GEOCHEMISTRY, PETROGENESIS AND TECTONO-MAGMATIC SIGNIFICANCE OF VOLCANIC AND SUBVOLCANIC ROCKS FROM THE KOZIAKAS MÉLANGE (WESTERN THESSALY, GREECE)

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ABSTRACT

The Koziakas ophiolites have previously been interpreted as a Jurassic complex generated at a mid-ocean spreading ridge, and have been subdivided into a lower volcanic unit and an upper ultramafic unit.

The data presented in this paper demonstrate that the lower volcanic unit is, in fact, a tectonic mélange consisting of a polygenetic, multiple packet of stacked thrust-bound slices, representing distinct volcanic sequences and radiolarian chert successions.

Volcanic sequences include pillowed and massive lava varieties, which are frequently crosscut by dikes of various nature, including boninitic dikes. Both lava varieties and dikes are predominantly aphyric, though a few samples display slightly porphyritic textures, where phenocryst assemblages include olivine and/or plagioclase in pillow and massive lavas, clinopyroxene, plagioclase and sanidine in trachytic dikes, clinopyroxene in boninitic dikes.

Geochemical data point out the presence of three compositionally distinct groups of lavas: (1) transitional to alkaline basalts, trachyandesites, trachytes; (2) tholeiitic basalts; (3) very low-Ti (boninitic) basaltic andesites and andesites.

The transitional to alkaline rocks displays high Nb/Y ratios and enriched incompatible element characteristics, similar to those of many intraplate oceanic island basalts (OIB).

The tholeiitic group display lower abundances of incompatible elements and includes rocks resembling normal mid-ocean ridge basalts (N-MORB) with light REE (LREE) depletion (La_N/Sm_N = 0.29) and very low Ce/Y, Ta/Hf, Th/Yb ratios, as well as rocks resembling enriched mid-ocean ridge basalts (E-MORB) showing moderate LREE enrichment (La_N/Sm_N = 1.26-1.52) and Ce/Y, Ta/Hf, Th/Yb ratios higher than those of N-MORBs.

The very low-Ti group displays greatly depleted incompatible element abundances and the U-shaped REE patterns typical of boninites generated in supra-subduction settings.

The Koziakas Mélange appears to have formed due to tectonic dismemberment and accretion of material generated in an oceanic environment (MORBs and OIB-type rocks), possibly in an intra-oceanic forearc setting. After their accretion these rocks were affected by widespread boninitic dikes generated by partial melting of depleted peridotites in the fore-arc setting.

The record of different lava types in the Koziakas Mélange is in accordance with the general geological evolution of the Neo-Tethyan Pindos oceanic basin, from the Permo-Triassic rifting to the Middle-Late Jurassic intra-oceanic convergence phase.

INTRODUCTION

During the past decades, great efforts have been made to define the geochemical and petrological features of Greek ophiolites (Bébien et al., 1980 and references therein). Nonetheless, a satisfactory interpretation of the geochemical characteristics and tectono-magmatic significance of several Greek ophiolitic complexes has not been reached yet: for instance, the Koziakas ophiolites, which are considered to be part of the Jurassic Subpelagonian Zone ophiolites of the Hellenides (Capedri et al., 1985), and are located between the two major ophiolitic complexes of the Pindos (to the north) and Othrys (to the south) mountains (Fig. 1).

Two major geological differences between the Koziakas and Pindos-Othrys ophiolitic sequences can be observed. (1) Both the Pindos and Othrys Subpelagonian (or Maliac, Ferrière, 1982) ophiolites are overthrust onto the Eocene flysch of the Pindos Zone (Smith et al., 1979; Ferrière, 1985; Jones and Robertson, 1991; Robertson, 1994), whereas the Koziakas ophiolites (Figs. 1, 2) are thrust over the Middle Triassic - Upper Cretaceous sedimentary formations of the Western Thessaly Unit (Beotian Zone?) (Papanikolaou and Lekkas, 1979; Jaeger, 1980). (2) In both Pindos and Othrys ophiolites the magmatic sequences are characterized by well-developed intrusive and extrusive successions, and by the occurrence of amphibolitic soles and sub-ophiolitic mélange at their bases (Jones and Robertson, 1991; Rassios and Smith, 2000). By contrast, according to Capedri et al. (1985), and in contradiction with the results of this study, in the Koziakas ophiolites intrusive rocks are very rare, and metamorphic soles and sub-ophiolitic mélange are lacking.

Previous works (Capedri et al., 1985) have attributed the formation of the Koziakas ophiolites to a mid-ocean ridge setting, but have also highlighted their great compositional variability, including normal, transitional, and enriched mid-ocean ridge basalts, as well as very low-Ti basaltic rocks (boninites).

In addition, recent biostratigraphical datings have demonstrated that radiolarian cherts associated with the Koziakas basaltic rocks (Fig. 2) display two different ages, that is: Middle Jurassic and Late Triassic (Chiari et al., in progress).

All these facts point out many doubts about the attribution of the Koziakas basaltic rocks to an ophiolitic sequence. They seem rather more similar to many Hellenides accretionary complexes, which consist of heterogeneous tectono-sedimentary associations of Triassic oceanic intraplate basalts (OIB) (Pe-Piper, 1998 and references therein), Middle Jurassic mid-ocean-ridge (MOR) and supra-subduction-zone (SSZ) ophiolitic sequences (Capedri et al., 1996, Jones and Robertson, 1991), Permian-Jurassic marginal and platform-related sediments, and Jurassic-Tertiary trench-type sediments (Robertson, 1994; Clift and Robertson, 1989;

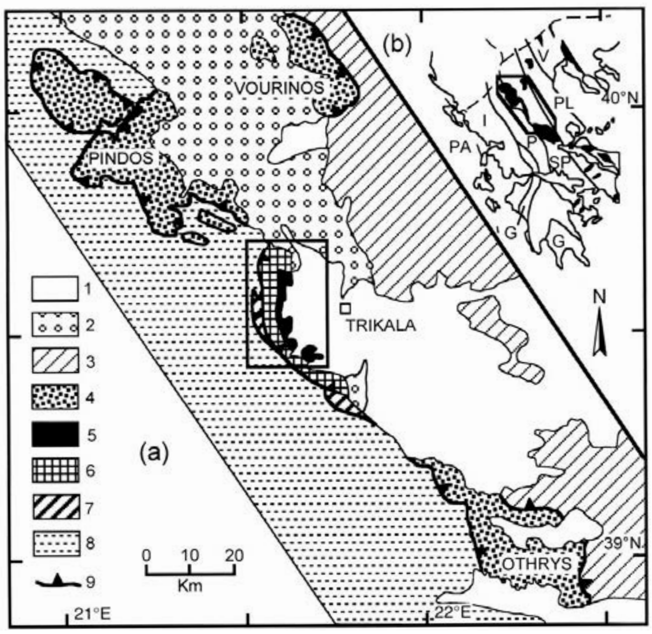

Fig. 1 - Geological sketch map of the Pindos-Koziakas-Othris area (**a**) and its location with respect to the main tectonic zones of Greece (**b**). Legend in (**a**): 1- Quaternary sedimentary rocks; 2- Oligocene-Early Miocene molasse deposits (Meso-Hellenic Trough); 3- Pelagonian Zone; 4- Subpelagonian (Maliac) ophiolitic complexes; 5- Koziakas series volcanic and ultramafic rocks; 6- Koziakas series sedimentary units; 7- Thymiama series (Beotian Zone); 8- Pindos Zone (mainly Tertiary Pindos Flysch formation); 9- Thrusts. Abbreviations of tectonic zones in (**b**): PA- Pre-Apulian; I- Ionian; G- Gavrovo; P- Pindos; SP- Subpelagonian (Maliac); PL- Pelagonian; V- Vardar.

Degnan and Robertson, 1998).

For all these reasons, the aim of this work is to present new geochemical and petrological data on volcanic and subvolcanic rocks from the Koziakas Massif in order to assess their tectono-magmatic environment in which they formed.

GEOLOGICAL SETTING

The Koziakas Massif (Figs.1, 2) is located at the western boundary of the Thessaly plain and extends, with a general N-S trend, from the Vitoumas village (to the north) to the Mouzaki village (to the south). This Massif is composed of a westward stack of thrust sheets tectonically implicated between the Subpelagonian ophiolites, at the top, and the Pindos flysch, at the bottom. Thrust sheets of the Koziakas Massif include the Cretaceous - Tertiary Thymiama series and the Triassic - Jurassic Koziakas series (Fig. 2).

The Koziakas Massif has been considered as a separate geotectonic unit in the Hellenic Alpine system due to a number of specific tectonic and lithostratigraphical characteristics (Magganas et al., 1997).

Aubouin (1959) and Scandone and Radoicic (1974) concluded that the Koziakas Massif should correspond to the Ultrapindic Zone, which was palaeogeographically located between the Subpelagonian (Maliac), or Parnassus, and the Pindos Zones. However, Papanikolaou and Sideris (1979), Papanikolaou and Lekkas (1979), and Lekkas (1988) have suggested that the Thymiama and Koziakas series represent a continuous sequence defined as "Western Thessaly Unit" that can be regarded as a paleographical heteropic equivalent to some Pindos Zone units. Alternatively, the "Western Thessaly Unit" has been ascribed to the Beotian Zone by several authors (Jaeger, 1980; Migiros et al., 1989; Karfakis et al., 1993). These authors also conclude that the Koziakas ophiolites, included at the top of the Koziakas series, represent Early Cretaceous olistolitic bodies.

