61	5,56	5,70	9,24	9,26	4,29	3,81	4,04
Lu	0,20	0,28	0,57	0,48	0,98	0,79	0,79	1,35	1,38	0,60	0,52	0,58
Hf	1,30	1,72	3,06	2,45	5,27	4,95	4,76	10,20	12,90	3,33	3,16	3,25
Ta	0,87	0,26	0,24	0,22	0,40	0,40	0,39	0,55	0,64	0,19	0,20	0,25
Th	0,28	0,22	0,42	0,46	0,75	0,67	0,51	6,92	8,84	0,31	0,25	0,46
U	0,16	0,31	0,19	n.d.	0,67	0,26	0,39	1,03	1,07	0,47	0,22	0,14

[1] Nb has been interpolated from Ta, assuming a Nb/Ta ratio of 17; Cum-Gabbro for cumulate gabbro.
Samples WH5, UJ3, TAR1, TOB6B, VAR1C and RW1 had their Sr, Rb, Sm and Nd concentrations analysed by isotope dilution.

pillow basalts. Gabbroic rocks varying from fine-grained, medium-grained, pegmatitic, up to cumulates type are the dominant types in the plutonic section. Plagiogranites are also found as small intrusions and nests, some of them with garnets and/or dark xenolithic enclaves. The Fe-Ti oxides-bearing ultramafic cumulates described as wherlites, pyroxenites, hornblendites, and minor peridotites are peculiar rocks of their odd mineralogy and geochemistry. All rocks underwent a low-grade prehnite-pumpellyite regional metamorphism during the Alpine orogeny (Arkai, 1983).

The mafic rocks are composed of clinopyroxene (mostly augite), plagioclase, Fe-Ti oxides, and minor amphibole. The ultramafic rocks have very high contents of ilmenite and olivine($fo_{0.50}$), various amounts of clinopyroxene, rare orthopyroxene, and Ti-rich tschermakitic amphibole. The leucocratic rocks (hereafter generally termed “plagiogranites”) are made of albite, quartz, biotite and locally garnet. Secondary products formed by a later metamorphic event are typically represented by chlorite, albite, prehnite, pumpellyite, calcite, and sericite. The petrography and mineral chemistry of these rocks is under study.

Concerning the age of the complex, Jurassic radiolarians have been related to the Szarvaskö complex (Balla, 1987; Balogh et al., 1984; Dosztály and Józsa, 1992). K-Ar age determinations from the intrusives and their contacts have also been made and range between 157 and 170 Ma (Harangi et al., 1996). Ages obtained on primary igneous amphiboles from the gabbros average 166 Ma (Downes et al., 1990) and are adopted here for isotopic systematics.

Previous geological and geochemical investigations (Balla and Dobretsov, 1984; Balla et al., 1983; Downes et al., 1990; Kubovics et al., 1990) recognized the complicated tectonic structure and the strong ocean floor affinity of the complex. Previous results (Aigner-Torres, 1996) have also sustained the oceanic signature of these igneous rocks, though the question of their origin has not yet been solved. Over the last decade, several authors have assigned the complex to an ophiolite, (e.g., Balla et al., 1983; Dal Piaz et al., 1995), but there are still some doubts regarding the nature of these rocks, (e.g., Downes et al., 1990; Kubovics et al., 1990).

ANALYTICAL METHODS

Whole-rock XRF analyses (55 samples) were made on glass discs fused with lithium tetraborate flux (for major element) or pressed discs of rock powder and a polyvinyl alcohol binder (for trace elements), carried out using a Philips PW 2400 spectrometer (Institute of Petrology, University of Vienna). The INAA analyses (12 samples) were made on bulk-rock powders split from samples previously analyzed by XRF. Irradiation was completed at the Atominstitut der

Table 2 - Nd and Sr isotopic compositions for the Szarvaskö igneous rocks

Sample	Rock type	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr_p$	$(^{87}Sr/^{86}Sr)_i$	ε_{Sr}	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd_p$	$(^{143}Nd/^{144}Nd)_i$	ε_{Nd}
RW1	Basalt	0,2017	0.704014 ± 13	0,703538	-10.90	0,19626	0.512978 ± 6	0,512765	+6.64
VAR1C	Basalt	0,4802	0.704251 ± 9	0,703118	-16.90	0,19799	0.512964 ± 6	0,512749	+6.33
TAR1	fine-grained Gabbro	0,1768	0.705332 ± 8	0,704915	+8.62	0,20016	0.512960 ± 6	0,512743	+6.21
WH5	ultramafic-"Wherlite"	0,4127	0.705102 ± 7	0,704128	-2,55	0,22929	0.512928 ± 32	0,512679	+4.97
UJ3	cumulate Gabbro	0,1743	0.704988 ± 4	0,704577	+3.82	0,21965	0.512915 ± 5	0,512676	+4.92
TOB6B	Plagiogranite	0,4075	0.706549 ± 7	0,705587	+18.20	0,14626	0.512798 ± 5	0,512639	+4.19

Initial ratios and ε values calculated for an age of 166 Ma. Subscript *p* is present-day measured ratios, subscript *i* means initial ratios.
Ratios are normalized $^{86}Sr/^{88}Sr=0.1194$ and to $^{146}Nd/^{144}Nd=0.7219$; errors are in 2σ mean. Model ages were calculated using: $(^{143}Nd/^{144}Nd)_{CHUR}=0.512638$; $(^{87}Sr/^{86}Sr)_{BSE}=0.7045$; $(^{147}Sm/^{144}Nd)_{CHUR}=0.1967$; $(^{87}Rb/^{86}Sr)_{BSE}=0.0816$, where CHUR and BSE stand for chondritic uniform reservoir and bulk silicate Earth, respectively.

