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ORIGIN OF THE ANTIPAROS ISLAND RHYOLITES, GREECE, BY SUBDUCTION-RELATED ANATEXIS OF A GRANULITIC SOURCE POSSIBLY LOCATED IN THE MIDDLE TO LOWER CONTINENTAL CRUST

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ABSTRACT

The high-silica Pliocene rhyolites of the Antiparos island, which is located about 50 km north of the central part of the South Aegean volcanic arc in Greece, have relatively high K and Na contents, are significantly enriched in Rb, Cs, U, Th and Nb and extremely depleted in Ba and Sr. According to the proposed model the rise of hot basic magma, generated by the subduction of the Aegean oceanic slab, near the base of the continental crust underneath Antiparos incited the partial melting of a felsic granulite source possibly located in the middle to lower crust. Subsequently the primary magmas rose to a shallow magma chamber where they evolved by significant fractional crystallisation involving dominant alkali-feldspar and subordinate Ab-rich plagioclase, quartz, titanite and hornblende.

KEY WORDS: rhyolite; continental crust anatexis; subduction; Antiparos; Greece.

1. INTRODUCTION

The region of the Aegean Sea in Greece is an area which has been affected by widespread volcanism, of a divergent geochemical character, from the Tertiary period up to the present. In the Southern Aegean Sea a calc-alkaline volcanic arc started to form around 4 Ma ago and it is still active (Fytikas et al., 1984). This volcanic arc is referred in the literature as the South Aegean or Hellenic volcanic arc.

On the inner side and at the north-western end of the South Aegean volcanic arc a series of small, K-rich volcanic centres were formed during the last 5.4 Ma. These centres are either slightly older (Antiparos rhyolites: 5.4-4 Ma) than or contemporaneous (Volos-N. Evia high-K basaltic andesites and andesites: 3-0.5 Ma) with the South Aegean arc (Fytikas et al., 1984). The Antiparos rhyolites were studied geochemically by Anastopoulos (1963), Innocenti et al. (1982) and Clapsopoulos (1991).This paper examines the nature and origin of the Antiparos rhyolites and their relationship with the tectonics of the area at the time of the eruption, by using new mineralogical and geochemical data in conjunction with previous work.

2. GEOLOGICAL SETTING

Antiparos belongs to the Cyclades group of islands of the South Aegean Sea and has a surface area of approximately 45 km^2. The volcanic outcrops are located in the southern part of Antiparos, extending over an area of about 4.5 km^2 (Fig. 1) and consist of two lava domes with associated lava flows and pyroclastic rocks. The domes are encircled by layered tuffs which are composed of lava fragments integrated with many obsidian or perlitic particles and fine pumice. Sporadically small fragments of the

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metamorphic country rocks are included in the tuffs. According to Anastopoulos (1963) these tuffs are of subaqueous origin (subaqueous pyroclastic flow deposits). Additionally to the layered tuffs outcrops of unlayered, welded tuffs, consisting of angular fragments of spherulitic lava, can also be found within the volcanic rocks.

The volcanic rocks are emplaced on schists of the metamorphic complex of Antiparos. This complex has been described in detail by Anastopoulos (1963) and is believed to be of Paleozoic age. The complex consists of alternating bands of biotite - muscovite gneisses, schists (micaceous, chloritic, talc and epidote-chloritic) and intercalated marbles.

The majority of the faults in Antiparos have a NW-SE to N-S strike (Anastopoulos, 1963; Innocenti et al., 1982), while the crust is continental and approximately 29 km thick (Makris, 1977, 1978).

3. PETROGRAPHY, MINERALOGY AND MINERAL CHEMISTRY

The chemical compositions of primary mineral phases from selected samples were determined by Electron Probe Microanalysis (EPMA). The analyses were obtained by Energy Dispersive Spectrometer (EDS) analysis, using a Modified Cambridge Instrument Company Geoscan microprobe and a Cameca Camebax microprobe fitted with a Link Systems 860-500 E.D.S. detector.

