PETROLOGY OF GABBROIC TO PLAGIOGRANITIC ROCKS FROM SOUTHERN TUSCANY (ITALY): EVIDENCE FOR MAGMATIC DIFFERENTATION IN AN OPHIOLITIC SEQUENCE

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ABSTRACT

This work deals with the petrogenesis of a composite ophiolite body composed of gabbroic to plagiogranitic rocks. In particular, the sequence consists of evolved gabbros, Fe-gabbros and Fe-diorites crosscut by ferrobasaltic dikes and intruded by dikelets and lenses of plagiogranites, mainly represented by diorites and trondhjemites. Complex relationships among the different lithotypes indicate multiple melt injections into both crystal-mush and solidified rocks. Whole-rock compositions show continuously increasing FeO/(FeO + MgO) and increasing REE, Th, Nb, Y, La_N/Sm_N and Eu depletion from gabbros to plagiogranites.

Major and trace element variations and modelling show that the whole intrusive sequence originated from the differentiation of a ferrobasaltic magma reservoir. Ferrobasaltic liquid compositions are compatible with fractional crystallization of N-MOR type basalts through olivine + plagioclase + clinopyroxene removal, whereas the evolved gabbros represent clinopyroxene + plagioclase cumulates from ferrobasaltic liquids with small amounts of trapped melt. Fe-gabbroic to Fe-dioritic rocks, characterized by abundant Fe-Ti oxides + apatite as liquidus phases, likely represent cumulates from evolved liquids, possibly segregated from the gabbroic crystal-mush. The attainment of Fe-Ti oxide saturation promoted silica enrichment yielding residual liquids of dioritic composition which were segregated and injected into fractures of solidified gabbroic rocks or infiltrated into the Fe-gabbroic crystal-mush. Locally, silica-rich and slightly LREE-enriched liquids of trondhjemite composition were produced through fractionation of plagioclase + clinopyroxene + amphibole + Fe-Ti oxides (+ apatite + zircon + allanite).

INTRODUCTION

The genesis of the felsic, evolved intrusive rocks in modern and fossil oceanic lithosphere is still poorly understood. In particular, the origin of small volumes of silica-rich K_2O-poor leucocratic rocks occurring in the uppermost part of ophiolitic plutonic sequence, i.e. the oceanic plagiogranites (Coleman and Peterman, 1975; Coleman and Donato, 1979), has received contrasting explanations. According to the most popular view they represent residual melts after extensive differentiation of MOR-type basaltic magmas (Coleman and Peterman, 1975; Coleman and Donato, 1979; Pallister and Knight, 1981; Pedersen and Malpas, 1984; Wilberg, 1987; Bébien, 1991; Floyd et al., 1998; Niu et al., 2002). On the other hand, liquid immiscibility with evolved basic melts (Dixon and Rutherford, 1979; Pognante et al., 1982), partial melting of hydrated mafic oceanic crust (Gerlach et al., 1981; Pedersen and Malpas, 1984; Flagler and Spray, 1991) and hydrous partial melting of gabbroic rocks (Koepke et al., 2004) have been also proposed.

Here we present a petrologic study of differentated intrusive rocks belonging to the oceanic lithosphere of the Jurassic Ligurian Tethys. The work focuses on a small ophiolite body from Bartolina quarry in southern Tuscany where an intrusive sequence ranging from evolved gabbros to trondhjemites, crosscut by ferrobasaltic dikes, is preserved. The wide compositional spectrum and the primary relationships among the different rock types represent a unique occurrence among the plutonic complexes of the Northern Apennine ophiolites, where evolved rocks are commonly scarce or absent. The complex relationships among the different rock types are illustrated and a detailed major and trace element fractional crystallization model is provided to explain the formation of the whole sequence. This ophiolitic sequence may therefore provide new insights into the late magmatic differentation processes taking place within the oceanic lithosphere.

GEOLOGIC AND PETROLOGIC FRAMEWORK

The ophiolitic complexes from the Northern Apennines (Fig. 1) are lithospheric remnants of the Ligurian Tethys oceanic basin which opened in Middle Jurassic (Lemoine et al., 1987; Bill et al., 2001). The ophiolites mainly occur in two distinct tectonic units, ascribed to different paleogeographic domains, i.e. an intraoceanic domain (Internal Liguride Units, Barrett and Spooner, 1977; Cortesogno et al., 1987) and a pericontinental one (External Ligurides, Marroni et al., 1998). The Internal Liguride ophiolites (IL) show stratigraphic relationships with their sedimentary cover (pelagic sediments and thick turbidite sequences ranging in age from Late Jurassic to Early Paleocene, Cortesogno et al., 1987). They are considered as analogues of oceanic lithosphere that originated at slow-spreading centres (Hebert et al., 1989; Lagabrielle and Cannat, 1990; Tribuzio et al., 2000a), or as embryonic oceanic crust, originated in an early stage of ocean formation, that was associated with older, unroofed subcontinental mantle (Rampone et al., 1998; Piccardo and Rampone, 2000). In the External Liguride (EL)

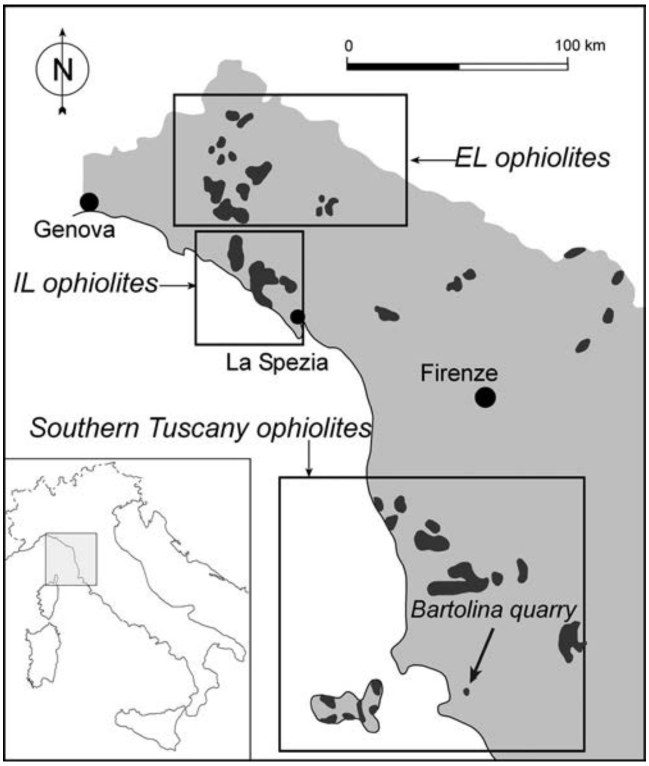

Fig. 1 - Location of the investigated rocks in the framework of the Northern Apennines, Corsica and Western Alps ophiolites (in black).

units, ophiolites consist of fertile lherzolites, MOR-type basalts and rare gabbroic rocks occurring as large slide-blocks within Cretaceous sedimentary mélanges, together with sparse continental crust bodies locally displaying primary contacts with the ophiolites (Pagani et al., 1972; Marroni et al.,1998). In southern Tuscany, the ophiolitic rocks form olistoliths embedded in flyschoid sequences of Late Cretaceous to Eocene age and are not associated with continental crust (Abbate et al., 1980). Their paleogeographic location within the Ligurian oceanic domain, however, is still unclear.

Previous studies have shown that the gabbroic complexes of the Northern Apennine - Tuscany ophiolites formed by low-pressure fractional crystallization of MOR-type liquids (Serri, 1980; Hebert et al., 1989; Tiepolo et al., 1997; Tribuzio et al., 2000a) in Early to Middle Jurassic times (Rampone et al., 1998; Tribuzio et al., 2004). In particular, late-stage closed-system fractional crystallization was originally proposed by Serri (1980) for the origin of the rare plagiogranitic rocks from some Ligurian and Tuscany ophiolite bodies. The Nd isotope composition (ε_{Nd} = 8.5-8.9) of ophiolitic gabbros from southern Tuscany (Tribuzio et al., 2004) testify for derivation from a depleted MORB-type asthenospheric source.

The gabbroic to plagiogranitic rocks under investigation crop out on a small area (0.2 km^3) mainly located in the Bartolina quarry, south of the village of Ribolla (Grosseto) and west of the Rosia-Monticiano-Roccastrada structural high. The ophiolitic body is partially buried under Quaternary alluvial sediments and shows tectonic contact with the Palombini Shales formation of Late Cretaceous age (Abbate et al., 1980 a; 1980b). The intrusive sequence consists of an association of evolved gabbros, Fe-gabbros and Fe-diorites characterized by significant grain-size and compositional variations; they are crosscut by ferrobasaltic dikes and intruded by centimeter-to decimeter-thick dikelets and lenses of plagiogranites that represent the latest magmatic event. The relative abundance of the lithotypes decreases from mafic to felsic rocks. The origin of some evolved Fe-rich gabbroids from this area was previously discussed on the basis of major and trace element mineral and whole-rock compositions by Tiepolo et al. (1997) and Tribuzio et al. (2000a).

^{40}Ar/^{39}Ar dating performed on amphiboles from two Fe-diorite samples yields 152±2 Ma (Bortolotti et al., 1990). Although this value is almost coincident with the emplacement age of the Ligurian plagiogranites testified by U-Pb zircon ages (Borsi et al., 1996), major element composition of the dated amphiboles (ranging in composition from Fe-actinolite to Al-poor Fehornblende) suggest that their ^{40}Ar/^{39}Ar age may record seafloor hydrothermal activity.

ANALYTICAL TECHNIQUES

Electron microprobe (EMP) analyses of minerals were carried out using a JEOL-6400 electron microprobe, equipped with a Oxford Instruments ISIS energy-dispersive microanalytic system at the University of Parma. Operating conditions were 15 kV accelerating voltage, 0.62 nA current, a spot diameter of 1-2 μm, and counting time of 100 s.

Whole-rock major and trace (Ba, Rb, Zr, Nb, Sr, Y) element analyses were determined by conventional X-ray fluorescence (XRF) techniques at the University of Pisa. The analyses of the other trace elements including rare earth elements (REE), Co, Hf, Ta, Th, and Sc were performed by instrumental neutron activation at the Department of Geology, Saint Mary’s University, Halifax (Canada). The precision and accuracy of the analytical methods were described in Dostal et al. (1986). In general, the precision is better than 5% for major element and 3-10% for trace elements.