However, Triassic andesitic lavas, radiolarites and cherty limestones occur at the base of the Koziakas series (Ferrière, 1974; 1982; Magganas et al., 1997). Furthermore, the presence of a clastic formation, rich in ophiolitic detritus locally overlying the sedimentary sequence of the Koziakas series and underlying the Koziakas ophiolites, support the hypothesis that the Koziakas series is palaeogeographically related to the Subpelagonian (Maliac) Zone (Ferrière, 1974; 1982; Celet et al., 1978). According to this hypothesis, the Koziakas ophiolites may represent dismembered ophiolitic slices accreted onto the adjacent continental margins, and can be correlated with the Pindos and Othrys Subpelagonian ophiolitic complexes of the Hellenides.

On the other hand, the presence of Tithonian-Berriasian flyschioidal formation, analogues to the Beotian flysch, which forms the base of the Thymiama series may suggest that both the Thymiama and Koziakas series pertain to the Beotian Zone (Aubouin and Bonneau, 1977; Celet et al., 1978; Jaeger and Chotin, 1978; Jaeger, 1980).

Based on our new geological field-mapping, the tectono-stratigraphic setting of the Koziakas series (Fig. 2) can be regarded as a stack of thrust sheets tectonically emplaced westward onto the Thymiama series (Beotian Zone), which in turn overthrusts the Eocene Pindos Flysch (Pindos Zone). The dismembered thrust units of the Koziakas series are unconformably overlain by the Oligocene-Miocene molasse deposits of the Mesohellenic trough (Savoyat and Lalechos, 1972), which in turn are covered by the Quaternary deposits of the Thessaly plain (Fig. 2).

The Koziakas series can be subdivided (Fig. 2) into the following main tectonic units (from bottom to top):

Triassic to Jurassic cherty limestones overlain by re-deposited and oolitic limestones with chert intercalations;

radiolarian red cherts bearing calcarenite intercalations, characterized in their upper part by a Middle-Late Jurassic manganiferous red chert succession (Skarpelis et al., 1992; Chiari et al., in progress). Scattered outcrops of ophiolite-derived turbidites and/or olistostromes are locally found in this sedimentary sequence.

tectonic mélange (Mélange Unit) mainly consisting of sequences of pillow and massive lava flows. Rare dolerite sheeted dike complexes are locally found at the base of the extrusive sequences (e.g., Mouzaki area). Various very low-Ti "boninitic" dikes crosscut both pillow and massive lava flows, as well as the sheeted dike complex. Bedded radiolarian chert successions are locally included in the Mélange Unit, and show both Triassic and Jurassic ages (Chiari et al., in progress). This unit has been previously interpreted (Capedri et al., 1985) as a mid-ocean ridge sequence, with variable geochemical signatures.

Ultramafic mantle tectonites (Ophiolitic Unit), tectonically overlying the Mélange Unit, are composed of serpentinized harzburgite slivers and sheared serpentinite. In places, the harzburgite masses also contain dunite bodies and are locally intruded by plagiogranite and boninite dikes (Hatzipanagiotou et al., 1995).

The present-day structure of the Koziakas Massif is characterized by the occurrence of recumbent folds, NNW axial direction, and 50-70° dipping to the east, and NW and NE trending faults.

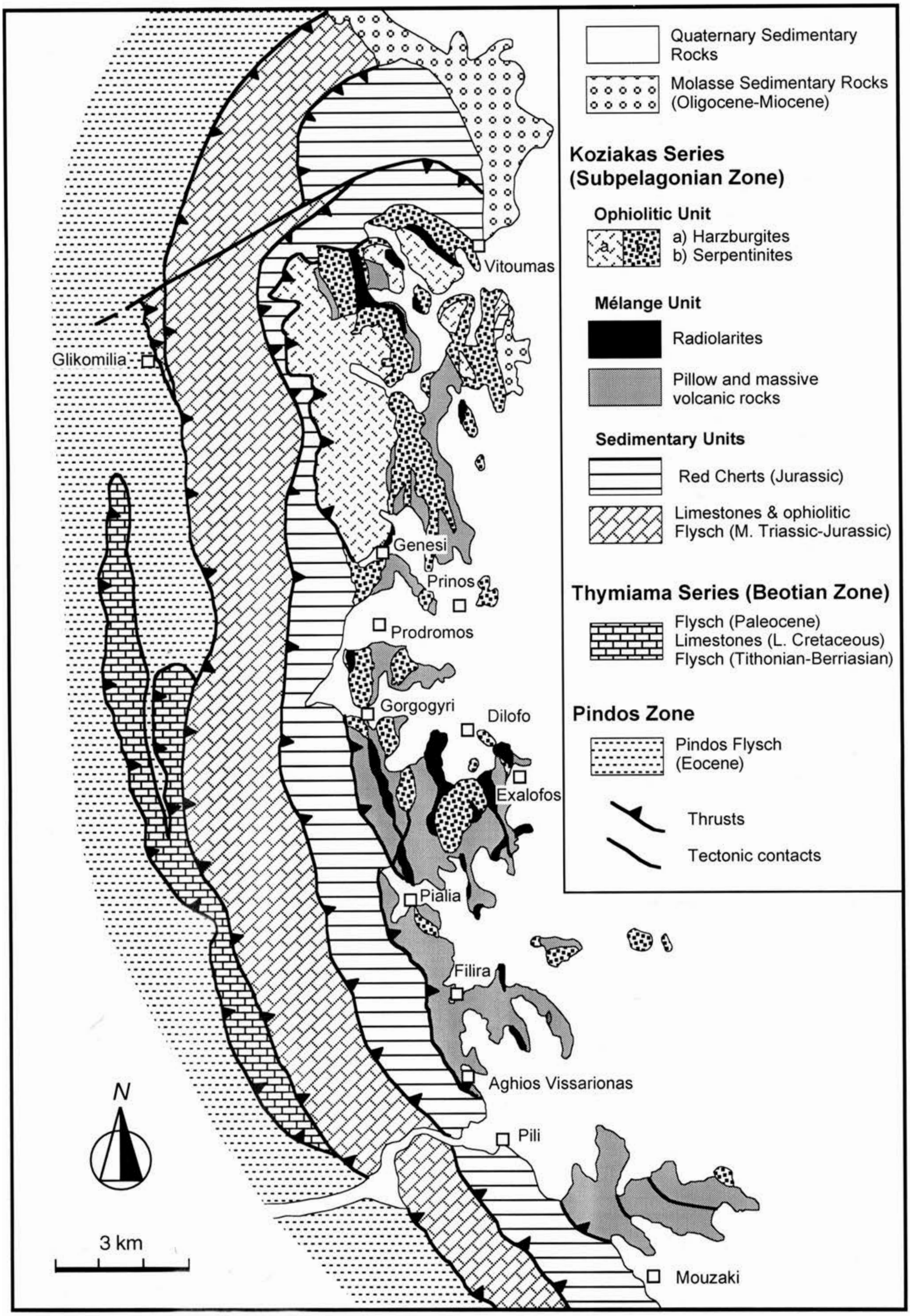

Fig. 2 - Simplified geological map of the Koziakas area. Modified after Savoyat and Lalechos (1972) and Karfakis et al. (1993).

During the post-Upper Jurassic compressive tectonic phase which affected the internal Hellenides, the mélange and the ophiolitic units tectonically overthrusted the Subpelagonian (Maliac) continental margin represented by the sedimentary units of the Koziakas series. During the post-Upper Eocene compressive tectonic phase, all the previous units were refolded and overthrust southwestward over the Beotian-derived Thymiama series and then, together with this, over the Eocene Pindos flysch.

SAMPLING AND METHODS

Sampling was carried out on the volcanic and subvolcanic rocks from the Mélange Unit in the central and southern areas of the Koziakas Massif (Fig. 2). Volcanic rocks occur as pillow and massive lava flows, both frequently cross-cut by dikes of variable nature.

In particular, the sampled volcanic and sub-volcanic rocks include pillowed (samples Nos. 8, 9, 22-26, 31, 32, 37, 41) and massive flows (samples Nos. 6, 11, 13, 16, 36, 38), as well as dikes (samples Nos. 5, 27-30). Dike samples Nos. 27 to 30 are included in pillow sequences corresponding to samples Nos. 22 to 26 and 31, 32, whereas dike No. 5 is intruded in the massive lavas of sample No. 6.

The twenty-two collected samples of volcanic and subvolcanic rocks (Table 1) were analyzed for major and some trace elements by X-ray fluorescence (XRF) on pressed-powder pellets, using a Philips PW1400 automated X-ray spectrometer (Department of Earth Sciences, Ferrara University). The matrix correction methods proposed by Franzini et al. (1972) were applied. Accuracy and detection limits were determined using results from International Standards. Accuracy is lower than 2% and 5% for major and trace element determinations, respectively, whereas the detection limits for trace elements range from 1 to 2 ppm. Trace element analyses were repeated using a Philips PW1470 automated X-ray spectrometer (University of Modena, Italy), giving a precision generally less than 5%.

Volatiles were determined as loss on ignition (L.O.I.) at 1000°C, whereas CO_2 was determined by simple volumetric technique (Jackson, 1958). This technique was calibrated using standard amounts of reagent grade $CaCO_3$, and checked by analyzing 20 reference samples with different CO_2 contents. The mean relative percentage error obtained was 2.8%.

17 representative samples (Table 1) were chosen for additional trace-element analyses, including rare earth elements (REE), Sc, Nb, Hf, Ta, Th, and U, using inductively coupled plasma-mass spectrometry (ICP-MS). The precision and accuracy of the data was evaluated using results for international standard rocks, duplicate runs of several samples, and the blind standards included in the sample set. Accuracy varies from 1 to 8%, whereas detection limits (in ppm) are: Sc, Y = 0.29; Nb, Hf, Ta = 0.02; REE < 0.11; Th, U = 0.011. ICP-MS analyses were performed at the Department of Earth Sciences, Ferrara University, using a VG Elemental Plasma Quad PQ2 Plus spectrometer.