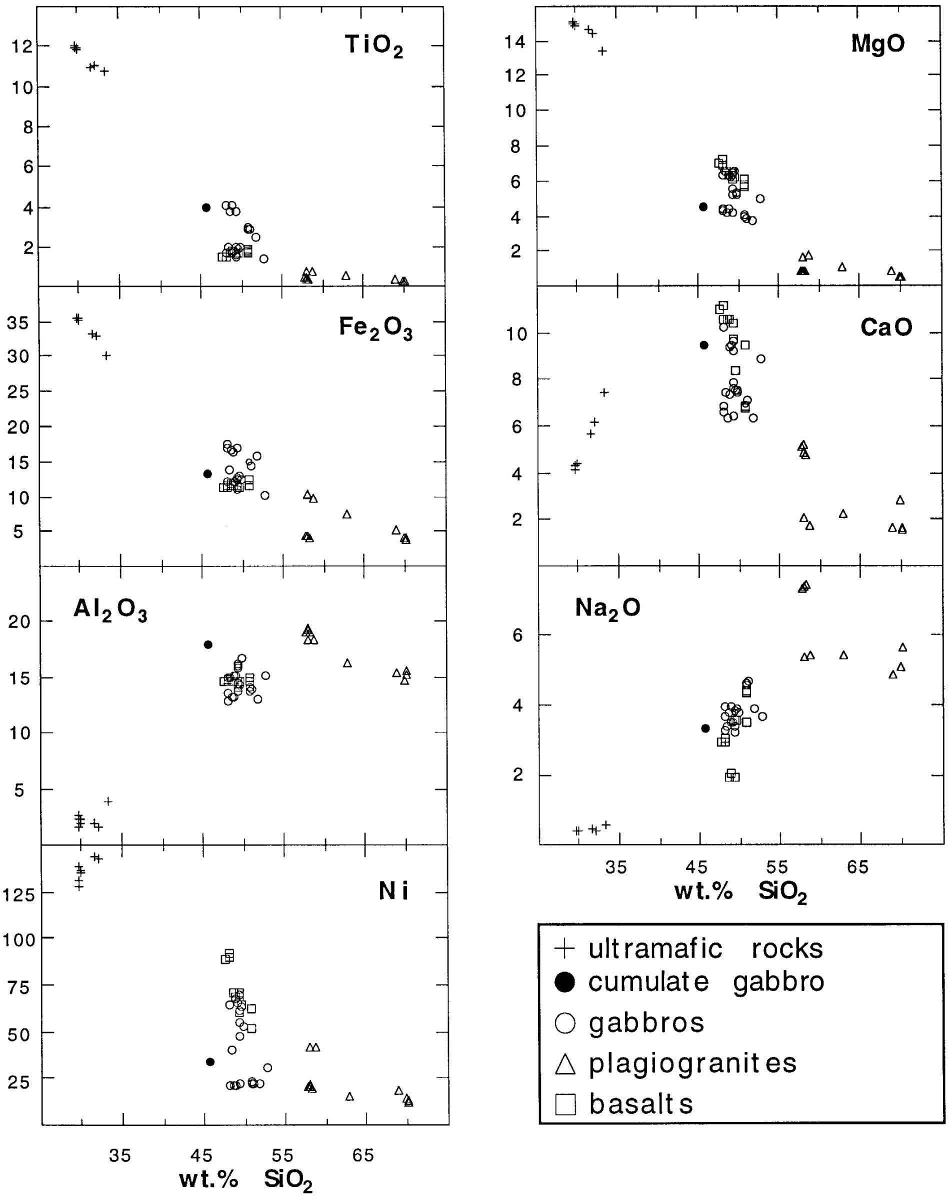

Fig. 2 - SiO_2 (wt%) versus major oxides (wt%) and trace elements (ppm) variation plots for the Szarvaskö samples. For simplicity, fine-grained gabbros and intermediate gabbros are all under the "gabbros" symbol; as quartz-diorites and trondjemites are under the "plagiogranites"symbol, unless otherwise stated.

Österreichischen Universität in Vienna using the TRIGA Mark II type reactor. Counting was done at the Institute of Geochemistry, University of Vienna. Procedures and equipment are discussed in Koeberl (1993). The isotope analyses (6 samples) were performed at the BVFA Arsenal - Abteilung Isotopengeologie und Geochronologie (Vienna), using standard cation exchange techniques as described by Thöni and Jagoutz (1992). Leached separates were spiked after separating aliquots of the solution, using a mixed REE (150Nd-149Sm) tracer for determining Nd and Sm concentrations. All Nd and Sm isotope ratios were carried out by a Finnigan MAT 262 multicollector mass spectrometer, while Sr and Rb concentrations were determined using a VG Micromass M 30.

RESULTS

Here we describe and compare the major geochemical characteristics of the Szarvaskö magmatic rocks. Major and trace element data sets are listed in Table 1. Isotopic ratios of representative samples are given in Table 2. Because of the altered nature of most samples, combined with the evidence of mineralogies and textures of cumulates, major and mobile trace elements data should be interpreted with care.

Major elements

In Harker variation diagrams (Fig. 2), the relationship among the major elements of the rocks can best be shown using SiO_2 as an index of differentiation. From these diagrams four groups can be outlined; the first one represents the most basic (non-cumulitic) rocks, i.e. *basalts* and *fine-grained gabbros*. They are characterized by small compositional ranges of SiO_2 (47-49 wt%) and show small variations in TiO_2, Al_2O_3, Fe_2O_3, MgO, and CaO. Na_2O (less) and K_2O (strongly; not shown) reveal scatters that are probably related to alteration processes. The basalts are predominantly hypersthene- and slightly quartz-normative tholeiites with minor olivine-normative samples. No sample has more than 7.2% MgO (5.16 - 7.16 wt%) and basalts are the least differentiated rocks from the Szarvaskö complex. Thus, all basic samples have experienced some level of fractionation.

The second group involves all the other gabbros further subdivided into two subgroups: The *cumulate gabbro* (sample UJ3) and the *intermediate gabbros*, including medium-, coarse-grained, and pegmatitic types. Both subgroups have similar contents of TiO_2, Fe_2O_3, MgO, and K_2O, but differ from each other in Al_2O_3, MnO, CaO, Na_2O, and P_2O_5 contents (Fig. 2, Table 1). These two groups are presumably not genetically related, as will be explained later, but have comparable modal contents of Fe-Ti oxides.

The third group comprises the most evolved rocks, ranging from quartz-diorite to tonalites (trondhjemites), hereafter *plagiogranites*. The most striking feature in these rocks is the very low K_2O content, less than 1% (with the exception of the garnet-bearing quartz-diorites, all samples have $K_2O < 0.5$ wt%), typical of the oceanic plagiogranites found in ophiolites (Coleman, 1977). The evolution of Al_2O_3 and Fe_2O_3 follow a normal trend, but some samples show higher Al_2O_3 and CaO values combined with lower Fe_2O_3 (Fig. 2).

The *ultramafic rocks* make up the fourth group; these rocks are extremely rich in Fe_2O_3 (up to 35.5 wt%) and TiO_2 (12 wt%), with a low content of SiO_2 (29.6 wt%) corresponding to their mineralogy, e.g. accumulation of ilmenite and Fe-rich olivines. Even in less altered samples CaO and Al_2O_3 contents vary to some extent among the samples, mostly because of the presence of minor An-rich plagioclase.

The wide SiO_2 variation from 47.59 to 70.17 wt% (basalt to plagiogranite), correlates negatively with the oxides of Ti, Fe, Mn, Mg, Ca, and P. There is a positive correlation with Al and Na. K values fluctuate a great deal probably due to alteration. The strong inflections seen in the SiO_2 vs. Al_2O_3, CaO, Na_2O, and TiO_2 trends combined with a weak trend for MgO and Fe_2O_3 can be related to crystallization of olivine, plagioclase, clinopyroxene and Fe-Ti oxides.

Trace elements

Compatible and less incompatible elements show great variation with respect to SiO_2 content. In the basalts and fine-grained gabbros they are moderately variable, e.g. Ni varies from 52 to 92 ppm, Cr from 62 to 278 ppm, which reinforces their differentiated nature.