PETROGRAPHY AND MINERALOGY

The studied rocks have been subdivided into three main groups: gabbros, Fe-gabbroic rocks and (oceanic) plagiogranites. Remarkably, the most primitive rocks of the intrusive oceanic sequence (dunites, troctolites and olivine gabbros) are totally lacking.

The rocks have been named following the IUGS nomenclature scheme. However, as for the Fe-rich gabbroic rocks we decided to adopt the terms Fe-gabbro and Fe-diorite (used in the literature on the Alpine-Apennine ophiolites) instead of oxide gabbro and diorite, respectively (as suggested by Dick et al., 2002 for rocks containing Fe-Ti oxide > 4 vol%).

The Bartolina ophiolitic sequence has been pervasively affected by a medium- to low grade ocean-floor metamorphism mainly developed under static conditions. However, the igneous textures are well preserved and relics of the primary igneous assemblage are widespread. The metamorphic event led to (i) growth of light green-blue green amphibole (Ti-poor hornblende to actinolite, Riccardi et al., 1994) ± chlorite as partial replacement of clinopyroxene and igneous amphiboles; (ii) alteration of plagioclase to albite + epidote ± prehnite ± chlorite aggregates, (iii) pseudomorphic replacement of olivine and orthopyroxene by chlorite (± opaques); (iv) replacement of Fe-Ti oxides by sphene. There is no evidence of former high-temperature shear zones characterized by recrystallization of clinopyroxene + Ti-rich amphibole as observed in the olivine-bearing gabbros of the Ligurian ophiolites (Molli, 1995; Tribuzio et al., 1995). Secondary amphiboles and chlorite, as well as calcite and stilpnomelane, may occur in veins.

These transformations are presumably related to the second metamorphic cycle recognized in the ophiolites from the Northern Apennines (Cortesogno et al., 1994), that was associated with the development of an ocean-floor hydrothermal system and dominant brittle deformations.

Gabbros

They are medium- to fine grained rocks with ophitic-subophitic texture. Grain size variations are similar to those described for the "patchy" or "varitextured gabbros" (Dick et al., 1991) from the intrusive sequence of the Southwest Indian Ridge. They have a primary assemblage of plagioclase (55-75 vol%), clinopyroxene (15-25 vol%), brown amphibole (~5 vol%), Fe-Ti oxides (2-3 vol%) and accessory amounts of apatite and exceedingly rare zircon. Clinopyroxene is locally poikilitic. Fe-Ti oxides are typically interstitial and frequently show brown amphibole rims; the amphibole can also occur as blebs in clinopyroxene. The common occurrence of subhedral-interstitial pseudomorphs of chlorite ± actinolite testifies for the former presence of small amounts of orthopyroxene (up to ~5 vol%), which is rarely preserved.

Fe-gabbroic rocks

Fe-gabbroic rocks include three different lithotypes, Fe-gabbros, apatite-rich Fe-diorites and Fe-diorites.

Fe-gabbros show hypidiomorphic texture and medium-fine grain size. The magmatic assemblages consist of plagioclase (65-70 vol%), clinopyroxene (20-25 vol%), Fe-Ti oxides (~ 5 vol%), yellow-brown and brown-green amphibole (5-10 vol%) occurring interstitially or as rims of Fe-Ti oxides, rare olivine and tiny apatite crystals enclosed in the main constituents.

A*patite-rich Fe-diorites* (Fig. 2a) are characterized by highly variable grain-size (very fine to coarse-grained) and by hypidiomorphic texture. They differ from Fe-gabbros by the presence of high amounts of apatite (> 5 vol%) occurring as large (up to 3 mm in length) euhedral crystals enclosed in clinopyroxene and plagioclase and small amount of interstitial, fresh biotite (Fig. 2b). Fe-Ti oxides (Ti-magnetite, > 5 vol%) occur as subhedral rounded grains; zircon may be an accessory phase. The amphibole is a relatively Ti-rich Fe-edenite (Table 1).

Fe-diorites have medium- to coarse-grained hypidiomorphic textures. The igneous assemblage includes plagioclase (65-85 vol%), clinopyroxene (10-30 vol%), brown-green amphibole (5-7 vol%, Fe-hornblende, Table 1), Fe-Ti oxides (5-10 vol%), Fe-rich olivine (3-5 vol%, Table 1) and accessory biotite (Table 1) and apatite.

Ferrobasalts

The ferrobasalts are aphyric to slightly porphyritic rocks characterized by a groundmass with intergranular texture. Their assemblage is formed by plagioclase + clinopyroxene + olivine + titanian pargasite (Table 1) + Fe-Ti oxides + apatite. Phenocrysts of plagioclase and olivine occur in places. Remarkably, oxides and apatite form subhedral microphenocrysts (Fig. 2c). Olivine is generally replaced by chlorite ± actinolite ± Fe-Ti oxides.

Plagiogranites

The plagiogranites include four different lithotypes: diorites, quartz-diorites, leucodiorites and trondhjemites.

Diorites are coarse-grained rocks with hypidiomorphic texture. More rarely, they have medium-grained porphyritic texture. They are composed of euehdral to subhedral plagioclase (70-75 vol%), subhedral brown-green amphibole (15-20 vol%), minor clinopyroxene (~5 vol%) and accessory amounts of Fe-Ti oxides, apatite, zircon and allanite. In the *quartz-diorites*, small amounts (2-3 vol%) of interstitial quartz are present.

Leucodiorites have the same mineralogical composition as the diorites with a lower colour index. They contain large amounts of plagioclase (85-95 vol%) and minor amphibole + clinopyroxene (5-15 vol%) and are characterized by trachytic texture with parallel orientation of plagioclase crystals ("igneous lamination", Fig. 2d).

Trondhjemites (Fig. 2e) commonly have a consertal allotriomorphic texture. They mainly consist of anhedral plagioclase (65-70 vol%) and quartz (~20 vol%) with minor green subhedral green amphibole (3-7 vol%); accessory minerals are Fe-Ti oxides, apatite, zircon. They may also occur as porphyritic rocks characterized by plagioclase + quartz phenocrysts in a microcrystalline groundmass.

Relationships among the different rock-types

The gabbros may show both sharp and irregular contacts with the diorite dikes. Along the gabbro-diorite interface, fractured crystals and injection of thin (2-3 mm in thickness) veins of felsic material have been observed. Centimeter-sized thick leucodiorite veins may also crosscut the gabbros.

The apatite-rich Fe-diorites contain millimetric lenses of diorites and leucodiorite veins displaying irregular margins and crystals with interpenetrated boundaries at the interface with the host. The ferrobasalts, in turn, include more felsic, medium-grained lenses formed by plagioclase + amphibole + Fe-Ti oxides + zircon characterized by fuzzy contacts with the host and may be crosscut by millimetre- to centimeter-thick plagiogranite veins (Fig. 2f).

Table 1 - Microprobe analyses of clinopyroxene, amphibole, biotite, olivine, magnetite in the Bartolina ophiolite

	FB CB9 Ti-prg	CB9 Cpx	Ap-FD B16 Fe-ed	FD B11 Fe-Hbl	FD CB14 Bi	CB14 Ol	CB14 Mag
SiO_2	43,83	52,40	45,48	46,80	34,81	39,73	-
TiO_2	2,63	0,33	1,77	1,09	2,94	0,01	7,94
Al_2O_3	9,45	1,07	6,69	5,01	13,42	0,49	0,82
Cr_2O_3	-	-	-	-	0,11	-	0,01
FeO	16,61	12,70	22,85	27,58	32,35	50,41	85,20
MnO	0,25	0,36	0,44	0,69	0,24	1,25	0,29
MgO	11,31	12,24	9,36	6,92	3,35	6,79	0,13
CaO	10,71	20,53	9,84	9,56	0,20	0,80	-
Na_2O	2,72	-	2,14	1,02	0,87	-	-
K_2O	0,30	-	0,29	0,53	7,51	-	-
Total	97,81	99,63	98,86	99,20	95,79	99,48	94,39

Ti-prg = titanian pargasite, Fe-ed = ferroedenite, Fe-hbl = ferro-hornblende, Bi = biotite, Ol = olivine, Mag = magnetite. Total iron as FeO, FB = ferrobasalt, Ap-FD = apatite-rich ferrodiorite, Fd = ferrodiorite.

GEOCHEMISTRY

Only major (Si, Ti, P, Al, Fe, Mg) and trace elements (Sc, Co, Cr, Ta, Th, Hf, Zr, Y, REE) that are generally believed to be relatively immobile during ocean-floor hydrothermal alteration have been taken into account for the petrogenetic discussion. The investigated samples have low to moderate loss on ignition (LOI), within the range 0.57-4.32 wt% (Table 2); the gabbroic rocks are characterized by higher LOI values (1.87-4.32 wt%) than the plagiogranites (0.57-1.91 wt%) in agreement with higher amounts of secondary hydrous phases observed in thin section.

Gabbros

The gabbros have FeO*/(FeO*+MgO) ratios varying between 0.46 and 0.64 positively correlated with TiO_2 and P_2O_5 (Fig. 3), whereas SiO_2 contents do not show significant variations. They are characterized by LREE depletion (La_N/Sm_N = 0.5-0.6, La_N/Yb_N = 0.6-0.9) and does not display a significant Eu anomaly (Eu/Eu* = 0.9-1.0) (Fig. 4). These rocks have intermediate composition between olivine gabbros of the Northern Apennine ophiolitic sequences and Fe-gabbroic rocks (Serri, 1980; Tiepolo et al., 1997).

