

NEOGENE TECTONIC ROTATIONS IN THE VICINITY OF THE NORTH AEGEAN TROUGH: NEW PALAEOMAGNETIC EVIDENCE FROM ATHOS AND SAMOTHRAKI (GREECE)

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Abstract

The present study focuses on two post-orogenic plutons, the Athos (Grigoriou) and Samothraki granites, as well as the Samothraki volcanics, located in the vicinity of the North Aegean Trough. A detailed palaeomagnetic study was carried out, with the aim of constraining the age and mechanism of tectonic rotations. In addition, anisotropy of low-field magnetic susceptibility (AMS) was studied and isothermal remanent magnetization (IRM) and thermomagnetic analyses were performed. Finally, a radiometric age for the Athos granite was obtained (43.3 ± 1.0 Ma K/Ar biotite). The measured declinations indicate clockwise rotations of the Athos (16.6°) and Samothraki (36.3°) blocks. The age of rotation is constrained to be <18 Ma at Samothraki, whereas the much smaller rotation of the Athos block can only be dated as younger than Eocene. Comparing the new palaeomagnetic data to the published dataset for Northern Greece, we suggest that the palaeomagnetically determined rotations in the vicinity of the North Aegean Trough are dominantly of post-Early Miocene age, and are controlled by major strike-slip faults and distributed "small" or minor faults.

Key words: palaeomagnetism, block rotations, geodynamics, Aegean.

Περίληψη

Μελετώνται δύο μετα-ορογενείς πλουτωνίτες, ο γρανίτης του Άθω (Γρηγορίου) και της Σαμοθράκης, και τα ηφαιστειακά πετρώματα της Σαμοθράκης, που βρίσκονται στην ευρύτερη περιοχή της Τάφρου του Βορείου Αιγαίου. Πραγματοποιήθηκε λεπτομερής παλαιομαγνητική μελέτη, με σκοπό τον καθορισμό της ηλικίας και του μηχανισμού των τεκτονικών περιστροφών. Επιπλέον μετρήθηκε η ασθενούς πεδίου ανισοτροπία της μαγνητικής επιδεκτικότητας (AMS) και έγιναν πειράματα ισόθερμης παραμένουσας μαγνήτισης και θερμομαγνητικές αναλύσεις. Τέλος, καθορίστηκε ραδιομετρικά η ηλικία του γρανίτη του Άθω (43.3 ± 1.0 Ma K/Ar σε βιοτίτη). Οι μετρηθείσες μαγνητικές αποκλίσεις αναδεικνύουν δεξιόστροφη περιστροφή για τον

Άθω (16.6°) και τη Σαμοθράκη (36.3°). Η ηλικία της περιστροφής εκτιμάται μικρότερη των 18 Ma για τη Σαμοθράκη, ενώ η πολύ μικρότερη περιστροφή του τεκτονικού τεμάχου του Άθω χαρακτηρίζεται νεώτερη από Ηώκαινική. Συγκρίνοντας τα νέα παλαιομαγνητικά αποτελέσματα με τα ήδη δημοσιευμένα για τον ευρύτερο βορειοελλαδικό χώρο, προτείνουμε ότι οι τεκτονικές περιστροφές στην περιοχή της Τάφρου του Βορείου Αιγαίου, που καθορίστηκαν με παλαιομαγνητικές μετρήσεις, χρονολογούνται μετά το Κάτω Μειόκαινο και ελέγχονται από κύρια ρήγματα οριζόντιας μετατόπισης και πληθώρα μικρότερων δευτερευόντων ρηγμάτων.

Λέξεις κλειδιά: Παλαιομαγνητισμός, περιστροφές τεμαχών, γεωδυναμική, Αιγαίο.

1. Introduction

The Aegean Sea and surrounding regions are a natural laboratory for the study of continental crust deformation. From the late Middle Miocene to Present, the area underwent dramatic N-S to NE-SW extension and crustal thinning (Mercier *et al.* 1979, Le Pichon and Angelier 1979, Papazachos and Kiratzi 1996). This was coeval with the westward displacement and bulk counterclockwise (CCW) rotation of the Anatolian-Aegean plate, which was accommodated by roll-back of the Hellenic subduction zone (McKenzie 1978, Jolivet *et al.* 1996). The active displacement data derived from focal mechanism solutions (Taymaz *et al.* 1991, Papazachos *et al.* 2000) and GPS or SLR positioning analyses (Le Pichon *et al.* 1995, McClusky *et al.* 2000) show that deformation of the Aegean continental crust strongly differs north and south of the North-Aegean Trough (NAT, Fig. 1). The NAT fault zone forms the boundary between these crustal domains and is a seismically active, ~ 30 km-wide, transtensional dextral fault zone along the western continuation of the North-Anatolian (NAF) fault system (Pavlidis *et al.* 1990, Taymaz *et al.* 1991, Roussos and Lyssimachou 1991, Hatzfeld 1999, Laigle *et al.* 2000, McNeill *et al.* 2004).

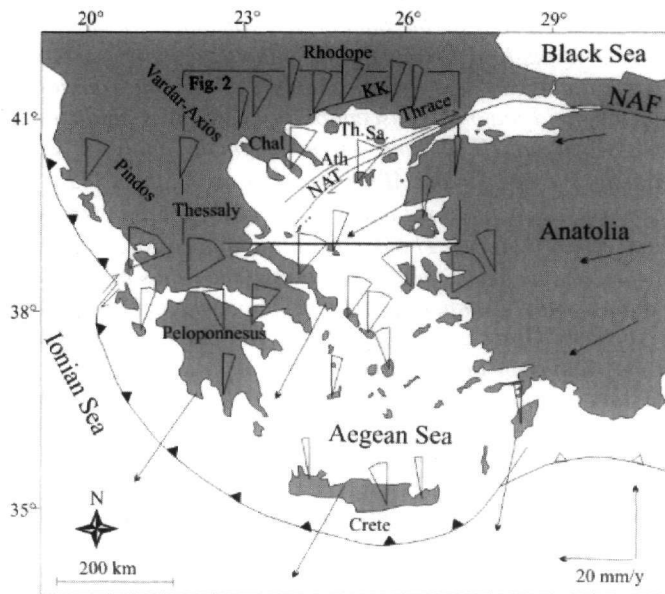


Figure 1 - Location map (framed: Fig 2) showing the regional plate boundaries and geodetic displacements (arrows) after McClusky *et al.* (2000). Pie figures represent previously published palaeomagnetic rotations. Bold line with teeth: thrust faults related to the Aegean subduction. NAF: North-Anatolian fault; NAT: North-Aegean Trough; KK: Kavala-Xanthi-Komotini fault. Ath: Athos peninsula; Chal: Chalkidiki; Sa: Samothraki island; Th: Thasos island

