

HORNBLENDITES WITHIN OPHIOLITES OF CRETE, GREECE: EVIDENCE FOR AMPHIBOLE-RICH CUMULATES DERIVED FROM AN IRON-RICH THOLEIITIC MELT

Jürgen Koepke* and Eberhard Seidel**

* Institut für Mineralogie, Universität Hannover, Germany.

** Institut für Mineralogie und Geochemie, Universität Köln, Germany.

Corresponding Author: Jürgen Koepke, e-mail koepke@mineralogie.uni-hannover.de

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ABSTRACT

Massive hornblendites within Jurassic ophiolites near the Kerames village (Central Crete, Greece) represent cumulates derived from an iron-rich tholeiitic magma intruding Iherzolitic peridotites. After intrusion, the rocks were affected by crystal-plastic deformation, which broadly overprinted the magmatic textures within a high-temperature shear zone. The dominant rock types are strongly foliated pargasite schists rich in accessory minerals, like ilmenite, titanite, apatite, occasionally bearing relics of primary plagioclase and clinopyroxene. Most of the pargasites are characterised by a marked zoning which indicates re-equilibration at lower temperature.

The rocks are characterised by a well defined correlation of $\text{FeO}^{\text{tot}}/(\text{FeO}^{\text{tot}}+\text{MgO})$ with TiO_2 , P_2O_5 , Zr, Y, and light rare earth elements (positive) and with Cr and Ni (negative) indicating a typical magmatic differentiation trend. Major and trace element geochemistry reveal a single-stage magmatic evolution to extreme iron-rich compositions consistent with a model of fractional crystallization with amphibole as the dominant phase. Some highly differentiated rocks show chondrite-normalised REE patterns characterised by a strong enrichment of LREE, which is attributed to an accumulation of monazite and/or allanite. From experimental data we conclude that the Cretan hornblendites were generated under strongly reducing conditions from an iron-rich melt with a water content >4 wt% at temperatures around 1000°C. The inferred high water content is in accordance with an origin of the melt in a back-arc basin.

In comparison with well-described examples of oxide gabbro occurrences from other ophiolites and from the modern oceanic crust, the evolution of a continuous sequence of ferrogabbroic composition with a thickness of some decameters is an outstanding feature of the investigated rocks from Crete. It seems probable that the hornblende sequence represents a frozen cumulate volume derived from a heavy late-stage melt, that was not able to erupt. No chemical gradient (e.g., in the MgO/FeO ratio) with height was observed in the Cretan intrusive complex. Therefore, we assume that the intrusion is not the result of an in-situ solidification of one magma body, but the product of multiple intrusions of small magma batches with different grades of differentiation mixed prior to complete solidification by deformation and compaction processes.

INTRODUCTION

Our knowledge on the mechanism how oxide gabbros (synonym: Fe-Ti gabbros, ferrogabbros) in the oceanic crust are formed is poor. Extreme differentiation of a tholeiitic magma can lead to either iron-rich, silica-poor or iron-poor, silica-rich products. Rocks of both trends occur within the oceanic crust, where they have been generated from small volumes of melt at a very late stage. The iron-poor, silica-rich rocks mostly composed by "oceanic plagiogranites" (Coleman, 1977; Nicolas, 1989) within the gabbroic section (layer 3) of the oceanic crust. Relevant melts are also able to erupt onto the surface of the oceanic crust and can occur as rhyodacites or rhyolites in suites accompanying MORB-type evolved basalts (e.g., Perfit and Fornari, 1983). On the other hand, the iron-rich, silica-poor differentiates were believed for a long time to exist exclusively in the intrusive section of the earth's crust forming cumulate oxide gabbros. Prominent examples of such rocks are known from tholeiitic layered intrusions in the continental crust, such as Skaergaard (Wager and Brown, 1968), and from gabbroic sequences in the oceanic crust, either from ophiolites (e.g., Stern, 1979; Serri, 1981; Pognante, 1982; Hopkinson and Roberts, 1995; Tiepolo et al., 1997) or from recent oceans (e.g., Dick et al., 2000; Natland and Dick, 2002). The classification of these oxide gabbros as cumulates led to the hypothesis that relevant iron-rich liquids, such as those postulated by Wager and Deer (1939) for the Skaergaard intrusion, do not exist (Hunter and Sparks, 1987). However, during the last two decades the

formation of iron-rich basaltic glasses near the spreading axes of modern oceans has been reported indicating the existence of iron-rich melts during a late stage of differentiation of a MORB magma (Ludden et al., 1980; Le Roex et al., 1982; Perfit and Fornari, 1983; Sinton, 1983; Melson and O'Hearn, 1986; Juster et al., 1989; Brooks et al., 1991). Because of their high density it is more probable that iron-rich magmas occur at depth, without the ability to reach the surface (Brooks et al., 1991). This is confirmed by a great number of oxide gabbroic intrusions within the plutonic foundation of the oceanic crust. The occurrences of equivalent extrusive rocks, by contrast, are very restricted (e.g., Beccaluva et al., 1977; 1999; Stern, 1979; Serri, 1981). Thus, the existence of melt lenses filled with highly fractionated, uneruptable liquids was postulated (e.g., Brooks et al., 1991; Natland and Dick, 1996; MacLeod and Yaouancq, 2000) and at least some oceanic oxide gabbros can be regarded as the crystallised products of such magmas (MacLeod and Yaouancq, 2000).

A different model for the generation of oxide gabbros was reported from the ultra-slow-spreading Southwest Indian Ridge, where a section of the lower oceanic crust was drilled during the Ocean Drilling Program, Leg 118/176, (Ozawa et al., 1991; Dick et al., 2000; Natland and Dick, 2002). The observed spatial relation between olivine gabbro, trondhjemite and oxide gabbro is not compatible with a model of upward differentiation of a large magma body. The whole section is strongly tectonised and it is assumed that deformation and compaction played an important role for the formation of the observed oxide gabbro associations.

According to Niu et al. (2002) the Fe-Ti oxide minerals within the oxide gabbros are not in equilibrium with the associated silicate phases. These authors assumed that the Fe-Ti oxides were mechanically impregnated, having been driven by strong tectonic forces within the high-temperature shear zones that are very frequent in the Southwest Indian Ridge (Shipboard Scientific Party, 1999).

Another model of the generation of iron-rich melts in tholeiitic systems takes into account a primary mantle source enriched in iron. This is indicated by MORB differentiation trends from the Galapagos propagating rift system in the oceanic crust (Christie and Sinton, 1986), but also by differentiation trends from Large Igneous Provinces (e.g. West Greenland) in the continental crust, where iron accumulation was assumed to be a consequence of iron-rich streaks in upwelling mantle plumes (Gibson et al., 2000).

We present here new petrographic and geochemical data from a suite of hornblendites occurring in the Jurassic ophiolites of Crete (Greece). These rocks, in their chemical composition, correspond to typical iron- and titanium-rich gabbros. Our results show that the hornblendites were generated by accumulation of amphibole in an iron-rich magma which intruded the surrounding lherzolitic peridotites in Jurassic times.

GEOLOGIC SETTING

Dismembered ophiolites of Crete represent the southernmost outliers of the Jurassic ophiolite belt of the Dinarides/Hellenides (Koepke et al., 2002). They form decameter- to kilometer- sized isolated bodies within the so-called "Uppermost Unit" of the Cretan nappe pile (Fig. 1). This unit is a composite nappe consisting of ophiolites on the top and a chaotic mixture (*mélange*) of various sedimentary, metamorphic, and magmatic rocks of different ages at the base (Creutzburg and Seidel, 1975; Langosch et al., 2000). The ophiolites occurring in this unit were described in detail by Koepke (1986) and Koepke et al. (1985; 2002). The Cretan ophiolites mainly consist of serpentinites with relatively high contents of Al_2O_3 and CaO and include relics of spinel lherzolite. As inferred from the chemistry of relict orthopyroxene and spinel (Koepke et al., 2002), these ultramafic rocks may represent primitive, undepleted mantle formed under a slow-spreading ridge (e.g., Nicolas, 1989; Niu and Hékenian, 1997). However, such lherzolite-type mantle occurs also in ophiolites of polygenetic origin, characterised by a long-lasting formation in an arc environment, which is indicated by isotope signatures (e.g., the Trinity ophiolite; Jacobsen et al., 1984; Gruau et al., 1995).

The studied hornblendites are exposed in an outcrop west of the village of Kerames in Central Crete (Fig. 1). They are closely associated with serpentinitised spinel lherzolites. K-Ar dating of the magmatic amphibole yielded model ages between 149 and 166 Ma (mean of 156 Ma; Koepke et al., 2002). This Middle-Upper Jurassic age is assumed as age of intrusion of the hornblendites into the lherzolites and indicates that the ophiolite was formed during the Jurassic, in accordance with the previous datings of the ophiolites from the Dinarides and Hellenides (see reviews of Spray et al., 1984; Robertson and Shallo, 2000, and Robertson, 2002).

The peridotites of Kerames, like those of other Cretan ophiolites, are intruded by gabbro-dioritic dikes mainly meter-sized, and ranging in composition from pyroxene gabbros to hornblende diorites. The dominance of hornblende as crystallizing phase in these rocks as well as the geochemical signature indicate a subduction-related origin for these dikes (Koepke et al., 2002). K-Ar amphibole dates obtained by Koepke et al. (2002) for the gabbro-dioritic dikes within the Kerames peridotites are around 20 Ma younger than those of the hornblendites. This time gap, in conjunction with the different geochemical, petrographic and structural features of the rocks, indicates a polygenetic origin of the Cretan ophiolites, probably in a subduction-related environment (Seidel et al., 1981; Koepke et al., 2002).

All the ophiolites of Crete show a more or less pronounced metamorphic overprint of the greenschist to epidote-amphibolite facies, evidenced by antigorite in the serpentinites and the paragenesis actinolite - epidote - albite in the gabbro-dioritic dikes (Koepke et al., 2002). In some locations, where this overprint is strong (Pefkos, Asterousia, Fig. 1), the serpentinites exclusively consist of antigorite, and the dike rocks are transformed to foliated metagabbros. Here, the K-Ar amphibole dates of the dike rocks are also strongly influenced leading to distinctly younger model ages around 95 Ma. Interestingly, the ophiolites near Gonies and Spili (Fig. 1) are associated with low-grade metamorphosed limestones which besides ophiolitic detritus contain Upper Cretaceous fossils (Koepke, 1986). Therefore, it is assumed that the metamorphic overprint of the Cretan ophiolites took place during the Late Cretaceous or later.