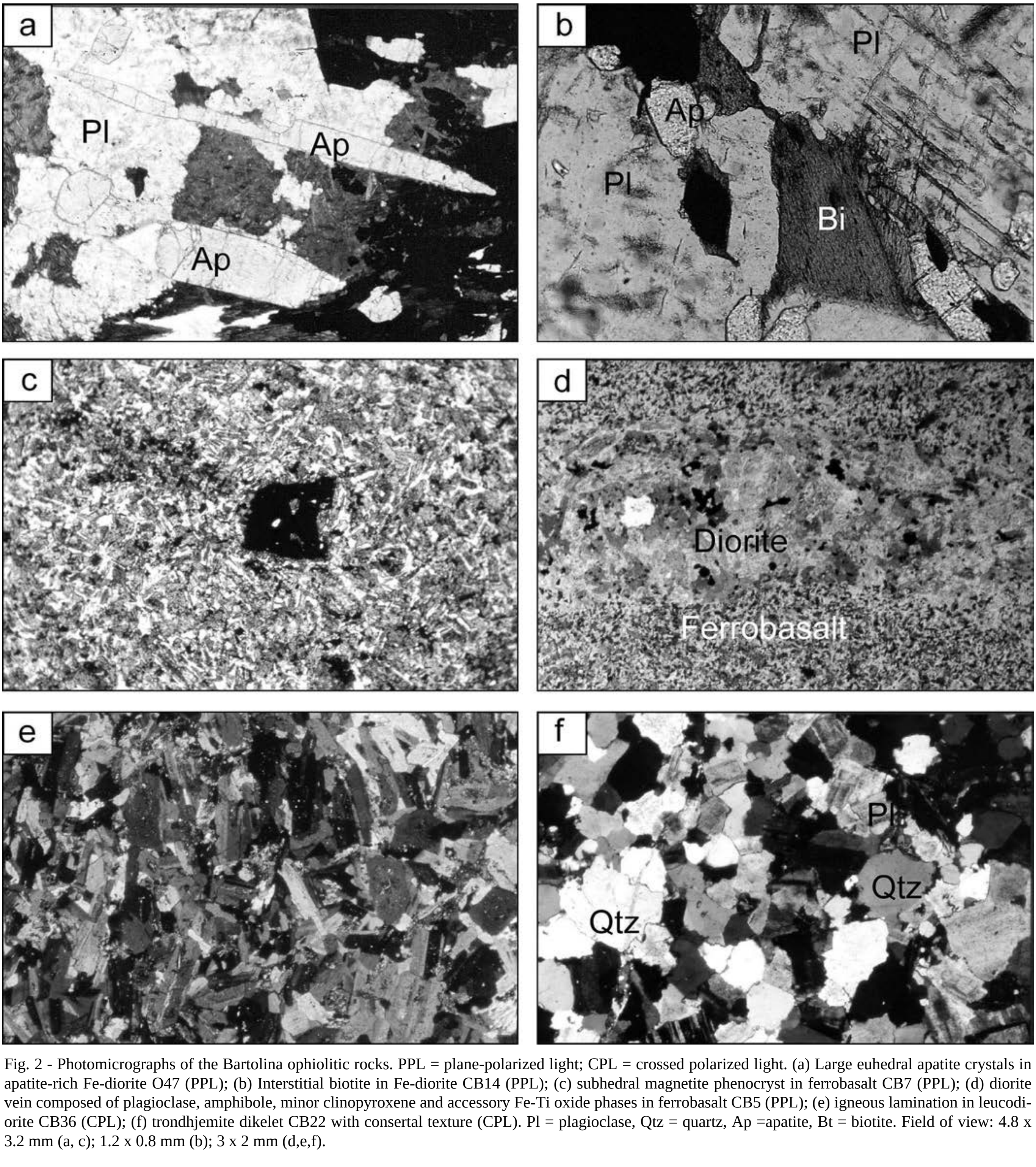

Fig. 2 - Photomicrographs of the Bartolina ophiolitic rocks. PPL = plane-polarized light; CPL = crossed polarized light. (a) Large euhedral apatite crystals in apatite-rich Fe-diorite O47 (PPL); (b) Interstitial biotite in Fe-diorite CB14 (PPL); (c) subhedral magnetite phenocryst in ferrobasalt CB7 (PPL); (d) diorite vein composed of plagioclase, amphibole, minor clinopyroxene and accessory Fe-Ti oxide phases in ferrobasalt CB5 (PPL); (e) igneous lamination in leucodiorite CB36 (CPL); (f) trondhjemite dikelet CB22 with consertal texture (CPL). Pl = plagioclase, Qtz = quartz, Ap =apatite, Bt = biotite. Field of view: 4.8 x 3.2 mm (a, c); 1.2 x 0.8 mm (b); 3 x 2 mm (d,e,f).

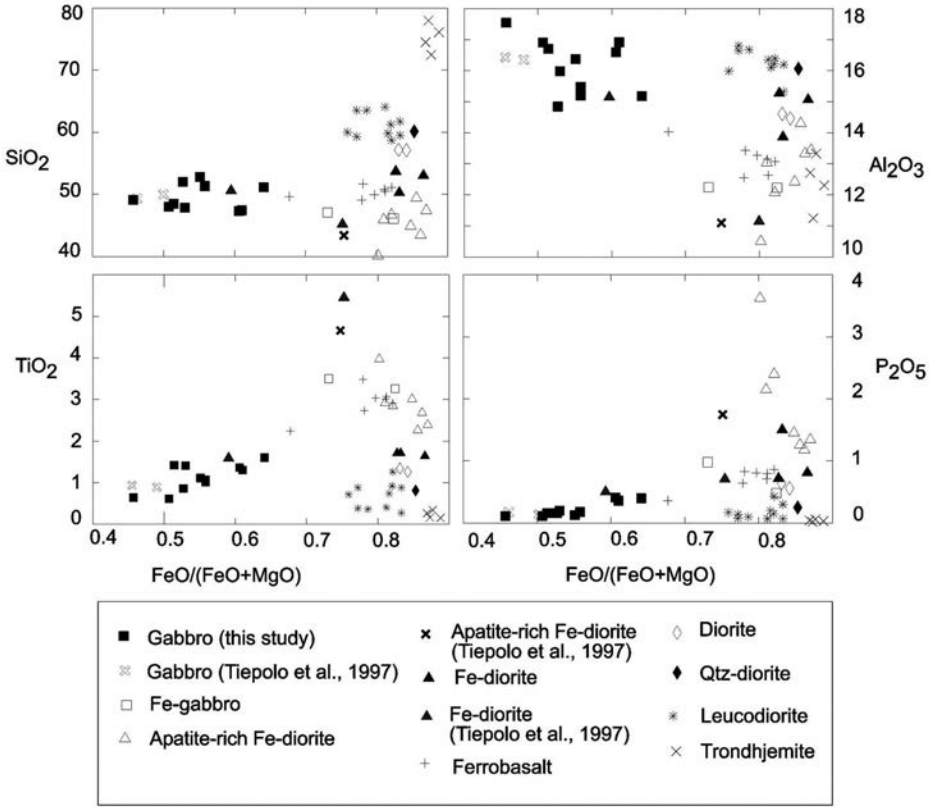

Fig. 3 - Variations of SiO_2, TiO_2, Al_2O_3, P_2O_5 (wt%) vs. FeO*/(FeO*+MgO) (wt%) in the ophiolitic rocks of the Bartolina quarry.

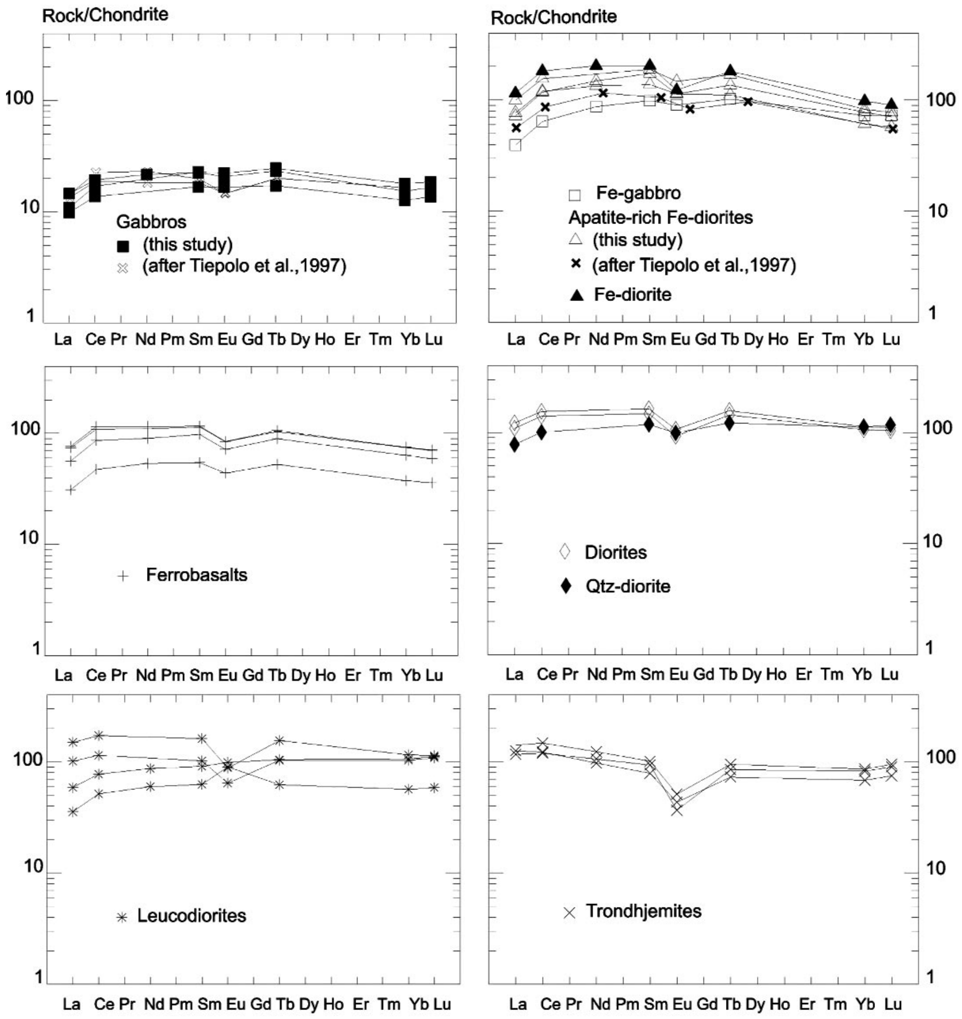

Fig. 4 - Chondrite-normalized whole-rock REE patterns (normalizing values after Andersen and Grevesse, 1989) for the rocks the Bartolina quarry.