PETROGRAPHY

Most of the rocks studied in this paper are affected by severe alteration, which resulted in the replacement of primary minerals, though primary igneous textures are well preserved.

Olivine is never preserved, and is always replaced by iddingsite. Fresh plagioclase (An ca 57%) is very rare, since it is usually replaced by albite and carbonate assemblages. Clinopyroxene is normally pseudomorphosed either by chlorite or actinolitic amphibole; however, very few samples display fresh clinopyroxene relics as both phenocrysts and groundmass minerals. The groundmass secondary phases mainly consist of albite, chlorite, zeolites, and clay minerals, as well as rare silica phases in the more differentiated products. Many volcanic rocks exhibit variolitic textures, with varioles filled by calcite and, subordinately, by chlorite and zeolite. In addition, small calcite veins are frequent.

Regardless of the secondary mineralogical transformation, the following petrographic description of the various rock-types of the Koziakas Mélange Unit has been made on the bases of the primary igneous phases.

Pillow lavas are predominantly aphyric with intersertal texture. However, sample No. 24 is characterized by a small amount (< 10%) of olivine phenocrysts, whereas sample Nos. 8 and 31 display ophitic and sub-ophitic to hyalopilitic textures, respectively. Groundmass mineral assemblages include olivine, clinopyroxene, plagioclase and opaque phases.

Massive lavas range in textures from aphyric to slightly porphyritic (porphyric index < 10). In the latter varieties, phenocrysts are exclusively represented by plagioclase. Groundmasses show a number of textures, including sub-ophitic, intersertal, and trachytic varieties (e.g., sample No. 16). They also show a wide range of variation in crystal sizes (from cryptocrystalline to doleritic). Groundmass mineral assemblages include plagioclase, clinopyroxene, and opaque minerals, as well as sanidine in sample No. 11.

Dikes are characterized by both aphyric and slightly porphyritic (P.I. < 10) textures. Aphyric basalts are characterized by coarse-grained doleritic textures with euhedral plagioclase and sub-euhedral clinopyroxene. In porphyritic mafic varieties, phenocrysts are exclusively represented by olivine, settled in microcrystalline to doleritic groundmasses characterized by sub-ophitic to intersertal textures. Groundmass mineral assemblages include olivine, plagioclase, clinopyroxene and Fe-Ti oxides. In the more evolved sample No. 5, the phenocryst assemblage includes plagioclase, sanidine and clinopyroxene settled in an intergranular, microcrystalline groundmass in which plagioclase, clinopyroxene, quartz, and opaques can be recognized.

Boninitic dikes are predominantly aphyric to moderately porphyritic (P.I. < 10), with clinopyroxene phenocrysts set in intergranular to sub-ophitic microcrystalline groundmasses. Only sample No. 30 displays a coarse-grained texture (microdiorite). The groundmasses consist dominantly of plagioclase and clinopyroxene assemblages, though Fe-Ti oxides and Cr-spinel can be recognized in some samples. Skeletic clinopyroxene is observed in the groundmass of sample No. 38.

GEOCHEMISTRY

Although most of the studied rocks from the Koziakas Mélange Unit are rather altered, incompatible elements (such as, Ti, P, Nb, Zr, Y, Hf, Th, Ta, REE), as well as transition metals (particularly, Ni, Co, Cr, V), can be used effectively to describe the primary chemical features of these rocks (Beccaluva et al., 1979; Pearce and Norry, 1979). Many samples are characterized by diffuse secondary calcite; consequently, CaO content has been recalculated without the CaO contained in calcite (Table 1).

The analyzed rocks show a wide range of geochemical characteristics (Table 1), however, three main geochemical groups can be recognized: 1) transitional to alkaline rocks; 2) tholeiitic basalts; and 3) boninitic basalts and basaltic andesites. The first two groups include, in turn, geochemically distinct sub-groups.

Transitional to alkaline rocks

This group includes both pillowed and massive lavas from various localities of the Koziakas Massif, as well as one dike sample (No. 5) from the Mouzaki area, and is mainly represented by basalts and basaltic andesites, and subordinate trachyandesites and trachytes (Table 1). The latter have silica ranging from 51.19 wt% to 62.49 wt%, and total alkali content from 5.81 wt% to 8.76 wt% (Table 1). The transitional to alkaline character of these rocks is evidenced in the Ti/Y vs. Nb/Y diagram of Fig. 3, as well as in the Zr/Ti vs. Nb/Y diagram of Fig. 4.

The incompatible element abundance of mafic volcanic rocks (Figs. 5a, c) is characterized by regularly decreasing patterns, from Rb to Yb, which are comparable to those of typical within-plate tholeiites, as also confirmed by the Nbx2-Zr/4-Y diagram (Fig. 6). These rocks are characterized by marked light REE (LREE) enrichment with respect to heavy REE (HREE) (Figs. 5b, d). La_N/Yb_N ratios range from 4.19 to 13.49.

The trachyandesitic and trachytic samples are characterized by strong Sr, P, Ti (Fig. 5e), and Eu (Fig. 5f) negative anomalies, implying that these rocks underwent conspicuous clinopyroxene, plagioclase, Ti-rich phases, and apatite fractionation prior to eruption, whereas early olivine and Cr-spinel fractionation is suggested by their low Ni (29-34 ppm) and Cr (6 ppm) contents, with Mg# ranging from 33.2 to 36.2 (Table 1).

Massive basalts and basaltic andesites curiously display slightly different geochemical features with respect to their pillowed analogues. The former display higher FeO_t, V, and Y enrichment coupled with lower P_2O_5 enrichment and decreasing Mg#, in comparison with the latter varieties (Fig. 7). Massive lavas are characterized by fairly constant CaO/Al_2O_3 ratios, whereas pillow lavas display decreasing CaO/Al_2O_3 ratios with decreasing Mg# (Fig. 7). This may reflect the early crystallization of plagioclase and clinopyroxene in the massive types, and the predominance of clinopyroxene over plagioclase crystallization in the pillow lavas. In addition, massive lavas are characterized by moderate Nb positive anomalies (Fig. 5a), which are not observed in pillow lavas.

The geochemical data from transitional to alkaline rocks indicate a transitional to alkaline affinity, and imply derivation from a mantle source distinct from the depleted MORB mantle.

Tholeiitic basalts

These include one pillow lava from the Exalofos area (sample No. 37), as well as one pillow lava (sample No. 24) and various dikes (samples Nos. 22, 23, 25, 26) from the Filira area.

All these rocks are basaltic in composition, with silica ranging from 45.43 wt% to 47.30 wt%. Their sub-alkaline affinity is exemplified in the Ti/Y vs. Nb/Y and Zr/Ti vs. Nb/Y plots of Figs. 3, 4. These basalts represent two geochemically distinct groups with respect to their incompatible and rare earth element abundance.

Sample No. 37 show low abundance in low field strength elements (LFSE) relative to typical N-MORB (Sun and McDonough, 1989), with marked U and Nb depletion, and slightly decreasing high field strength element (HFSE) patterns (Fig. 5g). REE abundance varies from 9 to 30 times that of chondrite, and displays a flat profile (Fig. 5h) with LREE patterns, typical of N-MORB (Sun and McDonough, 1989). This interpretation is supported by the Nbx2-Zr/4-Y discrimination diagram of Fig. 6.

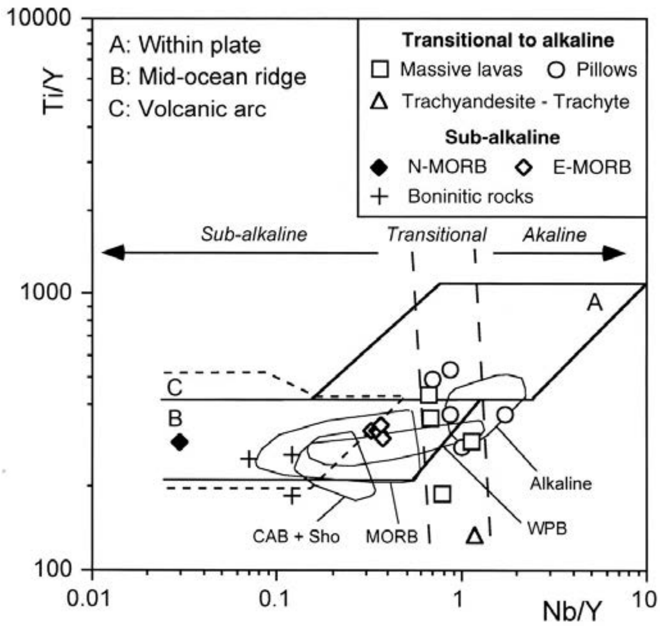

Fig. 3 - Ti/Y vs. Nb/Y diagram for volcanic and subvolcanic rocks from the Koziakas Mélange Unit. Modified after Pearce (1982). Variations of elemental ratios of alkaline, calc-alkaline (CAB), shoshonitic (Sho), mid-ocean ridge (MORB), and undistinguished within-plate (WPB) Triassic and Jurassic mafic volcanic rocks from various localities of the Hellenides are reported for comparison. Data sources: Capedri et al. (1985), Capedri et al. (1997) and Pe-Piper (1998).