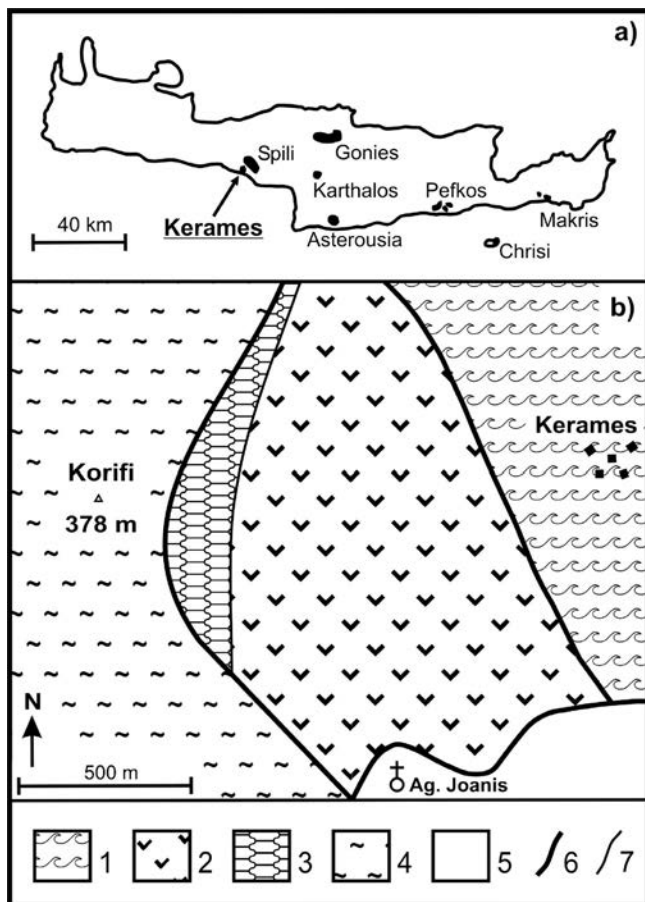


Fig. 1 - (a) Sketch map of Crete showing ophiolite outcrops and the location of the studied hornblendites. For a more detailed structural map of Crete see Creutzburg and Seidel (1975). (b) Geological map of the investigated area west of the Kerames village; modified after Bonneau and Lys (1978) and Skeris (1978). 1: Phyllite-Quartzite Unit; 2 and 3: Ophiolites - 2: Serpentinites with relics of lherzolites; 3: Hornblendites; 4: blueschist-facies metabasalts and metasediments (see text for details); 5: Holocene; 6: tectonic contact; 7: intrusive contact.

According to Bonneau and Lys (1978) and Bonneau (1985), the hornblendites and serpentinites west of Kerames rest on the Vatos Unit, recently named “Preveli subunit” (Thomson et al., 1998) which comprises high-pressure metamorphic metabasalts and metasediments (with Permian fossils) at the base, and flyschoid sediments at the top. According to Koepke et al. (1997) these rocks were metamorphosed under the epidote-blueschist facies condition. Metamorphic rocks corresponding to those of the Vatos Unit occur in Gavdos island south of Crete (Kalypso Unit; Seidel et al., 1977) and Gerakari in Central Crete (Vicente, 1970; Creutzburg and Seidel, 1975; Seidel et al., 1977). In Gavdos, these metamorphic rocks are covered by a transgressive sequence of Upper Cretaceous sediments. K-Ar dating on sodic-calcic amphiboles and phengites of high P/T metamorphic rocks from Gavdos and Central Crete yielded mainly Late Jurassic model ages around 150 Ma (Seidel et al., 1977).

FIELD RELATIONS

The hornblendites are foliated mafic rocks, mainly consisting of amphiboles exposed about 1 km W of Kerames (Fig. 1). They form a coherent sheet up to some tens of meters in thickness that lies beneath serpentinitised lherzolites. In places, the mafic rocks exhibit a marked layering due to alternating bands of black hornblendites and greenish-coloured clinozoisite/epidote albite rocks. A systematic variation in petrography or geochemistry of the whole suite with profile height is not discernible. This is confirmed by the rock chemistry (e.g., a decrease in MgO/FeO). Like the serpentinites, the mafic rocks are intensely deformed, and typically form meter-sized lenticular bodies in a cataclastic matrix of the same material.

The strongly tectonised contacts between hornblendites and serpentinitised peridotites are irregular and interfingering. Relatively small, decimeter-sized veins of hornblendite within the serpentinites are observable, showing contacts that are regarded as primary intrusive. Locally, the hornblendites form meter-thick large bodies which are surrounded by serpentinites. In general, intrusive contact relationships are documented for the smaller intrusives, but not for the whole large body. Since the small intrusives and the large hornblendite body consist of the same rock type showing identical geochemical and petrographic characteristics, we assume that the large body is also intrusive into the peridotite. Very similar decimeter- to decameter-scaled gabbroic intrusions into upper mantle peridotites were reported from transform faults in recent spreading centers (Cannat and Casey, 1995; Tartarotti et al., 1995; Constantin, 1999).

At the contact between hornblendite and serpentinite, decimeter-thick zones of rodingites are developed, so evidencing that intrusion was prior to the process of serpentinitization/rodingitization, which probably occurred in the high heat flow environment of a ridge system. Conversely, the 20 Ma younger gabbro-dioritic dikes were intruded into the peridotites after the serpentinitization process. This is indicated by the formation of hydrous high-temperature parageneses at the contact between gabbro-dioritic dikes and serpentinites (presence of anthophyllite; Koepke, 1986) and by the lack of rodingitization in the gabbro-diorite dikes.

A late-phase of brittle deformation was accompanied by hydrothermal activity resulting in the formation of veins that consist of prehnite, albite, potassium feldspar, and actino-

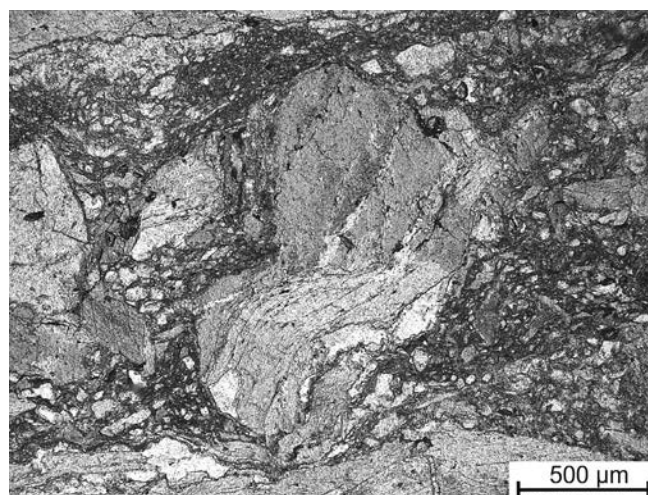
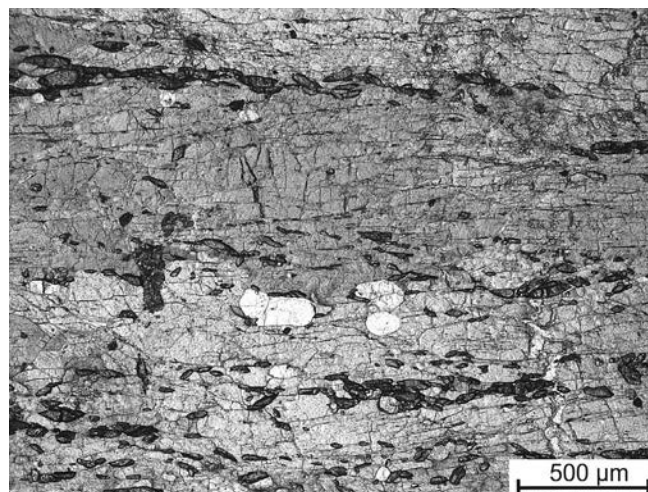
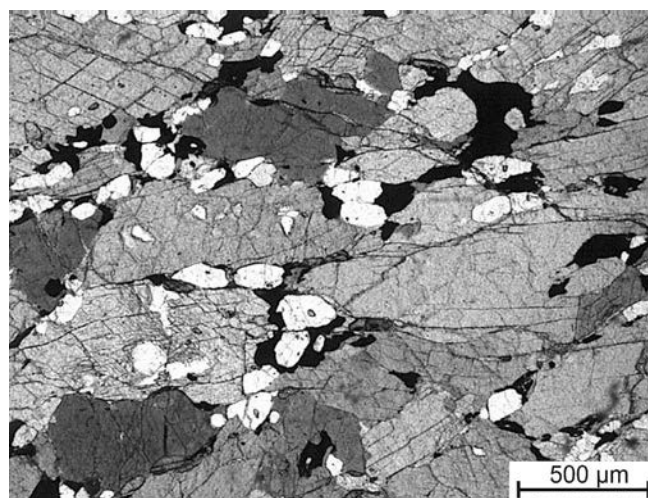


Fig. 2 - Petrography of the Cretan hornblendites. (a) Relics of a primary magmatic cumulate texture: Hypidiomorphic magnesiohastingsite with late-stage crystallised interstitial apatite and ilmenite; sample KER1. (b) Strongly sheared pargasite-schist exhibiting extreme flattening of the grains; primary ilmenites were changed into tiny lenticular titanites which were aligned in disrupted strings at the contact between the platy amphibole grains, accompanied by apatite; sample 96/35a. (c) Sigmoid ductile deformed pargasite porphyroclast in a brittle shear zone; the cataclastic matrix consists of disrupted pargasites (same composition as the porphyroclast), titanite, and prehnite; sample JK232.

lite. During this phase meter-thick lenses of albite rocks with a copper mineralisation (presence of malachite) were generated.

PETROGRAPHY

The hornblendites are mostly medium-grained and exhibit a strong foliation, sometimes with a typical flaser texture. The rocks show signs of ductile and brittle deformation. Often, they are mylonitic with rounded, augen-like amphiboles in a fine-grained, cataclastic matrix. The most common rock type is a platy, nearly monomineralic granoblastic pargasite schist rich in accessory minerals like ilmenite, titanite, and apatite. Another rock type with layers of altered plagioclase (now albite), clinopyroxene, and clinozoisite/epidote resembles typical amphibolites.

Magmatic minerals and textures, and postmagmatic shear features

Although the geochemistry of the hornblendites indicates that these rocks represent a magmatic suite of ferrogabbroic composition (see below), magmatic relics are rare: they are characterised by protogranular textures exhibiting a framework of hypidiomorphic amphiboles with late-stage crystallised, interstitial apatite, ilmenite, and rarely allanite (Fig. 2a); clinopyroxene with tiny lamellae of chlorite probably representing former orthopyroxene exsolutions; hypidiomorphic tabular plagioclase (now altered to albite). Early crystallised zircons and, very rarely, monazite occur as tiny inclusions within the amphiboles of the rocks strongly en-

riched in incompatible elements. Such textures (Fig. 2a) imply a cumulate origin for these rocks, according to the following scenario. Amphiboles, initially crystallised in an evolved, Fe-rich magma (see below), were accumulated probably on the floor of a magma chamber. Late-stage ilmenite and apatite filled the interstices. Then, the residual melt was squeezed out, probably due to compaction or high-temperature shearing processes within the newly formed lithosphere.

A post-magmatic, high-temperature shearing is often visible, indicated by an extreme flattening of the primary amphibole (predominantly pargasite) which did not change its composition (Fig. 2b). Some amphiboles show signs of plastic deformation like sigmoid deformation (Fig. 2c). During this process, most of the primary ilmenite was changed into tiny lenticular titanite which is aligned in disrupted strings at the contacts between the platy amphibole grains (Fig. 2b).