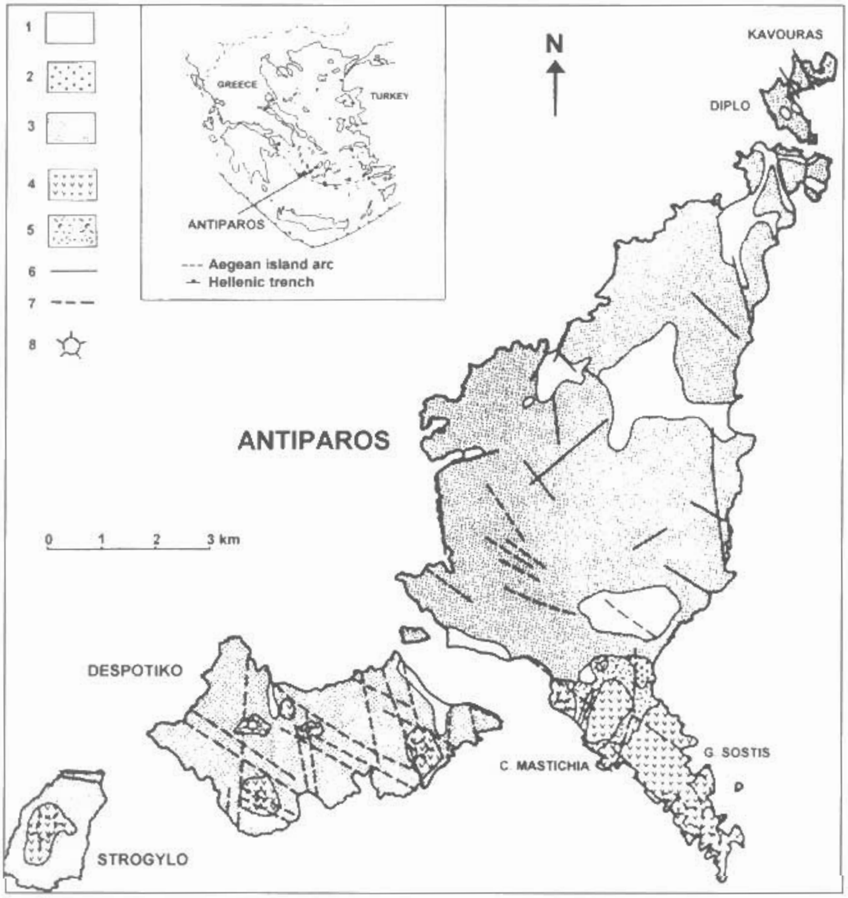

Fig. 1: Geological map of Antiparos (modified after Anastopoulos, 1963 and Innocenti et al., 1982). 1 = Quaternary deposits, 2 = Neogene sediments, 3 = Metamorphic complex, 4 = Rhyolitic lavas, 5 = Pyroclastic deposits, 6 = Faults, 7 = Inferred faults, 8 = Lava domes.

Most of the lavas are massive and aphanitic. The textures of the rocks are mainly vitrophyric to vitreous and occasionally can be hyalopilitic, spherulitic, perlitic, or vesicular. In the majority of the rocks the phenocrysts constitute about 5% of the rock volume and the groundmass about 95%, while some rocks have rare or no phenocrysts at all.

The phenocrysts are alkali feldspars and quartz accompanied by a few opaque grains and traces of biotite and occasionally hornblende. The alkali feldspars occur as euhedral, subhedral or anhedral (mainly) microphenocrysts. In a few samples many of the crystals, usually the larger ones, display a texture of granophyric intergrowth of alkali feldspar with quartz. They are sodic sanidines ranging from Or_{39} to Or_{50}, while their An content is very low ($An_{0-1.6}$). Quartz occurs as small anhedral and embayed phenocrysts. Rare needles of apatite are present as an accessory mineral (mainly as inclusions in phenocrysts of alkali feldspar). In addition, very few zircon crystals were observed. Anastopoulos (1963) and Innocenti et al. (1982) reported the rare occurrence of Ab-rich plagioclase (oligoclase), clinopyroxene and titanite in the Antiparos rhyolites, but no such crystals were found in the studied samples. In a number of samples spherulites consisting of intergrown radiating needles of alkali feldspar and quartz are present. In many samples the groundmass is devitrified to a microcrystalline aggregate of alkali feldspar and quartz or sometimes to a cryptocrystalline felsitic one. Representative microprobe mineral analyses are given in Table 1.

Intergrowths of alkali feldspar with quartz

a) Granophyric intergrowths

Granophyric intergrowths in the Antiparos rocks occur as both phenocryst and groundmass constituents. Two representative microprobe analyses are presented in Table 1. Their bulk normative compositions fall within the range of Q (≈30%), Ab (≈40%) and Or (≈30%), while in terms of their An-Ab-Or contents they overlap with the Antiparos sanidine phenocrysts.

b) Spherulites

The Antiparos lavas are spherulitic, while a representative analysis of a spherulite is given in Table 1. The spherulite has a higher Q (≈38%) content and a lower K_2O/Na_2O ratio than the granophyric intergrowths, while it has a much higher Ab (≈65%) content and a lower K_2O/Na_2O ratio than the sanidine phenocrysts (Table 1). The Antiparos spherulites have chemical compositions similar to the spherulites of the rhyolitic lavas from the Central Volcanic Region in New Zealand which were analysed by Ewart (1971).

