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No significant post-Eocene rotation of the Moesian Platform and Rhodope (Bulgaria): Implications for the kinematic evolution of the Carpathian and Aegean arcs

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ABSTRACT

The region located between the Carpathian–Balkan and Aegean arcs, the Moesian Platform and Bulgarian Rhodope, is generally assumed to have been stably attached to the East European craton during the Cenozoic evolution of these arcs. The kinematic evolution of this region is, however, poorly constrained by paleomagnetic analysis. In this paper we provide new paleomagnetic data (800 volcanic and sedimentary samples from 12 localities) showing no significant post-Eocene rotation of the Moesian platform and Rhodope with respect to Eurasia, therefore confirming the stability of this region. We compare this result to a provided review of paleomagnetic data from the South Carpathians (Tisza block) and the Aegean region. The Tisza block underwent 68.4±16.7° of middle Miocene (~15–10 Ma) clockwise rotation with respect to the Moesian Platform, in line with previous rotation estimates based on structural geology. The stability of the Moesian platform during middle Miocene eastward emplacement of the Tisza block into the Carpathian back-arc supports dextral shear along the Southern Carpathians recorded by 13–6 Ma clockwise strike-slip related rotations in foreland deposits. The new reference direction for the Moesian platform and Rhodope allows accurate quantification of the rotation difference with the west Aegean domain at 38.0±7.2° occurring between 15 and 8 Ma. To accommodate this rotation, we propose that the pivot point of the west-Aegean rotation was located approximately in the middle of the rotating domain rather than at the northern tip as previously proposed. This new scenario predicts less extension southeast of the pivot point, in good agreement with estimates from Aegean structural geology. Northwest of the pivot point, the model requires contraction or extrusion that can be accommodated by the coeval motion of the Tisza Block around the northwestern edge of the Moesian platform.

1. Introduction

The mountain belts that evolved in southeastern Europe have served as type localities for the development of general concepts for mountain building, orogenic collapse and subducted slab dynamics: the Aegean, Carpathian and Balkan orogenic systems (Fig. 1). Between the Carpathian and Balkan orogenic systems is the Moesian Platform, a Precambrian basement block that is generally regarded to belong to stable Europe. It is therefore used as an essential reference for kinematic reconstructions during the Cenozoic (e.g. Linzer, 1996; Morley, 1996; Linzer et al., 1998; Ricou et al., 1998; Schmid et al., 1998; Zweigel et al., 1998; Fig. 1). However, the tectonic stability of this block during the evolution of the Carpathian, Balkan and Aegean systems is questionable and still poorly constrained by reliable paleomagnetic studies of vertical-axis rotation.

In regions adjacent to the Moesian Platform, large vertical-axis block rotations have been determined paleomagnetically, allowing to constrain and quantify Cenozoic kinematics (Horner and Freeman, 1983; Balla, 1987; Kissel and Laj, 1988; Morris, 1995; Speranza et al., 1995; Kissel et al., 2003; Csontos and Voros, 2004; van Hinsbergen et al., 2005a). To the north in the Carpathian region, ~80° of clockwise rotations in the large Tisza Block since the Oligocene are generally interpreted to reflect its wholesale motion around the northwestern corner of the Moesian Platform during eastward roll-back of the Carpathian subducted slab (e.g. Patrascu et al., 1994; Panaiotu, 1998; Schmid et al., 1998). To the south, the entire domain formed by the western Aegean and Albanian regions, rotated 50° clockwise (away from the Northern Rhodopes and the Moesian Platform) largely between ~15 and 8 Ma. This has been interpreted to be caused by a combination of southward roll-back of the African slab and/or westward extrusion of Anatolia (Kissel and Laj, 1988; Kissel et al., 2003; van Hinsbergen et al., 2005a). These interpretations, however, rely on the assumption that the Moesian Platform has not rotated during the rotation of these domains. To date, paleomagnetic information from the Moesian Platform is scarce, but does suggest some Cenozoic rotation (Dolapchieva, 1994). Recently acquired results from the Moesian Platform itself suggest a ~15° clockwise rotation of