Ni decreases rapidly with increasing SiO_2 content (Fig. 2), and displays high values where MgO content is also elevated in response to olivine crystallization or accumulation, as in the case of the ultramafic rocks. Cr (not shown) exhibits wide variation among the most basic samples, which could be related to the presence of chrome spinel, although this phase was not petrographically observed.

Sc, Co and V (not shown) all show similar trends, decreasing in abundance as SiO_2 increases. This is best explained by pyroxene and amphibole crystallization (Sc and Co), but also by oxide crystallization, where all three elements are strongly partitioned into Fe-Ti oxides, behaving essentially like Fe and Ti. The high values in the ultramafic cumulates are in accordance with their mineralogy, i.e. the presence of ilmenite, clinopyroxene and amphibole.

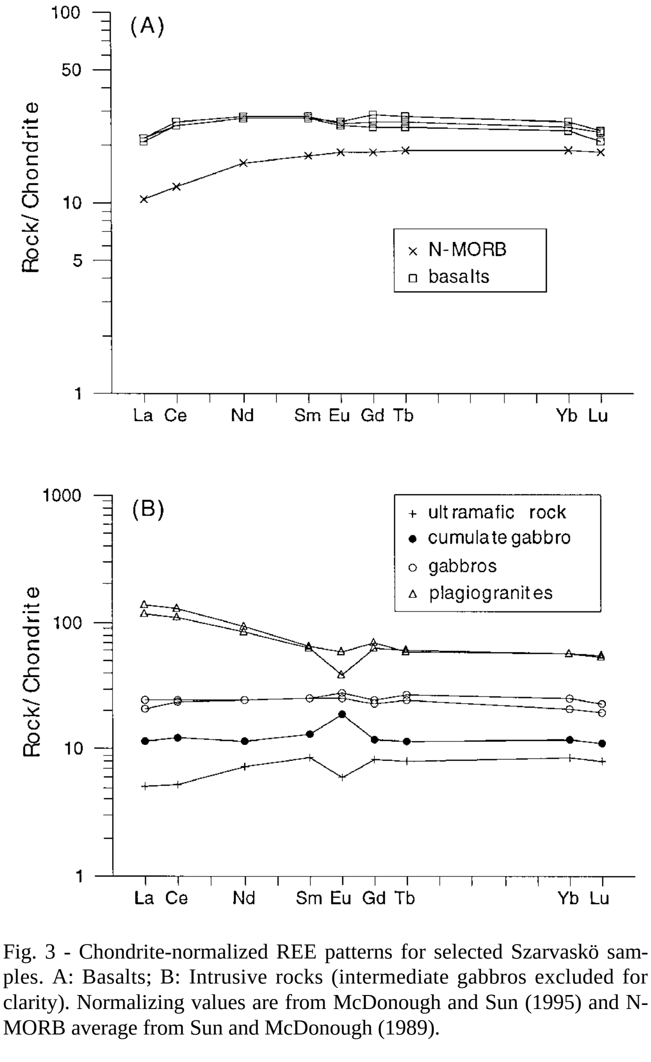

Fig. 3 - Chondrite-normalized REE patterns for selected Szarvaskö samples. A: Basalts; B: Intrusive rocks (intermediate gabbros excluded for clarity). Normalizing values are from McDonough and Sun (1995) and N-MORB average from Sun and McDonough (1989).

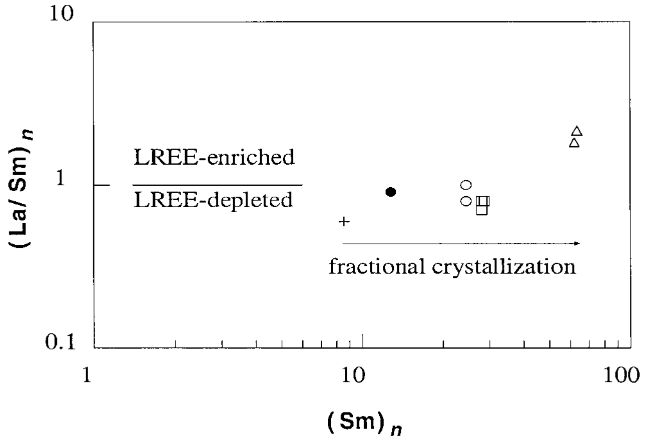

Fig. 4 - $(La/Sm)_n$ ratio plot showing the LREE-depleted trend of increasing Sm by little change in the $(La/Sm)_n$ ratio, probably related to fractional crystallization, and the high LREE contents of the evolved samples, likely to be caused by contamination. Symbols as in Fig. 3.

Sr is highly variable (not shown), indicating in part some mineralogical control as in the case of a the cumulate gabbro with high Sr values (plagioclase accumulation), but most of the scatter likely represents a secondary feature.

Among the evolved samples, some of the quartz-diorites show contrasting behavior with respect to the other rocks of this group with higher values of Ni, Cr and Sc and with a contrasting decrease of Fe_2O_3 and MgO in the same sample.

Chondrite-normalized concentrations of *rare-earth elements* (REE) in the Szarvaskö volcanic and intrusive rocks are illustrated in Fig. 3A-B, respectively. Basalts show a slight degree of REE fractionation [$(Ce/Yb)_n$ = 0.95-1.1] with small negative Eu anomalies (0.93-0.98). The $(La/Sm)_n$ ratios (subscript *n* indicates chondrite normalization) of around 0.8 give them upward convex patterns like those for N-MORB. Moreover a ratio < 1 is considered to be indicative of a source much more depleted than "typical" arc-source mantle (Fig. 3B), in which the ratio would commonly be near 1 or higher, i.e. enriched in LREE (Arculus, 1994; Hawkins and Allan, 1994). Nonetheless, with an overall elevation of REE abundance, e.g. Ce_n = 25-26 in contrast to the Ce_n = 12 for N-MORB, and their low Ni, Cr and MgO contents, these basalts have a strong fractionated nature and do not represent primary magmas.

Fine-grained gabbros present similar REE patterns as the basalts, but with very slightly positive Eu anomalies. The intermediate gabbros on the other hand (not shown), are all more REE enriched than the basic samples, with negative Eu anomalies. The cumulate gabbro shows lower REE abundance and a marked positive Eu anomaly (~1.5) consistent with its plagioclase-clinopyroxene cumulate origin. Also the ultramafic cumulate has a REE pattern that corresponds to the mineralogy of the rock, i.e. olivine, ilmenite, pyroxenes and the absence of plagioclase, giving the lowest REE abundance, with strong LREE depletion and a strong negative Eu anomaly (0.7).