Table 1. Representative electron probe analyses of sanidine phenocrysts, granophyric intergrowths of alkali feldspar and quartz and of a spherulite from the Antiparos volcanic rocks. $Fe_2O_3^*$ = total iron as $Fe_2O_3^*$. The CIPW normative compositions are recalculated to 100% Q-Ab-Or and 100% An-Ab-Or respectively.

	SANIDINES				**GRANOPHYRIC INTERGROWTHS**		**SPHERULITE**
Wt.%							
SiO_2	67.57	67.37	66.87		76.00	76.85	78.72
Al_2O_3	18.12	17.63	18.67		13.09	12.98	11.45
$Fe_2O_3^*$	0.00	0.33	0.25		0.32	0.29	0.25
MnO	0.09	0.22	0.00		-	-	-
CaO	0.27	0.16	0.32		0.00	0.20	0.25
Na_2O	6.54	5.37	5.93		4.63	4.91	4.88
K_2O	6.54	8.32	7.91		5.19	4.68	3.57
Total	99.13	99.40	99.95		99.23	99.91	99.12
Numbers of ions on the basis of 8 Oxygens				**Norm % (CIPW)**			
Si	3.030	3.039	3.000	**Q**	29.8	30.8	38.4
Al	0.958	0.937	0.987	**Ab**	39.0	41.2	40.0
Fe*	0.000	0.012	0.009	**Or**	31.2	28.0	21.6
Mn	0.003	0.008	0.000	**Norm % (CIPW)**			
Ca	0.013	0.008	0.016	**An**	0.0	0.0	0.0
Na	0.569	0.470	0.516	**Ab**	55.6	59.5	64.9
K	0.374	0.479	0.453	**Or**	44.4	40.5	35.1
Mol%							
An	1.4	0.8	1.6				
Ab	59.5	49.1	52.4				
Or	39.1	50.1	46.0				

4. GEOCHEMISTRY

Selected samples were analysed for major and trace elements. The major elements were determined by X-Ray Fluorescence (XRF) while the trace elements were determined by XRF (Ba, Rb, Sr, Nb, Zr, V, Ni, Cr, Cu, Zn, La, Ce) Inductively Coupled Plasma Atomic Emission Spectrometry (ICPAES: Pr, Nd, Dy, Y, Lu) and Instrumental Neutron Activation Analysis (INAA: Cs, Th, U, Ta, Hf, Sc, Sm ,Eu, Tb, Yb). Representative rock compositions are given in Table 2.

Major elements

All the rocks are rhyolites and display a restricted range of major element composition (Table 2). They are high in SiO_2, Na_2O, and K_2O/Na_2O ratio and low in MgO, CaO and TiO_2 compared with most rhyolites. The rhyolites are either slightly peraluminous (presence of corundum {c} in the norm and A/CNK > 1.00), or slightly metaluminous (presence of diopside {di} in the norm and A/CNK < 1.00).

Trace elements

The Antiparos rhyolites are enriched in the incompatible elements Rb, Cs, U, Th and Nb and extremely depleted in Ba and Sr relative to rhyolites from the active South Aegean arc. The concentration of Ni, Cr, V and Sc which are compatible with Fe-Mg minerals is low.

Table 2. Chemical compositions and CIPW norms of Antiparos rhyolites. Total iron as $Fe_2O_3^*$.

Sample	**Anp-1**	**Anp-2**	**Anp-3**	**Anp-4a**	**Anp-8**	**Anp-9**
wt.%						
SiO_2	77.58	76.27	77.18	77.47	76.34	77.34
TiO_2	0.11	0.10	0.13	0.11	0.11	0.11
Al_2O_3	12.33	13.28	12.36	12.38	12.91	12.59
$Fe_2O_3^*$	0.83	0.74	0.94	0.85	0.76	0.80
MnO	0.03	0.03	0.04	0.04	0.03	0.03
MgO	0.11	0.11	0.18	0.09	0.14	0.12
CaO	0.43	0.24	0.41	0.37	0.84	0.22
Na_2O	3.65	4.63	3.77	4.28	4.35	4.28
K_2O	4.93	4.63	5.00	4.47	4.56	4.56
P_2O_5	0.01	0.02	0.02	0.02	0.01	0.02
CIPW norms (wt.%)						
Or	29.14	27.37	29.56	26.42	26.96	29.96
Ab	30.90	39.27	31.91	36.23	36.82	36.23
An	2.07	1.06	1.90	1.37	2.23	0.96
Q	36.40	30.92	34.99	34.63	31.99	34.38
Hy	0.37	0.35	0.55	0.21	0.00	0.38
Di	0.00	0.00	0.00	0.27	0.89	0.00
Ap	0.02	0.05	0.05	0.05	0.02	0.05
Il	0.21	0.19	0.25	0.21	0.21	0.21
C	0.23	0.25	0.05	0.00	0.00	0.26
Mt	0.67	0.61	0.76	0.69	0.62	0.65
ppm						
Ba		10	13	10		10
Rb	481	462	456	451	453	461
Cs		21.2	50.1	41.3	20.1	22.4