During the last two decades, a large set of palaeomagnetic data (Kissel and Laj 1988, Kondopoulou 2000 and references therein; van Hinsbergen 2005), has been obtained from Cenozoic formations from all over the Aegean area with the aim of constraining the style and timing of crustal deformation. In the last 4 years additional palaeomagnetic data have been obtained in the northern, western and southern Aegean, summarized in van Hinsbergen *et al.* (2005). Palaeomagnetic measurements reveal dominant CW rotations north of the NAT fault zone (Vardar-Axios basin, Chalkidiki peninsula, Rhodope massif) with an apparent systematic increase in magnitude from almost 0° (no rotation) in Thrace to 20°-30° in the Vardar Zone, together with a northward decrease of the rotation angle (Atzemoglou *et al.* 1994 and references therein). However, the timing of the observed rotations and their tectonic interpretation still remain controversial there, as palaeomagnetic data north of the NAT fault zone were mostly deduced from Eocene-Oligocene formations, and only rarely from Miocene-Pliocene ones.

The present contribution aims to constrain the age and mechanism of tectonic rotations in Northern Greece through the palaeomagnetic study of two post-orogenic plutons, the Athos (Grigoriou) and Samothraki granites (Fig. 1), and of the volcanic-sedimentary formations intruded by the latter granite. The particular interest of these study areas is twofold: i) the plutons are of different age, dated as Eocene and Miocene, respectively; and ii) they are located at different distances to the north of the NAT, with the Samothraki granite on the very edge of the easternmost and narrower part of the Aegean trough. Our results suggest a link between block rotation and strike-slip faulting during late Neogene time.

2. Geological setting

The North Hellenic orogen mainly formed at the expense of the Variscan and of Jurassic oceanic crust which is preserved in scattered ophiolitic massifs (Fig. 2). These crustal elements were involved in a protracted Alpine orogeny and the northeastward subduction of the Rhodopian domain under the Eurasian margin during the Late Cretaceous-Eocene (Burg *et al.* 1990, 1996, Michard *et al.* 1994, 1998, Ricou *et al.* 1998, Kiliyas *et al.* 1999). From the Middle-Late Eocene to the Middle Miocene, the North Hellenic orogen underwent strong post-orogenic extension with coeval emplacement of granodioritic plutons, deposition of clastic sediments and calc-alkaline volcanism in half-graben basins (Maltezou and Brooks 1989, Jones *et al.* 1992, Yanev and Bardintzeff 1997, Harkovska *et al.* 1998, Soldatos *et al.* 2001). Late Miocene-Pliocene to Pleistocene extension eventually occurred in brittle conditions and resulted in graben and half-graben subsidence controlled by NW-SE and E-W trending normal faults (Pavlidis and Kiliyas 1987, Martin and Mascle 1989, Voidomatis *et al.* 1990, Roussos and Lyssimachou 1991). The normal fault array seems broadly controlled by the NAT strike-slip fault zone which progressively changes westward into an extensional horsetail system (Fig. 2).

The *Athos (Grigoriou)* granodiorite is a subcircular pluton (Fig. 3a) which intrudes the Serbo-Macedonian unit, i.e. the upper plate of the Rhodope Core Complex. The Athos pluton belongs to a group of calc-alkaline intrusions of southern Chalkidiki, emplaced either within the Serbo-Macedonian unit or the adjoining, lower grade "Circum-Rhodope" units (Sithonia). Prior to the present study, Eocene ages had already been obtained from three out of the five south Chalkidiki granodioritic intrusions: 40-50 Ma at Sithonia (Kondopoulou and Lauer 1984, Vergely 1984, De Wet *et al.* 1989, Christofides *et al.* 1990), 53 Ma at Ierissos (Frei 1996), 44-47 Ma at Ouranopolis (De Wet *et al.* 1989). As shown at Sithonia (Sakellariou 1993, Tranos *et al.* 1993), the emplacement of the south Chalkidiki plutons accompanied the latest stages of ductile deformation of the country rocks, when they were already exhumed at upper crustal levels. The foliation pattern in both the granodiorite and the dioritic screens of the Athos apex (Fig. 3a) suggests syntectonic emplacement, coeval with a normal-dextral throw of the fault which separates the intrusion from the overlying Mount Athos greenschist and marble unit.