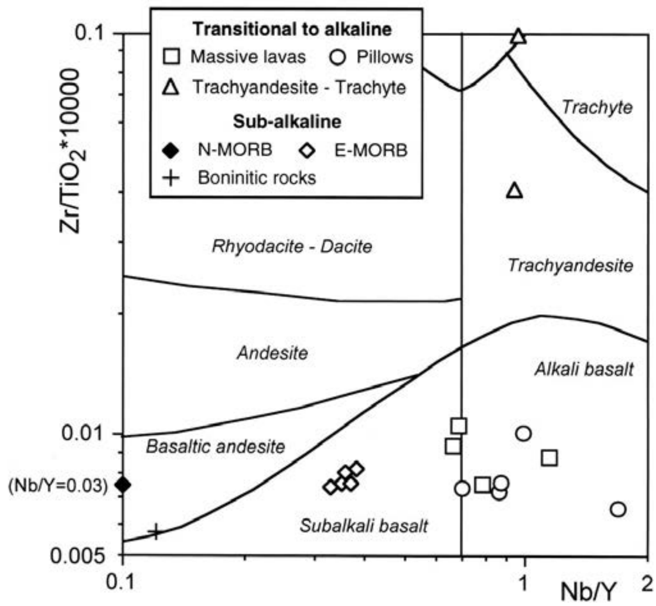

Fig. 4 - Zr/TiO_2*10000 vs. Nb/Y discrimination diagram for volcanic and subvolcanic rocks from the Koziakas Mélange Unit. Modified after Winchester and Floyd (1977).

Table 1 - Bulk rock major and trace element analyses of volcanic and subvolcanic rocks from the Koziakas Mélange Unit.

	Transitional to alkaline rocks										
Locality	Pialia	Mouzaki	Mouzaki	Mouzaki	Mouzaki	Filira	Prinos	Mouzaki	Filira	Mouzaki	Mouzaki
Sample	36	6	16	13	9	32	41	8	31	11	5
Rock	Bas	Bas	Bas	Bas	Bas	Bas And	Bas And	Bas	Bas And	Tra And	Tra
Type	MLF	MLF	MLF	MLF	Pillow	Pillow	Pillow	Pillow	Pillow	MLF	Dyke
XRF Analyses:											
SiO_2	44.53	44.50	44.73	45.56	48.80	37.72	37.17	34.23	38.05	51.19	62.49
TiO_2	0.81	1.72	2.18	2.24	1.13	0.94	1.41	1.54	1.34	1.68	0.82
Al_2O_3	16.92	16.21	14.23	15.92	15.97	14.06	13.59	12.43	14.50	15.82	14.08
Fe_2O_3	1.03	1.30	1.44	1.79	1.07	0.63	0.79	0.83	0.71	1.40	1.04
FeO	6.84	8.69	9.62	11.90	7.13	4.21	5.25	5.53	4.76	9.35	6.95
MnO	0.13	0.21	0.23	0.22	0.12	0.14	0.25	0.15	0.07	0.16	0.11
MgO	7.65	7.36	7.41	6.30	9.32	5.30	4.03	3.68	1.65	2.61	2.22
CaO	11.30	11.49	10.87	8.79	8.30	20.30	18.57	23.25	19.08	6.72	1.08
Na_2O	4.03	2.26	3.06	3.62	3.42	2.32	4.27	3.05	1.75	4.76	1.43
K_2O	0.15	0.67	0.18	0.35	1.25	0.12	0.75	0.84	3.53	1.05	7.33
P_2O_5	0.29	0.31	0.40	0.47	0.29	0.37	0.37	0.44	0.54	0.60	0.10
L.O.I.	6.32	5.28	5.65	2.84	3.20	13.88	13.54	14.04	14.01	4.64	2.35
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
CO_2	2.15	0.66	1.93	n.d.	n.d.	10.19	11.82	11.70	12.16	n.d.	n.d.
CaO*	8.56	10.65	8.41			7.33	3.53	8.35	3.60		
Mg#	66.6	60.1	57.9	48.6	70.0	69.2	57.8	54.3	38.2	33.2	36.2
Pb	n.d.	6	4	6	2	8	4	9	8	6	3
Zn	39	79	84	106	67	17	33	73	91	169	149
Ni	92	52	56	35	98	93	97	221	70	29	34
Co	28	47	45	47	34	22	27	36	17	10	8
Cr	385	130	100	28	270	274	263	420	151	6	6
V	216	251	280	307	278	200	159	124	170	66	55
Rb	2	8	n.d.	6	17	n.d.	12	16	60	12	52
Sr	315	305	465	289	466	297	384	413	188	312	53
Ba	70	153	52	138	167	68	365	121	302	262	413
Th	n.d.	3	n.d.	3	n.d.	3	2	n.d.	6	9	11
Nb	15	40	22	20	19	18	17	18	52	67	91
Zr	79	139	224	245	118	96	153	166	128	725	990
Y	18	59	33	39	25	22	22	27	32	78	91
La	12	25	18	20	15	17	12	15	36	63	63
Ce	22	60	49	34	33	36	34	30	63	143	126
Nd	15	37	21	26	14	15	20	18	27	66	70
ICP-MS Analyses:											
Sc	53.4	97.9		45.3	40.3	37.5	28.7	31.2	26.4	20.8	7.5
Y	21.2	70.8		36.9	31.0	22.7	22.9	32.6	31.0	68.8	70.7
Nb	20.9	46.2		26.4	24.9	18.7	19.5	19.0	53.8	73.2	89.5
La	12.3	28.2		17.0	14.9	14.6	16.7	12.4	38.9	58.7	57.2
Ce	24.0	59.8		37.6	29.6	29.7	34.1	25.5	67.1	119.3	141.2
Pr	3.12	8.12		5.15	3.61	3.57	4.27	3.42	7.19	15.1	14.8
Nd	13.0	35.2		22.4	15.2	15.9	19.7	14.7	27.6	61.4	61.4
Sm	2.86	8.01		5.47	3.35	3.61	4.49	3.49	4.97	13.2	12.9
Eu	0.93	2.80		1.81	1.13	1.19	1.65	1.14	1.55	3.60	2.49
Gd	3.03	8.14		5.92	3.60	3.62	4.63	3.59	5.56	13.8	13.0
Tb	0.50	1.36		0.99	0.61	0.53	0.67	0.63	0.77	2.27	2.18
Dy	3.03	8.31		5.80	3.60	3.01	3.40	3.82	4.42	13.37	12.6
Ho	0.65	1.72		1.16	0.76	0.60	0.67	0.84	0.88	2.76	2.57
Er	1.83	4.35		3.17	2.00	1.55	1.64	2.20	2.45	7.68	7.25
Tm	0.28	0.64		0.45	0.31	0.27	0.26	0.33	0.36	1.17	1.08
Yb	1.90	3.99		2.92	2.19	1.83	1.76	2.09	2.07	7.24	7.03
Lu	0.28	0.59		0.43	0.33	0.26	0.29	0.31	0.31	1.13	1.03
Hf	3.09	5.11		7.31	2.70	2.23	3.38	3.32	2.67	23.4	14.4
Ta	1.29	2.53		1.67	1.45	1.35	1.30	1.09	3.49	4.94	6.15
Th	1.57	2.71		1.88	1.54	1.66	1.87	1.36	4.47	8.14	7.99
U	0.36	0.57		0.40	0.45	0.59	0.43	0.70	0.83	1.76	1.71
Nb/Y	1.14	0.78	0.66	0.68	0.99	0.86	0.87	0.70	1.69	0.94	0.98
La_N/Sm_N	2.78	2.27		2.01	2.88	2.61	2.40	2.29	5.06	2.87	2.86
Sm_N/Yb_N	1.67	2.23		2.08	1.70	2.19	2.84	1.85	2.67	2.02	2.04