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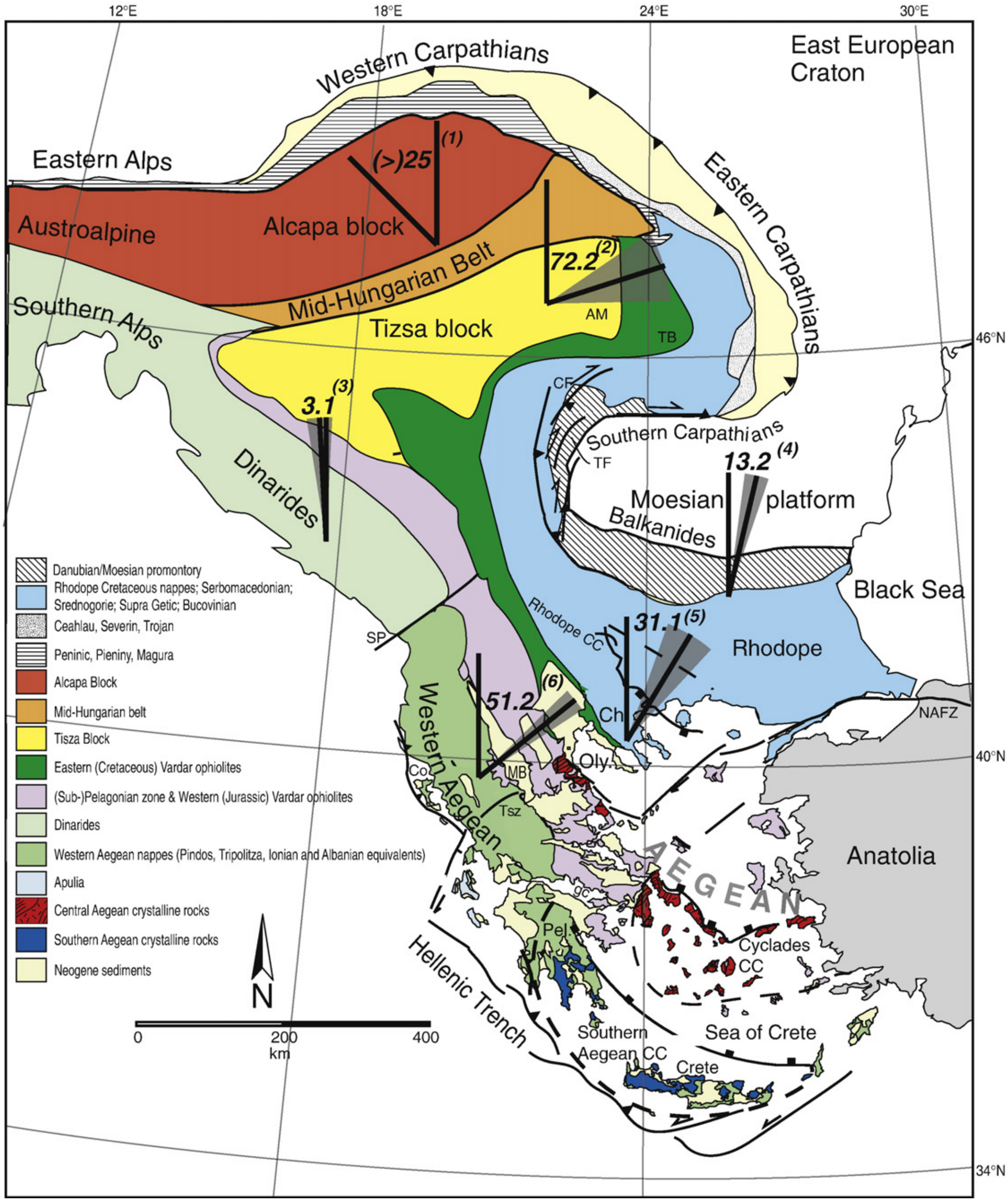


Fig. 1. Schematic geological map with the main tectonostratigraphic units of the southeastern Europe, with the declinations obtained from the Eo-Oligocene of the main blocks of the region. References for the composite declinations constructed from the available literature: 1. Alcapa-block—we adopt the ~25° counterclockwise rotation argued for on structural/palinspastic basis by Ustaszewski et al. (in press). See this reference for discussion and review of the paleomagnetic and structural data; 2. Tisza Block—see Appendix A; 3. Dinarides—data from Kissel et al. (1995), see Appendix A; 4. Moesian Platform and Bulgarian Rhodope—This study; 5. Chalkidiki Peninsula—data from Kondopoulou and Westphal (1986), see Appendix A; 6. Western Greece and Albania—see Appendix A for data and references. Map modified after van Hinsbergen et al. (2005c) and Schmid et al. (2008). AM = Apuseni Mountains; CC = crystalline complex; CF = Cerna-Jiu Fault; Ch = Chalkidiki peninsula; gc = Gulf of Corinth; MB = Mesohellenic Basin; SP = Scutari-Pec Transform Fault; TB = Transylvanian Basin; TF = Timok Fault; Tsz = Thesprotiko Shear Zone.