Table 2 - Whole-rock major and trace element compositions of the Bartolina ophiolitic rocks

	Gabbros										
Sample	BAR10	CB 2	CB10	CB11	CB43	CB44	BAR4	CB47A	CB50B	CB50C	CB 50A
Major (wt. %) and trace element (ppm) abundances by X-ray fluorescence analyses											
SiO_2	48,46	52,79	47,41	47,28	47,87	49,11	51,15	52,04	51,36	51,29	47,98
TiO_2	1,42	1,11	1,30	1,36	1,41	0,64	1,60	0,86	1,01	1,07	0,61
Al_2O_3	16,70	16,37	16,91	16,59	15,98	17,54	15,18	14,84	15,19	15,47	16,90
Fe_2O_3	3,60	2,18	3,57	2,98	5,70	4,51	4,46	3,63	4,69	5,03	4,82
FeO	5,14	6,07	8,16	8,34	4,56	3,75	7,03	5,28	5,36	5,07	5,36
FeO*	8,38	8,04	11,38	11,03	9,79	7,81	11,04	8,54	9,59	9,59	9,69
MnO	0,14	0,15	0,21	0,19	0,15	0,13	0,21	0,15	0,17	0,19	0,16
MgO	7,92	6,55	7,28	7,18	8,69	9,27	6,17	7,67	7,60	7,61	9,41
CaO	9,20	6,22	7,61	7,14	8,65	7,10	6,08	6,88	7,36	7,29	7,42
Na_2O	3,16	4,85	2,89	3,30	3,04	3,05	4,83	4,89	4,37	4,54	2,82
K_2O	0,85	1,04	1,25	0,90	0,72	1,39	0,78	0,62	0,65	0,70	1,19
P_2O_5	0,16	0,13	0,36	0,41	0,20	0,11	0,40	0,16	0,18	0,18	0,11
L.O.I.	3,24	2,55	3,05	4,32	3,00	3,40	2,11	3,00	2,15	1,87	3,41
FeO*/(FeO*+MgO)	0,514	0,551	0,610	0,606	0,530	0,457	0,641	0,527	0,558	0,558	0,507
Ba	64	80	108	73	50	105	71	56	48	58	84
Nb	3	< 3	3	6	3	3	8	3	4	3	3
Zr	89	68	90	167	92	63	257	122	120	123	61
Y	25	27	32	48	26	23	72	39	40	43	20
Sr	280	445	327	266	287	329	303	167	249	236	287
Rb	12	13	22	20	12	22	9	10	10	11	23
Zr/Nb	29,7	-	30,0	27,8	30,7	21,0	32,1	40,7	30,0	41,0	20,3
Zr/Y	3,6	2,5	2,8	3,5	3,5	2,7	3,6	3,1	3,0	2,9	3,1
Y/Nb	8,3	-	10,7	8,0	8,7	7,7	9,0	13,0	10,0	14,3	6,7

	Fe-gabbroic rocks											
	FG	FG	Ap-FD	Ap-FD	Ap-FD	Ap-FD	Ap-FD	Ap-FD	Ap-FD	FD	FD	FD
	CB1	B8	B 16	CB 4	O50	CB16	CB30	O47	O 51	B 11	CB 14	CB29
Major (wt. %) and trace element (ppm) abundances by X-ray fluorescence analyses												
SiO_2	46,09	46,78	44,87	43,47	46,68	49,40	47,39	45,90	40,11	50,25	53,02	53,68
TiO_2	3,26	3,50	3,01	2,68	2,84	2,26	2,39	2,91	3,97	1,71	1,64	1,71
Al_2O_3	12,23	12,24	12,42	13,33	12,07	14,30	13,44	13,03	10,50	13,86	15,06	15,27
Fe_2O_3	5,45	4,44	4,37	4,61	5,05	3,18	5,64	5,45	6,82	5,65	2,75	2,55
FeO	13,70	11,10	15,44	13,07	13,01	13,21	12,83	11,90	15,34	9,51	10,96	9,68
FeO*	18,60	15,06	19,38	17,22	17,56	16,07	17,91	16,80	21,47	14,60	13,42	11,97
MnO	0,39	0,28	0,35	0,36	0,31	0,33	0,36	0,25	0,32	0,28	0,26	0,21
MgO	3,96	5,55	3,47	2,76	3,82	2,70	2,68	3,95	5,30	2,94	2,07	2,51
CaO	8,21	8,49	7,54	6,80	7,01	6,66	6,92	7,47	7,85	6,79	4,99	4,90
Na_2O	4,02	4,12	4,30	5,40	4,12	5,46	5,11	4,60	2,98	5,63	6,06	6,50
K_2O	0,19	0,39	0,21	0,20	0,23	0,17	0,15	0,13	0,19	0,21	0,23	0,26
P_2O_5	0,48	0,98	1,45	1,18	2,40	1,26	1,34	2,15	3,62	1,50	0,81	0,72
L.O.I.	2,02	2,14	2,57	1,14	2,47	1,08	1,74	2,25	3,01	1,67	2,16	2,01
FeO*/(FeO*+MgO)	0,824	0,731	0,848	0,862	0,821	0,856	0,870	0,810	0,802	0,832	0,866	0,827
Ba	35	39	67	37	22	27	27	28	58	20	29	23
Nb	19	11	12	21	10	11	15	16	7	9	11	9
Zr	166	204	269	243	145	245	109	196	103	486	188	243
Y	108	100	139	162	158	135	163	172	152	187	114	150
Sr	167	166	147	207	158	184	180	160	114	166	199	159
Rb	4	5	4	< 3	3	8	3	<3	4	3	10	4
Zr/Nb	8,7	18,5	22,4	11,6	14,5	22,3	7,3	12,3	14,7	54,0	17,1	27,0
Zr/Y	1,5	2,0	1,9	1,5	0,9	1,8	0,7	1,1	0,7	2,6	1,6	1,6
Y/Nb	5,7	9,1	11,6	7,7	1,1	0,6	1,5	0,9	21,7	20,8	10,4	16,7

FeO* = total iron; FG = ferrogabbro, ap-FD = apatite-rich ferrodiorite, FD = ferrodiorite.

Table 2 (continued)

	Ferrobasalts							**Plagiogranites**			
								Diorites		Leucodiorites	
	B 6	CB 5	CB 6	CB 9	B15	CB7	CB15	CB 35	CB 32	CB 50D	CB 8
Major (wt. %) and trace element (ppm) abundances by X-ray fluorescence analyses											
SiO_2	49,08	49,94	51,13	49,63	51,69	50,48	50,82	57,16	57,08	59,28	64,09
TiO_2	3,48	3,04	2,91	2,24	2,73	3,07	3,01	1,34	1,26	0,88	0,40
Al_2O_3	12,55	13,27	13,07	14,03	13,42	12,62	13,14	14,60	14,46	16,80	16,35
Fe_2O_3	3,75	4,19	2,21	1,77	2,43	2,50	1,92	2,51	2,41	2,48	0,22
FeO	11,37	10,78	12,60	10,70	11,37	12,73	13,07	7,90	8,34	4,26	4,08
FeO*	15,35	14,55	15,48	12,29	13,55	14,98	14,80	10,16	10,51	6,49	4,28
MnO	0,29	0,24	0,27	0,23	0,22	0,28	0,25	0,15	0,15	0,09	0,07
MgO	4,36	3,70	3,37	5,86	3,80	3,47	3,44	2,07	1,97	1,92	0,99
CaO	6,23	6,00	5,88	7,57	5,36	6,39	6,15	4,11	4,13	3,48	1,85
Na_2O	4,89	5,52	5,58	4,44	5,74	5,33	5,30	7,78	7,60	8,47	9,87
K_2O	0,13	0,15	0,17	0,47	0,21	0,17	0,27	0,11	0,13	0,14	0,11
P_2O_5	0,64	0,81	0,86	0,36	0,83	0,80	0,71	0,64	0,56	0,14	0,07
L.O.I.	2,64	2,36	1,96	2,70	2,20	2,15	1,91	1,63	1,90	2,05	1,90
FeO*/(FeO*+MgO)	0,779	0,797	0,821	0,677	0,781	0,812	0,811	0,831	0,842	0,772	0,812
Ba	23	35	33	50	30	25	34	62	43	32	46
Nb	11	13	12	6	10	11	10	16	16	10	9
Zr	456	682	584	207	517	533	410	1058	1063	2594	877
Y	104	120	120	59	110	111	91	175	182	95	144
Sr	156	106	136	235	126	131	139	112	98	170	47
Rb	<3	<3	8	15	<3	7	10	3	<3	3	14
Zr/Nb	41,5	52,5	48,7	34,5	51,7	48,5	41,0	66,1	66,4	259,4	97,4
Zr/Y	4,4	5,7	4,9	3,5	4,7	4,8	4,5	6,0	5,8	27,3	6,1
Y/Nb	9,5	9,2	10,0	9,8	11,0	10,1	9,1	10,9	11,4	9,5	16,0