	Sub-alkaline tholeiitic basalts						Boninitic rocks				
Locality	Exalofos	Filira	Filira	Filira	Filira	Filira	Gorgogyri	Filira	Filira	Filira	Filira
Sample	37	22	24	26	23	25	38	27	29	30	28
Rock	N-MORB	E-MORB	E-MORB	E-MORB	E-MORB	E-MORB	Bas And	Bas And	Bas And	And	Bas And
Type	Pillow	Pillow	Pillow	Pillow	Pillow	Pillow	MLF	Dyke	Dyke	Dyke	Dyke
XRF Analyses:											
SiO_2	45.87	45.43	47.30	45.50	45.53	45.49	55.06	52.81	51.69	55.94	53.61
TiO_2	2.03	2.21	2.12	2.64	2.20	2.24	0.20	0.31	0.42	0.45	0.58
Al_2O_3	17.98	15.55	15.34	14.51	15.44	15.29	14.26	15.17	15.67	15.92	15.48
Fe_2O_3	1.46	1.80	1.80	2.01	1.79	1.75	1.16	1.12	1.35	1.20	1.43
FeO	9.73	12.03	12.03	13.40	11.94	11.69	7.75	7.45	8.97	8.01	9.55
MnO	0.17	0.21	0.21	0.33	0.21	0.20	0.09	0.15	0.16	0.16	0.16
MgO	6.37	5.91	5.86	6.42	5.37	5.24	7.44	6.27	7.01	5.50	3.87
CaO	11.02	11.62	8.38	10.11	12.22	12.55	7.90	10.05	8.42	8.75	8.27
Na_2O	3.48	2.44	3.76	2.19	2.59	2.65	3.29	4.28	3.31	2.19	5.18
K_2O	0.03	0.33	0.67	0.27	0.30	0.33	0.02	0.15	0.34	0.29	0.12
P_2O_5	0.23	0.35	0.32	0.47	0.35	0.36	0.06	0.1	0.09	0.12	0.09
L.O.I.	1.63	2.10	2.20	2.16	2.08	2.21	2.74	2.12	2.54	1.45	1.63
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
CO_2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CaO*											
Mg#	53.9	46.7	46.5	46.1	44.5	44.4	63.1	60.0	58.2	55	41.9
Pb	6	11	2	7	8	4	11	n.d.	5	5	9
Zn	77	106	98	125	92	110	55	58	n.d.	68	74
Ni	154	50	47	46	60	50	36	23	30	24	18
Co	47	47	48	44	43	43	37	31	38	29	33
Cr	297	69	78	71	88	72	117	51	97	81	19
V	265	386	370	384	388	397	300	231	304	263	351
Rb	n.d.	4	6	4	3	2	n.d.	n.d.	4	n.d.	n.d.
Sr	140	152	275	126	156	157	197	91	161	108	76
Ba	10	103	151	81	123	100	21	33	50	34	27
Th	n.d.	n.d.	3	n.d.	2	n.d.	n.d.	n.d.	n.d.	n.d.	3
Nb	n.d.	15	14	21	15	14	n.d.	2	n.d.	n.d.	n.d.
Zr	157	171	174	221	169	169	12	14	15	26	28
Y	43	41	42	55	43	43	7	8	10	13	14
La		14	17	24	13	15	n.d.	n.d.	n.d.	n.d.	n.d.
Ce	10	28	26	44	31	27	n.d.	4		6	6
Nd	14	18	19	21	19	18	n.d.	n.d.	n.d.	n.d.	3
ICP-MS Analyses:											
Sc	49.3	45.5	55.5			51.2	47.3	47.4	57.1		
Y	40.6	47.4	51.9			45.2	5.33	7.38	7.61		
Nb	1.10	15.3	14.9			14.2	0.78	0.55	1.20		
La	2.01	11.3	12.5			12.0	0.63	0.91	1.45		
Ce	9.42	25.9	28.8			28.3	1.24	2.24	2.51		
Pr	2.12	3.57	4.09			3.97	0.19	0.31	0.38		
Nd	12.9	16.7	22.5			19.6	0.85	1.67	2.25		
Sm	4.56	4.81	6.37			5.50	0.33	0.57	0.88		
Eu	1.65	1.67	2.03			1.82	0.14	0.23	0.37		
Gd	5.62	5.77	5.81			5.63	0.55	0.65	1.46		
Tb	1.10	1.13	1.10			1.08	0.13	0.13	0.27		
Dy	6.93	7.12	6.81			6.95	1.00	0.96	2.00		
Ho	1.50	1.56	1.58			1.64	0.27	0.24	0.54		
Er	4.21	4.21	4.32			4.55	0.82	0.69	1.66		
Tm	0.63	0.64	0.73			0.77	0.15	0.14	0.27		
Yb	4.13	3.90	4.94			4.76	1.05	0.86	2.05		
Lu	0.63	0.61	0.80			0.70	0.17	0.15	0.35		
Hf	6.43	4.40	5.68			4.97	0.47	0.62	1.39		
Ta	0.12	0.97	1.13			1.03	0.07	0.11	0.57		
Th	0.06	1.14	1.49			1.47	0.16	0.27	0.82		
U	0.02	0.30	0.49			0.39	0.08	0.15	0.73		
Nb/Y	0.01	0.37	0.36	0.38	0.35	0.33	0.12	0.07	0.12		
La_N/Sm_N	0.29	1.52	1.26			1.41	1.21	1.03	1.06		
Sm_N/Yb_N	1.23	1.37	1.43			1.28	0.36	0.73	0.48		

Fe_2O_3- FeO X 0.15; Mg#- 100 X Mg/(Mg+Fe^{2+}), where Mg- MgO/40.32 and Fe- FeO/71.85; CaO*- CaO content recalculated without the CaO contained in calcite. Abbreviations: MLF- massive lava flow; n.d.- not detected.

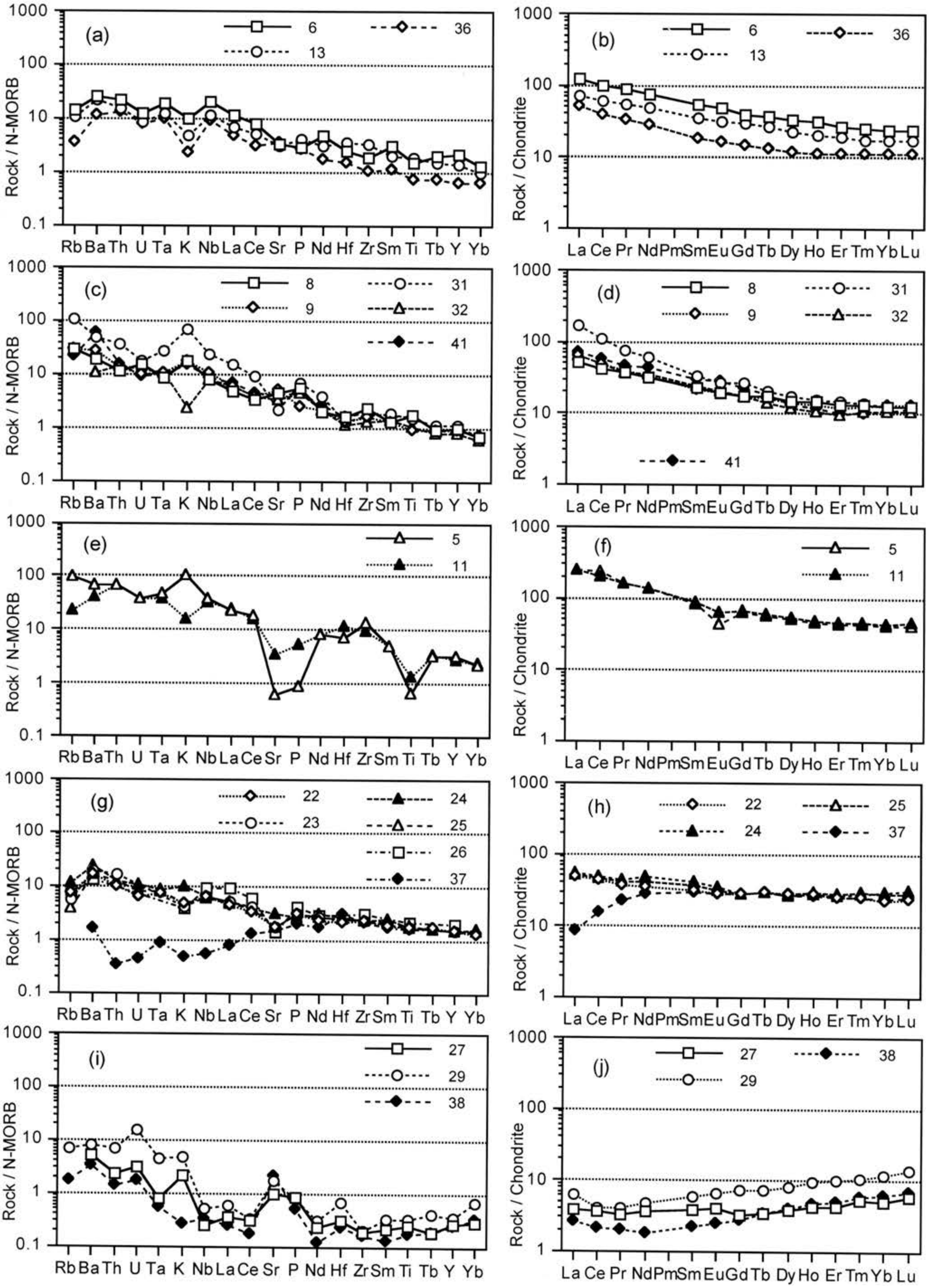

Fig. 5 - N-MORB normalized incompatible element and chondrite-normalized REE patterns for volcanic and subvolcanic rocks from the Koziakas Mélange Unit. Normalizing values are from Sun and McDonough (1989). a, b: transitional to alkaline massive lavas; c, d: transitional to alkaline pillow lavas; e, f: trachyandesite (Nr. 11) and trachyte (Nr. 5); g, h: sub-alkaline tholeiitic basalts (MORBs); i, j: boninitic rocks.

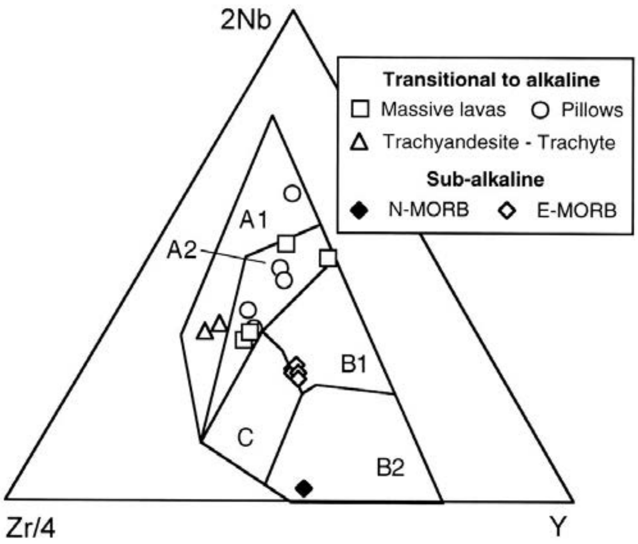

Fig. 6 - Zr/4, Y, 2*Nb discrimination diagram for volcanic and subvolcanic rocks from the Koziakas Mélange Unit. Modified after Meschede (1986). A1- within-plate alkali basalts, A2- within-plate alkali and tholeiitic basalts, B1- transitional mid-ocean ridge basalts, B2- normal mid-ocean ridge basalts, C- within-plate tholeiitic and calc-alkaline basalts.