Sr	5	4	10	4	17	4
Th		59.2	57.4	59.5	60.9	61.4
U		20.19	25.2	28.27	16.89	23.72
Nb	42	39	45	45	43	41
Ta		1.85	1.78	1.81	1.88	1.8
Zr	144	142	153	156	141	146
Hf		5.01	4.86	4.81	5.03	4.8
V	6	5	6	6	8	6
Sc		0.76	0.97	0.81	0.86	0.77
Ni	15	16	17	19	18	17
Cr	3	4	4	7	6	4
Cu	2	2	1	4	5	2
Zn	22	17	26	22	15	15
La	56	56	63	57	52	56
Ce	84	71	73	78	59	64
Pr		6.2		5.4		7.1
Nd						10
Sm		3.15	3.94	1.89	2.96	1.83
Eu		0.15	0.19	0.13	0.19	0.18
Tb		0.2	0.19	0.18	0.28	0.2
Dy		1.12		1.17		1.39
Y		12.9		10.4		13.3
Yb		2.32	2.13	2.13	2.01	2.23
Lu		0.38		0.38		0.51

REE

The chondrite-normalised patterns of representative rhyolites from Antiparos are presented in Fig. 2. These patterns show prominent negative Eu anomalies (Eu/Eu* ratios range from 0.17 to 0.32) and a depletion of the middle REE (MREE). Because Eu is compatible with plagioclase and alkali-feldspar the observed Eu depletions in these rocks could be explained by fractional crystallisation of plagioclase and alkali-feldspar. Two minerals that the MREE are compatible with, in dacitic and rhyolitic rocks, are titanite and amphibole (Hanson, 1978). The occurrence of hornblende phenocrysts and titanite microphenocrysts (Anastopoulos, 1963) in the Antiparos rhyolites can explain the observed MREE depletions. Similar REE patterns to the Antiparos ones have been observed in high silica rhyolites from the Mineral Mountains in Utah, U.S.A. and were explained by fractional crystallisation of titanite (Lipman et al., 1978; Christiansen et al., 1983a, 1986). The $(La/Yb)_N$ and the $(Ce/Yb)_N$ (chondrite normalised) ratios range from 16-20 and 7-9 respectively, showing relative depletion of the HREE relative to the LREE. Zircon and garnet have high distribution coefficients for the HREE, while hornblende has higher ones for the HREE than the LREE in dacitic and rhyolitic melts. Thus the HREE depletion could be a possible result of zircon and hornblende fractionation or of the presence of residual zircon and/or garnet in the source of the rhyolites.

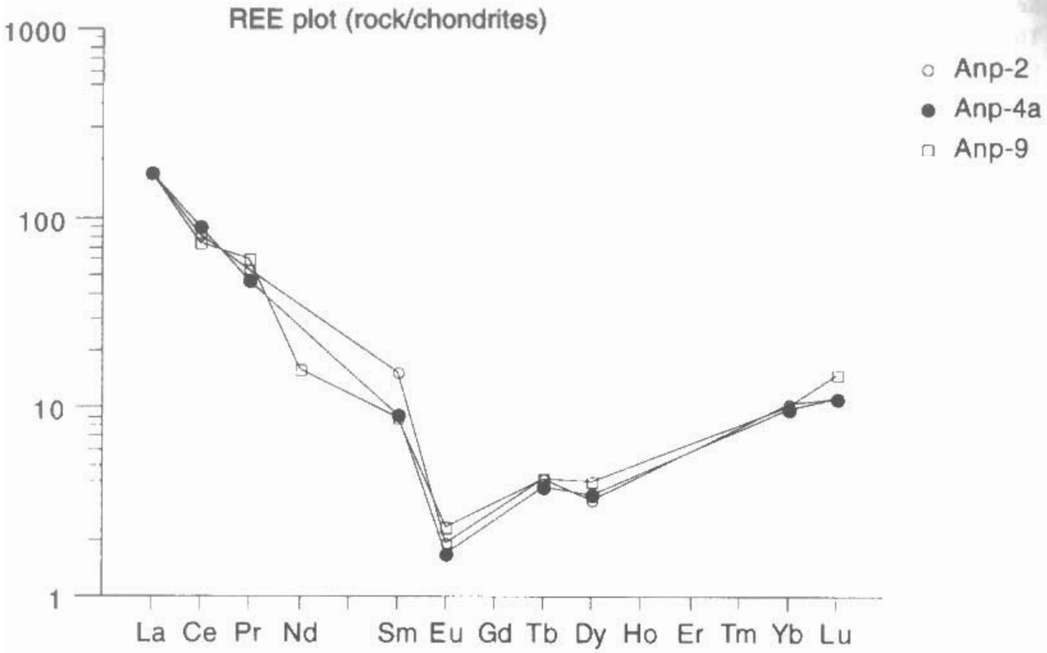

Fig. 2: Chondrite normalised REE pattern for a selection of representative samples from the Antiparos rhyolites.