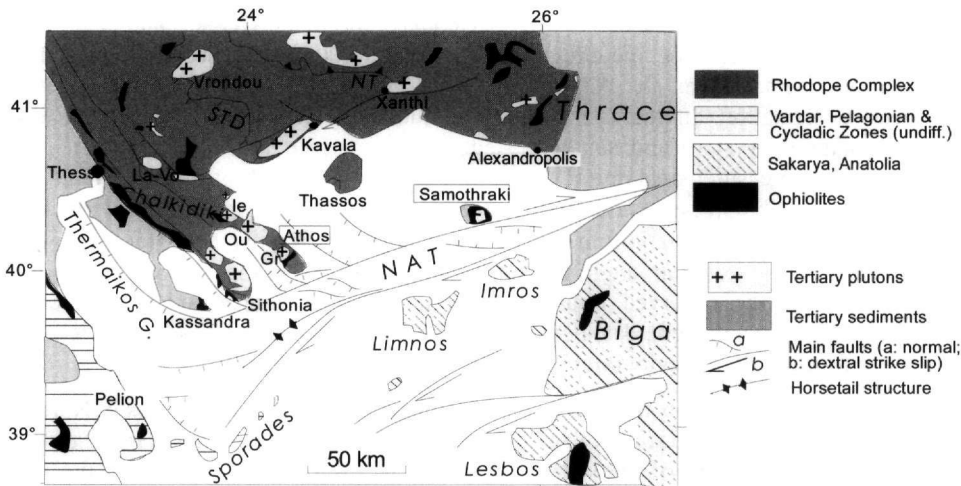


Figure 2 - Structural setting of the Athos and Samothraki study areas. NAT strike-slip and normal fault network after Pavlides *et al.* (1990), Voidomatis *et al.* (1990), Taymaz *et al.* (1991), Roussos and Lyssimachou (1991), Papazachos *et al.* (1999), Laigle *et al.* (2000), McNeill *et al.* (2004). Horsetail structure south of Sithonia after Martin and Mascle (1989). Gr: Grigoriou; Ie: Ierissos; La-Vo: Langada-Volvi graben; NT: Nestos thrust; Ou: Ouranopolis; STD: Strymon-Thassos detachment; Thess: Thessaloniki

Samothraki Island is located in the north-eastern Aegean (Fig.2) and geotectonically belongs to the Circum-Rhodope Belt (Kauffmann *et al.* 1976). From bottom to top, the island consists of, i) a low-grade meta-ophiolitic complex intruded by a calc-alkaline granite; ii) tilted volcano-sedimentary formations of Late Eocene-Early Miocene age; iii) almost horizontal Upper Miocene-Pliocene sediments (Fig. 3b; Tsikouras and Hatzipanagiotou 1998, Christofides *et al.* 2000). The volcanics are divided into an older and a younger volcanic rock series (Eleftheriadis *et al.* 1994). The younger volcanics (YVRS; sampled here) consist of latites, trachytes and high-K dacites dated at 22.3-18.9 Ma (K-Ar in biotite and whole rock; Eleftheriadis *et al.* 1994). Their chemical characteristics show that they represent a calc-alkaline volcanic rock series (Eleftheriadis *et al.* 1993). The *Samothraki pluton* emplaced at shallow depth, about 4-5 km according to the Al-in-hornblende barometer, contemporaneously with the youngest overlying volcanics, both being dated at 18.5±0.2 to 18.1±0.2 Ma (Rb-Sr in biotite and whole rock, Christofides *et al.* 2000). The Eocene-Early Miocene formations dip NW and SE in the western and eastern parts of the island, respectively. Their basal contact is no longer an unconformity, but has been reworked by detachment faults which place the granite close to the coeval volcanics (Fig. 3b). Hence, the granite was emplaced and exhumed in extensional conditions (ENE-trending direction of extension), like the coeval Kavala (Symvolon) granite beneath the Strymon Valley detachment (Sokoutis *et al.* 1993, Dinter and Royden 1993, Lips *et al.* 2000). The western part of Samothraki Island shows outcrops of gently folded Upper Miocene-Pliocene sediments. They are separated from the uplifted part of the island by a west-dipping normal fault, related to the recent, North Aegean extensional tectonics.

3. Sampling and laboratory measurements

In the Athos peninsula, 12 sites have been sampled in the Grigoriou batholith (Fig. 3a). Sampling was carried out by core drilling and orientation with standard compass techniques in nine out of the 12 studied sites, whereas 8-10 oriented hand samples, later cored in the laboratory, were taken from sites GR, GS and GT. As each of the neighbouring GR and GS sites yielded an insufficient number of samples, they were grouped together into a GRS site. Therefore we finally have 11 sites

in Athos. By contrast, most of the 13 Samothraki sites (Fig. 3b) were sampled as oriented hand samples. The Athos samples come from a limited range of lithologies (porphyroid granite, granodiorite, dioritic enclaves) within the Grigoriou batholith (Table 1). The Samothraki sites are more varied and include 5 sites in granodiorite (GPA, GPK, GPL, GSA, GSB), 7 sites in the coeval acidic lavas (AUN, AP, ALO-ALN, HO: lower level, HR: upper level, TRL, VRE), and one site (TK) in the Eocene sediments faulted above the plutonic body (Table 2).

A total of 162 standard specimens from the Athos area were measured for both palaeomagnetic analysis and anisotropy of magnetic susceptibility, whereas 109 specimens from Samothraki were measured for palaeomagnetic analyses. Natural remanent magnetizations (NRM) were measured with AGICO JR5 and MOLSPIN spinner magnetometers. Predominantly thermal and in several cases also alternating field (AF) demagnetizations were performed. Directions were analysed on orthogonal plots and selected with a least-square routine (Kirschvink 1980). The variation of magnetic susceptibility with temperature was monitored with a KLY-3 Kappabridge and a Bartington MS2B Meter. Acquisition of isothermal remanent magnetization (IRM) curves were obtained using a ASC pulse magnetizer. Finally, anisotropy of low-field magnetic susceptibility measurements (AMS) were performed with the KLY-3 Kappabridge apparatus. All measurements were carried out in the Palaeomagnetic laboratories of the Ecole Normale Supérieure (Paris) and the Aristotle University of Thessaloniki.