By contrast, samples from the Filira area display regularly decreasing incompatible element patterns (Fig. 5g) with no U or Nb depletions, and are characterized by LFSE abundance ranging from 2 to 20 times typical N-MORB composition. They show REE abundance ranging from 25 to 50 times that of chondrite (Fig. 5h), and display LREE enrichment similar to E-MORB (Sun and McDonough, 1989). Filira basalts also have higher LFSE abundance than typical N-MORB, appearing in general similar to that of average E-MORB (Sun and McDonough, 1989). Their E-MORB affinity is also reflected in the Nbx2-Zr/4-Y diagram (Fig. 6). These rocks are characterized by highly uniform chemical compositions, and possibly represent fairly evolved, cogenetic rocks, as suggested by their high FeO_t (13.57-15.54 wt%), V (370-397 ppm), and Y(41-55 ppm) contents with Mg# ranging from 46.7 to 44.4 (Table 1, Fig. 7). The increasing CaO/Al_2O_3 ratios with decreasing Mg# (Fig. 7) is compatible with plagioclase fractionation at the initial stages of crystallization, though no negative Eu anomalies are observed (Fig. 5h).

These sub-alkaline tholeiitic rocks may represent the pillow basalts and related dikes of an ophiolitic sequence.

Boninitic basaltic andesites and andesites

This group include exclusively dike varieties, mainly sampled in the Filira area with the exception of sample No. 38 (Gorgogyri area).

The sampled rocks range from basaltic andesites to andesites, and have silica contents varying from 51.69 wt% to 55.94 wt%. The very low TiO_2 contents (0.20-0.58 wt%) and very low Ti/V ratios (4.2-10.5) are comparable with those of typical boninitic rocks from various ophiolitic complexes (Beccaluva et al., 1984; Beccaluva and Serri, 1988). The sub-alkaline character of these rocks is evidenced in the Ti/Y vs. Nb/Y diagram of Fig. 3.

HFSE are generally included between 0.1-0.8 times N-MORB compositions (Fig. 5i), whereas LFSE concentrations are rather variable and enriched with respect to typical N-MORB, possibly reflecting the effects of secondary alteration. REE concentration varies from 1.8 to 13.8 times chondritic abundance, and displays the U-shaped patterns typical of boninites (Fig. 5j), with La_N/Sm_N and Sm_N/Yb_N ratios ranging from 1.03 to 1.21 and 0.36 to 0.73, respectively.

P_2O_5 (0.06-0.12 wt%), Zr (12-28 ppm) and Y (7-14 ppm) contents, with Mg# from 63.1 to 41.9 (Table 1, Fig. 7), are very low, in accordance with typical boninitic compositions. Compatible elements, such as Ni (18-36 ppm), Co (29-38 ppm), and Cr (19-117 ppm) concentrations (Table 1), are generally lower than would be expected for rocks of this type, possibly reflecting the fractionation of mafic phases (olivine, Cr-spinel, and clinopyroxene) at the initial stages of crystallization.

The overall geochemical features of these rocks suggest close similarities with the very low-Ti (boninitic) lavas found in the forearc regions of oceanic island arcs (Crawford et al., 1989; Falloon and Crawford, 1991), as well as in many Hellenides ophiolitic complexes (Beccaluva et al., 1984; Beccaluva and Serri, 1988), and imply magma source(s) modified by subduction processes (Pearce, 1982).

PETROGENESIS AND TECTONO-MAGMATIC INTERPRETATION

The studied rocks from the Koziakas Mélange Unit show a wide variation in their chemical composition which may be related to different source characteristics, since the variation in lava chemistry is a function of mantle composition rather than shallow-level crustal processes (Pearce and Norry, 1979; Pearce, 1983). Since the genetic relationships between rocks within each single chemical group is not proven, it is not possible to evaluate their petrogenetic evolution in detail. We will therefore restrict our petrogenetic discussion to an identification of the possible mantle sources of the distinct lava groups. The use of geochemical indicators for inferring the tectonic setting of eruption of mafic extrusive rocks has been suggested by many authors (Pearce and Norry, 1979; Beccaluva et al., 1979). Analogously, we use the geochemical characteristics of the volcanic and subvolcanic rocks from the Koziakas Mélange Unit to deduce the tectonic setting in which they formed.

Transitional to alkaline rocks

The transitional to alkaline rocks display OIB-Iike trace-element and REE characteristics, suggesting that they represent seamounts formed by magma generation associated with mantle plumes. The chemistry of these rocks is comparable with that of various Pacific seamounts (Moore et al., 1982; Frey and Clague, 1983; Lipman et al., 1989; Haase et al., 1997). In particular, a close similarity with the transitional to alkaline rocks from the Easter seamount chain in the SE Pacific ocean (Haase et al., 1997) is observed, can be seen from the Ce/Y vs. La_N/Yb_N (Fig. 8) and Ta/Hf vs. La_N/Sm_N (Fig. 9) plots. The rocks studied in this paper were generated from an enriched mantle-type source without any detectable influence of continental crust contamination, as indicated by the Th/Yb and Ta/Yb ratios, which are included in the MORB - OIB array (Fig. 10), and are emplaced in a within-plate oceanic setting. This is also confirmed by the Zr/Y ratios (4-11) which, according to Pearce (1983), are the typical values for within-plate ocean island basalts.

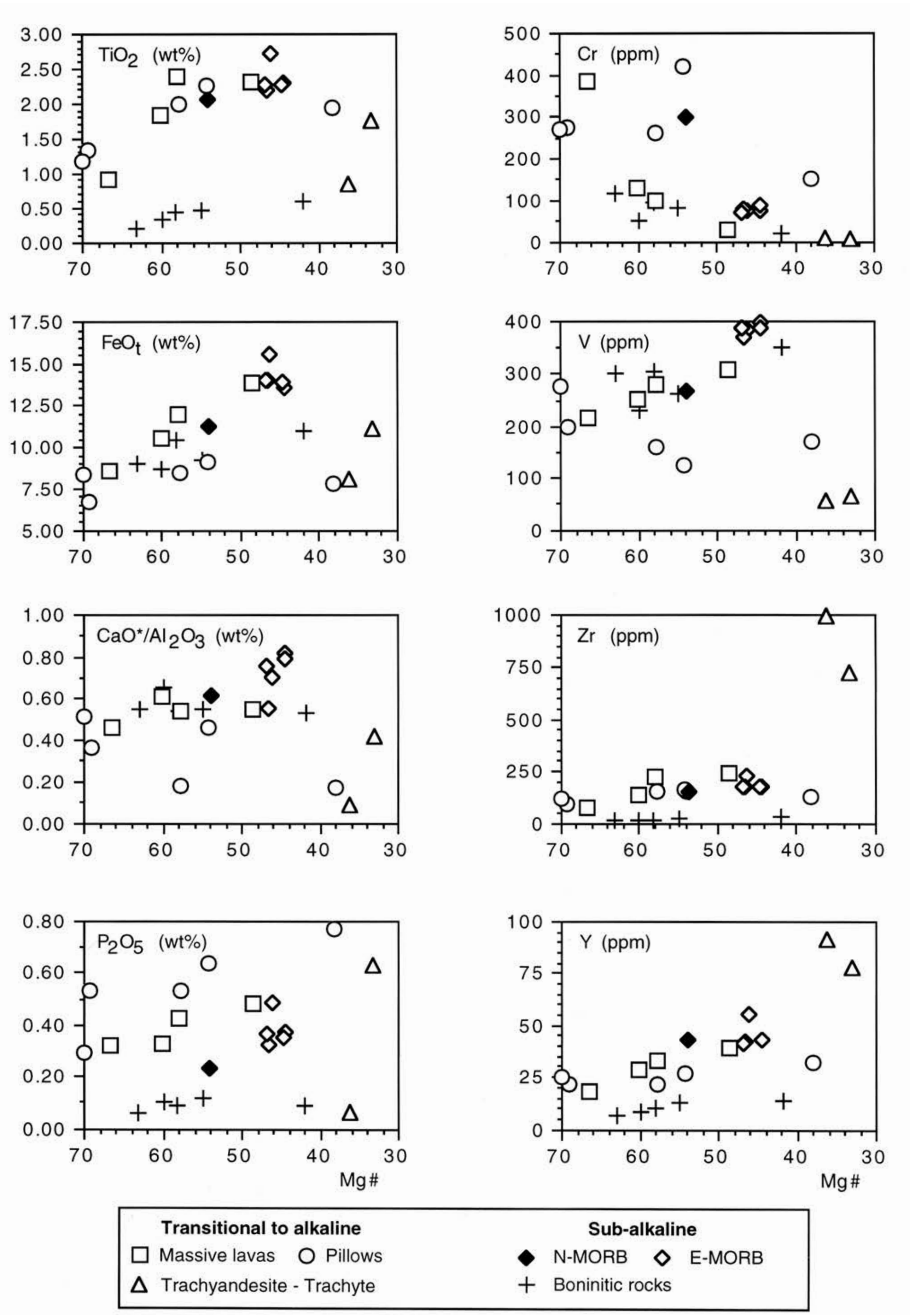

Fig. 7 - Variation of selected major and trace elements vs. Mg# for volcanic and subvolcanic rocks from the Koziakas Mélange Unit. CaO*-CaO content recalculated on calcite-free basis. All major elements are recalculated on anhydrous basis and considering the CaO* content.