One sample (96/38a) contains relics of a probably magmatic clinopyroxene, now included in magnesiohornblende. Its X_{Mg} of 0.78 ($X_{Mg} = Mg/(Fe^{2+} + Mg)$) is more typical of "normal" oceanic olivine-clinopyroxene gabbros than of oxide gabbros. The high X_{Mg} in comparison with the surrounding hornblende (0.68) suggests that these clinopyroxenes represent trapped minerals of the main-stage crystallization of olivine-clinopyroxene gabbros rather than relics of early phases within the late-stage iron-rich magma.

Ilmenite is the typical Fe-Ti oxide phase within those hornblendites that were only weakly affected by postmagmatic shearing processes or by the later metamorphic overprint. Representative ilmenite analyses are given in Table 1. Magnetite was not observed as a primary magmatic phase.

A striking feature of these rocks are shear zones showing

Table 1 - Representative analyses of primary ilmenites and secondary zoisite, epidote, garnet, and diopside of hornblendites from Kerames, Crete

Sample	KER1	96/143	96/35b	JK739	96/35b	96/35b	96/35g
	ilmenite	ilmenite	zoisite	epidote	garnet ^{a)}	garnet ^{a)}	diopside
No. analyses	(10)	(9)	(2)	(8)	(1) core	(1) rim	(4)
SiO ₂	0.03	0.01	39.11	38.04	38.06	37.97	50.33
TiO ₂	49.96	52.65	0.16	0.15	0.10	0.22	0.32
Al ₂ O ₃	b.d.l.	b.d.l.	31.57	25.12	19.87	20.57	4.16
FeO ^{tot}	47.85	44.73	1.97	9.80	20.08	20.87	12.78
MnO	1.74	3.11	0.04	0.25	3.04	0.63	0.36
MgO	0.06	0.06	0.07	0.03	2.27	2.36	8.92
CaO	0.07	0.10	24.70	23.81	16.03	16.47	20.73
Na ₂ O	-	-	-	-	-	-	1.88
Total	99.71	100.66	97.62	97.20	99.45	99.09	99.48

a) Single analysis taken from a profile through a garnet grain
-, not analyzed; b.d.l., below detection limit

signs of both high-temperature plastic and subsequent low-temperature brittle deformation. Very often mylonitic zones, visible down to the micro-scale, are characterised by relics of plastically deformed minerals incorporated into a fine-grained cataclastic matrix (Fig. 2c). The brittle stage was accompanied by hydrothermal activity leading to the formation of minerals like prehnite, albite, potassium feldspar, epidote, and actinolite.

Metamorphic overprint

The primary magmatic rocks have been affected by a metamorphism of the epidote-amphibolite to greenschist facies with variable deformation. Foliated rocks with extremely flattened minerals occur as well as rocks which are only statically recrystallised with granoblastic polygonal textures. Typical metamorphic parageneses are actinolitic hornblende - albite - clinozoisite/epidote - titanite $\pm$ garnet $\pm$ zoisite $\pm$ diopside in the foliated rocks, and actinolite - albite - clinozoisite/epidote - chlorite - titanite ($\pm$ magnetite, $\pm$ hematite) in the statically recrystallised rocks. The amphiboles show dark green or brown-green cores consisting of magmatic pargasite surrounded by green rims which are distinctly lower in Na, Al, K, and Ti representing hornblendes or actinolites. Poikilitic garnets show inclusions of amphibole, clinozoisite/epidote, albite, titanite, and calcite and are very rich in Ca (up to 17 wt% CaO). Their MnO-concentrations decrease from the core (~ 4.0 wt%) to the rim (0.8 wt%; Table 1), a typical feature of garnets crystallised under greenschist facies metamorphism (e.g., Miyashiro, 1975). During the metamorphic phase the primary ilmenites were replaced by titanite and sometimes also by rutile. The observed metamorphism is characterised by strong oxidizing redox conditions. While ilmenite indicates reducing redox conditions during the magmatic stage, the coexistence of magnetite and even hematite with epidote rich in Fe^{3+} points to highly oxidizing conditions during the overprint. Representative analyses of garnet, zoisite, epidote, and diopside are listed in Table 1.

MINERAL CHEMISTRY OF THE MAGMATIC AMPHIBOLES

The minerals were analysed using Cameca electron microprobes (Cameca Camebax equipped with a SAMX operating system and Cameca SX100; 15 kV acceleration potential, a current of 18 nA, 10-20 seconds counting time per el-

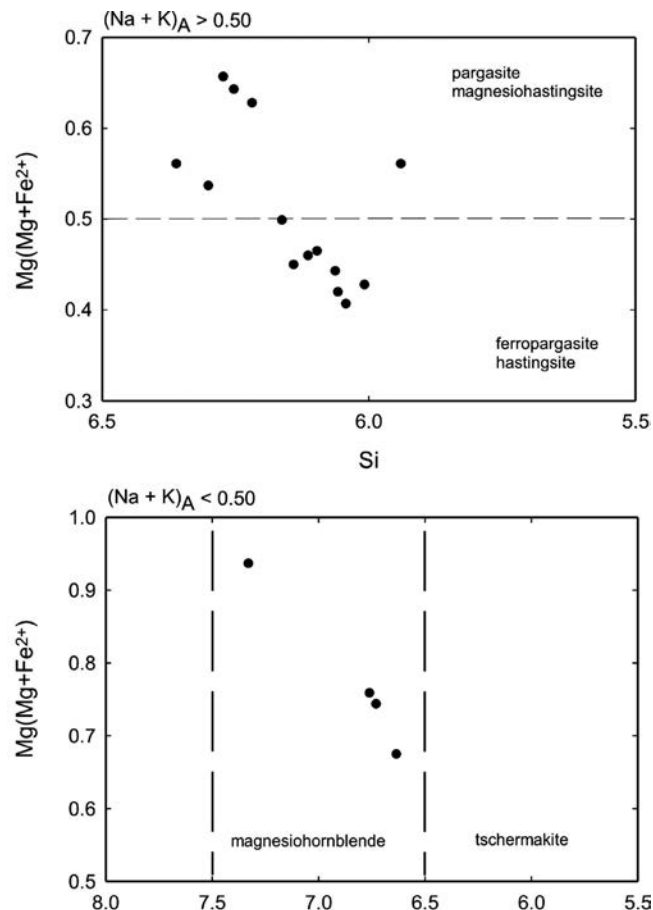


Fig. 3 - Composition of the amphiboles of the Cretan hornblendites (classification after Leake et al., 1997).

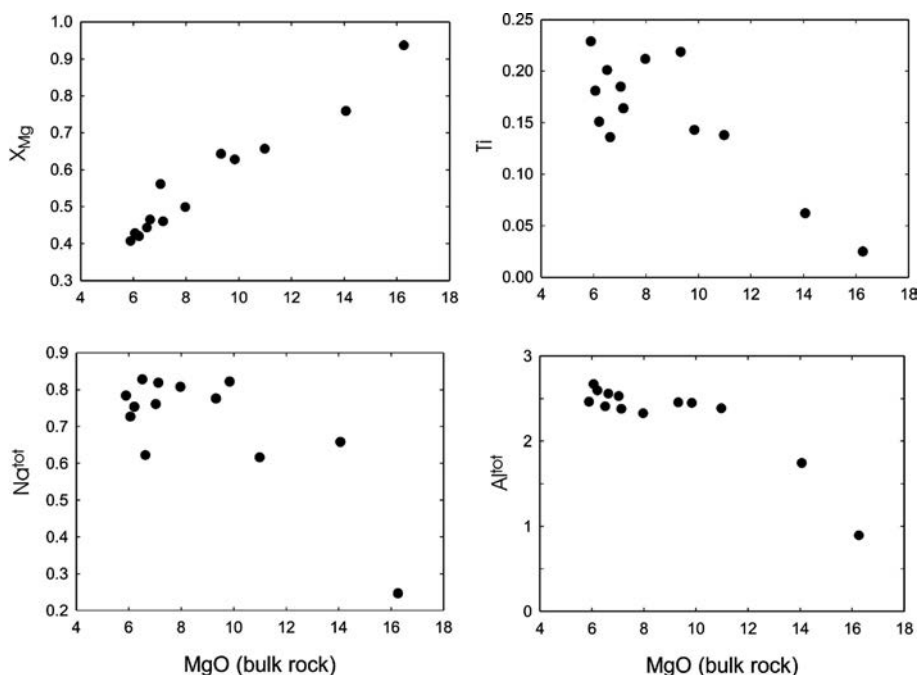


Fig. 4 - X_{Mg} , Ti, Na, Al of the amphiboles (atoms per formula based on 23 oxygen) versus MgO of the bulk rocks. The two samples with the highest MgO represent early cumulates. Only core compositions of the amphiboles are plotted.