the Platform during or after early to mid-Eocene contractional tectonics and remagnetisation (Jordanova et al., 2001). Along the northern margin of the Moesian Platform, the foreland basin sediments of the southern Carpathians, record a phase of 30° clockwise rotation between 13 and 6 Ma (Dupont-Nivet et al., 2005). This rotation phase is justifiably interpreted to relate to dextral shear associated with the eastward emplacement of the Tisza block, assuming the Moesian Platform did not rotate. However, another speculative interpretation—not excluded by existing constraints—is that the Tisza block and the Southern Carpathians rotated clockwise together with the Moesian Platform between 13 and 6 Ma. The fact that this rotation phase is contemporaneous with the west Aegean rotations, further suggests that the Tisza block and the western Aegean region may have (partly) rotated together. This would have major implications for the loci of kinematic (extensional) accommodation of the Aegean and Carpathian block rotations (i.e. the west-Aegean rotations may have been partly accommodated in the Carpathian back-arc). Assessing the rotation of the Moesian Platform is therefore a key element that is still missing to test these different hypotheses for the southeastern European geodynamics. In this study, we provide new paleomagnetic from the Moesian Platform and Bulgarian Rhodope to quantify its rotation since Eocene time and discuss the implications of our new findings for the kinematic evolution of the Carpathian and Aegean arcs.

2. Geological setting

The Moesian Platform extends over large parts of Bulgaria and Romania, from the southern Carpathians in the north, to the Balkan fold-and-thrust belt (Balkanides) in the south (Fig. 1). Its metamorphic basement is covered by >10 km of lower Paleozoic to Recent sediments and some volcanics (Carboniferous–Permian, Triassic and Miocene) (Tari et al., 1997). Alpine deformation over the southern margin of the Moesian Platform was associated with the northward emplacement of the Balkanides and Srednogorie nappes in a back-thrust system associated with northward subduction during Africa–Europe convergence (Boccaletti et al., 1974; Ricou et al., 1998; van Hinsbergen et al., 2005b), leading to an upper Cretaceous to Paleogene foreland basin stratigraphy covering the Platform (Tari et al., 1997). The contractional deformation along the southern edge of the Platform ended during the middle Eocene along most of its length, although some local contraction in the southeastern part continued in the Oligocene (Sinclair et al., 1997). The Rhodope forms a complex structure including exhumed high-grade metamorphic rocks buried and exhumed during the Cretaceous to Paleogene (and in northern Greece into the Oligo-Miocene) (Ricou et al., 1998; Brun and Sokoutis, 2007), overlain by Eocene to Oligocene sedimentary basins and volcanic fields (Lilov et al., 1987; Yanev and Pecskay, 1997; Yanev et al., 1998; Fig. 2). Significant contraction south of the Moesian Platform ended in the Eocene (e.g. Ricou et al., 1998), after which the loci of deformation shifted to the south as accretion migrated southward (van Hinsbergen et al., 2005c). By this time, the Rhodope had accreted to the Moesian Platform as indicated by the absence of major post-Eocene structures that could have accommodated significant motion (rotation or translation) between these blocks. Post-middle Eocene exhumation in the Rhodope was largely accommodated by tectonic denudation in core complexes occurring south of the volcanic fields of the Bulgarian Rhodope (Dinter and Royden, 1993; Ricou et al., 1998; Krohe and Mposkos, 2002; Brun and Sokoutis, 2007). This extensional history has formed a series of post-middle Eocene grabens and half-grabens in the southern Rhodope (e.g. the middle-late Miocene Struma (or Strimon) and middle Eocene Mesta grabens) cross-cutting the pre-Eocene nappe stack (Tzankov et al., 1996; Nakov, 2001; Burchfiel et al., 2003).