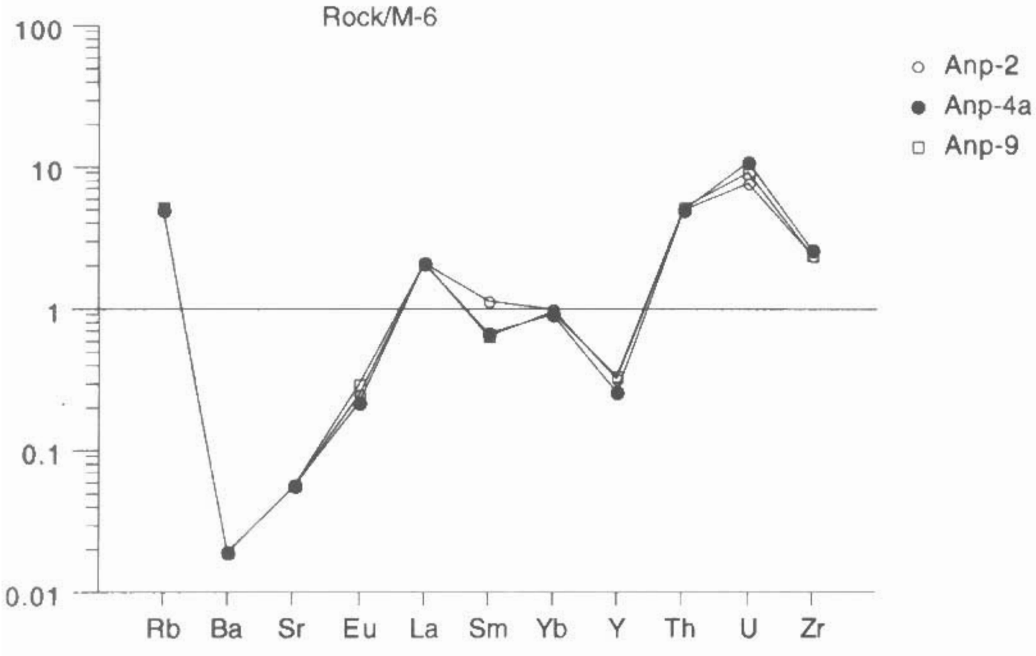

Fig. 3: Representative rhyolites from Antiparos normalised to trace element concentrations of M-6 (Barton et al., 1983),a calc-alkaline high-silica rhyolite from the island of Milos, Greece.

5. DISCUSSION

Comparison with other rhyolite suites

The Antiparos rhyolites, compared with the metaluminous to slightly peraluminous A-type (anorogenic) granites compiled by Collins et al. (1982), are, in terms of major element contents, similar to them, but they differ significantly as far as their trace element contents are concerned. Didier et al. (1982) proposed a modification in the classification scheme of Chappell and White (1974) by dividing granites into C-type ones (of crustal origin) and M-type ones (of mantle or mixed mantle and crust origin). The Antiparos rhyolites are much more similar in their major element chemical compositions to the "leucogranites" of Didier and Lameyre (1969), which are C-type granites, than to either S- or I-type granites (with the exception of the high-silica I-type granites reviewed by Collins et al., 1982). However, the alphabetical classification of granitoid rocks is presently considered to be a problematic, incomplete and aged classification method (Clarke, 1992; Pitcher, 1993)and its designations (I-, S-, A- type etc.) are used here as a mere device for ease.

In terms of their major element composition the Antiparos rhyolites are fairly similar to rhyolites from bimodal mafic-silicic volcanic associations, reviewed by Ewart (1979). According to Christiansen et al. (1986) most authors believe that these rhyolites were produced by fractional crystallisation of parental magmas derived by partial melting of the lower continental crust.

The Antiparos rhyolites are very similar in terms of both their major and trace element contents to the Cenozoic topaz-bearing rhyolites from the Western U.S.A. reviewed by Christiansen et al. (1983a, 1986). These rhyolites are connected with various tectonic environments (subduction-related, extension-related and transitional between the two). Their parental magmas are thought to have been generated by small degrees of partial melting of a residual felsic granulite source in either the lower or middle crust and subsequently evolved by extensive fractional crystallisation of sanidine, quartz, plagioclase, biotite and Fe-Ti oxides (Christiansen et al., 1986).