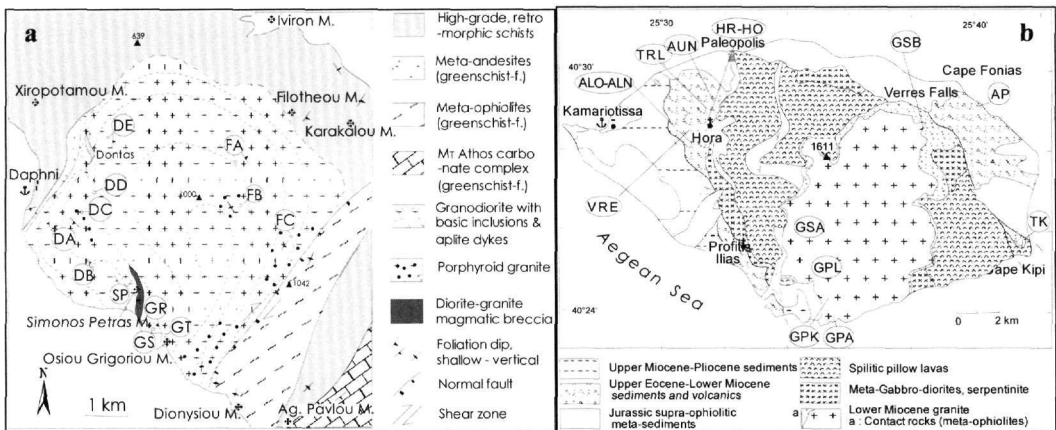


Figure 3 - a) Location of the sites sampled in the Grigoriou pluton of Athos peninsula; geological map modified after Kockel and Mollat (1977), b) Location of the sites sampled in the Samothraki granite and juxtaposed metasediments and lavas; geological map after Tsikouras and Hatzipanagiotou (1998)

4. Results

4.1. Athos peninsula

The nature of the main magnetic carriers varies with lithology at these sites. In the more acidic lithologies, sampled in the Filotheou area (sites FA, FB, FC), the magnetic carriers mainly consist of magnetite, with a smaller amount of haematite, according to the observed thermomagnetic and IRM curves (Fig. 4). By contrast, in the granodioritic parts of the Grigoriou pluton, the main magnetic mineral is likely to be maghemite, with an unquantified amount of pyrrhotite (indicated by unsuccessful AF cleaning of these samples and a drastic drop of magnetic remanence about 350-400 °C: Fig. 5). Two out of 11 sites (DE and SP, not reported here) were rejected after complete demagnetization, due to the absence of stable components of magnetization. The remaining 9 sites yielded stable and reasonably well-grouped palaeomagnetic components (Table 1). Some of the selected sites display a very low temperature (< 200 °C) overprint. At higher temperatures, samples from different lithologies show different behaviour during demagnetization

as follows. The Filotheou "acidic" sites exhibit an intermediate temperature (200-450 °C) unblocking component with a SW declination and a reverse inclination, sometimes followed by NE-directed normal component at 550-580 °C, likely carried by magnetite. The remaining granodioritic sites show a low temperature (150-250 °C) NE-directed normal component, again followed by an intermediate temperature (300-450 °C) SW-directed reverse component (Fig. 5).

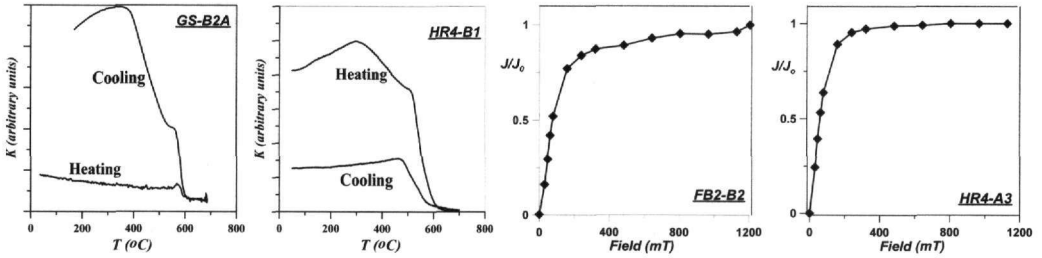


Figure 4 - Representative thermomagnetic curves (left) showing the variation of magnetic susceptibility (k) with temperature and IRM acquisition plots (right) from representative samples from Athos and Samothraki

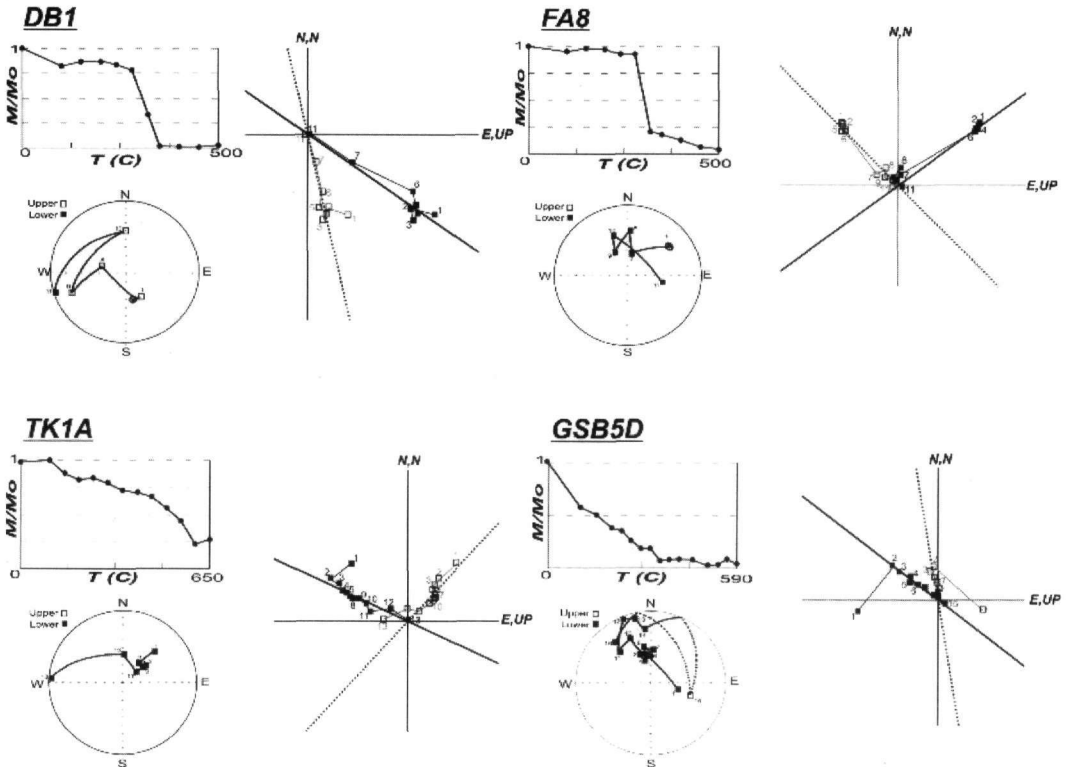


Figure 5 - Representative demagnetization decay curves, lower-hemisphere stereographic projections and orthogonal demagnetisation plots in geographic coordinates, from the Athos granite (DB1 and FA8) and the Samothraki CW (TK1A) and CCW (GPA5) sites