Table 2 - Analyses of amphiboles in hornblende from Kerames, Crete

	KER1	JK233	JK241	JK244	JK739	96/35a	96/35b	96/35c	96/35e	96/35g	96/35h	96/38a	96/39	96/40	96/142	96/143	96/144	96/145	97/25
SiO ₂	38.70	39.55	44.69	46.98	40.10	42.67	42.15	52.54	39.90	40.11	42.62	45.75	40.67	46.31	42.25	41.12	39.47	39.43	42.80
TiO ₂	1.62	1.31	1.12	0.57	1.19	1.18	1.08	0.24	1.76	1.43	1.25	0.84	1.47	0.77	1.29	1.88	1.99	1.58	1.99
Al ₂ O ₃	14.10	14.38	13.12	10.27	14.27	13.93	14.52	5.42	13.45	13.25	13.76	12.46	13.52	10.61	14.12	13.18	13.65	14.86	14.26
Cr ₂ O ₃ ^{a)}				0.14				0.17											
FeO ^{tot}	19.69	19.91	12.57	11.33	18.85	15.00	15.15	7.68	20.79	20.20	14.31	11.37	19.93	10.63	13.73	19.09	20.87	19.81	14.27
MnO	0.26	0.41	0.12	0.20	0.52	0.13	0.09	0.24	0.32	0.33	0.22	0.16	0.38	0.21	0.22	0.29	0.32	0.21	0.13
MgO	8.17	6.97	11.97	14.13	7.70	9.78	9.49	18.32	7.26	7.59	11.00	12.70	7.64	14.08	11.26	8.36	6.80	7.00	10.62
CaO	10.20	11.27	11.92	11.20	11.21	11.66	11.94	11.51	10.58	10.79	11.09	12.07	11.29	11.70	11.68	10.58	10.84	11.06	10.42
Na ₂ O	2.58	2.54	1.77	2.36	2.11	1.66	1.52	0.91	2.81	2.77	2.16	1.60	2.23	2.08	2.88	2.78	2.64	2.46	2.74
K ₂ O	0.89	0.94	0.36	0.46	1.63	1.03	1.38	0.23	1.11	0.87	0.95	0.86	1.32	0.43	0.51	1.03	1.26	1.11	0.56
Total	96.21	97.28	97.64	97.64	97.58	97.04	97.32	97.26	97.98	97.34	97.36	97.81	98.45	96.82	97.94	98.31	97.84	97.52	97.79
Si	5.891	6.058	6.495	6.761	6.097	6.361	6.301	7.330	6.063	6.114	6.273	6.635	6.141	6.730	6.219	6.163	6.043	6.008	6.253
Al	2.530	2.596	2.247	1.742	2.557	2.447	2.558	0.891	2.409	2.380	2.387	2.130	2.406	1.817	2.450	2.328	2.463	2.669	2.455
Ti	0.185	0.151	0.122	0.062	0.136	0.132	0.121	0.025	0.201	0.164	0.138	0.092	0.167	0.084	0.143	0.212	0.229	0.181	0.219
Cr				0.015				0.019											
Fe ³⁺	1.056	0.349	0.241	0.400	0.385	0.166	0.070	0.651	0.576	0.552	0.499	0.058	0.416	0.244	0.225	0.520	0.408	0.398	0.458
Mg	1.854	1.592	2.593	3.032	1.745	2.173	2.115	3.810	1.645	1.725	2.413	2.746	1.720	3.050	2.471	1.868	1.552	1.590	2.313
Fe ²⁺	1.450	2.201	1.287	0.963	2.012	1.704	1.824	0.245	2.066	2.022	1.263	1.321	2.101	1.048	1.466	1.873	2.264	2.126	1.286
Mn	0.034	0.053	0.015	0.025	0.067	0.016	0.011	0.028	0.041	0.043	0.027	0.020	0.049	0.026	0.027	0.037	0.041	0.027	0.016
Ca	1.664	1.850	1.856	1.727	1.826	1.862	1.912	1.721	1.722	1.762	1.749	1.875	1.827	1.822	1.842	1.699	1.778	1.806	1.631
Na	0.761	0.754	0.499	0.658	0.622	0.480	0.441	0.246	0.828	0.819	0.616	0.450	0.653	0.586	0.822	0.808	0.784	0.727	0.776
K	0.173	0.184	0.067	0.085	0.316	0.196	0.263	0.041	0.215	0.169	0.178	0.159	0.254	0.080	0.096	0.197	0.246	0.216	0.104
Total	15.598	15.798	15.422	15.470	15.765	15.538	15.616	15.008	15.765	15.750	15.544	15.484	15.734	15.488	15.760	15.704	15.808	15.748	15.512
X _{Mg} ^{b)}	0.56	0.42	0.67	0.76	0.47	0.56	0.54	0.94	0.44	0.46	0.66	0.68	0.45	0.74	0.63	0.50	0.41	0.43	0.64
name ^{c)}	mg-ha	fe-par	mg-hb	mg-hb	fe-par	par	par	mg-hb	hast	hast	par	mg-hb	fe-par	mg-hb	par	hast	fe-par	fe-par	par

Averages of 10-30 analyses from amphibole cores. Formula calculation and stoichiometric Fe³⁺ estimation on the basis of 23 oxygens and 13 cations excluding Ca, Na, and K. F and Cl < 100 ppm. a) Cr₂O₃ concentration below 0.02 wt% except for two amphiboles (JK244 and 96/35c) from samples representing early cumulates. b) X_{Mg} = Mg/(Fe²⁺ + Mg); c) classification after Leake et al. (1997): mg-ha - magnesiohastingsite; fe-par - ferropargasite; mg-hb - magnesiohornblende; par - pargasite; hast - hastingsite.

ement; natural and synthetic standards; a "PAP" matrix correction according to Pouchou and Pichoir, 1991).

In this section only magmatic amphiboles occurring in hornblendites generally unaffected by the metamorphic overprint are considered. In addition, cores of magmatic amphiboles from metamorphosed rocks, which can easily be detected by microscope or microprobe were taken into account.

Most of the analysed amphiboles are pargasites with $\text{Na}+\text{K} > 0.5$ in the A-site and with Al^{VI} between 1.6 and 2 (Fig. 3; Table 2). Some amphiboles with typical pale green colours in thin section are magnesiohornblendes with high contents of a tschermakite component. The relevant rocks containing the magnesiohornblendes are free of accessory minerals like ilmenite or apatite. Their chemical compositions with high concentrations of Mg, Cr, and Ni as well as low concentrations of Ti, P, and REE indicate an origin as early cumulates. This is supported by distinct chromium contents of the magnesiohornblendes in these rocks (Table 2). By contrast, the Cr_2O_3 concentrations in pargasites of the common hornblendites rich in accessory minerals are below the detection limit of the microprobe. This feature, the high Fe concentrations (up to 21 wt% FeO^{tot}), and inclusions of zircons within the pargasites of many hornblendites indicate that the amphibole accumulation that produced these rocks was related to an evolved magma, rich in Fe and obviously saturated in Zr.

In Fig. 4, some chemical parameters of the amphiboles are plotted versus the MgO whole-rock concentration. The X_{Mg} -values for the amphiboles are strongly correlated with the whole-rock MgO . The two amphiboles with the highest X_{Mg} are found in the MgO -richest rock samples which are interpreted as early cumulates. Only poor correlations can be observed for Ti, Na, and Al versus MgO (Fig. 4), although the corresponding plots for the whole-rock compositions define reasonably good trends.

Niida and Green (1999) and Ernst and Liou (1998) investigated the dependence of the concentration of some components in pargasites on temperature and pressure. In the following, their findings were applied to the pargasites of this study.

Al: The Cretan pargasites show remarkably constant Al^{tot} -values with an average of 2.5 atoms per formula unit (a.p.f.u.; Fig. 4), in spite of a great variation in the Al_2O_3 whole-rock composition (~8 – 17 wt%). According to Ernst and Liou (1998) the Al^{tot} content of the pargasites is a function of both pressure and temperature, according to Niida and Green (1999) mainly a function of temperature. This is in contrast with the common hornblende barometers which are based on the pressure-dependent aluminum incorporation into the hornblendes. Despite this disagreement, we interpret the constant Al^{tot} values of the pargasites as an indication for more or less constant P-T conditions during their magmatic formation. Using the semiquantitative amphibole thermobarometer of Ernst and Liou (1998), most of the Cretan pargasites reveal magmatic temperatures around 1000°C at assumed low pressures due to the formation in the shallow level of the oceanic crust. The application of common hornblende barometers (e.g., Schmidt, 1992) to the Cretan hornblendites seems to be inappropriate, since Sato et al. (1999) have experimentally proved that their use with respect to pargasites leads to incorrect pressure values.

Na: Niida and Green (1999) have shown that the sodium content of pargasite is a function of pressure and temperature. The Cretan pargasites vary only slightly in Na with a mean value of ~0.8 Na^{tot} a.p.f.u. for most of them (Fig. 4). We interpret the constant Na contents of the pargasites, de-

spite a great variation in the Na_2O concentration of the whole rocks (1.0 – 4.4 wt%), as an indication for more or less constant P-T conditions during their magmatic formation.

Ti: According to Niida and Green (1999) and Ernst and Liou (1998) the titanium content in pargasite depends mainly on temperature. Ti concentrations in the Cretan pargasites vary considerably, correlating neither with MgO (Fig. 4) nor with the bulk TiO_2 content. This is ascribed to a complex buffering role of ilmenite and magnetite, probably due to varying oxygen fugacities and temperatures during differentiation which have a strong influence on the stability of these phases (e.g., Toplis and Carroll, 1995). The semiquantitative Ti-thermometer of Ernst and Liou (1998) for amphiboles coexisting with a Ti-phase yields a broad range of temperatures from 650° to 850°C which seems to be too low for the magmatic stage. There are several reasons for this discrepancy: (1) loss of Ti by exsolution of a titanium phase during the high-temperature shearing (formation of titanite); (2) compositional differences between the experimentally investigated system (MORB) and the Cretan hornblendites; (3) significant differences between the X_{Mg} of the amphiboles used in experiments and those of the Cretan hornblendites; (4) different oxygen fugacities between the laboratory system and the natural rocks; (5) inappropriate use of the thermometer due to the probable presence of high amounts of melt during the formation of the Cretan amphiboles (for details see Ernst and Liou, 1998).

In contrast to the cores, the rims of the pargasites vary considerably in Al^{tot} . Their contents of Al^{tot} , Na, and Ti are significantly lower than those of the cores. According to Niida and Green (1999) and Ernst and Liou (1998) this is interpreted as the result of re-equilibration at lower temperatures.

Tribuzio et al. (1995) reported the mineral chemistry of pargasites from gabbros and oxide gabbros within Apennine ophiolites. There, the pargasites show high F/Cl ratios (F up to 1700 ppm, Cl up to 113 ppm), that are attributed to the presence of an igneous silicate liquid. On the other hand, hornblendes within these rocks show distinctly lower F and higher Cl contents, indicating a fluid phase contaminated by seawater. The F and Cl concentration of the amphiboles in the Cretan hornblendites are below the detection limit of the microprobe (cores and rims < 100 ppm).

BULK-ROCK GEOCHEMISTRY

Major and trace elements

In Table 3 bulk-rock analyses of the Cretan hornblendites are given. Two thirds of the analysed samples ($n = 28$) have SiO_2 contents below 45 wt% and high concentrations of iron (up to 21.9 wt% $\text{Fe}_2\text{O}_3^{\text{tot}}$) and titanium (up to 5.33 wt% TiO_2). In the AFM diagram (Irvine and Baragar, 1971) and in the $\text{FeO}/(\text{FeO}+\text{MgO})$ diagram (Miyashiro, 1974) the Cretan hornblendites show a distinct tholeiitic trend. In Fig. 5 some major and trace elements of the Cretan hornblendites are plotted against $\text{FeO}^{\text{tot}}/(\text{FeO}^{\text{tot}}+\text{MgO})$. The rocks are characterised by well-defined positive correlations for TiO_2 , P_2O_5 , Zr, Y and negative correlations for Al_2O_3 and Ni indicating typical magmatic differentiation trends. With increasing differentiation the rocks show a strong increase of Ti and Fe. This is characteristic of the oxide gabbros that often are found within the plutonic sequence of the oceanic crust.