	Leucodiorites							Q-Diorite	Trondhjemites		
	CB 17	CB 31	B12	CB50E	O49	CB24	B1	CB36	CB 41	CB 20	CB 21	CB 22
Major (wt. %) and trace element (ppm) abundances by X-ray fluorescence analyses												
SiO_2	59,81	59,51	61,25	60,00	58,71	61,73	63,51	63,50	60,12	74,49	72,48	76,11
TiO_2	0,74	0,88	0,92	0,71	1,26	0,27	0,36	0,38	0,81	0,26	0,34	0,15
Al_2O_3	16,10	15,32	16,22	15,98	16,39	16,19	16,67	16,65	16,06	12,71	13,32	12,30
Fe_2O_3	1,99	1,90	1,62	1,83	3,29	2,43	1,31	1,48	2,89	0,89	1,17	0,49
FeO	5,99	6,29	4,57	2,82	5,55	3,93	3,23	3,13	5,74	1,87	2,51	1,15
FeO*	7,78	7,99	6,03	4,47	8,51	6,12	4,41	4,46	8,34	2,66	3,56	1,59
MnO	0,10	0,15	0,08	0,09	0,14	0,09	0,05	0,05	0,14	0,03	0,04	0,02
MgO	1,75	1,60	1,32	1,42	1,86	1,23	1,20	1,32	1,44	0,40	0,50	0,20
CaO	3,46	3,57	3,26	4,98	2,93	2,49	2,45	2,66	2,49	0,98	1,21	1,27
Na_2O	8,42	8,47	9,02	9,01	7,89	8,96	9,86	9,85	8,32	7,64	7,67	7,65
K_2O	0,10	0,11	0,13	0,12	0,23	0,14	0,04	0,05	0,22	0,11	0,10	0,06
P_2O_5	0,20	0,30	0,15	0,17	0,43	0,07	0,10	0,08	0,25	0,04	0,06	0,03
L.O.I.	1,34	1,91	1,46	2,89	1,31	2,47	1,23	0,86	1,52	0,58	0,60	0,57
FeO*/(FeO*+MgO)	0,816	0,833	0,82	0,759	0,821	0,833	0,786	0,772	0,853	0,869	0,877	0,888
Ba	16	62	50	45	-	44	10	5	62,4	25	23	27
Nb	5	15	7	10	13	7	15	14	13	12	10	6
Zr	1793	1351	312	1584	2157	492	891	911	1847	393	583	364
Y	136	178	68	93	105	129	119	117	150	137	137	123
Sr	106	108	97	208	149	58	229	278	91	61	54	58
Rb	<3	3	4	<3	4	5	<3	<3	3	4	3	<3
Zr/Nb	358,6	90,1	44,6	158,4	165,9	70,3	59,4	65,1	142,1	32,8	58,3	60,7
Zr/Y	13	8	5	17	21	4	7	8	12	3	4	3
Y/Nb	27,2	11,9	9,7	9,3	8,1	18,4	7,9	8,4	11,5	11,4	13,7	20,5

FeO* = total iron, Q-diorite = quartz-diorite.

Fe-gabbroic rocks

The analyzed Fe-gabbro samples (CB1, B8) show low SiO_2 (lower than gabbros) and marked enrichment in TiO_2, FeO*, MnO, P_2O_5, Sc, Nb, Ta (Table 2-3). In particular, the sample CB1 shows the highest MnO (0.39 wt%), Sc (53.1 ppm), Nb (19 ppm) and Ta (1.07 ppm) abundances found in the investigated ophiolitic sequence. The same sample (Fig. 4) display strong LREE depletion (La_N/Sm_N = 0.4, La_N/Yb_N =0.5) and a slight Eu negative anomaly (Eu/Eu* = 0.9).

The apatite-rich Fe-diorites show the lowest SiO_2 concentrations (40.1-49.4 wt%) and high TiO_2 (2.3-4.0 wt%), P_2O_5 (1.2-3.6 wt%), MnO (0.25-0.36 wt%) and FeO* (16.1-21.5 wt%) values. They are highly enriched in Y (135-172 ppm), Nb (7-16 ppm) and Ta (0.4-1.1 ppm). With regard to the transition elements, Cr is strongly depleted (< 5 ppm), whereas Sc (29.0-46.2 ppm) and Co (23.8- 60.5 ppm) show relatively high and variable values. REE patterns (Fig. 4) are LREE-depleted ($(La/Sm)_N$ = 0.4-0.6), and display HREE fractionation (Tb_N/Yb_N = 1.8-2.0) and a slight negative Eu anomaly (Eu/Eu* = 0.8). Absolute REE values reach 100 times chondrite.

The Fe-diorites are relatively enriched in SiO_2 (51.1-54.8 wt%) and have lower contents of TiO_2, P_2O_5, FeO*, MnO and MgO, Co, Sc, Ta, Nb (Tables 2-3; Figs. 3-5) than the apatite-rich Fe-diorites, whereas Y (114-187 ppm) and Zr are still high. The samples analyzed for REE (Fig. 4) show different patterns. Sample B11 is depleted in Eu (Eu/Eu* = 0.6) and has the highest total REE concentrations. In contrast, sample CB 14 has a slight positive Eu anomaly (Eu/Eu* = 1.11) and considerably lower REE contents; both rocks are LREE-depleted (La_N/Sm_N = 0.6).

Ferrobasalts

These rocks represent evolved mafic liquids characterized by remarkably high FeO*/(FeO*+MgO) ratios (0.68-0.81). Incompatibile trace elements (Zr, Hf, Nb, Ta, Th, Y, REE) increase with increasing FeO*/(FeO* + MgO) ratios (Fig. 5). Total REE contents are similar to those of the Fe-gabbroic rocks; they are LREE-depleted (La_N/Sm_N = 0.6-0.7) and have slightly fractionated HREE (Tb_N/Yb_N = 1.4, Fig. 4).

Plagiogranites

The plagiogranites show large variation of major elements (Table 2, Fig. 3). A steady increase of SiO_2 from diorites (57.1 wt%) to trondhjemites (78.0 wt%) is associated with considerable decrease of TiO_2 (1.34-0.15 wt%), FeO* (10.5-1.6 wt%), MgO (2.07-0.2 wt%), P2O5 (0.64-0.02 wt%) and MnO (0.15-0.02 wt%). All the plagiogranitic rocks are strongly depleted in compatible trace element (Co, Cr, Sc, Table 3), and relatively enriched in incompatibile trace elements. In particular, the trondhjemites have the highest Th contents (up to 3.2 ppm) whereas the highest Zr (up 2600 ppm) and Hf (42 ppm) concentrations are reached in the leucodiorites.

The diorites have moderate REE abundances, LREE depletion (La_N/Sm_N = 0.7), slightly fractionated HREE (Tb_N/Yb_N = 1.4) and negative Eu anomalies (Eu/Eu* = 0.6-0.7) (Fig. 4) The quartz diorite sample CB41 has a similar pattern with lower total REE concentrations. The leucodiorites have different REE compositions. Samples CB8 and CB31 have similar characteristics, i.e. almost flat REE patterns (La_N/Yb_N = 1.0-1.3) with no LREE depletion

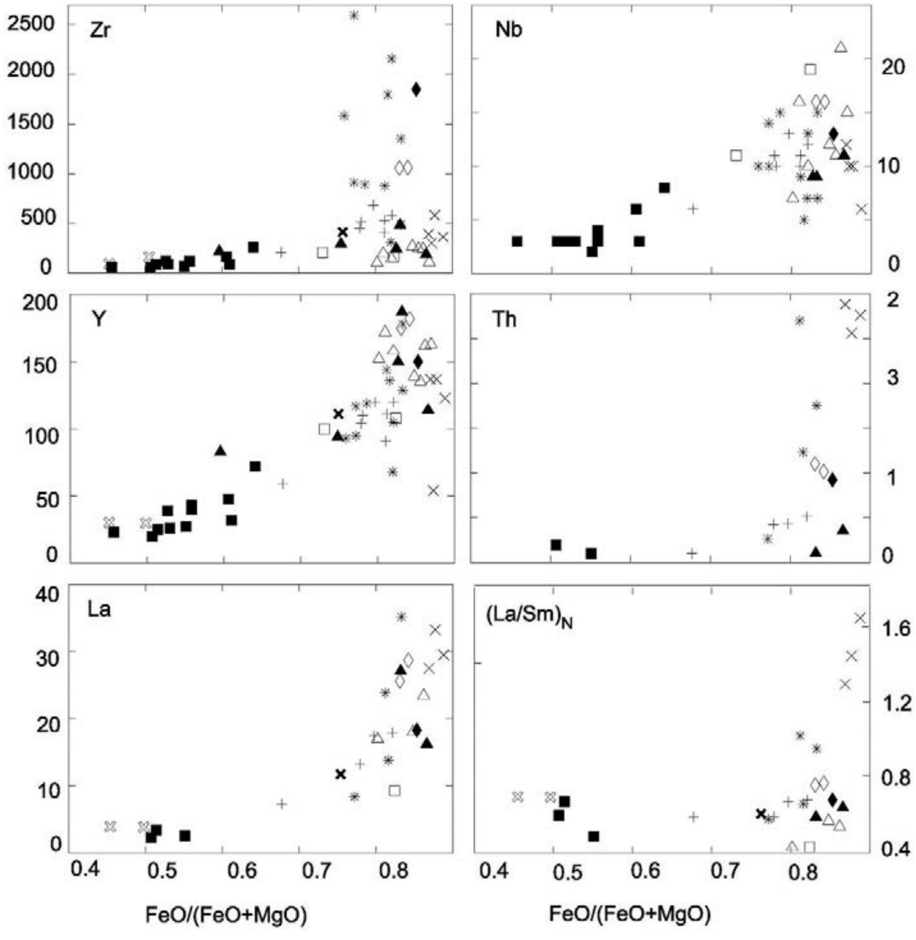

Fig. 5 - Variations of Cr, Ta, Hf, Th, La, Yb (ppm) vs. FeO*/(FeO*+MgO) (wt%) in the ophiolitic rocks of the Bartolina quarry. Symbols as in Fig. 3.

(La_N/Sm_N = 0.9-1.0) and a deep negative Eu anomaly = 0.5-0.6). In contrast, samples CB17 and CB50D are LREE-depleted (La_N/Sm_N = 0.6) and the sample CB50D, in particular, display an evident positive Eu anomaly (Eu/Eu* = 1.4).

The trondhjemites are characterized by high REE contents, slight LREE enrichment (La_N/Sm_N = 1.3-1.6) and a strong Eu depletion (Eu/Eu* = 0.4-0.6).

PETROGENETIC CONSTRAINTS FOR THE ORIGIN OF GABBROIC ROCKS

The relatively high FeO*/(FeO* + MgO) ratios, the generally low contents of compatible trace elements (Co, Cr) in the gabbroic rocks and the lack of associated more primitive cumulates (e.g. troctolites and olivine gabbros) indicate that the Bartolina intrusive sequence represent a differentiated portion of the oceanic crust of the Jurassic Ligurian Tethys. The ferrobasalts, on the other hand, are the most evolved basic magmas found in the ophiolitic sequences from the Northern Apennines.