Table 1 - Mean directions of high- and mid-temperature components from the Athos sites. (N/N1=samples used/total, ΔD =confidence limit on the declination, α_{95} =semi-angle of cone of confidence, K=precision parameter)

Site	Rock type	Age	N/N1	Dec $\pm \Delta D$	Inc	α_{95}	K
DA	Granodiorite	L. Eocene	11/16	168.5 \pm 8.8	-63.6	3.9	135
DB	Granodiorite	L. Eocene	10/10	189.6 \pm 17.9	-56.3	9.8	25
DC	Porphyroid granite	L. Eocene	14/17	202.2 \pm 22.6	-59.4	11.3	13
DD	Diorite/granite	L. Eocene	5/14	206.5 \pm 19.2	-64.0	8.3	85
FA	Granodiorite	L. Eocene	10/12	052.9 \pm 9.1	27.0	8.1	36
FB	Porphyroid granite	L. Eocene	7/11	194.1 \pm 43.8	-65.0	17.0	14
FC	Leucocratic granite	L. Eocene	10/20	209.4 \pm 13.4	-08.2	13.3	14
GRS (GR+GS)	Granodiorite	L. Eocene	10/14	209.4 \pm 9.2	-49.2	6.0	59
GT	Granodiorite	L. Eocene	10/14	198.1 \pm 25.7	-54.7	14.5	12
<i>Mean (FC excluded - low Inc)</i>			<i>8/11</i>	<i>203.6 \pm 21.1</i>	<i>-56.4</i>	<i>11.5</i>	<i>24</i>

Table 2 - Mean directions of high-temperature components from the Samothraki sites. (N/N1=samples used/total, ΔD =confidence limit on the declination – n.a.=not available, α_{95} =semi-angle of cone of confidence, K=precision parameter)

Site	Rock type	Age	N/N1	Dec $\pm \Delta D$	Inc	α_{95}	K
AUN	Lava	L. Miocene	7/7	223.9 \pm 23.2	-43.5	16.6	14
GPK	Granite	L. Miocene	9/10	216.3 \pm 37.2	-62.0	16.5	11
GSA	Granite	L. Miocene	8/10	230.7 \pm 18.1	-45.5	12.6	20
TK	Sediment	U. Eocene	9/10	041.6 \pm 13.0	49.6	8.4	39
VRE	Pyroclastic	Eocene	9/11	221.0 \pm 7.8	-17.8	7.4	49
HOR (HO+HR)	Lava	L.Miocene	5/9 CW	217.7 \pm 19.4	-42.1	14.3	42
			4/9CCW	316.3 \pm 56.8	66.0	19.9	22
GPL	Granite	L. Miocene	7/8	212.4 \pm 25.3	-16.5	24.2	7
AP	Volcanic	L. Miocene	6/6	323.9 \pm 23.7	51.8	14.4	23
GSB	Granite	L. Miocene	5/8	337.3 \pm 24.0	43.7	17.1	21
ALO-ALN	Lava	L. Miocene	4/7	157.1 \pm n.a.	-79.3	21.9	19
TRL	Lava	L. Miocene	12/13	037.3 \pm 12.9	-32.8	10.8	17
GPA	Microgranite	L. Miocene	7/10	224.0 \pm 19.1	29.0	16.6	14
<i>Mean (AUN, GPK, GSA, TK, HOR, VRE)</i>			<i>6/12</i>	<i>222.1 \pm 17.2</i>	<i>-43.6</i>	<i>12.4</i>	<i>30</i>
<i>Mean (HOR-CCW, AP, GSB, ALO-ALN)</i>			<i>4/12</i>	<i>328.6 \pm 41.4</i>	<i>60.5</i>	<i>19.0</i>	<i>24</i>

4.2. Samothraki Island

The main magnetic mineral in both the granitic and dacitic lithologies is mostly magnetite, as the remanence shows a low coercivity and is completely destroyed at 580 °C (Fig. 5). In the reddish sandstone at site TK, the magnetic carrier consists of fine-grained haematite, as suggested by the demagnetization curve and a lack of success of AF demagnetizations. Thermomagnetic analyses and IRM measurements of volcanics and granites show a clear predominance of magnetite but with small amounts of haematite in some cases (Fig. 4). The granite and lava samples generally show a low temperature (LT) component with normal polarity unblocked at T<300 °C, and a high temperature (HT) component with reverse polarity unblocked between 380-580 °C. The HT component is consistently oriented towards the SSW direction, despite some scattering in both declination and inclination. In contrast, the sandstone samples show a good clustering of a NE-directed normal component, defined between 100-550 °C, with a roughly antiparallel HT

component which only appears above 650 °C. The direction of this HT component cannot be precisely derived, due to its weak intensity. Despite the almost systematic trend of HT components towards a NE-SW direction, a more detailed analysis is needed. A. The three granitic sites (GPK, GPL and GSA), two lava sites (AUN, HOR), the pyroclastic site (VRE) and the sedimentary site (TK) yielded stable high temperature (HT) directions (Table 2), which converge towards the expected direction for the area and for this time-span (Miocene). The mean magnetization for these sites is 222.1/-43.6 ($\alpha_{95} = 12.4$, $K = 30$). B. Another group of directions is seen at sites AP and ALN with westward declinations and antipodal polarities, but with a very high inclination and a big scatter for site ALN. Site AP is located in strongly faulted and weathered lavas and yielded components oriented westwards. The same westward declination is seen in five out of eight samples from site GSB, located at the very border of the granitic stock (at about 5 m from the hornfels aureole). We should note that the remaining three samples at this site fit to the directions of the first group. A similar case is represented by sites HO-HR where two main tendencies can be identified; one fitting the first, expected for the area, direction and a second strongly deviated westwards with scattered inclinations. C. Finally, we draw the attention to a peculiar grouping obtained both in volcanics and granites (sites TRL and GPA) with eastward declinations but negative inclinations. Any error of sampling is excluded as these directions have been obtained also by other groups in more than three sites (Beck *et al.* personal communication) and were also reported in volcanics from Lemnos Island (Westphal and Kondopoulou 1993). A specific series of measurements, including paleointensities, is under progress in order to investigate this strange behaviour.