The evolved samples show clear negative Eu anomalies

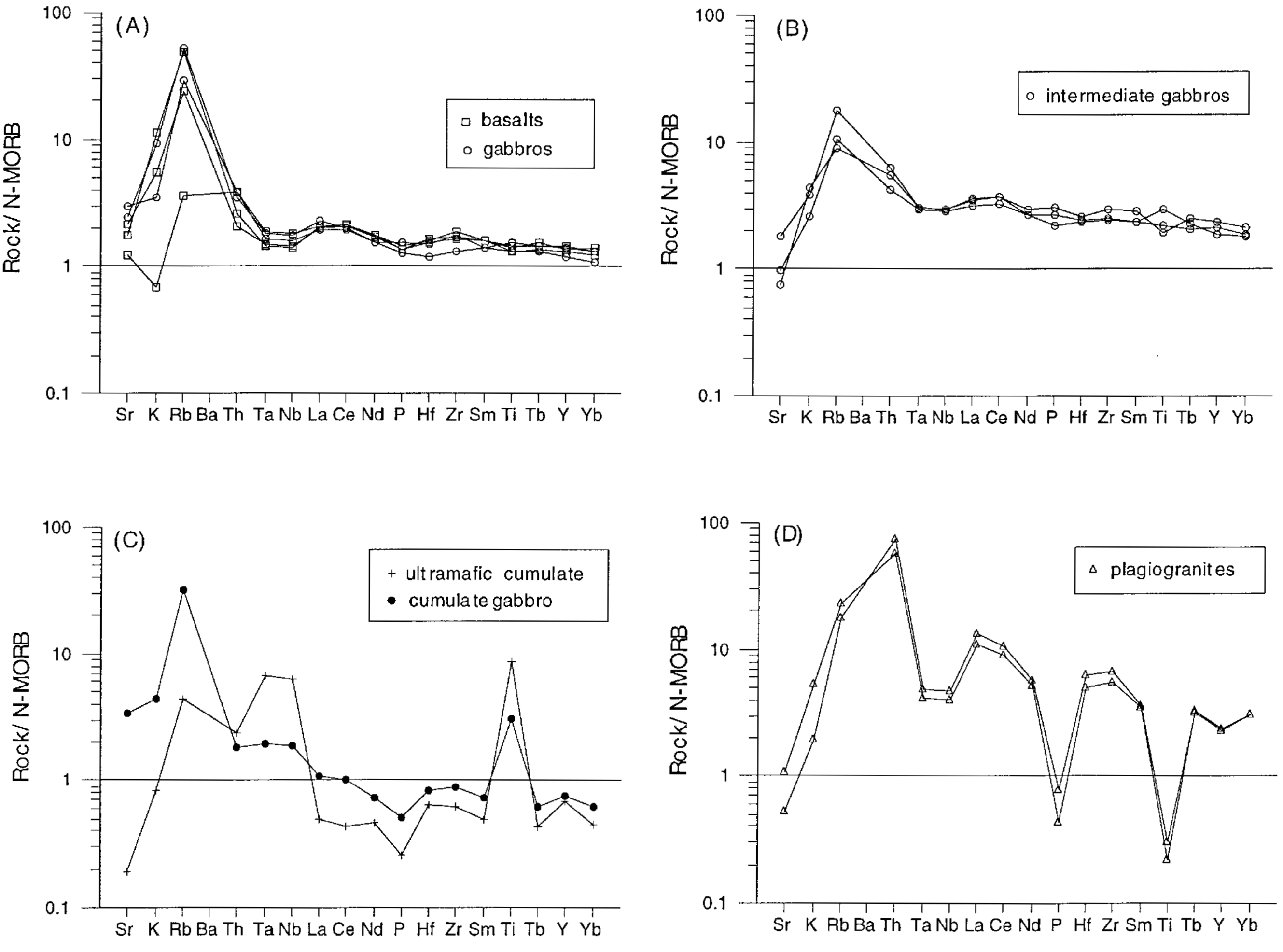

Fig. 5 - MORB-normalized incompatible elements patterns for representative Szarvaskö samples. Normalizing factors after Sun and McDonough (1989). Symbols as in Fig. 3.

(0.6-0.9) characteristic of earlier plagioclase fractionation and higher overall REE enrichment (Ce_n >100) with a remarkable LREE enrichment (Fig. 4). Generally speaking, these rocks have similar REE contents to plagiogranites dredged from the ocean crust (Aldiss, 1981) and ophiolites (Elthon, 1991).

The *incompatible element* contents are represented in Fig. 5. The basalts and fine-grained gabbros (Fig. 5A) define a fairly narrow band that resembles MORB for those high-field strength elements (HFSE) that plot to the right of Ta, with a general enrichment from 1.2 to 1.5 times relative to N-MORB. Large ion lithophile elements (LILE) on the other hand, show a distinct enrichment, with variations that differ among the elements and samples. For instance, La and Ce values are about 2 times those of MORB; Th 2-4 times; Sr 2-3 times; but Rb values are 20 to 50 times higher than MORB values. Moreover, the magnitude of the Nb-Ta anomaly is insufficient to place it outside the compositional spectrum of N-MORB from ocean basins unrelated to subduction, and is probably related to the sharp enrichment in the surrounding LILE.

Intermediate gabbros (Fig. 5B) have comparable LILE/HFSE patterns to the basic samples, although the HFSE show an overall slight enrichment. Ti differs only little among the samples, possibly due to different amounts of Fe-Ti oxides involved in fractionation. Sr is low (plagioclase fractionation), Rb and Th high.

Cumulate rocks (Fig. 5C) have a strong and positive Ti-anomaly, which indicates the accumulation of Fe-Ti oxides in these rocks. The cumulate gabbros also show high concentration of Sr due to the presence of plagioclase. Potassium and Rb are also high, but Th, on the other hand, is low.

Plagiogranites (Fig. 5D) are characterized by strong negative anomalies for Ta, Nb, P, and Ti, all others are strongly enriched. The LILE show a strong increase from Sr towards Th.

Sr and Nd isotope geochemistry

Representative Sr and Nd isotopic ratios and parameters for samples spanning the compositional range of the Szarvaskö magmas are reported in Table 2. All samples are corrected to an age of 166 Ma (Downes et al., 1990, p. 202) obtained by the K-Ar method on primary amphiboles from gabbros. This age is in good agreement with some previous K-Ar ages of hornblende from gabbros ranging from 157 to 171 Ma (Balla, 1984), as well as with paleontological data (Dosztály and Józsa, 1992). All these data suggest that the Szarvaskö igneous rocks are of Middle Jurassic age.

ε_{Nd} values from the Szarvaskö complex show a narrow range (Fig. 6A), from +6.6 for the basic samples, through +4.9 for the cumulates, to +4.2 for the evolved sample. All positive values imply that the magma, from which these rocks originated, was derived from a source with a greater Sm/Nd ratio than CHUR, i.e. from a depleted mantle source region. On the other hand, the initial $^{87}Sr/^{86}Sr$ ratios exhibit a great variation ranging from 0.7031-0.7035 for the basalts to 0.7049 for the fine-grained gabbro, and to 0.7056 for the plagiogranite sample. The cumulate gabbro and the ultramafic cumulate have ratios varying from 0.7041-0.7045, respectively.