Since the SiO_2 content of the hornblendites are in general very low, we assume that most of these rock represent cumulates rather than frozen liquids, and that the observed differentiation trends are dominated by the accumulation of

Table 3 - Whole-rock analyses of hornblendites from Kerames, Crete

	KER 1	JK 219	JK 234	JK 241	JK 244	JK 736	96/35c	96/35d	96/35e	96/35g	96/35h	96/35 i	96/35j	96/142
type ^{a)}	mon+	amph	amph	amph	mon-	amph	mon-	mon+	mon+	amph	mon+	mon+	mon+	mon+
SiO ₂	34.10	43.02	46.26	45.3	46.92	38.13	50.52	37.60	37.98	40.40	42.27	37.24	39.90	42.95
TiO ₂	5.33	1.15	1.55	2.39	0.64	2.38	0.31	5.04	4.71	3.39	1.62	4.34	2.65	1.97
Al ₂ O ₃	12.32	16.60	15.22	13.53	10.01	16.07	7.85	11.99	12.23	12.04	13.57	12.57	13.35	14.71
Fe ₂ O ₃ ^{tot}	21.90	10.79	11.55	13.67	11.19	17.51	9.07	21.00	20.75	20.20	15.19	21.83	20.08	13.62
MnO	0.35	0.17	0.17	0.2	0.21	0.47	0.24	0.30	0.29	0.27	0.22	0.35	0.29	0.18
MgO	7.03	8.11	9.38	5.21	14.06	6.24	16.26	6.47	6.51	7.13	10.98	7.21	8.64	9.84
CaO	12.34	15.45	10.44	13.34	11.87	13.86	12.07	12.68	12.68	12.00	11.47	11.23	10.51	10.93
Na ₂ O	2.13	1.16	2.49	2.25	1.92	1.32	0.97	2.26	2.26	2.53	1.71	2.37	2.37	2.87
K ₂ O	0.79	0.60	0.92	1.98	0.68	1.10	0.59	1.05	1.07	0.88	1.05	1.10	1.08	0.93
P ₂ O ₅	2.15	0.16	0.18	0.34	0.02	0.33	0.14	0.73	0.66	0.42	0.45	0.71	0.25	0.23
L.O.I	0.84	2.08	1.30	0.83	1.49	1.39	1.94	0.62	0.66	0.67	1.61	0.68	0.81	1.86
Total	99.28	99.29	99.46	99.04	99.01	98.80	99.96	99.74	99.80	99.93	100.14	99.63	99.93	100.09
^{b)} FeO ⁺ / (FeO ⁺ +MgO)	0.61	0.40	0.38	0.57	0.29	0.59	0.22	0.62	0.62	0.59	0.41	0.60	0.54	0.41
Sc	31	29	28	31	26	59	16	41	44	43	49	44	40	30
V	318	226	204	297	159	272	91	469	466	431	290	400	410	272
Cr	5	289	488	36	1044	519	1744	100	65	48	594	40	16	329
Co	42	41	37	29	43	58	56	44	35	42	84	44	48	75
Ni	b.d.l.	150	299	29	534	213	664	67	40	30	231	30	22	214
Zn	175	95	89	114	112	177	52	151	157	176	132	152	122	96
Ga	26	36	34	29	b.d.l.	36	9	22	24	26	19	22	22	18
Rb	5	10	14	51	8	10	17	9	5	10	20	7	8	11
Sr	226	423	262	260	40	207	18	96	113	90	61	89	69	193
Y	72	25	21	35	9	65	10	93	89	60	32	69	55	26
Zr	508	51	86	166	64	114	33	313	322	242	131	361	144	138
Nb	58	b.d.l.	8	21	7	22	b.d.l.	27	31	30	8	12	b.d.l.	10
Ba	41	89	93	190	72	55	60	13	24	111	63	43	47	133
La	90.5	4.5	9.57	19.6	1.59	22.7	2.37	23.5	22.3	21.1	61.0	13.8	6.78	13.2
Ce	193	11.2	22.0	42.7	4.51	49.5	5.72	63.8	61.9	48.9	124	39.3	22.5	29.9
Nd	103	9.07	14.0	24.7	3.66	29.5	4.23	45.9	45.3	31.7	51.9	32.3	21.4	19.2
Sm	20.1	2.51	3.50	5.69	1.09	7.25	1.08	11.7	11.2	7.87	8.66	8.95	6.14	4.75
Eu	6.17	1.01	1.29	1.94	0.36	2.91	0.65	4.05	3.92	2.70	2.32	2.95	2.36	1.59
Gd	16.0	2.88	3.36	5.50	0.95	8.34	1.10	13.1	12.8	8.27	6.11	10.2	7.40	4.44
Dy	14.3	3.94	3.73	6.61	1.45	10.2	1.53	15.5	14.8	10.0	6.11	11.9	9.26	4.80
Er	6.26	2.65	2.13	3.46	1.01	5.97	1.04	9.19	8.85	5.85	2.87	6.87	5.58	2.54
Yb	5.37	2.41	1.85	3.16	0.84	5.16	1.07	8.87	8.65	5.49	2.61	6.33	5.2	2.23
Lu	0.69	0.33	0.25	0.43	0.11	0.68	0.15	1.22	1.20	0.74	0.32	0.88	0.70	0.30
ΣREE	455.4	40.5	61.7	113.8	15.6	142.2	18.9	196.8	190.9	142.6	265.9	133.5	87.3	83.0

pargasite. The two samples (JK 244, 96/35c) with the highest concentrations of MgO, Ni and Cr and the lowest contents of Al₂O₃ and TiO₂ are interpreted as early cumulates.

For most major and trace elements, the observed trends are similar to those of gabbros and oxide gabbros from Apennine ophiolites (Tiepolo et al., 1997) that are plotted in Fig. 5. According to Tiepolo et al. (1997), and the differentiation was controlled by continuous fractionation of olivine, plagioclase, clinopyroxene, and Ti-oxides (+ apatite + pargasite). In contrast, the fractionation trends of the Cretan

hornblendites were obviously controlled mainly by pargasite. While the Apennine oxide gabbros show opposite trends for CaO and Na₂O indicating the fractionation of clinopyroxene and plagioclase, no such relationship is observed for the Cretan hornblendites, which is in accord with the rarity of magmatic clinopyroxene and plagioclase in these rocks. In the case of Na₂O, however, secondary alteration processes could have obscured the primary trends, which is also indicated for the mobile trace elements Rb and Ba. Within the Apennine oxide gabbros, the formation of

Table 3 (continued)

	96/143	96/144	96/145	96/146	97/20	97/21	97/25	97/26	97/27	97/97	97/98	97/99	97/100	97/10
type ^{a)}	mon+	mon+	mon+	mon+	mon-	amph	mon+	amph	amph	amph	amph	amph	amph	mon-
SiO ₂	39.46	36.43	37.10	39.05	43.80	46.61	40.11	43.35	41.40	48.33	45.32	47.25	46.14	44.8
TiO ₂	2.72	4.82	4.48	2.68	1.30	2.10	3.09	1.36	1.94	1.64	2.19	2.10	1.34	3.8
Al ₂ O ₃	13.00	12.07	13.55	13.78	14.11	13.74	15.50	15.09	13.37	14.35	14.08	14.14	15.39	14.0
Fe ₂ O ₃ ^{tot}	21.02	19.93	18.87	20.28	10.15	12.50	17.33	9.85	12.08	10.69	12.86	12.60	10.79	14.6
MnO	0.29	0.28	0.28	0.27	0.18	0.19	0.22	0.17	0.23	0.18	0.21	0.21	0.17	0.2
MgO	7.97	5.89	6.06	7.81	12.82	6.51	9.32	11.38	7.72	6.89	7.30	6.46	10.16	4.0
CaO	10.61	13.93	13.96	11.08	12.67	11.51	8.22	13.37	17.53	11.52	11.98	11.02	10.13	10.2
Na ₂ O	2.37	2.11	1.86	2.27	1.88	3.57	2.63	1.75	1.21	3.53	3.03	3.73	2.88	4.3
K ₂ O	1.18	1.20	0.99	1.17	0.33	0.37	0.87	0.26	0.30	0.73	0.66	0.53	0.53	0.7
P ₂ O ₅	0.24	2.05	1.44	0.27	0.18	0.22	0.31	0.11	0.21	0.18	0.22	0.21	0.15	1.2
L.O.I	1.12	0.86	1.00	1.01	1.88	2.02	1.85	2.29	2.46	1.36	1.34	1.22	1.76	1.0
Total	99.98	99.57	99.59	99.67	99.30	99.34	99.45	98.98	98.45	99.40	99.19	99.47	99.44	99.3
^{b)} FeO ⁺ / (FeO ⁺ +MgO)	0.57	0.63	0.61	0.57	0.29	0.49	0.48	0.30	0.44	0.44	0.47	0.50	0.35	0.6
Sc	49	31	34	49	23	35	39	27	39	36	36	36	30	2
V	46	309	329	471	173	288	392	192	271	243	296	291	212	22
Cr	10	5	39	10	463	56	219	453	60	118	69	70	428	
Co	64	37	49	62	59	21	33	22	22	23	18	25	53	
Ni	78	77	29	11	295	29	122	307	33	49	34	27	275	b.d.l
Zn	125	197	151	107	62	101	127	68	92	78	97	97	71	13
Ga	21	27	24	23	9	40	18	33	26	36	41	40	8	1
Rb	2	4	3	2	4	10	7	8	7	14	14	8	6	
Sr	61	176	194	67	53	135	208	40	51	217	134	172	168	11
Y	53	82	72	52	24	40	44	25	36	29	40	39	25	5
Zr	108	552	467	151	117	134	195	91	129	98	148	133	104	42
Nb	7	82	65	7	10	4	21	4	8	b.d.l.	6	b.d.l.	10	6
Ba	60	118	106	61	29	47	79	8	46	102	61	107	53	7
La	7.41	92.6	69.4	6.39	6.15	8.15	10.5	5.20	7.44	6.90	9.40	7.54	4.42	58.
Ce	23.9	199	151	21.0	16.1	21.8	29.6	13.2	19.9	17.8	24.4	20.4	12.2	12
Nd	21.1	108	84.0	19.5	11.7	16.7	22.8	9.85	15.5	13.2	18.4	16.3	10.1	72.
Sm	5.98	21.3	17.1	5.67	3.03	4.55	6.04	2.21	4.18	3.56	4.81	4.46	2.85	14.
Eu	2.22	6.83	5.56	2.06	1.16	1.74	2.11	1.11	1.58	1.39	1.86	1.73	1.19	4.7
Gd	7.05	17.4	14.2	6.82	3.23	5.31	6.71	2.59	4.91	4.07	5.60	5.19	3.30	11.
Dy	8.89	15.9	13.6	8.68	4.00	6.50	7.78	3.99	5.98	4.89	6.77	6.48	4.21	11.
Er	5.39	7.02	6.44	5.32	2.60	4.05	4.50	2.27	3.69	2.99	4.19	4.03	2.60	5.2
Yb	5.10	5.99	5.86	4.93	2.43	3.86	4.14	2.47	3.45	2.75	3.94	3.79	2.40	4.6
Lu	0.69	0.76	0.78	0.68	0.34	0.54	0.56	0.31	0.48	0.37	0.54	0.52	0.32	0.6
ΣREE	87.7	474.8	367.9	81.1	50.7	73.2	94.7	43.2	67.1	57.9	79.9	70.4	43.6	311.

Major elements (wt%) and trace elements (ppm) analyzed with XRF, REE including Y with ICP-AES; b.d.l., below detection limit.

a) Petrographic characterization: mon+, monomineralic amphibole schist rich in accessory minerals; mon-, monomineralic amphibole schist poor in accessory minerals; amph, amphibolite-type, often enriched in granoblastic epidote/clinozoisite ± albite ± garnet ± diopside.

b) FeO* corresponds to FeO^{tot}; molar values.

early cumulates was controlled by the fractional crystallization of olivine (highest Ni and lowest CaO concentrations at lowest FeO^{tot}/(FeO^{tot}+MgO); see Fig. 5). In the Cretan hornblendites, however, amphibole dominated the early stage of differentiation indicated by consistently high CaO concentrations even at low FeO^{tot}/(FeO^{tot}+MgO). High Cr contents

of the amphiboles from early cumulate samples support this view. As in the Apennine oxide gabbros, the observed trends for TiO₂, V (identical to TiO₂; not shown in Fig. 5), P₂O₅, Y, and Zr are attributed to the accumulation of the accessory phases ilmenite, apatite, and zircon which in the more evolved rocks are significantly enriched.