4.3. Inferred age of magnetization

In the framework of the present study, we measured a K/Ar biotite age on a sample from the Grigoriou pluton. The Athos granitoid pluton crossed the 300 °C isotherm around 43 Ma (Table 3). Taking into account that the characteristic component of magnetization is carried by magnetite, we assume an equivalent *Eocene* age for the acquisition of remanence in our samples from Athos, provided that the pluton did not stay long at temperatures above 300 °C.

Table 3 - K/Ar dating of the Grigoriou granite, Athos peninsula (R. Montigny, Laboratoire de Geochimie, EOST Strasbourg)

Sample number and location	Rock type	Mineral	K ₂ O (wt%)	⁴⁰ Ar rad (10 ⁻¹¹ mol/g)	$\frac{100 \times {}^{40}\text{Ar rad}}{{}^{40}\text{Ar}}$	Calculated age Ma ($\pm 2\sigma$)
G93 168b Grigoriou	granodiorite	biotite	8.90	56.22	91	43.3 \pm 1.0
$\lambda_e = 0.581 \times 10^{-10} \text{ a}^{-1}$; $\lambda_\beta = 4.962 \times 10^{-10} \text{ a}^{-1}$; ${}^{40}\text{K/K total} = 1.167 \times 10^{-4} \text{ mol/mol}$						

The Samothraki samples are more varied. Three out of the five sites taken from the granitic intrusion were dated at 19.8 Ma (Christofides *et al.* 2000). Site AUN was sampled in the upper volcanics, coeval with the underlying granite (Christofides *et al.* 2000). In both cases, the magnetic carrier is magnetite, and the magnetization is likely to have an Early Miocene age. The latitic samples from site TRL gave an age of 18.9 \pm 0.5 Ma (Eleftheriadis *et al.* 1994). The main magnetic carriers are both magnetite and hematite, but the origin of this magnetization is not fully understood as it is directed eastward and up. An age of 22.3 \pm 0.8 Ma is assigned to the dacitic site of Alonia (Eleftheriadis *et al.* 1994). Its magnetization is carried by magnetite and is directed westward and diverges from the general trend of the other sites. In contrast, the sediments of the site TK are dated as Late Eocene and their main magnetic carrier is hematite. In this case, the magnetization can be either diagenetic and primary, or related to a complete remagnetization due to the granite emplacement at shallow depth. We consider the second hypothesis as more likely (see next sub-section), and retain an Early Miocene age of magnetization.

It can be deduced from the above considerations that the age of magnetization for most volcanic and granitic sites in Samothraki is *Miocene*.

5. AMS measurements

AMS measurements yield the magnitudes of the three principal and mutually orthogonal axes of the AMS ellipsoid ($K_1 \geq K_2 \geq K_3$), as well as their declinations and inclinations with respect to the geographical reference frame. The mean magnetic susceptibility is given by $K_m = (K_1 + K_2 + K_3)/3$. The corrected anisotropy degree $P' = \exp \sqrt{2 \left[\left(\ln \frac{K_1}{K_m} \right)^2 + \left(\ln \frac{K_2}{K_m} \right)^2 + \left(\ln \frac{K_3}{K_m} \right)^2 \right]}$,

and the shape parameter of Jelinek (1981) $T = (\ln F - \ln L)/(\ln F + \ln L)$, where $F = K_2/K_3$ and $L = K_1/K_2$, were also calculated. The mean AMS data are presented in Table 4.

Table 4 - Mean AMS measurements for various locations from Athos and Samothraki

Location	Rock Type	Km ($\times 10^{-6}$ SI)	P'	T	K ₁ (Dec/Inc)	K ₂ (Dec/Inc)	K ₃ (Dec/Inc)
Athos (West)	Granite	271	1.089	-0.184	171/02	242/79	076/10
Athos (East)	Granite	229	1.092	0.237	029/10	263/35	137/01
Athos (South)	Granodiorite	345	1.045	0.117	169/05	065/77	266/12
Samothraki	Lavas	14842	1.043	0.246	016/05	272/31	096/68
Samothraki	Granite	8758	1.109	0.446	081/02	353/12	148/49

In the paramagnetic granitoid of Athos the magnetic susceptibility is quite low, with a mean of 263 μ SI. The anisotropy degree shows a high variation from 1.010 to 1.542, with a mean of 1.080 (8 %). The majority of the samples lie in the 1.010-1.200 interval. The magnetic fabric is generally well-defined; the K_1 axes, defining the magnetic lineation are almost horizontal with a mean direction of 356/01, while the magnetic foliation plane is nearly vertical (004/74 E). On the contrary, in Samothraki rocks the magnetic susceptibility is very high, with a mean value of 11800 μ SI, indicating the presence of magnetite, which has also been confirmed by other laboratory tests. The anisotropy degree is quite low for the Samothraki volcanics (4 %) while it reaches medium levels within the granitoids (11 %). The magnetic fabric shows a clear trend of the K_1 axes, which in all but one site are almost horizontal. The magnetic foliation, contrary to that of Athos rocks, is steep, ranging from 50-70°.