The elevated initial $^{87}Sr/^{86}Sr$ ratios together with the low-grade metamorphic overprint of the samples, raise concern that these values may reflect an incorporation of radiogenic Sr through reaction with seawater and/or high-temperature aqueous fluids. This has already been demonstrated in studies on altered samples from ocean floor and ophiolites (e.g., Jacobsen and Wasserburg, 1979). Alternatively, the Szarvaskö rocks may have higher initial $^{87}Sr/^{86}Sr$ ratios due to a process involving contamination by the abundant surrounding sediment during the magmatic fractionation.

The range of Nd-Sr isotopic compositions in the Szarvaskö igneous rocks lies well within or close to the mantle array. It is somewhat lower when compared to those of average N-MORB (ε_{Nd} =+10 and $^{87}Sr/^{86}Sr$ =0.7025 at 166 Ma) and lies within the fields for island-arc basalts and oceanic island basalts. A trend in initial $^{87}Sr/^{86}Sr$ ratios and nearly constant ε_{Nd} values among the basic samples can be observed (the lowest values are from basalts, the highest from the fine-grained gabbro). This trend can be best explained by some sort of secondary alterations, such as seawater alteration. Although the interpretation of Sr isotopic data remains problematic, the Nd isotopes are thought to have retained their original magmatic signatures, unaffected by alteration or metamorphism, since the REE appear to have been left immobile (Fig. 6B). The low eNd for plagiogranite, com-

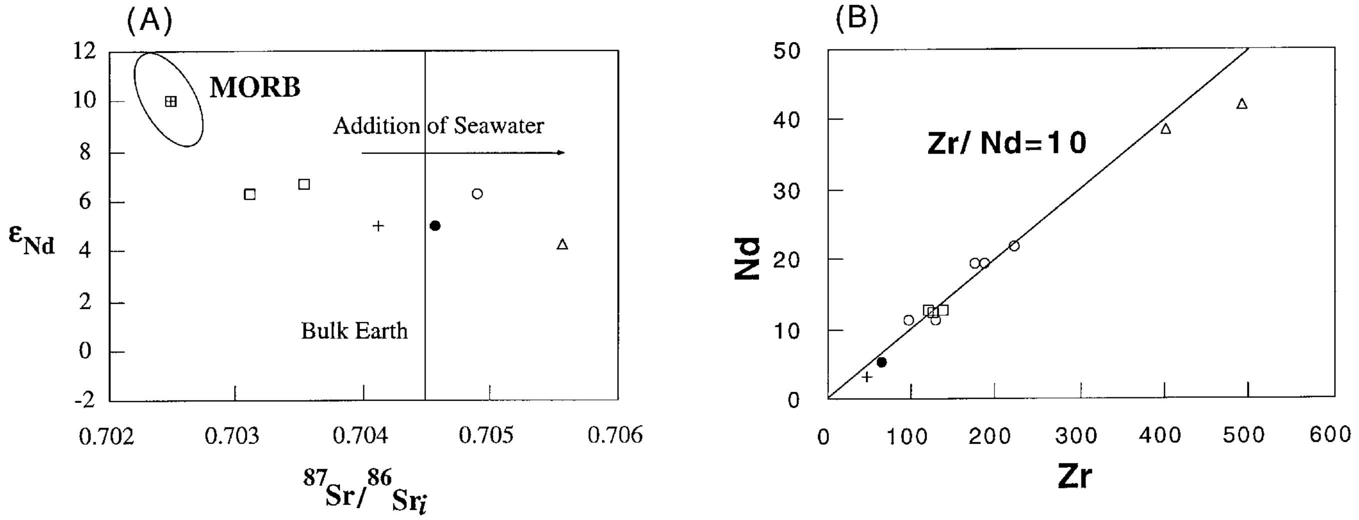

Fig. 6 - (A) ε_{Nd} and Sr_i calculated for 166 Ma. MORB field and average N-MORB after Sun and McDonough (1989) at 166 Ma are also shown; (B) Covariation of Nd with Zr (ppm) suggests that Nd (also La and Ce) was not significantly altered, although one plagiogranite has higher Zr values. Symbols as in Fig. 5.

bined with its high contents of LREE, Th and LILE (Figs. 4 and 5D), was attributed to a linked assimilation-fractional crystallization (AFC) process, possibly by a shale component (Downes et al., 1990). The cumulates have an intermediate initial ε_{Nd} that casts doubt on an origin linking them with the basic samples. Moreover, their differences in initial $^{87}Sr/^{86}Sr$ values are reasonably related to the differences in mineralogy between the two rocks and to the degree of alteration.

DISCUSSION

Magmatism in ocean basins occurs mainly at mid-ocean ridges and at convergent plate margins where two fundamentally different mantle portions give rise to partial melts, each having distinctive chemical signatures (Wilson, 1989). Mid-ocean ridge basalts (MORB) exhibit a fairly restricted range in chemistry and isotopic composition being depleted in magmatophilic elements such as LILE and HFSE. It is inferred that they are derived from a previously relatively depleted mantle source that gives them distinctive trace element abundances and ratios, also supported by isotopic signatures such as for the Sm-Nd, Rb-Sr, and U-Pb systems, serving to define a “normal” or N-MORB, (e.g., Sun and McDonough, 1989; Viereck et al., 1989). On the other hand, the source for supra-subduction zone (SSZ) magmas at convergent oceanic plate margins has been depleted even more than the MORB source and is considered to resemble the residue left after extraction of MORB (i.e., it is a multiple depleted mantle source). Many SSZ magmas are relatively enriched in LILE and water. The subduction zone setting permits the re-enrichment of the depleted mantle by fluids derived from the subducted ocean crust, by sediment involvement, either directly or through an intervening fluid phase or by the addition of partial melts derived from the subducted ocean crust. Although backarc basins are situated above SSZ mantle, recent studies on ODP Leg 135 samples of the Lau Basin (Hawkins and Allan, 1994) have demonstrated and confirmed that backarc basin basalts (BABB) have a heterogeneous crust that span the range from MORB to arc compositions.

One of the main petrologic objectives for the investigations of the Szarvaskö igneous rocks was to evaluate the nature of its magma source, to try to model a process that could explain the variation among the samples, and assign with more confidence the paleotectonic environment in which these rocks were generated. Since the work of Balla et al. (1983), the Szarvaskö magmatic rocks have been regarded as remnants of a dismembered ophiolite, with a marginal, backarc basin character (Balla et al., 1983; Downes et al., 1990; Kubovics et al., 1990). Recent studies on the genesis of many ophiolites have favoured a marginal basin or arc-related setting rather than a major ocean basin (Taylor et al., 1992). A starting point for comparative studies can be found in the present-day association of marginal basin/arc setting and spreading-generated crust, e.g. in the active and remnant back-arc basin and arc/fore-arc terrains of the Western Pacific (Hawkins, 1995).

Nature of the magma source

In this section we discuss several of the chemical and isotopic tracers that help to delineate the nature of the Szarvaskö igneous rocks and give insight about the magma sources. The HFSE and REE offer one of the best ways to constrain the nature of the original mantle source.