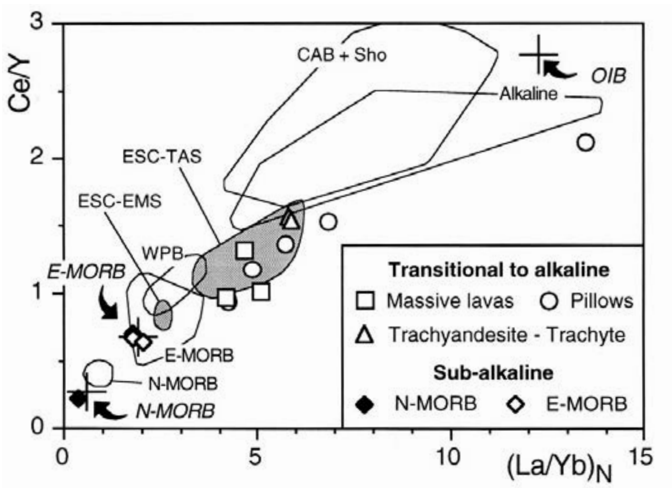

Fig. 8 - Ce/Y vs. $(La/Yb)_N$ diagram for volcanic and subvolcanic rocks from the Koziakas Mélange Unit. Compositions of Modern N-MORB, E-MORB and OIB (large crosses) are from Sun and McDonough (1989). Variations of elemental ratios of alkaline, calc-alkaline (CAB), shoshonitic (Sho), normal-type mid-ocean ridge (N-MORB), enriched-type mid-ocean ridge (E-MORB), and undistinguished within-plate (WPB) Triassic and Jurassic mafic volcanic rocks from various localities of the Hellenides are reported for comparison. Data souces: Capedri et al. (1985), Capedri et al. (1997), and Pe-Piper (1998). Compositional fields for transitional to alkaline series (-TAS) and enriched-MORB series (-EMS) from the Easter seamount chain (ESC-) are from Haase et al. (1997).

A similar tectono-magmatic environment of formation has been proposed for analogous Triassic volcanic rocks from various localities of the Hellenides (Capedri et al., 1997; Pe-Piper, 1998 and references therein). This conclusion is also supported by Nd and Pb isotopic data (Pe-Piper, 1998). Figs. 4, 8, 9, 10 show the close chemical analogy between the studied rocks and the Triassic alkaline rocks widespread in the Hellenides.

The possible mantle source of transitional to alkaline rocks from the Koziakas Mélange Unit is modelled in the $(Dy/Yb)_N$ vs. $(Ce/Yb)_N$ diagram of Fig. 11, in accordance with the model proposed for similar rocks from the Easter seamount chain by Haase and Dewey (1996). This diagram shows that the intraplate rocks studied in this paper are consistent with low degree (ca. 5-8%) partial melting of a theoretical plume source.

Tholeiitic basalts

The chemistry of the tholeiitic basalts suggests melt generation in a mid-ocean ridge tectonic setting. However, chemical data indicate that they were probably derived by partial melting of both N-MORB and E-MORB type asthenospheric mantle sources.

Ratios of highly incompatible trace elements, such as Ce/Y and La/Yb, are generally little influenced by small extents of fractional crystallization, and are believed to reflect either the source characteristics or the degree of partial melting (Saunders et al., 1988). Ce and La are more incompatible than Y, and Yb, respectively; thus, rocks representing the smallest degree of partial melting or derived from more enriched sources exhibit both the highest Ce/Y and La_N/Yb_N ratios. The elemental ratios plotted in Fig. 8 for basalt No. 37 are very similar to those of modern primitive N-MORB (Sun and McDonough, 1989), and are generally compatible with a genesis from primary magmas originating from depleted N-MORB type sub-oceanic mantle sources, with no influence of enriched OIB-type material, as also confirmed by the Ta/Hf vs. La_N/Sm_N and Th/Yb vs. Ta/Yb ratios plotted in Figs. 9 and 10. In addition, apart from low Nb abundance, the N-MORB studied in this paper display many similarities with N-MORBs from various localities of the Hellenides (Capedri et al., 1997; Pe-Piper, 1998), as shown in Figs. 8, 9, 10. In the model shown in Fig. 11 (Haase and Dewey, 1996), the N-MORB studied in this paper is compatible with 10% partial melting of a depleted MORB mantle source.

Basalts from the Filira area display elemental ratios (Figs. 8, 9) very similar to those observed in modern primitive E-MORB (Sun and McDonough, 1989), and possibly represent melts derived from more enriched sources or, alternatively, from lower degrees of partial melting when compared to sample No. 37. The enriched nature of the mantle sources can be inferred on the basis of the Th/Yb vs. Ta/Yb diagram of Fig. 10, where Filira basalts plot in the middle of the MORB - OIB array. The general enrichment in LILE and LREE observed in Figs. 5g, h) is in accordance with this conclusion. Analogous basalts (Figs. 8, 9, 10) from Hellenide ophiolitic mélanges (Pe-Piper, 1998), as well as from the Easter seamount chain (Haase et al., 1997) and Easter microplate rift (Haase, 2002), have been interpreted as originating from depleted mantle sources variably metasomatized by OIB-type components. In the model presented in Fig. 11 (Haase and Dewey, 1996), although some samples display quite low $(Dy/Yb)_N$ ratios, the E-MORBs from the Koziakas Mélange Unit are generally compatible with at least 10% partial melting of a theoretical mixed plume/MORB mantle source.

Similar E-MORBs from the Hellenides have been related

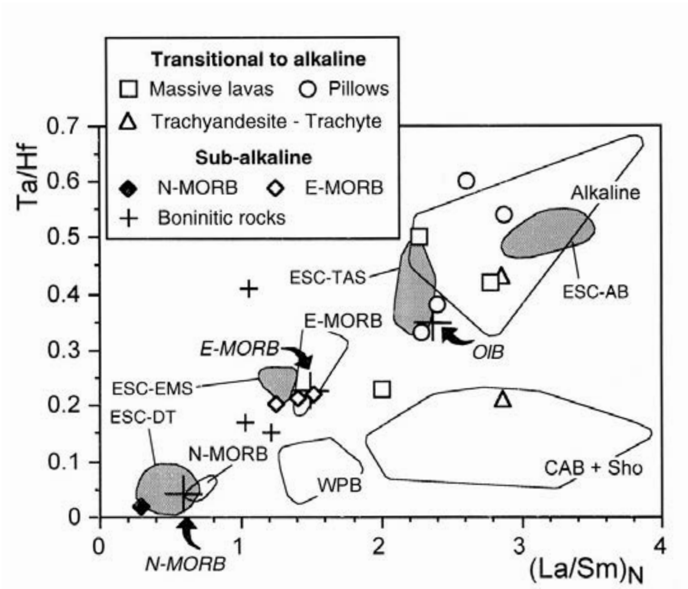

Fig. 9 - Ta/Hf vs. $(La/Sm)_N$ diagram for volcanic and subvolcanic rocks from the Koziakas Mélange Unit. Compositions of Modern N-MORB, E-MORB and OIB (large crosses) are from Sun and McDonough (1989). Variations of elemental ratios of alkaline, calc-alkaline (CAB), shoshonitic (Sho), normal-type mid-ocean ridge (N-MORB), enriched-type mid-ocean ridge (E-MORB), and undistinguished within-plate (WPB) Triassic and Jurassic mafic volcanic rocks from various localities of the Hellenides are reported for comparison. Data souces: Capedri et al. (1985), Capedri et al. (1997), and Pe-Piper (1998). Compositional fields for transitional to alkaline series (-TAS), enriched-MORB series (-EMS), alkali basalts (-AB), and depleted tholeiites (-DT) from the Easter seamount chain (ESC-) are from Haase et al. (1997) and references therein.

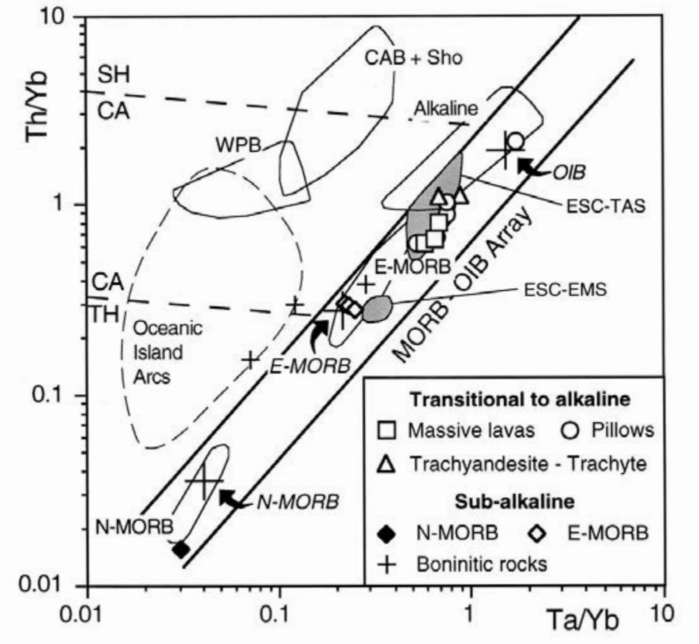

Fig. 10 - Th/Yb vs. Ta/Yb diagram for volcanic and subvolcanic rocks from the Koziakas Mélange Unit. Modified after Pearce (1982). Compositions of Modern N-MORB, E-MORB and OIB (large crosses) are from Sun and McDonough (1989). Variations of elemental ratios of alkaline, calc-alkaline (CAB), shoshonitic (Sho), normal-type mid-ocean ridge (N-MORB), enriched-type mid-ocean ridge (E-MORB), and undistinguished within-plate (WPB) Triassic and Jurassic mafic volcanic rocks from various localities of the Hellenides are reported for comparison. Data sources: Capedri et al. (1985), Capedri et al. (1997), and Pe-Piper (1998). Compositional fields for transitional to alkaline series (-TAS) and enriched-MORB series (-EMS) from the Easter seamount chain (ESC-) are from Haase et al. (1997).

by many authors to the interaction between uprising depleted oceanic asthenosphere and OIB-type plume material during the initial stage of oceanic spreading (Jones and Robertson, 1991; Robertson and Karamata, 1994; Pe-Piper, 1998). Although the data presented in this paper do not allow a clear distinction of the original tectono-magmatic setting, a similar conclusion can reasonably be postulated for the E-MORBs from the Koziakas Mélange Unit.