The ubiquitous LREE depletion (Fig. 4, Table 3) suggest that the parental liquids had a N-MOR affinity. High Zr/Nb (17-52) and Y/Nb (9.1-11) ratios in the ferrobasalts coupled with low Zr/Y (4-5) and La_N/Yb_N (0.8-1.0) are consistent with this notion (Le Roex et al., 1995). Mass balance calculations and trace element modelling (Table 4) show that the ferrobasalts could be the result of advanced fractionation (residual melt fraction ~ 20 wt%) of a N-MORB primitive melt through the removal of olivine, plagioclase and clinopyroxene. They would therefore represent residual liquids after troctolite and olivine gabbro accumulation. The attainment of the Fe-Ti oxide saturation, testified in the ferrobasalts by euhedral microphenocrysts, likely promoted the reversal of the Fe-Ti enrichment trend and the beginning of SiO_2 enrichment in the liquids, as shown by a number of experimental studies on the differentiation of tholeiitic liquids carried out at low pressure (e.g. Spulber and Rutherford, 1983; Juster et al., 1989; Toplis and Carroll, 1996; Berndt et al., 2005).

Whole-rock chemical compositions of gabbroic rocks can be generally explained as mixtures of trapped melt and cumulus crystals. Typically, these compositions are not representative of magmatic liquids. Serri (1980) proposed that the gabbronorites from the Northern Apennines, which are chemically analogous to the gabbros of the present study, represent cumulitic rocks from ferrobasaltic liquids formed just before the saturation in Fe-Ti oxides. To test this hypothesis, we used the Bedard (1994) equation to calculate the composition of the liquids in equilibrium with the gabbros. The modelling results (Fig. 6) for incompatible elements are consistent with an origin as cumulates from liquids similar to the ferrobasalts occurring as dikes in the Bartolina ophiolite with small amounts (< 10 %) of trapped melt.

The Fe-gabbroic rocks from ophiolites and modern oceanic settings are generally regarded as having crystallized from evolved liquids (e.g. Serri, 1980; Tiepolo et al., 1997; Natland and Dick, 2002). Several hypotheses may be formulated, however, for the origin of the apatite-rich Fe-diorites. These rocks, characterized by unusually high P_2O_5 contents, could be explained as (i) the result of prolonged crystal-liquid fractionation from a MOR-type basaltic magma under low f_{O2} conditions, (ii) liquids with extreme Fe, Ti and P enrichment formed through liquid immiscibility processes, (iii) apatite- and Fe-Ti oxide-rich cumulates with variable amounts of trapped liquids. Although representatives of differentiated liquids of fairly similar compositions are known in literature (e.g. the dioritic chilled pillow from the Nain plutonic complex, Morse, 1990), the exceedingly low SiO_2 (< 45 wt%) in most samples, the scattering of

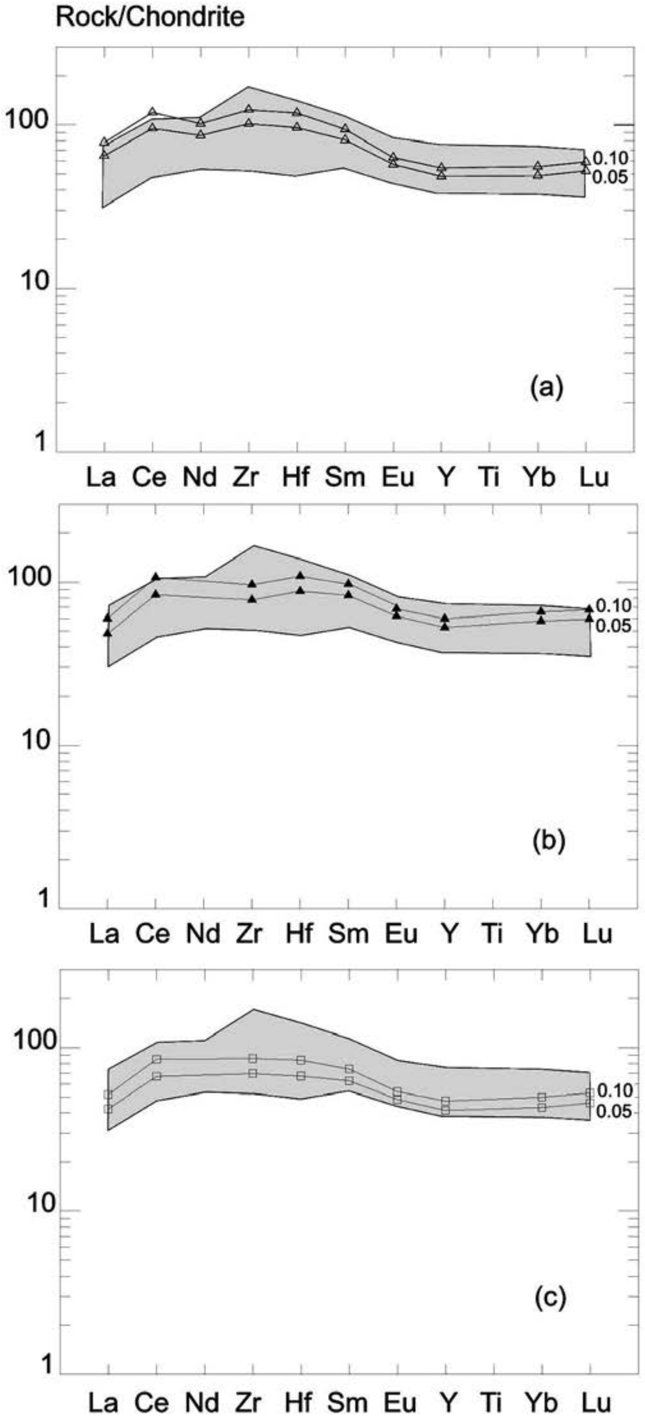

Fig. 6 - Calculated incompatible element compositions of parental liquids of gabbros BAR10 (a), CB2 (b), CB50A (c) considered as plagioclase + clinopyroxene cumulate rocks and assuming clinopyroxene/plagioclase ratio in the solid = 1. Numbers on the patterns refer to trapped melt percentages used for the calculations. Calculations were carried out according to the procedure of Bédard (1994), on the basis of whole-rock trace element concentrations. Partition coefficients for plagioclase and clinopyroxenes after Bindeman et al. (1998), Dunn and Sen (1994) and Hart and Dunn (1993). Shaded fields: ferrobasalt compositions.

Table 3 - Co, Sc, Cr, Hf, Th, Ta, REE abundances for selected samples

	Gabbros			**Fe-gabbroic rocks**						**Ferrobasalts**			
				FG	Ap-FD	Ap-FD	Ap-FD	FD	FD				
Sample	BAR10	CB 2	CB 50A	CB1	B 16	CB 4	O 51	B 11	CB 14	B 6	CB 5	CB 6	CB 9
Co	37,7	33,6	45,2	30,9	29,8	23,8	60,5	21,7	16,0	39,6	31,8	29,9	39,1
Sc	29,6	43,6	22,1	53,1	46,2	29,0	33,2	35,2	24,6	35,2	31,7	31,1	36,3
Cr	642	18,3	300	36,4	2,4	2,8	4,4	2,6	0,6	13,9	3,9	4,3	135
Hf	2,28	2,07	1,51	4,66	7,13	5,29	2,88	11,5	5,50	10,4	14,9	13,5	5,05
Th	< 0.1	0,1	0,19	< 0.1	< 0.1	< 0.1	< 0.1	0,1	0,35	0,42	0,43	0,51	0,1
Ta	0,10	0,14	0,08	1,07	0,64	1,08	0,35	0,46	0,62	0,55	0,82	0,73	0,39
La	3,41	2,55	2,30	9,25	18,0	23,4	16,9	27,1	16,2	13,2	17,4	17,9	7,29
Ce	11,6	10,2	8,24	38,7	72,3	93,7	70,8	110	55,5	52,2	65,6	69,3	28,6
Nd	9,78	-	-	39,4	60,6	-	67,2	91,8	46,6	40,8	49,9	52	24,3
Sm	3,29	3,35	2,46	14,5	20,3	27,8	25,5	29,9	16,4	14,4	16,8	17,1	8,03
Eu	1,16	1,24	0,93	5,06	6,36	8,23	6,34	6,87	6,34	4,02	4,72	4,73	2,46
Tb	0,84	0,89	0,62	3,71	4,94	6,14	4,04	6,58	3,50	3,25	3,76	3,84	1,91
Yb	2,49	2,91	2,06	11,8	12,6	13,4	9,90	15,90	11,1	10,3	12,1	12,2	6,11
Lu	0,40	0,45	0,33	1,77	1,72	1,88	1,39	2,19	1,70	1,44	1,71	1,72	0,87
$(La/Sm)_N$	0,67	0,49	0,60	0,41	0,57	0,54	0,43	0,59	0,64	0,59	0,67	0,68	0,59
$(La/Yb)_N$	0,98	0,63	0,80	0,56	1,02	1,25	1,22	1,22	1,05	0,92	1,03	1,05	0,86
$(Tb/Yb)_N$	1,53	1,39	1,37	1,43	1,78	2,08	1,85	1,88	1,43	1,43	1,41	1,43	1,42
Eu/Eu*	0,90	0,91	0,97	0,88	0,82	0,83	0,85	0,64	1,11	0,77	0,78	0,77	0,82