The magnetic anisotropy in minerals can deviate from the ambient field the direction of thermo- or chemical- remanent magnetization acquired by minerals. This deviation is expected to be negligible for small anisotropy degrees, but can be substantial for larger anisotropies. This problem has been treated at length by Stacey (1960), Uyeda *et al.* (1963), Hroudá (1982), and Stephenson *et al.* (1986). Given that the anisotropy degrees of the present study are relatively low (means of 8% and 4% for Athos and Samothraki respectively) the possible deflections of the palaeomagnetic vectors due to the anisotropy effect can be assumed to be negligible, enhancing thus the reliability of the calculated rotations.

6. Discussion and Conclusions

6.1. Tectonic corrections

Palaeomagnetic directions show rather scattered inclination values, either at the site level (especially in Samothraki; Table 2), or at the regional scale of the massif (Athos; Table 1, Fig. 6). In the latter case, NW-trending normal faults have been observed along the SW and NE sides of the peninsula (Pavlidis and Caputo 1994). Such faults may have split the peninsula into tilted fault blocks, and therefore may have caused the shallowing or steepening of inclinations in the NE or

SW blocks, respectively. This working hypothesis was not checked in the field, due to extensive forest cover and poor outcrops, and no angular values could be measured for untilting.

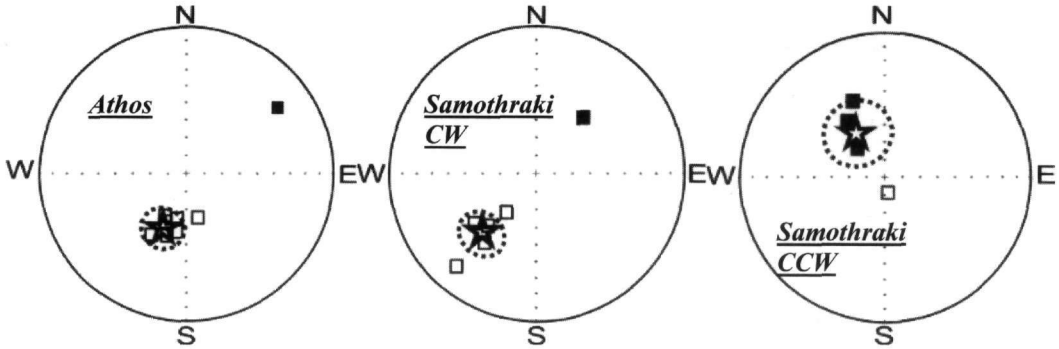


Figure 6 - Site mean directions in geographic coordinates for Athos and Samothraki. Open symbols represent negative inclinations, filled symbols indicate positive inclinations. The star indicates the calculated mean direction; a95 is shown around mean

In Samothraki, the *in situ* directions show a better cluster after transferring the TK mean direction into the upper hemisphere (Fig. 6). The granitic pluton does not appear to be affected by faulting (except at its very SW border, outside of the sampling area; Fig. 3b). In the sedimentary (TK) and volcanic (AUN, TRLO) sites, indications of the palaeohorizontal are available, with respective strike and dip values N180, 40E, N60, 50NW, and N190, 70. Application of these tectonic corrections results in poorly significant stratigraphic directions (AUN: dec. = 270°, inc. = 37°; TK: dec. = 341°, incl. = 59°) and scatter increases. In the case of site TK, magnetic mineralogy (haematite) and structural setting suggest a complete remagnetization during Lower Miocene granite emplacement and likely coeval tilting of the Late Eocene rocks. We note here that TK_{corrected} is partly supported by directional group B, but the resolution is very poor in order to conclude about a widely spread remagnetization. At site AUN, the dome-shaped acidic extrusives emplaced on the western slope of the Oligocene-Miocene volcanic centre were affected by further tilting related to the emplacement of the granitic body at shallow depth, likely before complete cooling. This particular setting would explain the lack of significance of the tectonic correction. Accordingly, all results are presented *in situ* (Tables 1, 2).

6.2. Rotation age and mechanisms

The observed declinations indicate clockwise rotations of the Athos and Samothraki blocks around vertical axes. The rotation angle is calculated (Butler 1992) by subtraction of the expected declination given by the reference poles of Westphal *et al.* (1986) and Besse and Courtillot (2002) for the Eocene (7.0°) and Early Miocene (5.8°), respectively. The resulting rotation angles are 16.6°±17.2° for the Athos (Grigoriou) area and 36.3°±14.0° for Samothraki island. The age of rotation is constrained to be <18 Ma at Samothraki, whereas the smaller rotation of the Athos block can only be dated as younger than Eocene. In the same way the corresponding inclination flattenings were calculated: -3.6°±9.6° for Athos and 12.2°±10.2° for Samothraki.

Previous research aiming to establish a chronology of rotations in Northern Greece was hampered by the lack of magnetizations of different ages in a single block, except in the Vrontou block of Central Rhodope, which includes Early Oligocene and Early Miocene intrusions. In that case, Dimitriadis *et al.* (1998) inferred the occurrence of a minor Oligocene CW rotation (8°) followed by additional CW rotation (13°) after the Early Miocene. In the broader area surrounding the Athos peninsula and Samothraki a number of published palaeomagnetic results allow the establishment of a general pattern for the occurrence of rotations after the Eocene (Table 5). For the Symvolos area published data by Atzemoglou (1997), Zananiri (2004) and van Hinsbergen (2005) show moderate angles of CW rotation. For Chalkidiki, including Sithonia (Kondopoulou

and Westphal 1986) and Kassandra (Haubold *et al.* 1997), the rotation angles do not vary significantly between the Eocene (29°-33°) and the Miocene sites (20°-30°). Rotation angles varying between 15°-30° CW have been also measured in Upper Pliocene sediments of Chalkidiki (two sites, one of which is located at the beginning of Athos peninsula; Westphal *et al.* 1991). Therefore, all the rotation angles measured in Athos and Samothraki follow this regional trend of recent (Neogene) rotations, despite their scattered values. The rotation angle for Athos (16.6°) is one of the smallest measured in the area (but converges with the ones in Symvolos), whereas the Samothraki rotation (36.3°) is the largest measured in Miocene formations. Post-Eocene anticlockwise rotation was not observed in the vicinity of the study area. The few and ill-defined CCW directions in four Samothraki sites do not constitute a solid set for valid interpretations. Hence, the weak Athos block rotation points to a contrasting behaviour during the Neogene, not only with Samothraki but with most other studied formations in the surrounding area (except the broader Symvolos region).