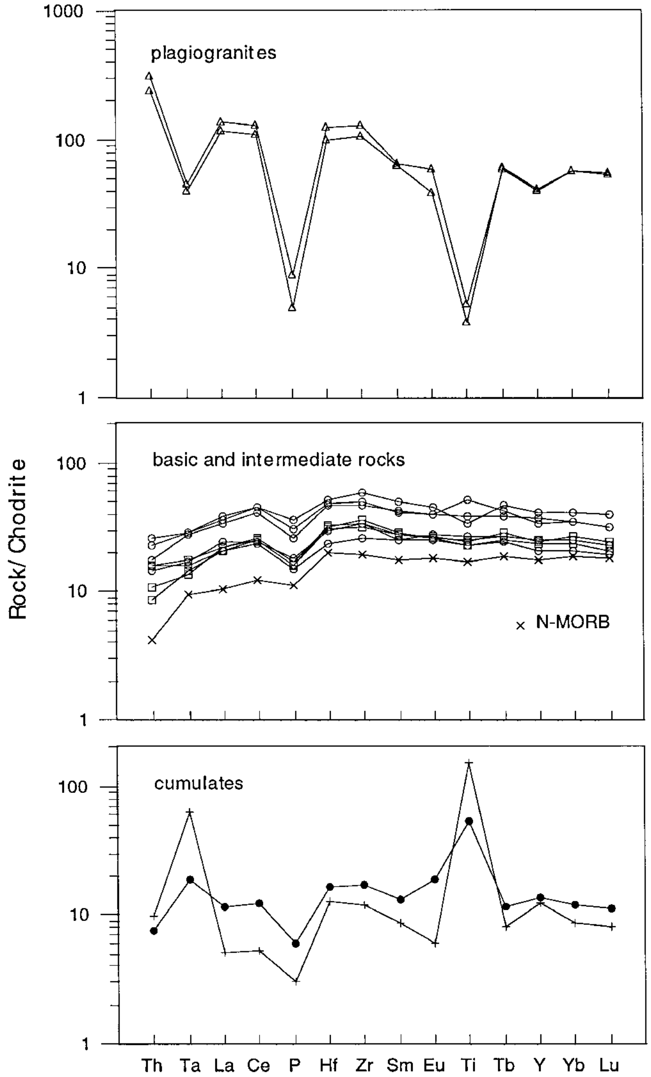

Fig. 7 - Chondrite-normalized (McDonough and Sun, 1995) trace element diagrams for representative samples from the Szarvaskö igneous rocks. Note the different Ta anomaly among the rocks. Symbols as in Fig. 5.

Our argument will be primarily based on the abundance of seven trace elements: Th, Ta, La, Ce, Zr, Ti, and Y. All these *immobile elements (HFSE and HREE)* are incompatible in basaltic systems, and they are listed in decreasing order of incompatibility, so Th is the most incompatible, Y the least incompatible. Th, La and, to a lesser degree, Ce are LIL elements, a group which also includes K, Rb and Cs, and which tends to have low charge/radius ratio ionic states. Most LILE are highly mobile during secondary alteration, which makes the interpretation of whole-rock data difficult. Th has long been considered to be a relatively immobile element, but there is a high probability that $Th(OH)_4$ may be stripped from sediments at elevated temperature and transported by the dehydration fluids in a similar way as B and Be (Hawkins and Allan, 1994). La and Ce are less mobile than Rb, K, and Cs, and their primary igneous abundances may be preserved in slightly altered samples. Ta, Zr, Ti and Y are HFS elements, whose ions have a high charge/radius ratio. They are considered as immobile in even strongly altered samples under hydrous conditions up to the greenschist facies of metamorphism (Bienvenu et al., 1990; Ludden and Thompson, 1979) and therefore they can be assumed to represent magmatic values with reasonable confidence.

Table 3 - Incompatible trace element ratios for the Szarvaskö basic rocks and compiled mantle reservoirs.

	Ti/Zr	Y/Zr	La/Ta	Th/Ta
Szarvaskö rocks [1]	79.6-102.8	0.29-0.35	20-27.5	1.25-2.09
MORB source [2]	141,9	0,44	11,4	0,39
N-MORB [3]	90-110 (102.7)	0.3-0.4 (0.38)	10 - 20 (18.9)	0.75-2.0 (0.90)
IOB [4]	61,4	0,1	13,7	1,48
SSZ [5]	>100	>0.4	30 - 40	3 - 5
East Scotia Sea [6]	67.3-91.4	0.23-0.35	11.1-12.6	0.90-1.58
Mariana Trough [6]	84.2-88.9	0,31	18.0-32.2	2.18-3.94
Brainsfield Strait [6]	62-66	0.17-0.18	n.d	n.d.
Lau Basin [7]	69-153	0.25-0.51	20-31.9	1.67-2.29

[1] values for the most basic rocks found in the Szarvaskö complex.
[2] Depleted mantle abundances are from McKenzie & O'Nions (1991, 1995).
[3] N-MORB ranges are from Hawkins et al. (1994) and Pearce (1991), values in parentheses are averages from Sun & McDonough (1989).
[4] IOB are averages from Sun & McDonough (1989).
[5] SSZ ranges for Ti/Zr and Y/Zr ratios are from Hawkins et al. (1994); La/Ta and Th/Ta ratios are from the Bay of Islands ophiolite (Elthon, 1991).
[6] are selected BABB from Saunders & Tarney (1991).
[7] ranges from ODP Leg 135-sites 834 through 839, from Hawkins et al. (1994)

Data for relatively immobile trace elements, normalized to CI chondrite abundances, are shown in Fig. 7, and ranges of values currently in use for some incompatible trace element ratios are compared to those of the most basic rocks of the Szarvaskö complex in Table 3. The presence of negative anomalies for tantalum, niobium and/or titanium are thought to be indicative of SSZ magmatism (Elthon, 1991; Pearce et al., 1984).

The basic rocks, i.e. basalts and fine-grained gabbros, have nearly flat patterns and mime the N-MORB trend, with no Ta anomaly. The intermediate rocks also have no Ta depletion and show an overall enrichment characteristic of a fractionation from the basic samples. A negative Ta anomaly occurs in the plagiogranites. There also have additional negative anomalies for P, Eu, Ti and Y due to previous substantial crystallization of plagioclase, Fe-Ti oxides and clinopyroxene and/or amphibole, respectively. The marked positive anomalies in Ta, Ti and Y for the cumulate rocks are in agreement with their mineralogy, involving the accumulation of Fe-Ti oxides (especially ilmenite for Ta), clinopyroxene and amphibole. The cumulate gabbro also has a positive Eu anomaly (Fig. 3B) related to the plagioclase accumulation, which was observed in this sample. Both the cumulate gabbro and the plagiogranites have a strong negative P anomaly, probably intrinsic to the original liquids with which they coexisted. This anomaly can also be observed in both the basic and intermediate samples in a lesser extent (Fig. 5). Such a negative P anomaly can be caused apatite fractionation, but no samples with P enrichment were found in the whole sequence so far.