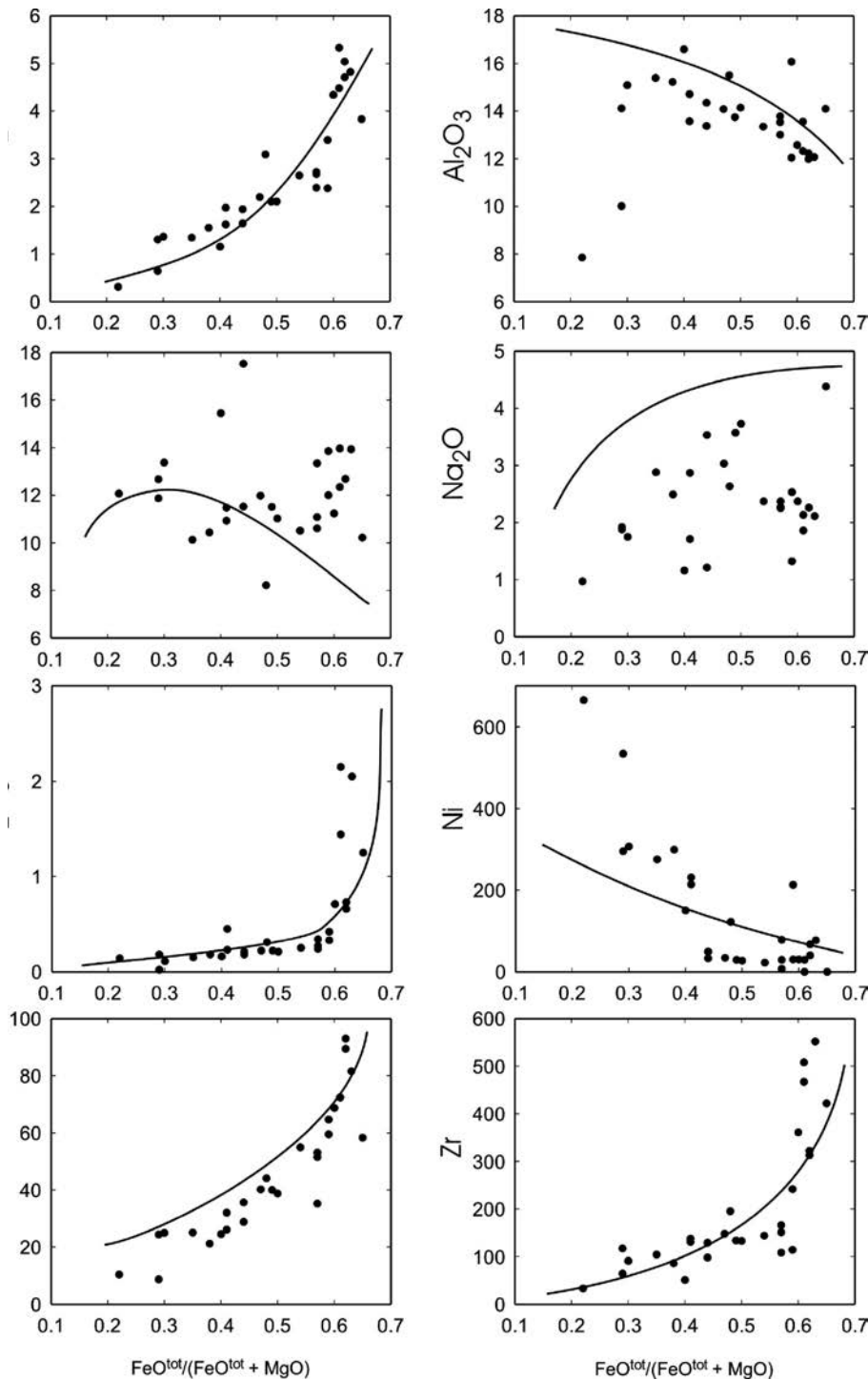


Fig. 5 - Selected major (wt%) and trace elements (ppm) against $\text{FeO}^{\text{tot}}/(\text{FeO}^{\text{tot}} + \text{MgO})$ (molar) for the Cretan hornblendites. The lines represent the trends of a suite of oxide gabbros from the Apennine ophiolites, after Tiepolo et al. (1997).

Rare earth elements

The total REE contents of the Cretan hornblendites vary considerably between 16 and 475 ppm (Table 3). The chondrite-normalised REE patterns of these rocks (Fig. 6a) can be divided broadly into three distinct groups as a function of $\text{FeO}^{\text{tot}}/(\text{FeO}^{\text{tot}} + \text{MgO})$ (Fig. 6b). Most of the rocks (21 samples) representing the main stage of differentiation show flat patterns with weak or no enrichment of the LREE (La_N/Yb_N 0.9 - 4.2). In contrast, four samples, which are highly fractionated ($\text{FeO}^{\text{tot}}/(\text{FeO}^{\text{tot}} + \text{MgO}) = 0.61 - 0.65$) and very rich in REE (SREE 312 - 475 ppm), have patterns characterised by a strong enrichment of the LREE (La_N/Yb_N 8.0 - 11.4).

The two samples regarded as early cumulates

($\text{FeO}^{\text{tot}}/(\text{FeO}^{\text{tot}} + \text{MgO}) = 0.22$ and 0.29) have low REE concentrations (SREE 19 and 16 ppm) and exhibit only very flat REE patterns (La_N/Yb_N 1.5 and 1.3). The more differentiated rocks show no or slightly positive Eu anomalies, while the less differentiated rocks have generally positive Eu anomalies ($\text{Eu}/\text{Eu}^* = 1.13 - 1.42$). This indicates that plagioclase fractionation was involved in the earlier stages of differentiation which is compatible with the presence of some relict plagioclase in the less differentiated rocks. The sample with the strongest cumulate character ($\text{FeO}^{\text{tot}}/(\text{FeO}^{\text{tot}} + \text{MgO}) = 0.22$, Cr = 1744 ppm, Ni = 664 ppm) shows the most pronounced positive Eu anomaly due to the presence of some grains of plagioclase (< 5 % in the mode), now altered to albite.

Since amphibole prefers the middle REE during crystallization (summarised in Rollinson, 1996), a convex shape of the REE patterns would be expected, because most of the investigated rocks are monomineralic amphibole schists. However, the Cretan hornblendites show patterns with LREE enrichment.

In Fig. 6b, the REE patterns of two representative oxide gabbros from the Apennine ophiolites (Tiepolo et al., 1997) which plot within the field of most of the Cretan hornblendites are shown. They display an enrichment in the middle REE which was ascribed to the fractionation of apatite incorporating the middle REE during crystallization (Fujimaki, 1986). In contrast, the Cretan hornblendites do not show this behavior even in those samples which are rich in apatite (up to 2.15 wt% P_2O_5) indicating a different fractionation mechanism. The observed enrichment of the LREE could also be generated by the fractionation of garnet. However, this mechanism can be excluded since there is no indication for a pressure sufficiently high to stabilize garnet in equilibrium with basaltic melt. The garnets in some hornblendites strongly affected by the later low-grade metamorphism are clearly of metamorphic origin. The presence of primary clinopyroxene, plagioclase, ilmenite, and zircon cannot explain the observed LREE enrichment since these phases do not favor the LREE during crystallization. We assume that the observed trends are due to the accumulation of monazite which prefers the incorporation of the LREE in a late stage of the cumulate-forming process. This is indicated by (1) a well-defined correlation between La and Ce with P_2O_5 and a very weak correlation between Y and Yb with P_2O_5 , and (2) the presence of rare tiny monazites in highly differentiated ($FeO^{tot}/(FeO^{tot}+MgO) = 0.61 - 0.65$) samples. A few samples with exceptionally high amounts of Ce (e.g., sample KER1, 193 ppm Ce) bear a late-stage crystallised allanite showing complex Ce zoning in backscattered electron images.

DISCUSSION

Alternative models for the genesis of the Cretan hornblendites

Our results based on the major and trace element geochemistry clearly point to a cumulate origin of the Cretan hornblendites, generated from a iron-rich tholeiitic source. Since the contacts between hornblendites and serpentinites are heavily tectonised, we can not exclude that they are of tectonic nature. This may imply alternative models for the generation of the Cretan hornblendites, which we want to discuss here briefly, before focusing on our preferred model (next chapter).

Metamorphic sole: Metamorphic soles are formed during the detachment of paleo-ridges in an initial state of obduction, when the still hot oceanic lithosphere overrides adjacent oceanic crust (summarised in Nicolas, 1989). Thus, typical protoliths of such metamorphic soles would be oceanic MORB-type basalts and oceanic sediments. If the Cretan hornblendites are regarded as results of a metamorphism, their source rocks must be oxide gabbros. Such rocks, however, are normally not located in the upper part of the oceanic crust. A potential metamorphism of variegated oxide gabbros from typical oceanic crust containing olivine, plagioclase, clinopyroxene, and Fe-Ti oxides should result in the formation of amphibolites, and not of massive monomineralic hornblendites like the Cretan ones. In addition, one would

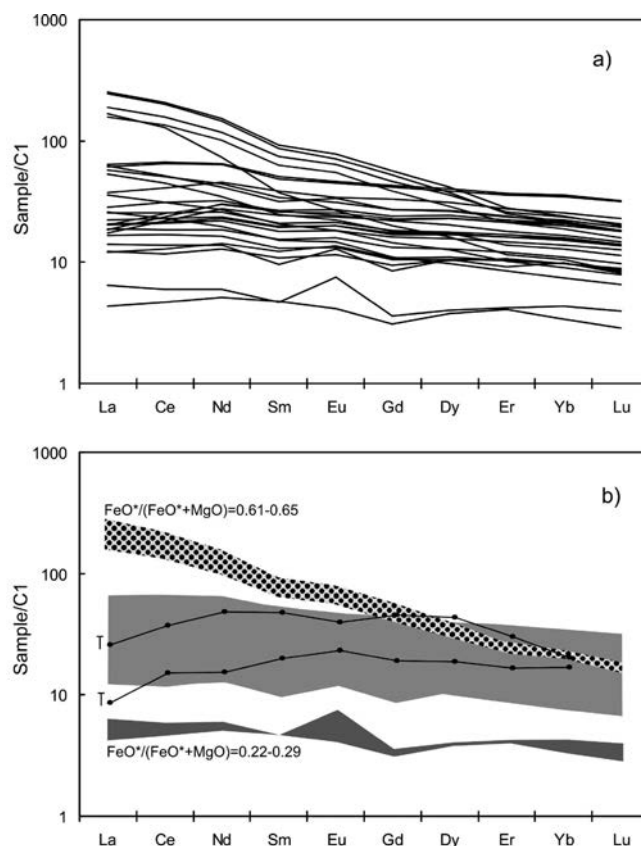


Fig. 6 - Chondrite-normalized REE patterns for the Cretan hornblendites. (a) all samples; (b) grouping as a function of the grade of differentiation expressed by $FeO^{tot}/(FeO^{tot} + MgO)$; see text for details. T: representative REE patterns for oxide gabbros from Apennine ophiolites after Tiepolo et al., (1997). Normalizing chondrite is from Taylor and McLennan (1985; Appendix 2).

expect further rock types derived from protoliths forming the present-day layer 1 of the oceanic crust, which is dominated by pelitic or calcareous sediments. Metapelites are lacking within the Cretan hornblendite suite, but are present in the metamorphic soles of many ophiolites worldwide (e.g., Gnos and Peters, 1993). A further strong argument against this model is the presence of cumulate structures in the Cretan hornblendites (Fig. 2a), which can hardly be generated by a prograde metamorphism. Moreover, typical rocks of the metamorphic sole of ophiolites often show a strong and reverse metamorphic zoning (Gnos and Peters, 1993; Hacker and Mosenfelder, 1996) which is not shown by the Cretan hornblendites. Briefly summarised, it is very unlikely that the Cretan hornblendites represent parts of a metamorphic sole.