Boninitic basalts and basaltic andesites

Observing the incompatible elements and REE abundance (Figs. 5i, j), a marked depletion of HFSE and REE coupled with a relative enrichment in LILE can be noticed in boninitic rocks. In addition, these rocks exhibit a mild LREE enrichment relative to medium-REE, and a relative depletion of medium-REE with respect to HREE that is, the typical U-shaped REE patterns of boninitic rocks.

The supra-subduction zone (SSZ) signature in basaltic rocks is commonly characterized by a depleted signature as a consequence of previous melt extraction in the mantle source (i.e., MORB generation), as well as LILE enrichment produced by subduction-derived fluids (Beccaluva et al., 1984; Beccaluva and Serri, 1988; Pearce, 1983). The latter is, however, difficult to identify in ophiolitic rocks since they are commonly altered. In particular, variable amounts of LILE mobilization have been recognized in the studied rocks when observing the relative variation of these elements with respect to Mg# and Zr variation (not shown in this paper).

According to Beccaluva and Serri (1988), the La_N/Sm_N ratios displayed by Koziakas boninites (Fig. 9) are compatible with high degrees of hydrous (but not necessarily water-saturated) melting of mantle sources that have experienced previous extraction of basaltic melts with variable subduction-derived enrichment events.

Trace element patterns similar to those displayed by Koziakas boninites are observed in many very low-Ti basalts from various ophiolitic complexes of the Albanides (Beccaluva et al., 1994; Bébien et al., 2000; Bortolotti et al., 2002) and Hellenides (Beccaluva et al., 1984; Kostopoulos, 1988; Capedri et al., 1996), which are interpreted as having originated during the Middle-Upper Jurassic intra-oceanic subduction phase(s) that characterized the Mirdita-Pindos oceanic basin. A similar tectono-magmatic formation setting can reasonably be postulated for the Koziakas boninites, as also suggested by the incompatible element variations shown in Fig. 10, where the studied boninites are generally shifted toward the oceanic island arc compositions (Pearce, 1982).

DISCUSSION AND CONCLUSIONS

The Mélange Unit of the Koziakas series consists of several, relatively intact volcanic sequences and radiolarian chert successions which are commonly in contact with each other by either faults or thrusts. A previous work (Capedri et al., 1985) interpreted this Unit as a mid-ocean ridge type, Jurassic ophiolitic sequence. Nonetheless, the volcanic rocks studied in this paper show different magmatic affinities, and possibly originated in distinct tectono-magmatic settings, i.e.: mid-ocean ridge, seamount, and island arc. Moreover, radiolarian assemblages indicate two distinct age ranges: Middle-Late Triassic and Middle-Late Jurassic (Chiari et al., in progress). Although radiolarian cherts are not stratigraphically related to any of the volcanic sequences, it is tempting to assume that the three distinct magmatic groups might be different in age. The lithologies dredged and drilled in many recent accretionary complexes

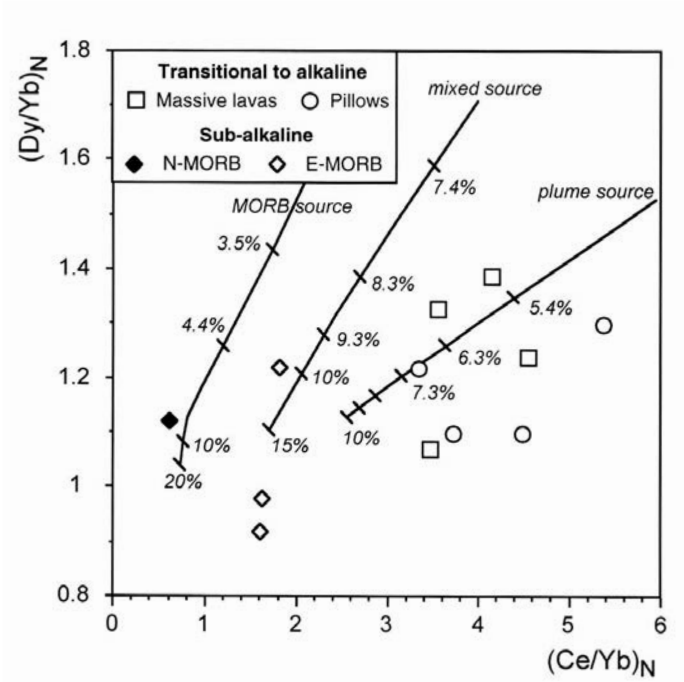

Fig. 11 - $(Dy/Yb)_N$ vs. $(Ce/Yb)_N$ diagram for volcanic and subvolcanic rocks from the Koziakas Mélange Unit. Melt model is from Haase and Dewey (1996).

such as, for example, the Mariana forearc, frequently include arc-related boninites and fragments of mid-ocean ridge and oceanic island basalts (Bloomer, 1983; Johnson et al., 1991). Similar lithologies are also common in many Alpine mélange complexes, from Greece (Capedri et al., 1996; Jones and Robertson, 1991; Danelian and Robertson, 2001) to India (Ahmad et al., 1996), via Turkey (Floyd, 1993; Tankut et al., 1998) and Iran (Arvin and Robinson, 1994). The presence of rocks generated in distinct tectonic settings (island arc, mid-ocean ridge, and seamount) in both modern and ancient accretionary complexes is commonly interpreted as a result of the incorporation of fragmented oceanic lithosphere by either accretion or obduction. A discussion on the formation and tectonic evolution of the Koziakas Mélange Unit is beyond the scope of this paper, nonetheless, a tectonic evolution similar to those of many alpine accretionary complexes can reasonably be postulated.

As a result, the Koziakas Mélange Unit, in contrast to a previous interpretation (Capedri et al., 1985), likely represents a tectonic mélange generated in an accretionary complex rather than a distinct mid-ocean ridge type ophiolitic sequence.

The Upper Mesozoic-Cenozoic accretionary complexes scattered along the Mirdita-Subpelagonian (Maliac) zone include different magmatic associations that document the igneous development of the Albanide-Hellenide sector of the Neo-Tethyan oceanic basin, from the Permian-Triassic rifting stage and Triassic-Jurassic oceanization to the Middle-Late Jurassic intra-oceanic subduction (Pe-Piper, 1998; Jones and Robertson, 1991; Danelian and Robertson, 2001; Robertson, 2002). Accordingly, the different igneous lithologies found in the Koziakas Mélange Unit suggest three distinct episodes of volcanism.

(1) The N-MORB and E-MORB signatures displayed by tholeiitic basalts suggest that these are dismembered fragments of oceanic crust, which presumably represent remnants of oceanic lithosphere developed at an oceanic spreading centre within the Pindos sector of the Neo-Tethyan ocean. This is consistent with the regional reconstruction of the Neo-Tethyan realm in the eastern Mediterranean, which implies the existence of the Mirdita-Pindos oceanic basin between the Apulia (in the west) and Pelagonia (in the east) continental margins during the Late Triassic-Middle Jurassic (Robertson et al., 1991; Jones and Robertson, 1991; Beccaluva et al., 1994; Robertson, 2002).

(2) The occurrence of transitional and alkaline rocks, mostly of Middle-Late Triassic age, has previously been described in several Hellenides ophiolitic mélanges (Jones and Robertson, 1991; Pe-Piper, 1998; Danelian and Robertson, 2001). They are interpreted as fragments of seamounts related to widespread plume activity that developed in the Pindos ocean from its Early Triassic rifting (Pe-Piper, 1998). A similar tectono-magmatic setting can be postulated for the transitional and alkaline rocks rocks included in the Koziakas Mélange Unit. The preservation of these intraplate volcanic bodies in the accretionary wedge at convergent margins is facilitated by their relative buoyancy, which favoured their incorporation into ophiolitic mélanges rather than being subducted at the convergent margin, and testify for the subduction of large portions of oceanic crust.

(3) The presence of island arc boninitic dikes is in accordance with the development in the Hellenide sector of the Neo-Tethyan oceanic basin of subduction zone-related magmatism, well-documented in many ophiolitic complexes of the Albanides and Hellenides (Beccaluva et al., 1994; Bortolotti et al., 2002; Beccaluva et al., 1984; Capedri et al., 1996).

However, in the Hellenides mélanges, boninites usually occur as blocks or clasts deriving from the intra-oceanic island arc (Capedri et al., 1996). By contrast, in the Koziakas Mélange Unit some boninitic dikes crosscut different tectonic slices which can be referred to both MORB and OIB sequences. This implies that tectonic incorporation into the accretionary wedge of MORB-type and OIB-type material occurred prior to the development of boninitic magmatism. The presence of boninitic dikes in the accretionary wedge suggests that boninitic magmatism was generated in a portion of the supra-subduction setting which was exceptionally close to the trench. In this framework, the genesis of boninitic magmas close to the trench (Falloon and Danyushevsky, 2000) suggests partial melting of depleted peridotites during the early stage of subduction due to a particularly high thermal regime in the mantle source or, alternatively, a very steep subduction.

The relative chronology of the three magmatic events recorded in the Koziakas Mélange Unit cannot be sustained by biostratigraphical datings; nonetheless, according to the general tectonic evolution of the Hellenide sector of the Neo-Tethys, based on literature data, they can be referred respectively to: (1) a Middle-Late Triassic to Middle Jurassic seafloor spreading characterized by early, short-lived eruptions of E-MORBs followed by N-MORBs; (2) a Triassic oceanic plume activity; (3) a Late Jurassic development of intra-oceanic subduction zone characterized by island arc tholeiitic and boninitic magmatisms.

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