	Plagiogranites									
	Diorites		Leucodiorites				Q-Diorite	Trondhjemites		
Sample	CB 35	CB 32	CB 50D	CB 8	CB 17	CB 31	CB 41	CB 20	CB 21	CB 22
Co	14,4	12,0	7,41	3,57	9,99	10,8	6,73	3,14	3,83	6,31
Sc	19,5	20,0	16,9	6,04	13,5	13,3	12,9	3,40	4,62	2,68
Cr	1,7	0,9	1,9	3,7	1,0	13,2	2,2	1,9	1,1	0,8
Hf	23,9	23,5	42,3	22,2	35,8	29,5	37,9	15,0	19,0	13,2
Th	1,10	1,01	0,26	2,70	1,23	1,75	0,92	2,88	2,56	2,76
Ta	0,94	0,88	0,39	0,61	0,23	0,74	0,62	1,12	0,62	0,57
La	25,6	28,7	8,34	23,8	13,8	35,2	18,3	27,5	33,2	29,5
Ce	84,5	93,3	31,1	69,1	46,8	104	60,7	72,1	89,2	73,9
Nd	-	-	27,1	-	39,4	-	-	48,0	55,9	44,1
Sm	21,7	24,1	9,22	15,0	13,4	23,8	17,4	13,8	14,9	11,6
Eu	5,18	5,98	5,05	3,60	5,54	4,94	5,56	2,06	2,87	2,43
Tb	5,22	5,71	2,26	3,75	3,80	5,64	4,42	3,09	3,45	2,65
Yb	17,2	18,4	9,20	16,8	17,3	18,8	18,2	13,4	14,0	11,1
Lu	2,52	2,69	1,42	2,64	2,70	2,76	2,85	2,19	2,31	1,82
$(La/Sm)_N$	1,07	1,12	0,65	1,02	0,57	1,34	0,72	1,47	1,70	1,91
$(La/Yb)_N$	1,38	1,41	1,12	1,01	1,00	1,36	1,10	1,05	1,12	1,09
$(Tb/Yb)_N$	0,63	0,66	1,43	0,62	0,98	0,55	0,81	0,41	0,52	0,57
Eu/Eu*	0,63	0,66	1,43	0,62	0,98	0,55	0,81	0,41	0,52	0,57

FG = ferrogabbro, ap-FD = apatite-rich ferrodiorite, FD = ferrodiorite Q-diorite = quartz-diorite.

P_2O_5 contents and, in general, of most major and trace element contents plotted against differentation index (Figs. 3-5) are not compatible with a liquid line of descent obtained by a fractional crystallization process. Accordingly, mass balance calculations (not reported), fail to reproduce the composition of the apatite-rich Fe-diorites as residual liquids after fractional crystallization of any reasonable assemblage. The hypothesis of liquid immiscibility can be ruled out on the basis of trace element contents (see section below). Most likely, these rocks originated through cumulus processes from evolved liquids, probably Fe-andesites to dacites as proposed by Tribuzio et al. (2000b) on the basis of trace element mineral compositions, where apatite and Fe-Ti oxides are liquidus phases. Similar conclusions were reached by Koepke and Seidel (2004) for amphibole- and apatite-rich rocks from the Crete ophiolites. The observed chemical heterogeneity may be due to variable amounts of cumulus minerals and trapped liquid. The very high Y contents, the increase in ΣREE with increasing P_2O_5 (Tables 2-3) associated with stronger LREE depletion (La_N/Sm_N = 0.4-0.6) and HREE fractionation (Tb_N/Yb_N = 1.8-2.0) than in the other Fe-gabbroic rocks are consistent

Table 4. Fractional crystallization models

Parental magma	N-MOR basalt *		Ferrobasalt CB6		Diorite CB32	
Daughter magma	Ferrobasalt B6		Diorite CB32		Trondhjemite CB20	
Fractionating assemblage						
	a		b		c	
Pl	$0.662^{(1,2)}$		$0.470^{(1,2)}$		$0.585^{(2)}$	
Cpx	$0.226^{(3,4)}$		$0.3116^{(3,4,5)}$		$.063^{(6)}$	
Ol	$0.113^{(2,5,6)}$					
Ilm			$0.155^{(7,8,9)}$			
Mag			$0.051^{(7)}$		$0.155^{(7)}$	
Amph					$0.185^{(10,11)}$	
Ap			$0.013^{(12)}$		$0.020^{(12)}$	
Zir					$0.0015^{(12,13)}$	
Aln					$0.0002^{(7,14)}$	
F	0,224		0,386		0,409	
Sr^2	0,027		0,009		0,002	
C(liq) - ppm	CALC	OBS	CALC	OBS	CALC	OBS
La	13,6	13,2	33,2	28,7	30,4	27,5
Ce	43,7	52,2	124,1	93,3	91,7	72,1
Nd	33,0	40,8	76,5	-	-	48
Sm	10,4	14,4	21,2	24,1	15,8	13,8
Eu	2,7	4,02	6,0	5,98	4,0	2,06
Tb	-	3,25	6,9	5,71	6,2	3,09
Yb	8,2	10,3	22,0	18,4	17,4	13,4
Lu	-	1,44	3,0	2,69	2,6	2,19
Hf	9,4	10,4	29,6	23,5	13,8	15
Ta	1,0	0,55	1,1	0,88	1,9	1,12
Th	0,6	0,42	1,0	1,01	1,5	2,88
Y	80	104	208	182	138	137
Zr	450	456	1334	1063	-	-

F = fraction of residual liquid, Sr2 = sum of squares of residuals for the major element mass balance calculations. Fractionating minerals from (a) olivine gabbro (Riparbella, Tuscany: Tiepolo et al., 1997), (b) ferrobasalt (clino-pyroxene, magnetite, Table 1) and apatite ferrodiorite (plagioclase: Tiepolo et al., 1997), (c) apatite ferrodiorite (plagioclase and clinopyroxene: Tiepolo et al., 1997), ferrobasalt (magnetite, Table 1), ferrodiorite (amphibole,Table 1). Apatite composition assumed as pure stechiometric calcium phosphate. Superscripts 1-13 indicate values of partition coefficients taken from:1: Bindeman et al. (1998), 2: Dunn & Sen (1994), 3: Hart & Dunn (1994),4: La Tourrette & Burnett (1992), 5: Beattie (1994), 6: Fujimaki et al. (1984), 7: Ewart & Griffin (1994), 8: Green & Pearson (1987), 9: Nielsen et al. (1992), 10: Sisson (1991), 11: Brenan et al. (1995), 12: Fujimaki (1986), 13: Bea et al., (1994), 14: Mahood & Hildreth (1983). CALC: calculated, OBS:observed * Internal Liguride N-MOR basalt R9/5 after Rampone et al. (1998).

with the presence of cumulus apatite. The lower TiO_2, FeO and P_2O_5 contents coupled with higher SiO_2 of the Fe-diorites likely reflect lower amounts of (cumulus) apatite and oxides.

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Overlapping values of FeO*/(FeO* + MgO) ratios in Fe-gabbro-dioritic rocks and most plagiogranites contrasting with significantly lower SiO_2 content in the former group of rocks (Fig. 3), suggest that Fe-gabbro-dioritic rocks and plagiogranites represent cumulates from relatively evolved liquids and liquid compositions, respectively. This is also consistent with the plagiogranite occurrence as dikelets and veins intruding the Fe-gabbroic rocks.

Several petrogenetic models have been proposed to explain the origin of plagiogranitic rocks in the upper part of the ophiolitic sequences and in the oceanic crust, i.e. liquid immiscibility, partial melting of mafic rocks and fractional crystallization. We discuss below the possible role of these processes in the petrogenesis of the Bartolina plagiogranites.

Liquid immiscibility

The origin of silica-rich rocks by a immiscibility process mainly stem from the common lack of liquids of intermediate compositions in the oceanic and ophiolitic intrusive sequences. In the framework of the Ligurian Tethys ophiolites, immiscibility occurring in the latest stages of fractionation has been proposed by Pognante et al. (1982) to explain the small volumes of plagiogranites associated with Fe-Ti gabbros from the Lanzo-Rocciavrè massif in the Western Alps.

Experimental studies carried out at 1 atm on tholeiitic magmas from the Galapagos Spreading Center by Dixon and Rutherford (1979) have shown that after 95% crystallization of olivine + plagioclase + clinopyroxene + ilmenite, the residual liquid is a Fe-rich basalt that undergoes unmixing at ~1000 °C; the two resulting liquids have rhyolitic and Fe-Ti-P enriched compositions. In Fig. 7, major element compositions of immiscible liquids obtained experimentally by Dixon and Rutherford (1979) and observed as spherulitic globules in tholeiitic basalt groundmass (Philpotts, 1982) have been compared with the most silica- and Fe-rich rocks of the present study, i.e. trondhjemites and apatite-rich ferrodiorites. Alkalis have not been taken into account because of their high mobility during hydrothermal processes. The trondhjemites display slightly lower FeO, TiO_2, MnO and P_2O_5 and higher SiO_2 than the experimental granitic liquids. On the other hand, larger disagreement between apatite-rich Fe-diorites and the mafic liquids of Dixon and Rutherford (1979) is observed: the Bartolina rocks show higher Al_2O_3 associated with significantly lower TiO_2, FeO, MnO and P_2O_5. A similar trend is displayed by the Fe-rich liquids found in groundmass ocelli of tholeiitic basalt from several localities (Philpotts, 1982), whereas the complementary silica-rich liquids have major element composition indistinguishable from the Bartolina trondhjemites.

We have also taken into account liquid/liquid trace element partition coefficients obtained through theoretical and experimental models of conjugate immiscible liquids (Watson, 1976; Ryerson and Hess, 1978; Ellison and Hess 1989; Vicenzi et al., 1994). The behaviour of highly charged cations like Zr, Hf, Nb, Ta, Y and REE is controlled by the relative states of polymerization of the coexisting magmas so that they are preferentially partitioned into relatively depolymerized basic liquids characterized by low SiO_2. Enrichment factors for such elements in the silica-poor Fe-rich liquids with respect to the silica-rich end-members mostly range between 3-6. However, in the investigated rocks, an opposite fractionation occurs as they are variably enriched in the trondhjemites (enrichment factors = 1.5-3.0) with respect to the apatite-rich Fe-diorites, which is the expected trend generated by a fractional crystallization process.

In summary, both major and trace element distribution preclude that liquid immiscibility played a role in the origin of plagiogranitic rocks of the ophiolite sequence.