Compared with calc-alkaline rhyolites from all over the world compiled by Ewart (1979), the Antiparos rhyolites generally have lower Al_2O_3, TiO_2, $Fe_2O_3^*$, MgO and CaO while they are richer in total alkalies; in terms of their trace element abundances they are significantly enriched in incompatible elements like Rb, Cs, U, Th and Nb and strongly depleted in Ba, Eu and Sr. These relationships are very clear in the M-6 diagram (Fig. 3), where the Antiparos rhyolites are compared with a typical high-silica calc-alkaline rhyolite from the island of Milos. The Milos calc-alkaline high-silica rhyolites have major element compositions similar to the Antiparos rhyolites and are considered to be derived by considerable fractional crystallisation of a mafic, mantle derived parental magma (Barton et al., 1983). The island of Milos belongs to the currently active S. Aegean arc and is located only about 50 km southwest of Antiparos. Sr and Eu are compatible with both plagioclase and alkali-feldspar, while Ba is compatible only with alkali-feldspar (Cox et al., 1979, Rollinson, 1993). The very low abundances of Ba and Sr and the considerable negative Eu anomalies (Fig. 2) indicate that fractional crystallisation of both alkali-feldspar and plagioclase may have played an important role in the evolution of the primary magmas of the rhyolites. The enrichments in Rb, Cs and Nb could indicate the presence of nonresidual biotite in the source of the rhyolites and those of U and Th that the source was enriched in these elements. The feldpar-dominated fractional crystallisation processes would enrich the resulting melts in these incompatible elements.

The low contents of the compatible elements (Ni, Cr, V and Sc) probably reflects their low abundances in the source rock of the lavas (felsic source) rather than any crystal fractionation involving minerals with which these elements are compatible.

Magma genesis and evolution

Water-saturated conditions in the generation of granitic magmas are thought to be uncommon (Huppert and Sparks, 1988; Hall, 1996) and they are probably reached only at a rather late stage of magmatic evolution in most cases (Luth et al., 1964; Conrad et al., 1988). Wyllie (1977, 1984) also showed that when granitic liquids are generated from various compositions of continental gneisses the temperature interval for the generation and stability of water-saturated magmas is quite limited, while water-undersaturated liquids are stable and coexist with residual minerals over a broad range of temperatures.

When the amount of free water is either limited or there is no free water available during crustal anatexis, then the water for the melting is supplied by the breakdown of OH-bearing minerals, such as muscovite, biotite and hornblende, present in the source (Brown and Fyfe, 1970). Melts which are in equilibrium with muscovite would produce strongly peraluminous liquids with 3 to 8% normative corundum (Thompson and Tracy, 1977; Clemens and Wall, 1981). Therefore, the presence of muscovite in the source of the Antiparos primary magmas is considered unlikely. Additionally, the decomposition of hornblende starts at about 900° C (Brown and Fyfe, 1970) a temperature which is considered very high (very difficult to be attained at the site of the melting) for the generation of granitic liquids (Winkler, 1979). Instead, melting of a biotite-bearing granite or gneiss, which according to Winkler (1979) is the most analogous one to natural processes, maybe more likely for the generation of the primary magmas of the Antiparos rhyolites.

The beginning of melting of a biotite-bearing granite or gneiss, when all the water is supplied by the decomposition of biotite within the 5 to 10 kb pressure interval, which is the most typical one for crustal anatexis (Conrad et al., 1988), takes place from about 780-820° C to about 855-870° C respectively; the observed range of the biotite dehydration (at the same pressure) depends on the biotite composition (Hyndman, 1981). The primary liquids produced by this process have granitic to granodioritic compositions and are mostly water-undersaturated (Winkler, 1979; Hyndman, 1981; Wyllie, 1984). The pressures between 5 and 10 kb (16 to 33 km) cover a range from middle to lower crust. In Antiparos since the crust has a thickness of approximately 29 km the applicable pressure interval would be from 5 to about 9 kb.

Fig. 4 is the Thompson plot for the Antiparos rhyolites. Compared with such plots of granites and rhyolites by Thompson et al. (1984) they are similar to spiderdiagrams of continental, non-peralkaline, ultra leucocratic acid rocks, while they have several differences from plots of S-type granites. The extreme depletions of Ba and Sr indicate that alkali-feldspar and plagioclase fractional crystallisation processes played a crucial role in the evolution of the primary magmas of the lavas. A typical difference from subduction-related magmas is the absence from the Thompson plots of the Antiparos lavas of any significant negative Nb or Ta anomalies.