Assignment to the crystalline rocks of the "Preveli subunit": West of the hornblendite complex, a further metamorphic rock unit consisting of blueschist-facies metabasalts and metasediments is exposed, which belongs to the ophiolitic mélange of the Uppermost Unit of Crete: "Vatos series" according to Bonneau and Lys (1978) or "Preveli subunit" according to Thomson et al. (1998, see introduction and Fig. 1). The contacts between these rocks and the hornblendites are of tectonic nature. Since both rock units show signs of low-temperature metamorphism (in a broad sense) and contain basic components it may be suggested that they constitute a single unit. Arguments against this hypothesis are: (1) The crystalline rocks of the Preveli subunit are regionally metamorphosed at high pressure as result from the presence of glaucophane (Koepeke et al.,

1997), while the hornblendites do not show any sign of high-pressure metamorphism. (2) Both crystalline units show marked differences in their style of deformation. (3) Ferrogabbroic cumulate rocks are not present within the high-pressure metamorphic unit, neither in the rocks of the Preveli subunit adjacent to the hornblendites nor in equivalent series from Gavdos or Central Crete.

Part of the gabbro-dioritic dike intrusions: Since both the hornblendites and the gabbro diorite dikes form intrusions within the lherzolites of the Kerames area, it may be suggested that these rocks are genetically related. However, there are several pieces of evidence against this assumption: (1) K-Ar dating yielded a significant difference in the age of formation (Koepke et al., 2002). (2) The dike rocks never show signs of high-temperature shearing. (3) At the contact between hornblendites and peridotites intense rodingitization developed, whereas none of the numerous gabbro diorite dikes show signs of rodingitization. This indicates that the phase of ocean floor alteration resulting in serpentinites and rodingites was prior to the dike intrusion event.

Preferred model: Accumulation of amphibole in an iron-enriched MORB

Major and trace element geochemistry reveals that the Cretan hornblendites result from the fractional crystallization and accumulation of amphibole from an iron-enriched, evolved tholeiite magma, probably derived from a MORB source. Since the magma chambers within the oceanic lithosphere are located in a shallow crustal level, low confining pressures (500 to 2000 MPa) can be assumed for the genesis of the hornblendites. The average bulk composition calculated from all analysed samples shows a very low SiO_2 content of 43.0 wt%. Compared to other iron-rich tholeiitic systems, in which ferrogabbroic or ferrobasaltic derivatives were produced by differentiation of olivine and clinopyroxene (e.g. Galapagos spreading center, Juster et al., 1989), the Cretan hornblendites are strongly impoverished in SiO_2 . This is the result of high amounts of cumulus pargasite, which is very low in SiO_2 (~40 wt%, see Table 2) and thus drives those magmas with a high amount of cumulate to very low silica contents.

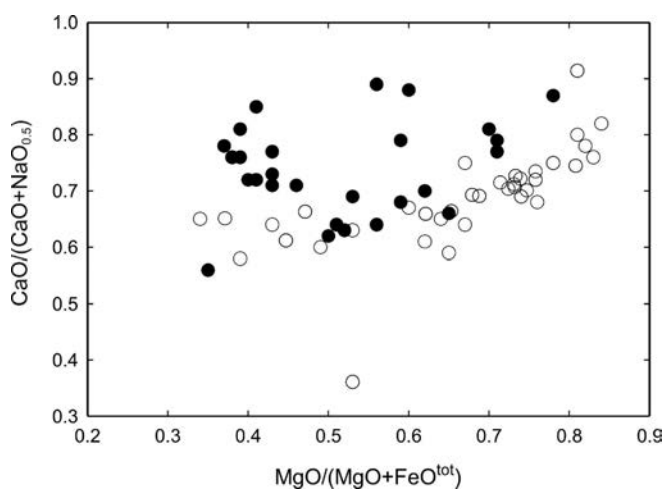


Fig. 7 - $\text{MgO}/(\text{MgO} + \text{FeO}^{\text{tot}})$ versus $\text{CaO}/(\text{CaO} + \text{NaO}_{0.5})$ (molar) for the Cretan hornblendites (filled circles). For comparison, data points for oceanic gabbros including olivine gabbros, olivine clinopyroxene gabbros, and oxide gabbros from the Southwest Indian Ridge (ODP hole 735B) are plotted (Online Background Dataset of Dick et al., 2000, open circles).

In the $\text{MgO}/(\text{MgO} + \text{FeO}^{\text{tot}})$ versus $\text{CaO}/(\text{CaO} + \text{NaO}_{0.5})$ diagram (Fig. 7) the Cretan hornblendites are plotted together with oceanic gabbros from the Southwest Indian Ridge (ODP hole 735B, Dick et al., 2000). While the latter show a trend typical for oceanic gabbros expressed by a slight positive slope due to the co-crystallization of olivine, plagioclase and clinopyroxene, the hornblendites from Crete form a horizontal trend parallel to $\text{MgO}/(\text{MgO} + \text{FeO}^{\text{tot}})$ which is a consequence of their somewhat uncommon differentiation trend dominated by the accumulation almost exclusively of pargasite.

The dominance of cumulative amphibole within the Cretan hornblendites is a striking feature when compared with typical oxide-rich gabbros from the plutonic layer of the oceanic crust. The latter bear beside olivine, plagioclase, and clinopyroxene, that represent cumulate phases of the main stage of crystallization in typical MORB magmas, high amounts of late-stage magnetite and ilmenite (Lippard et al., 1986; Ozawa et al., 1991; Dick et al., 2000). Recently, Niu et al. (2002) stated that the Fe-Ti oxides of oxide-rich gabbros from the Southwest Indian ridge are not in chemical equilibrium with olivine, clinopyroxene, and plagioclase, but precipitated later at much lower temperature, either from trapped Fe-Ti-rich melts, or were expelled as a result of compaction or filter pressing. On the other hand, within the Cretan hornblendites the high bulk Fe content (up to ~22 wt% $\text{Fe}_2\text{O}_3^{\text{tot}}$) is concentrated in pargasite containing up to ~23 wt% $\text{Fe}_2\text{O}_3^{\text{tot}}$, while the high Ti content of these rocks is incorporated in ilmenite. Magnetite as primary phase is missing at all. Since the geochemical trends suggest a single-stage evolution of these rocks, and since very Fe-rich amphiboles crystallised as liquidus phase (presence of tiny inclusions of zircons), it is very probable that the Cretan hornblendites represent the settled crystallization products of an iron-enriched, late-stage melt, probably not able to erupt due to its high density (Brooks et al., 1991; Natland and Dick, 1996; MacLeod and Yaouancq, 2000). As no chemical gradient with height was observed, we infer that the intrusive complex is not a result of in-situ solidification of a single magma batch, but a product of multiple intrusions of smaller magma volumes with different grades of differentiation, redistributed prior to complete solidification by deformation and compaction.

Constraints from experimental data on iron-rich tholeiite systems

Relevant experimental phase equilibria studies in natural tholeiitic ferrobasaltic systems concerning the iron-enrichment as a function of the ambient oxygen fugacity, are restricted to dry systems (e.g., Juster et al., 1989; Snyder et al., 1993; Toplis and Carroll, 1995). The main conclusions with respect to the generation of iron-rich melts are: (1) Effective iron-enrichment ($\text{Fe}_2\text{O}_3^{\text{tot}} \geq 19$ wt%) is only possible at strongly reducing conditions ($f\text{O}_2 \ll \text{FMQ}$ oxygen buffer) since the magnetite saturation temperature is significantly depressed. Then, iron can increase in the melt instead of being incorporated into magnetite. (2) The more reducing the system, the stronger is the iron enrichment. (3) At reducing conditions ilmenite reaches saturation prior to magnetite ($f\text{O}_2 \ll \text{FMQ}$), but the order is reversed at more oxidizing conditions. (4) The saturation temperatures for the Fe-Ti oxides strongly depend on the bulk composition (e.g., a shift to 30°C higher temperatures close to the saturation of olivine and plagioclase by increasing the iron content of

the system from ~16 to ~19 wt% $\text{Fe}_2\text{O}_3^{\text{tot}}$, Snyder et al., 1993). Direct application of these results to the Cretan hornblendites is hampered, however, by the fact that all these studies do not take into account the role of amphibole during differentiation, which in this case was a major fractionating phase.

In the experimental ferrobasic systems mentioned above, plagioclase is a major phase in the crystallizing assemblage that is stable from near-solidus conditions to the highest temperatures. In contrast, in the Cretan hornblendites plagioclase in general occurs only in small amounts and is absent in the high-differentiated rocks at all. There are several possible reasons for this discrepancy. (1) Amphibole was separated from a plagioclase-bearing iron-rich magma by accumulation at the floor of the magma chamber due to its high density, while the lighter plagioclase stayed in the main magma body, or even floated. Possibly the separation was enhanced by tectonic processes in a late-magmatic stage, since strong plastic deformation was observed (Fig. 2c). (2) A high water activity, which is necessary for the crystallization of amphibole, has the potential to destabilize plagioclase (e.g., Baker and Eggler, 1983, 1987; Danyushevsky, 2001; Berndt, 2002; Koepke et al., 2003). (3) Another possible explanation for the rarity of plagioclase in the Cretan hornblendites consists in the capability of pargasite of incorporating appreciable amounts of major and minor elements such as Al, Ti, Na, K, F, Cl. This eventually makes possible the formation of monomineralic rocks in hydrous ferrobasic systems, whereas in corresponding dry systems, generally, would crystallise olivine, plagioclase and clinopyroxene.