Table 5 - Published rotation angles from the vicinity of the North Aegean Trough

Area	Formation	Age	D/I (°)	α^{95}	Rot	Reference
Riza	Sediments	Up. Pliocene	195/-45	8	15°	Westphal et al. 1991
Ierissos	Sediments	Up. Pliocene	212/-17	38	32°	Westphal et al. 1991
Kassandra	Sediments	Miocene	025/54	5	19.5°	Haubold et al. 1997
Lemnos	Volcanics	Miocene	034/48	15	28.5°	Westphal and Kondopoulou 1993
Ouranoupolis	Granite	Oligocene	219/ -24to-58	9	32.8°	Kondopoulou and Westphal 1986
Sithonia	Granite	Oligocene-Eocene	037/31	9	30.8°	Kondopoulou and Westphal 1986
Kassandra	Sediments	Jurassic (post-depositional magnetization, Eocene?)	043/43.6	8.3	36.0°	Feinberg et al. 1994
<i>D_{exp} : Miocene=5.5°, Oligocene=6.2°, Eocene=7.0° (Calculated using the Europe Apparent Polar Wander Path; Besse and Courtillot 2002)</i>						

The difference between the measured Neogene rotations at Athos and Samothraki could be related to the different location of these blocks with respect to the North Aegean Trough (NAT). Samothraki island stands directly on the northern edge of the NAT, close to the Saros basin (Figs 2 and 3). In contrast, the Athos (Grigoriou) block is c. 25 km from the western part of the NAT, i.e. the area where the dextral movement is increasingly distributed. Fault mechanics suggest that before the onset of a concentrated displacement regime along the NAF-NAT fault system a wide dextral shear zone should have accommodated the incipient Anatolia-Europe relative movement, starting at c. 13 Ma (Le Pichon *et al.* 2003). We assume that this early shear zone included the future Samothraki island and caused its large rotation as early as the Late Miocene. The Athos peninsula suffered less rotation at that time since it was more distant from the NAT shear zone, and was already located in the extremity of the extensional horsetail. This scenario seems plausible but one cannot disregard that the Sithonia and Ouranoupolis plutons, which are synchronous in age and with equal distance from the NAT display a rotation angle twice as large as the Athos one, i.e. 30° vs 17°, respectively. The ca. 15 of differential rotation between these peninsulas is likely to originate from the conjugate movements between the Athos and Sithonia fault zones and the southwestward increase of horizontal slip along these faults (Koukouvelas and Aydin 2002). This increase could result in the opening of Sigitikos basin with a small scissor component. An additional evidence for such a pattern is given in Papanikolaou *et al.* (2006). Based on detailed analysis of air gun lithoseismic profiles, the existence of a separate Athos block is clearly shown in

their fig.4, and a gradual increase of opening and deepening of the neighboring basin supports our scenario as exposed in the previous paragraph.

6.3. Regional implications

Geodetic determination of horizontal velocities in a Eurasia-fixed reference frame indicates coherent motion of 30 mm/yr towards the SW in central and southern Aegean, *versus* inhomogeneous motions of 5 to 10 mm/yr towards the S or SW in the coastal area of northern Greece (Chalkidiki, south Rhodope) and adjoining islands (Thassos, Samothraki) (McClusky *et al.* 2000). These domains are separated by the NAF-NAT (Fig. 1).

The tectonic interpretation of the palaeomagnetically observed rotations from Northern Greece, and of their relationship with those observed in central Aegean (south of the NAT active boundary) is still an open issue. Kondopoulou *et al.* (1996) suggested that the area was tectonically similar to the present-day south Aegean subduction zone as early as the Early Tertiary. Based on the trend of ductile stretching lineations in the Aegean basement rocks, as well as on the palaeodeclination pattern, Walcott and White (1998) suggest that Northern Greece, central Aegean, mainland Greece and Peloponnesus altogether belong to a "Central Aegean Block" which sequentially formed at the expense of an Oligo-Miocene, E-W trending orogen, and rotated from ~25 Ma onward. Dimitriadis *et al.* (1998) also claim that parts of the Rhodope massif underwent minor CW rotations during the Late Oligocene, whereas additional CW rotations occurred after the Early Miocene. They propose that at least the post-Early Miocene rotations in Northern Greece were the result of plate tectonic motion in the Aegean, and of brittle crustal detachment, translation and rotation on top of a stretched ductile lithosphere. In contrast, Goldsworthy *et al.* (2002) argue that the rotations observed from Chalkidiki to the central Aegean and mainland Greece could be explained by a model of distributed strike-slip faulting in western Anatolia-Central Aegean, ending in a normal fault system in central-western Greece, i.e. a modified version of the Taymaz *et al.* (1991) "broken slats" model.

Comparing the new palaeomagnetic data to the published dataset for Northern Greece (Kondopoulou 2000), we suggest that the palaeomagnetic rotations in Northern Greece are dominantly related to post-Early Miocene block rotations, and are controlled by major strike-slip faults and distributed "small" or minor faults as suggested by Duermeijer *et al.* (1998) and consistent with a generic theoretical model for rotations proposed by Peacock *et al.* (1998).

During the rotation phase of the western Aegean domain, extension in the Aegean was accommodated among others by extensional detachments of the Cyclades and the Rhodope (Altherr *et al.* 1982, Hejl *et al.* 2002, Lips *et al.* 2000). Therefore, at least part of the rotation difference between Sithonia and Athos, as well as Samothraki and Athos could be explained by the activity of these detachments.

The present palaeomagnetic dataset, combined to the one of surrounding areas and the compiled regional kinematic information, constitutes an important step towards distinguishing local rotations from the regional ones. Furthermore, the limits of smaller blocks are better constrained through the study of the here referred rotations which show interactions between local and regional deformation.

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