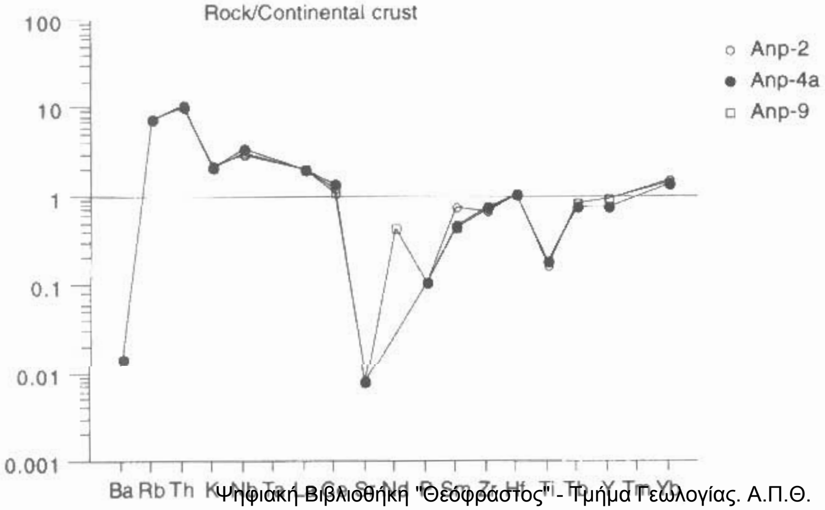

Fig. 4: Thompson plot (Thompson et al., 1984) for a selection of representative samples from the Antiparos rhyolites.

Additional evidence for the genesis and evolution of the Antiparos lavas was provided by Clapsopoulos (1991) by plotting the lavas in the Bachelor and Bowden (1985) multicationic diagram. In this diagram the Antiparos rhyolites plot within the field of anatectic granites (PM) which represents liquids, produced by partial melting of a felsic crustal source (e.g. gneiss, meta-pelite or intermediate meta-igneous rocks), which would have compositions corresponding to alkali feldspar and quartz ± sodic plagioclase ± biotite.

The possibility that the rhyolites could be differentiates of mantle-derived melts is contradicted by the fact that less silicic (less differentiated) rocks than the rhyolites are not present in Antiparos; Additionally, differentiates of melts of mantle origin (a) would have much higher CaO contents (e.g. about 1.50 wt.% CaO in rhyolites of the active South Aegean volcanic arc (Innocenti et al., 1982; Fytikas et al, 1984) compared with a maximum of 0.84 wt.% in Antiparos) and (b) normally their $^{87}Sr/^{86}Sr$ initial ratios should fall in the range 0.704 to 0.706 (Cox et al., 1979), as the ratios from the rhyolites of the S. Aegean volcanic arc do (Gale, 1981; Barton et al., 1983), instead of their observed range of 0.7113 to 0.7124 (Innocenti et al., 1982). These initial $^{87}Sr/^{86}Sr$ ratios suggest that either the source rocks had a higher Rb/Sr ratio than the 0.05 typical of lower crustal granulite or that the primary magmas were contaminated en route to the surface by upper crustal continental rocks, which had higher Rb/Sr and $^{87}Sr/^{86}Sr$ ratios.

Because from the arguments presented in the previous sections it is implied that the Antiparos rhyolites had a crustal origin, three representative samples were plotted in a spiderdiagram (Fig. 5) where they were normalised against the average continental crust composition of Weaver and Tarney (1984). The plot has characteristic large troughs for Ba and Sr accompanied by smaller ones for K, P and Ti. Another feature of the diagram is the much lower degree of enrichment of the REE (La, Ce, Yb) compared with the very high enrichments of other incompatible elements like Rb and Th. Christiansen et al., (1986) suggested that if monazite or some other REE-rich phase was present in the source of the topaz-bearing rhyolites from the Western U.S.A. then it would possibly be residual in the melting process, resulting in lower REE contents in the primary melts than otherwise. The relatively small troughs in P and Ti can indicate either a limited fractionation of apatite, titanite and Fe-Ti oxides or a deficiency of P_2O_5 and TiO_2 in the source.

Innocenti et al. (1982) interpreted the Antiparos lavas as a manifestation of bimodal volcanism and suggested that the alkaline basaltic magmas were formed as a result of the extensional tectonism affecting the area at that time, while the melting of the crust which produced the rhyolitic magmas occurred as result of a rise of the asthenosphere in this area which created a thermal anomaly and resulted in the melting of relatively shallow parts of the crust. The model of Innocenti et al. (1982) does not seem very likely for the generation of the Antiparos rhyolites because the alkaline basalts, which could provide the heat source for the initiation of crustal melting, are located far away (≥ 150 km) from Antiparos which is located very near to the currently active South Aegean volcanic arc.