The effect of water on phase equilibria in ferrobasic systems is not yet well constrained experimentally and the question on the role and stability of pargasite in such systems remain to be established. The presence of pargasite in the Cretan hornblendites indicates a maximum crystallization temperature of ~1030°C, since experimental studies in hydrous basaltic systems, performed at pressures < 0.5 GPa, show that pargasite crystallises only at relatively lower temperatures < 1030°C (e.g., Holloway and Burnham, 1972; Kawamoto, 1996). In partial melting experiments on a typical oxide gabbro, performed by Koepke et al. (2003), pargasitic amphibole was stable at 1000°C but became unstable at 1030°C. This is consistent with the estimated crystallization temperatures of interstitial pargasites in natural oxide gabbros which are generally lower than 1000°C (e.g., Larsen and Brooks, 1994; Tribuzio et al., 2000). Experimental phase equilibria studies in hydrous basaltic systems at low pressures involving the crystallization of amphibole show that relatively high water concentrations in the melt, about 4 wt% according to Johnson et al. (1994, and references herein), are necessary to stabilize the amphibole. This is consistent with the estimated water contents in natural liquids leading to the formation of pargasite-bearing oxide gabbros in the Northern Apennines (Tribuzio et al., 2000: 4.4 – 6.1 wt%; from H_2O partitioning data between pargasite and silicate liquid). Since water added to a dry system can have an oxidizing effect (for instance, an increase in $f\text{O}_2$ of two log units by increasing the water activity from 0.1 to 1, Berndt, 2002), it is inferred that the $f\text{O}_2$ of the system before the enrichment of water due to the crystallization of dry silicate phases must have been very low, as much as 2–3 log units below the FMQ buffer. Further support for very reducing conditions is given by the hematite content in ilmenite solid solutions which is a function of oxygen fugacity (at given T,

P). By using the oxygen barometer of Buddington and Lindley (1964), four out of six analysed ilmenites with very low hematite contents (0 – 2.4 mol%) reveal a $f\text{O}_2$ around FMQ – 1–2 (assumed temperature of 1000°C). These values correspond to the oxygen fugacity at a late stage of solidification of the water-enriched melt, since ilmenite crystallised as an interstitial phase between the amphiboles. However, this conclusion should be interpreted with caution since it is well known that T – $f\text{O}_2$ equilibria concerning Fe-Ti oxides could easily reset down to very low temperatures.

The discussion of the available experimental data applied to the formation of the Cretan hornblendites can be briefly summarised as follows: Cumulate pargasite rocks within the ophiolites of Crete crystallised from an iron-rich melt with water contents > 4 wt% at temperatures around 1000°C under reducing conditions (probably at $f\text{O}_2$ defined by the FMQ buffer or lower). Such high amounts of water are uncommon for magmas from normal mid-ocean ridge settings and indicate a subduction-related environment (see next section).

Geotectonic setting

The field relations indicate that the Kerames hornblendites were intruded into the lherzolites included in the dismembered ophiolites of Crete which are linked to the Jurassic ophiolites of the Dinarides/Hellenides (Koepke et al., 2002). The K-Ar dating of amphiboles from the Kerames hornblendites revealed Middle to Late Jurassic ages (Koepke et al., 2002), and SHRIMP-dating of zircons from a hornblendite from Kerames also yielded Jurassic ages (167 Ma ± 2.8; Liati et al., 2004). Therefore, it is very probable that intrusion of the cumulate hornblendites occurred during the formation of the Jurassic oceanic crust in the vicinity of a ridge system, where heat flow and magmatic activity were high. It should be noted that near-ridge gabbroic intrusions into mantle rocks are a common phenomenon, well-known from present-day oceanic crust (e.g., from the East Pacific Rise, Constantin, 1999, or from the Mid-Atlantic Ridge, Cannat and Casey, 1995), and from many ophiolites (summarised in Nicolas, 1989). Such magmatic activity is restricted to a high structural level within the young oceanic crust and often related to transform faults.

The lherzolites associated with the hornblendites at Kerames are characterised by high concentrations of Al_2O_3 and CaO. These rocks represent primitive, undepleted mantle material, which may suggest an origin at a slow-spreading ridge (e.g., Nicolas, 1989; Niu and Hékenian, 1997). A slow-spreading ridge environment is also suggested by the crystal-plastic deformation observed in some hornblendites (Fig. 2c). Such features are regarded as results of ductile deformation of the young, still hot oceanic lithosphere in high-temperature shear zones, probably related to near-ridge transform faults. High-temperature ductile shearing was observed in slow-spreading ridges (Mével, 1988; Mével and Cannat, 1991; Gaggero and Cortesogno, 1997; Cortesogno et al., 2000; Dick et al., 2000), but is absent in fast-spreading ridges (Manning and MacLeod, 1996).

As outlined in the previous section, the generation of magmatic hornblendites requires a melt with high water contents. This is not compatible with generation at a “normal” mid-ocean ridge, where the MORB liquids show low water contents, varying between 0.1 and 0.5 wt% (e.g., Michael and Chase, 1987; Sobolev and Chaussidon, 1996;

Danyushevsky, 2001; Dixon et al., 2002). Thus, it is hardly possible to generate large batches of liquids rich in water by fractionation of anhydrous crystal phases. Much higher water contents are reported from MORBs related to back-arc basins: up to 2 wt% (Danyushevsky et al., 1993; Danyushevsky, 2001) or up to 2.5 wt% (Kovalenko et al., 2000). Here, the crystallization of 50 to 80 % of anhydrous phases would drive the residual liquids in MORB magmas to water contents of several wt%, high enough for the crystallization of amphibole. This aspect strongly suggests a subduction-related environment (e.g., back-arc basin) for the genesis of the Cretan hornblendites.

As mentioned above, about 20 Ma after the hornblendite formation, the Cretan peridotites were intruded by gabbro-dioritic dikes. In these dikes the dominance of hornblende and the geochemical signature indicate a subduction-related origin. These results strongly suggest a complex evolution for the Cretan ophiolite in a slow-spreading ridge system related to the environment of a subduction zone. A supra-subduction zone setting (SSZ) has also been proposed for many well-studied ophiolites world-wide (e.g., Oman, Troodos, Vourinos, Pindos; e.g., Pearce et al., 1984; Bizimis et al., 2000). It should be noted that these ophiolites typically contain extremely depleted harzburgites as mantle tectonites, while the Cretan ophiolites bear fertile lherzolites.

Origin of pargasites in oceanic gabbros

Typical gabbros from mid-oceanic ridges are formed by the accumulation of co-crystallised olivine and plagioclase $\pm$ clinopyroxene, representing the main stage of crystallization in a MORB magma at shallow pressure. Amphibole as a companion crystal phase is lacking, since the temperatures are considerably above the thermal stability of amphiboles, and the water activities in such magmas are much too low (see previous section). Amphiboles occurring in such gabbros are exclusively related to late-stage or postmagmatic metamorphic processes. Their petrographic appearance is characterised by (1) overgrowths around clinopyroxenes, (2) fillings of high-temperature fractures, (3) interstitial clots often locally intergrown with Fe-Ti oxides.

(1) The amphiboles replacing clinopyroxene are mostly of actinolitic composition, and are regarded as products of hydrothermal activity initiated by the infiltration of seawater (e.g., Stakes et al., 1991). (2) High-temperature microscopic veins which often occur in the upper part of the gabbroic section of the oceanic crust are interpreted as due to fluid-rock interaction during the transition between magmatic and hydrothermal processes. Amphiboles of such veins vary in their composition from actinolitic to pargasitic, representing temperatures of formation up to 750°C for gabbros from the fast-spreading East Pacific Rise (Manning and MacLeod, 1996) or up to 1000°C for gabbros from the slow-spreading Southwest Indian Ridge (Maeda et al., 2002). (3) Interstitial amphiboles in oceanic gabbros are very often intergrown with Fe-Ti oxides, indicating a late magmatic evolution, normally decoupled from the gabbro main crystallization phase (see above). Their compositions range from hornblende to pargasites (e.g., Gillis, 1996; Tribuzio et al., 2000; Maeda et al., 2002). Tribuzio et al. (2000) interpreted such amphiboles in gabbros ranging from Mg-rich types to oxide gabbros as magmatic products finalizing the MORB-type crystallization sequence (olivine - plagioclase - clinopyroxene) from a percolating late-stage melt enriched in water and LREE.

The formation of pargasitic amphiboles in typically oceanic gabbros, as mentioned above, contrasts with the features in Cretan hornblendites in some ways. In the Cretan rocks, pargasite is a major crystallizing phase. Inclusions of zircons and monazites in pargasites from the most evolved rocks indicate that amphibole was crystallised from a melt saturated with Zr and some REE.

It should be noted that pargasite-rich rocks (commonly called hornblendites) occur within the mantle rocks, well-known from alpine-type peridotites (e.g., Alpe Arami, Möckel, 1969), in xenoliths (e.g., Kerguelen-Plateau, Grégoire et al., 1998) or in mantle protrusions in oceanic rift systems (e.g., St. Paul's rocks at the Mid-Atlantic Ridge, Melson et al., 1972). The genesis of such rocks is generally attributed to mantle metasomatism in combination with ambient high water activities. Although the Cretan hornblendites are in intimate contact with peridotites, a mantle origin can be excluded for the following reasons: (1) The absence of typical high-pressure phases (e.g., pyrope-rich garnet); (2) The chemical composition of the pargasites, which are significantly lower in TiO_2 , Na_2O , and X_{Mg} compared to the amphiboles in some mantle-derived rocks (e.g., hornblendite xenoliths from the Kerguelen Plateau, Grégoire et al., 1998); (3) A well-defined tholeiitic low-pressure differentiation trend as outlined above.

CONCLUSION

The hornblendites of Kerames in Central Crete represent a suite of cumulates intruded in Jurassic time into lherzolitic peridotites belonging to the Eastern Mediterranean ophiolite. Major and trace element geochemistry reveal a magmatic evolution to extreme iron-rich liquids with high amounts of incompatible elements, consistent with a model of fractional crystallization of magmas with amphibole as dominant phase. The magmas were generated under strongly reducing conditions with water contents > 4 wt% permitting the onset of amphibole crystallization probably at temperatures around 1000°C. The assumed high water contents in the initial melts indicate a subduction-related environment for the formation of the hornblendites.

The petrographic features reveal a complex interplay of magmatic and tectonic events. As evidenced by the textural relationships, after complete solidification, the rocks were affected by high-temperature crystal-plastic deformation, which commonly overprinted the magmatic textures, followed by brittle deformation at lower temperatures. A similar structural development, including high-temperature deformation, has been observed in gabbros from slow-spreading ridges, often in spatial association with transform faults (Ito and Anderson, 1983; Honnorez et al., 1984; Stakes et al., 1991; Cannat and Casey, 1995; Gaggero and Cortesogno, 1997; Cortesogno et al., 2000; Dick et al., 2000; Natland and Dick, 2002). The same style of metamorphic overprint related to high-temperature shear zones was also reported from some ophiolites (e.g., Pedersen and Malpas, 1984; Flagler and Spray, 1991). Thus, it is very probable that the location for the intrusion of the hornblendite crystal mush occurred in a high-temperature shear zone environment probably related to a transform fault in a subduction-related setting.

Geological and stratigraphical studies and correlations with the adjacent ophiolite units are needed to precise the geotectonic environment where these rocks formed.

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