It is important to note that the extent of the Ta anomaly remains uniform (chondrite-normalized Ta/[(Th+La)/2] = 0.97 ± 0.1) throughout the basic/intermediate series of rocks, but differs in the plagiogranites (0.22 ± 0.01). This is in agreement with the strong enrichment seen in the LILE, e.g. Th and the LREE in these samples (Figs. 3A-B, and 4). The practically non-existent Ta anomaly in the basic and intermediate rocks argues against the involvement of a SSZ component.

Further evidence that the Szarvaskö rocks have a MORB signature rather than a SSZ origin is the depletion of Ta compared with Hf and Th, and of Y compared to Zr. Although the use of discrimation diagrams based on these elements has been extensively employed to help evaluate the geodynamic settings for ophiolites, it has some pitfalls (Hawkins, 1995; Klein and Karsten, 1995; Wang and Glover III, 1992). However, it can be said that the Szarvaskö samples clearly plot within the N-MORB field. Furthermore, the Ti/Zr, Y/Zr, La/Ta and Th/Ta ratios (Table 3) in the Szarvaskö basic rocks are within the MORB and BABB ranges. It is important to recall that the absence of SSZ magmas does not rule out the possibility that these rocks have been formed in a backarc basin environment, since BABB may span between the SSZ/MORB signatures (Hawkins and Allan, 1994; Saunders et al., 1991).

Subduction-derived “dehydration fluids” (Hawkins, 1995) would cause enrichment in the LILE but would have no effect on HFSE or HREE abundance in the mantle wedge because of the immobility of these elements in aqueous fluids. However, partial melts of the subducted (amphibolitized) basaltic crust would be able to impregnate the SSZ mantle with “immobile” elements (Pearce and Peate, 1995). Consideration of distribution coefficients (Kelemen et al., 1990) suggests that small amounts of melt may be enriched selectively in the relative order of LREE > Ta and Nb > Zr > Ti > Hf > HREE. The discussion of HFSE partitioning in the genesis of arc basalts applies to back-arc magma genesis as well. Collectively, the data displayed in Figs. 5 and 7 show no marked depletion in Nb and Ta, a characteristic for non-SSZ involvement, but would also correspond to patterns observed in the back-arc basalts. The same is also observed for the REE data for the Szarvaskö samples that have MORB-like patterns, with $(La/Sm)_n < 1$ (Figs. 3A-B), and an enrichment up to 20x higher than chondritic values.

The mobile elements (LILE) are useful diagnostic markers with their very low amounts of K, Rb, Cs, Ba, and LREE in N-MORB’s (Viereck et al., 1989). Both LILE trace element abundance and Sr isotopic ratios increase as low temperature alteration increases. K, Rb, and Cs increase much more readily than Ba, Sr, or LREE (Staudigel et al., 1981). Unlike isotope ratios, abundances and trace elements ratios can be modified by fractional crystallization. Since the LILE are highly incompatible elements, one expects to find positive correlations between K, Rb, Cs, Ba and La, and SiO^2 or with HFS elements such as Zr. However, the LILE should show scatter if they are the result of fluid transfer into the rock system, and are also highly mobile under hydrous low-temperature conditions of alteration. The presence of a LILE enrichment in unaltered glass in SSZ settings reflects a subduction fluid component in the melt source. Back-arc basin basalts show variable degrees of enrichment in LILE that may be related to the width of the basin and to the maturity of the adjacent subduction zone (Saunders et al., 1991). For the case of the Szarvaskö samples, it is inferred from Fig. 5 that the strong variations seen in Cs, Rb, K and in some degree in Sr, are of secondary origin, e.g. addition of seawater, low-grade metamorphism, and not a primary feature. La, Ce and Th on the other hand, do not show strong enrichment in the basic samples as discussed previously, showing ratios comparable to MORB values. This is not the case for the evolved rocks, which are accompanied by a decrease in ε_{Nd} (Fig. 6). This may imply that an another process besides simple fractional crystallization may play an important role during the evolution from the gabbros to the plagiogranites (e.g. AFC, mixing).

The Sr and Nd - isotope data provide important insights into the origin of magma systems; in particular they are important in identifying potential end members and for constraining possible mixing between them. The small data set for Sr and Nd isotopes for the Szarvaskö rocks has already been presented and their important features highlighted. In general the data show high $^{87}Sr/^{86}Sr$ ratios and lower ε_{Nd} values compared to the range of MORB compositions. Some BABB also have high $^{87}Sr/^{86}Sr$ ratios, e.g. in the East Scotia Sea and Lau Basin. In the case of the Szarvaskö rocks, alteration might have offset the original Sr isotope values without affecting Nd (Fig. 6). There must therefore be involvement of a mantle component with low $^{143}Nd/^{144}Nd$ ratios.

Mixing between MORB-source mantle and other components seems to occur in many back-arc basins. For example the Mariana Trough has both a MORB and an arc component and the Lau basin requires multiple mantle sources involving MORB- and OIB-like sources (Hawkins, 1995; Saunders et al., 1991). For the Szarvaskö igneous rocks there is (1) a MORB-like component depleted in incompatible elements, and (2) a distinct component with high $^{87}Sr/^{86}Sr$ and low $^{143}Nd/^{144}Nd$. Thus the source of this component could be IOB-like, an enriched MORB source mantle, but also a shale-like component, similar to the material into which the complex is intruded (Downes et al., 1990), cannot be ruled out.

The ophiolite dilemma

Ophiolites are thought to represent “fragments of oceanic crust and upper mantle detached and uplifted from their original deep-ocean setting by global tectonic processes” (Pearce, 1991), but are not necessarily formed at mid-ocean ridges (MORB-ophiolites); in fact, most of them are fragments of marginal basin crust (SSZ-ophiolites). There are several modern reviews dealing with the concept of ophiolites, e.g. (Cann, 1991; Gass, 1989; Nicolas, 1989; Pearce, 1991; Robertson and Xenophontos, 1993; Spray, 1991).

The idea that the Szarvaskö magmatic rocks represent parts of an ophiolite complex is not new in the literature. Balla et al. (1983), for example, has written a historical summary on the geological development of the Szarvaskö area. Several authors have also assigned a back-arc basin tectonic setting to the complex (Balla, 1987; Balla et al., 1983; Downes et al., 1990; Kázmér and Kovács, 1989; Kubovics et al., 1990). Here, we summarize the problems related to the ophiolitic nature of the Szarvaskö complex.

The Szarvaskö complex has an incomplete lithologic sequence as compared to the typical ophiolite sections, as shown in Table 4. This may result from its dismembered nature and/or from the limited degree of exposure of its rocks. There is no mantle sequence or sheeted dikes in this complex, which makes it difficult to link the parent melt to tholeiitic pillow basalts. Moreover, these magmatic rocks consist of an evolved assemblage of basalt-gabbro-(ultramafite)-plagiogranite bodies, an association not very typical to the crustal sequences seen in ophiolite complexes. Comparisons to the Alpine and Dinaride ophiolites are thus highly uncertain, and even the strong oceanic signature cannot be taken for granted. Downes et al. (1990) for instance have even considered the possibility of these rocks being strongly depleted continental tholeiites.