Innocenti et al. (1982) proposed that these rhyolites were generated by partial melting of relatively shallow parts of the continental crust. Based on the data presented and reviewed here it is suggested that it is quite possible that the primary liquids were generated by rather small degrees of partial melting of a felsic granulite source located in the lower or middle continental crust underneath the island of Antiparos. Subsequently the primary liquids rose to shallow crustal levels where fractional crystallisation processes, dominated by alkali-feldspar fractionation, and perhaps with some interaction with, and contamination by, upper crustal rocks modified their initial composition to the one observed today in the erupted volcanics. This model of evolution is very similar to the one that Christiansen et al. (1983a, 1986) proposed for the topaz-bearing rhyolites of the Western U.S.A. which have major and trace element compositions very similar to the Antiparos ones.

Although the rhyolites have very high Rb/Sr ratios and Rb and U contents it is possible to generate such rocks by partial melting of a granulitic source which has a very low Rb/Sr ratio and low Rb and U contents as Christiansen et al. (1983b) demonstrated when, by using bulk distribution coefficients for a granulite

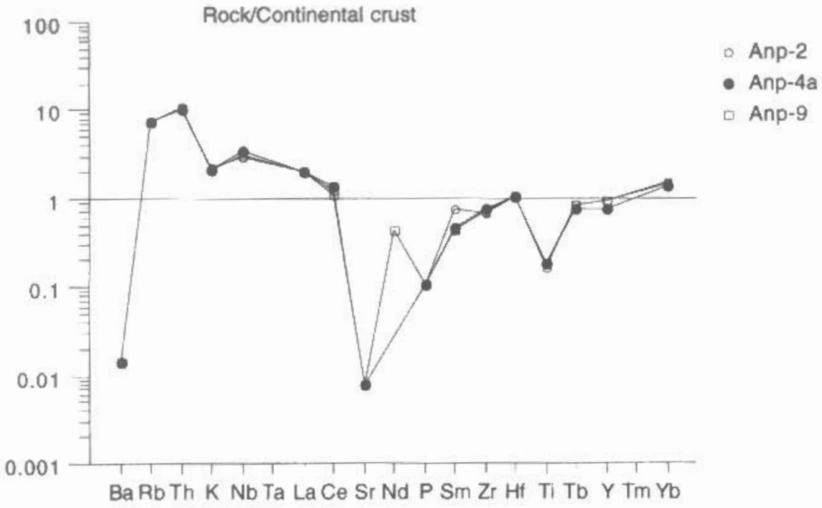

Fig. 5: Representative samples from the Antiparos rhyolites normalised to average continental crust composition. The normalisation values are from Weaver and Tarney (1984).

restite, they suggested that about 10% batch partial melting of a low Rb (30-50 ppm) and U (1.5 ppm) source would produce liquids that could fractionate towards compositions typical of the topaz-bearing rhyolites of the Western U.S.A.

The proposed heat source for the beginning of the melting process is the injection of basic magma at the base of the crust which raised the temperature sufficiently for the decomposition of hydrous silicates like biotite. The subduction of the Aegean oceanic slab under the Cyclades continental crust was well underway at the time of the eruption of the Antiparos lavas. As Antiparos is located only about 50km north of the currently active S. Aegean volcanic arc and the upper part of the subducting slab is situated at around 115 km beneath Antiparos (Meulenkamp et al., 1988), it is quite possible that basic magmas, generated by partial melting of either the upper part of the sinking slab or the mantle wedge overlying it, could have risen into the base of the continental crust. The generation of granite magmas by this process has been experimentally constrained by Huppert and Sparks (1988) who suggested that once crustal melts have been formed they act as a density barrier (Herzberg et al., 1983) and obstruct basic magmas to rise towards the surface. According to Christiansen et al. (1986) the resulting rhyolitic partial melts would be small and in extensional tectonic environments they would rise to higher crustal levels, fractionate and erupt as high-silica rhyolites. The presence of an extensional tectonic regime in the Antiparos area during the eruption period of the volcanic rocks is well documented and established (Mercier, 1981; Angelier et al., 1982).

6. CONCLUSIONS

The primary melts of the Antiparos high-silica rhyolites were possibly generated by small degrees of partial melting of a felsic granulite source (either under dry or water-undersaturated conditions) which was probably located in the middle to lower crust.

The melting of the crust occurred as a result of heating provided by basic magma, generated by the subduction of the Aegean oceanic slab beneath the continental crust, which subsequently rose to higher levels and reached the base of the continental crust.

After their generation the primary liquids rose to a shallow crustal magma chamber where before their eruption to the surface they underwent significant fractional crystallisation of alkali-feldspar (mainly), Ab-rich plagioclase, quartz, titanite and hornblende.