Partial melting of mafic lithologies

Amphibolite partial melting has been assumed as a genetic model for plagiogranites in the upper part of some ophiolitic sequences (Gerlach et al., 1981; Pedersen and Malpas, 1984; Flagler, 1991). Likewise, partial melting of oceanic crust altered by hydrothermal processes has been proposed to explain the origin of the high volumes of Icelandic rhyolites and the bimodal distribution of extrusive basic and acidic magmas (e.g. Sigurdsson, 1977; Gunnarsson et al., 1998). However, experiments on mafic rocks at conditions appropriate for the plagiogranite genesis, e.g. melting of MOR-type gabbroic rocks at low pressure in the presence of a H_2O-rich fluid simulating infiltration of seawater-derived fluids into hot oceanic crust (Beard and Lofgren, 1991; Koepke et al., 2004), yield peraluminous silica-rich liquids characterized by low FeO and TiO_2 contents, unlike the rocks of the present study.

Fractional crystallization

The previous discussion has shown that liquid immiscibility and partial melting of mafic rocks do not seem to be viable explanations for the origin of the wide spectrum of plagiogranite compositions in the Bartolina ophiolite. In contrast, the paucity of silicic rocks, their close association with intermediate Fe-Ti-rich gabbroic rocks and the absence of compositional gaps support an origin as a late product of MORB differentiation via crystal-liquid fractionation.

Overall depletion in TiO_2, Al_2O_3, P_2O_5 associated with SiO_2 and incompatible trace element (REE, Zr, Hf, Th, Ta) enrichment in the plagiogranites suggest that fractionation of the Fe-gabbroic rocks assemblage, i.e. plagioclase + clinopyroxene + Fe-Ti oxides + apatite ± amphibole may have been the dominant process. In particular, the role of plagioclase fractionation is clearly displayed by the increasing Eu depletion from the gabbroic rocks to plagiogranites.

Petrogenetic hypotheses of simple closed-system fractional crystallization for the plagiogranitic rocks have been tested adopting the conventional mass-balance procedures for major elements and Rayleigh fractionation modelling for trace elements (Table 4). Trace element modelling is critically dependent upon the choice of partition coefficients. To take into account the effects of the compositional dependence of crystal/liquid partitioning, we have tried to use as far as possible partition coefficients determined for the liquid comparable to those involved in the modelling. It is noteworthy, however, that partition coefficients for zircon and allanite are available only for peraluminous and high-silica granitic liquids, respectively (see references in Table 4).

The results show that liquids corresponding to diorites may be obtained through removal of plagioclase + clinopyroxene + magnetite + apatite starting from a ferrobasalt parental melt. A good fit is observed for most major and trace element for about 50% crystallization of the starting liquid (Table 4). Calculations also show that trondhjemite major element composition is compatible with fractionation of plagioclase + clinopyroxene + amphibole + Fe-Ti oxides + apatite from a liquid of dioritic composition. Trace element concentrations calculated on the basis of this model show, however, unrealistically high enrichment in REE, Hf, Y (e.g. La = 49.8 ppm, Hf = 47 ppm, Y = 208 ppm). At this stage the abundances of such elements is strongly controlled by the removal of accessory phases other than apatite, i.e. zircon and, possibly, allanite, that have been observed as crystallizing phases in the diorites. Hence, the addition of small amounts of these accessory minerals to the fractionating assemblage, the resulting liquids become more akin to the compositions observed for the trondhjemitic rocks (Table 4). In addition, the observed discrepancies may be partially due to the lack of appropriate parti-

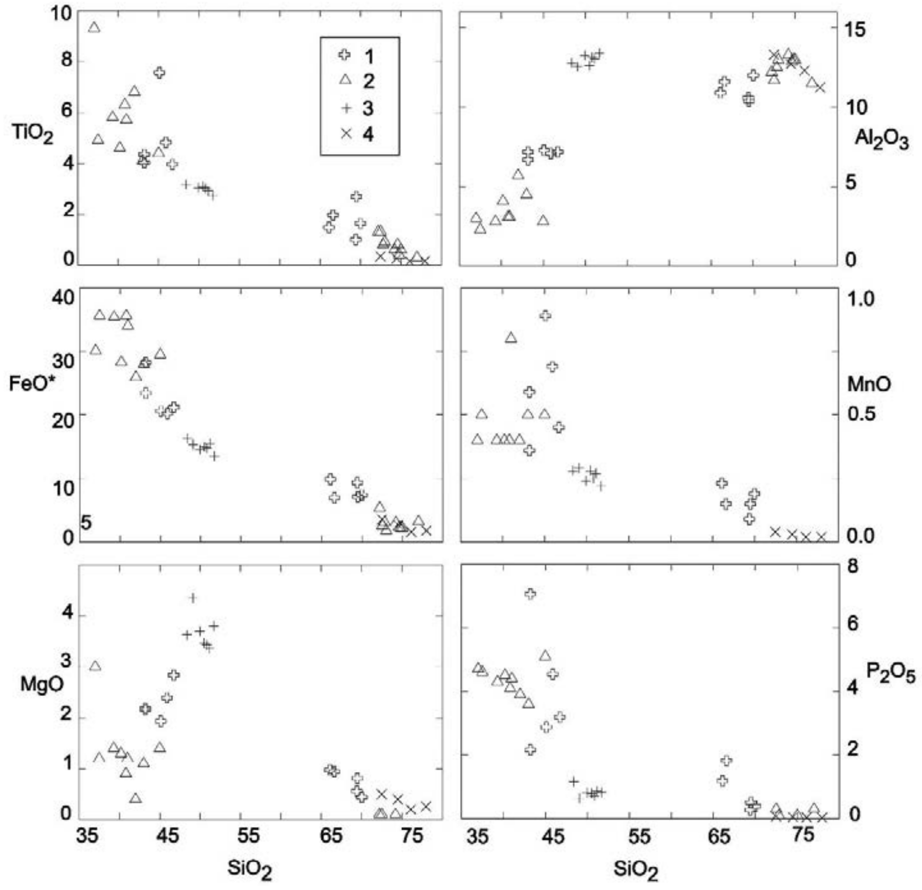

Fig. 7 - TiO2, Al2O3, MnO, FeO*, MgO, P2O5 vs. SiO2 diagrams for (1) experimental immiscible liquids (Dixon and Rutherford, 1979), (2) natural immiscible liquids in tholeiitic basalts (Philpotts, 1982), (3) apatite-rich Fe-diorites and (4) trondhjemites of the present study.

tion coefficients for REE, Hf, Ta, Th in allanite and Tb, Y, Th in zircon crystallizing from liquids with intermediate silica contents.

Fractional crystallization models do not yield reasonable results for the leucodiorites. Actually, the leucodiorite textures, the high modal proportions of plagioclase and the high Al_2O_3 contents suggest that these rocks are dominated by plagioclase segregation and expulsion of trapped melt and do not represent liquids intermediate between diorites and trondhjemites as expected from their silica contents.

Borsi et al. (1996) proposed that the slight LREE enrichment in the plagiogranites from the Ligurian ophiolites was related to T-MORB parental melts. However, T-MORB magmas are incompatible with the LREE depletion observed for the Bartolina ferrobasalts, whose compositions may be satisfactorily reproduced by differentiation of a primitive N-MORB type magma (Table 4). Furthermore, our data and calculations indicate that La_N/Sm_N may slightly increase in the latest stage of the crystallization as a result of the combined removal of amphibole (MREE-enriched) and LREE-depleted accessory minerals like apatite and zircon.

The accessory amounts of interstitial biotite associated with apatite and Fe-Ti oxides (Fig. 2b) in the Fe-dioritic rocks suggest a likely igneous origin from an evolved interstitial melt. This requires a relative K_2O enrichment in the melt, in contrast with the exceedingly low K_2O contents of the rocks corresponding to the most evolved liquids, i.e. the trondhjemites (Table 2). However, the extreme K depletion coupled with very high Na/K ratios is probably the result of seafloor alteration processes (see also Serri, 1980), similarly to many oceanic plagiogranitic rocks worldwide. The rarity of primary biotite in fossil or modern oceanic crust (see Pedersen and Malpas, 1984, Dick et al., 1991; Niu et al., 2002) could be thus related to its unlikely preservation in rocks extensively affected by ocean-floor hydrothermal processes.

CONCLUSIONS

Field relationships, petrographic and geochemical data for the composite Bartolina ophiolitic body are consistent with fractional crystallization of small batch(es) of ferrobasaltic magma evolving towards silica-rich liquids of trondhjemitic composition. Geochemical data and modelling indicate that the ferrobasaltic liquid was derived from a N-MORB parental magma through olivine + clinopyroxene + plagioclase fractionation. At this stage, the cumulates (not observed in the dismembered Bartolina ophiolite) should be represented by troctolites and olivine gabbros. Further removal of clinopyroxene + plagioclase (i.e. the main constituents of the gabbros) yielded differentiated liquids (Fe-andesites?) that segregated from the gabbroic crystal-mush in pods and lenses. Here, clinopyroxene + plagioclase + Fe-Ti oxides (magnetite) + Fe-rich olivine + apatite crystallization formed the most differentiated cumulates, i.e. the Fe-gabbroic rocks. The achievement of Fe-Ti oxide (+ apatite) saturation promoted silica enrichment in the liquid interrupting the Fe-Ti-P enrichment trend. These residual liquids were eventually segregated and injected into fractures of solidified cumulates forming diorite veins in the gabbros or infiltrated the Fe-gabbroic crystal-mush as revealed by the sutured margins between diorites and apatite-rich Fe-diorites. Locally, further fractionation of plagioclase + clinopyroxene + amphibole + Fe-Ti oxides (+ apatite + zircon + allanite) led to silica-rich and slightly LREE-enriched liquids that were locally squeezed out yielding the trondhjemite dikelets. Textural observations indicate that melt migration chiefly occurred under a brittle regime.

The general absence of primary association between olivine gabbros and Fe-gabbroic rocks in the Northern Apennine ophiolites suggests that evolved (ferro)basaltic magmas became isolated within the oceanic crust. We propose that the Bartolina intrusive body originated from a ferrobasalt magma reservoir derived from compaction of an underlying troctolite-olivine gabbro crystal-mush coupled with vertical melt migration of the differentiated liquid, similarly to what was envisaged for the ephemeral magma chambers at slow-spreading ridges (Thy and Dilek, 2000), where Fe-Ti-rich rocks are invariably found near the top of gabbroic intrusions (Robinson et al., 2000). The occurrence of ferrobasaltic dikes into already solidified gabbroic rocks, should represent new injections of these evolved basic liquids.

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