The chemical signature of the most basic rocks, already discussed in the last section, displays a strong MORB affinity. We have to keep in mind, though, that the “least differ-

Table 4 - Comparison of Szarvaskö complex with two distinct types of ophiolites.

Szarvaskö complex	Name	Eastern Alps [1]	Troodos [2]
terrigenous	sediment cover	terrigenous and pelagic	pelagic
N-MORB	chemical signature	N-MORB	IAB
not present	sheeted dikes	not present	present
evolved cumulates	cumulate sequence	ol-plag-cpx	ol-cpx-plag
not present	mantle sequence	harzburgite	harzburgite
???	ophiolite type	MORB	SSZ

Data from [1]Koller and Höck (1990) and [2]Pearce et al. (1984).

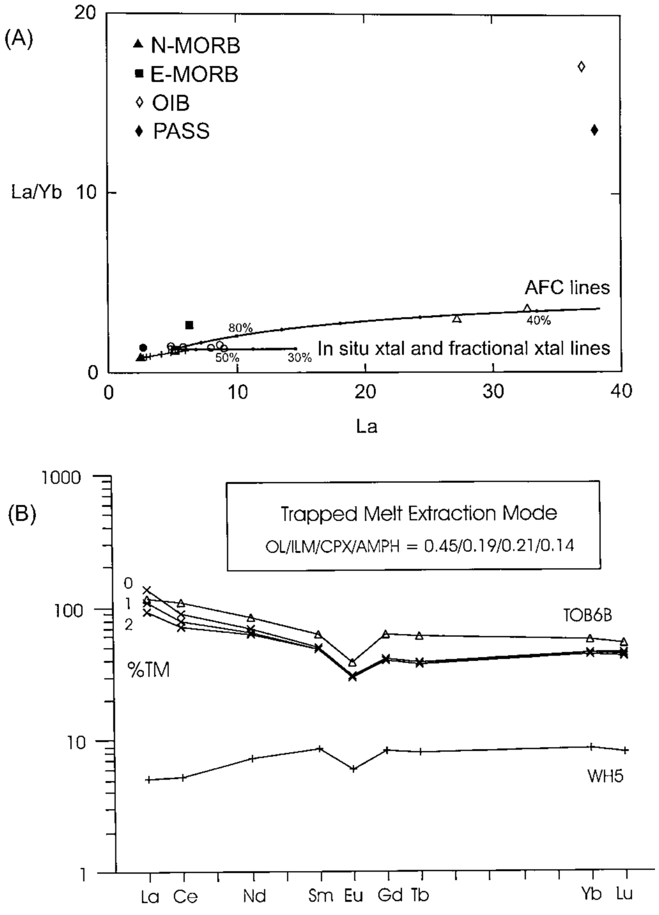

Fig. 8 - Petrogenetic models for the Szarvaskö samples (symbols as in Fig. 2, otherwise see caption); (a) N-MORB, E-MORB, and OIB sources from Sun and McDonough (1989); PASS source from Taylor and McLennan (1985); distribution coefficients are from Smith et al. (1998). In situ “modal” crystallization lines (line with crosses) assuming 10% *f* returned of N-MORB initial magma: model parameters = gabbroic source (ol20-cpx30-plag50). Fractional crystallization lines (line with open dots; ticks in % of parent): parent = sample RW1 using a fractionated gabbroic assemblage = ol30-cpx40-plag30. Assimilation-fractional crystallization (AFC) lines (ticks in % of parent) using r = 0.3 and mixing parent RW1 basalt with assimilant PASS; (b) REE liquid profiles, including the effects of trapped melt fraction (TMF) after Bédard (1994). Liquids calculated to be in equilibrium with ultramafic rock WH5 at different % TM. Those profiles show the possibility of a related origin between the ultramafic cumulates (WH5) and the high-silica melts (TOB6B) on the complex. Mineral abbreviations: OL = olivine, ILM = ilmenite, CPX = clinopyroxene (with orthopyroxene included), AMPH = amphibole.

entiated" rocks (i.e. basalts and fine-grained gabbros) of the complexes are themselves fractionated samples when compared to a typical N-MORB. This fact does not invalidate the incompatible element ratio arguments, but does impose some limitations for developing a petrogenetic model for this rock suite.

Previous work on the genesis of the Szarvaskö igneous rocks suggested that the chemical variation observed among these rocks could be explained by a combination of crystallization processes. For instance Balla and Dobretsov (1984) and Downes et al. (1990) evoked fractional crystallization of a single parent melt (from a basaltic, MORB-like magma) that, during magmatic fractionation in a shallow magma chamber, became contaminated by terrigenous sediments. The interpretation of the cumulate rocks varies a little; Balla and Dobretsov (1984) and Balla (1984) favour the idea that these rocks represent differentiated products of basalts after strong interaction with the surrounding sediments, while Downes et al. (1990) implied that they could have been derived from fractionation of an intermediate magma, i.e. with composition between the basic and evolved samples. From our previous discussion based on the geochemical signatures, it is clear that there will be no final model, which would perfectly describe all the petrogenesis of these rocks. In Fig. 8, we have considered two major operating processes, assimilation - fractional crystallization (AFC) and crystal accumulation, using REE patterns for modeling the melt evolution to present some possible petrogenetic models for the Szarvaskö igneous rocks. The first model, based on a N-type MORB composition, shows evolution trends for fractionated crystallization involving a gabbroic composition and a combined assimilation-fractional crystallization (AFC) path with mixing of a basalt (RW1 of Table 1) and a sediment component like PAAS (post-Archean average Australian shale, after Taylor and McLennan, 1985). The latter fits quite well the compositional range of the investigated plagiogranites and would explain easily the common occurrence of garnet in this rocks. The second model is based on the effect of a trapped melt (TMF), after Bédard (1994). A low rate of TMF (Fig. 8B) allows calculation of a possible dioritic melt between the ultramafic cumulates (WH5) and the high-silica melts (TOB6B) as starting compositions for the plagiogranites. Both the AFC and the TMF models can rather simply explain the origin and the geochemisty of the Szarvaskö plagiogranites.

CONCLUSIONS

The Szarvaskö igneous rocks with ultramafic and mafic cumulates, various gabbro types, evolved plutonic rocks equivalent to plagiogranites, and pillow lavas still provide an exceptional window into a branch of the "Vardar" ocean, even though their former tectonic setting cannot be assigned with confidence. According to nearby sediments of the Szarvaskö igneous rocks an intra-oceanic basin character of the sequence seems to be proved (Balla et al., 1983). The basalts of the Szarvaskö complex show a strong N-MORB component regarding their trace element signatures, while isotopic compositions tend to show an additional enriched source component. Although the complex has been viewed as a dismembered ophiolite sequence with back-arc basin basalt affinity, we can still not establish a secure link to other Jurassic Tethyan ophiolites.

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