Along the northern margin of the Platform—south of the Carpathians—late Cretaceous nappe stacking is followed by significant post-Eocene deformation (Fügenschuh and Schmid, 2005). Although contractional deformation along the southern margin of the Moesian Platform ended in the Eocene, the western and northern margins were subjected to overthrusting and right-lateral wrenching associated with northward and eastward propagation of the Carpathian fold-and-thrust belt around the Moesian Platform from Eocene to Pliocene time (ending in the late Miocene in the Southern Carpathians, and Pliocene in the Eastern Carpathians) (Schmid et al., 1998; Bertotti et al., 2003; Matenco et al., 2003; Dupont-Nivet et al., 2005; Vasiliev et al., 2005). This north-eastwards migration of the Carpathian arc is believed to result from eastward slab roll-back of (oceanic?) lithosphere that occupied the present Carpathian–Pannonian region (Carpathian embayment) prior to subduction (Linzer et al., 1998; Schmid et al., 1998; Wortel and Spakman, 2000; Csontos and Voros, 2004). Arc migration was stopped by subsequent continental collision between the accreted allochthonous terranes and the East European cratonic margin (Morley, 1996; Linzer et al., 1998; Zweigel et al., 1998; Matenco and Bertotti, 2000; Fig. 1). In this context, the southern Carpathians have been interpreted as a subduction transform edge propagator (STEP) fault (Govers and Wortel, 2005). The formation of the Carpathians resulted in a distinctive depositional pattern of thick Miocene to Pliocene foreland deposits on the Moesian Platform that are exposed in thrust sheets along the Southern Carpathians. Basin analysis and modeling indicates accelerated subsidence during the early to middle Miocene time followed by a climax of subsidence between ~13 and 9 Ma (Bertotti et al., 2003; Matenco et al., 2003; Cloetingh et al., 2004). In summary, the Moesian platform was, in the south, overthrusted and deformed until mid-Eocene times due to subduction/collision processes in the Aegean arc, and, along its northern margin, transpressionally deformed by eastward propagation of the Carpathian arc mainly during Miocene times.

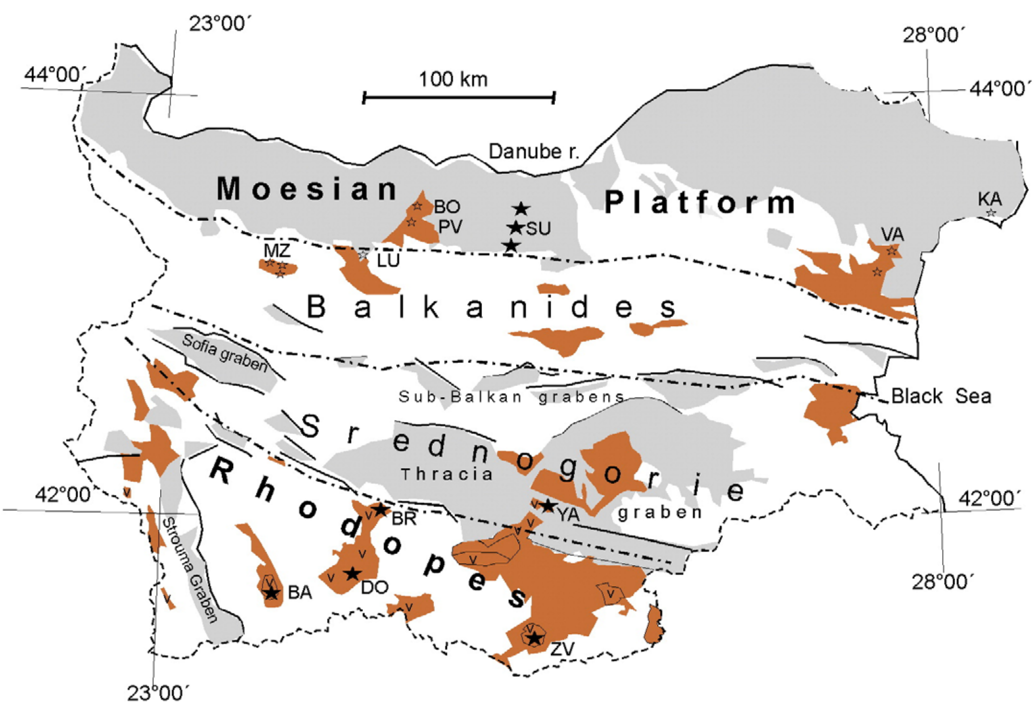
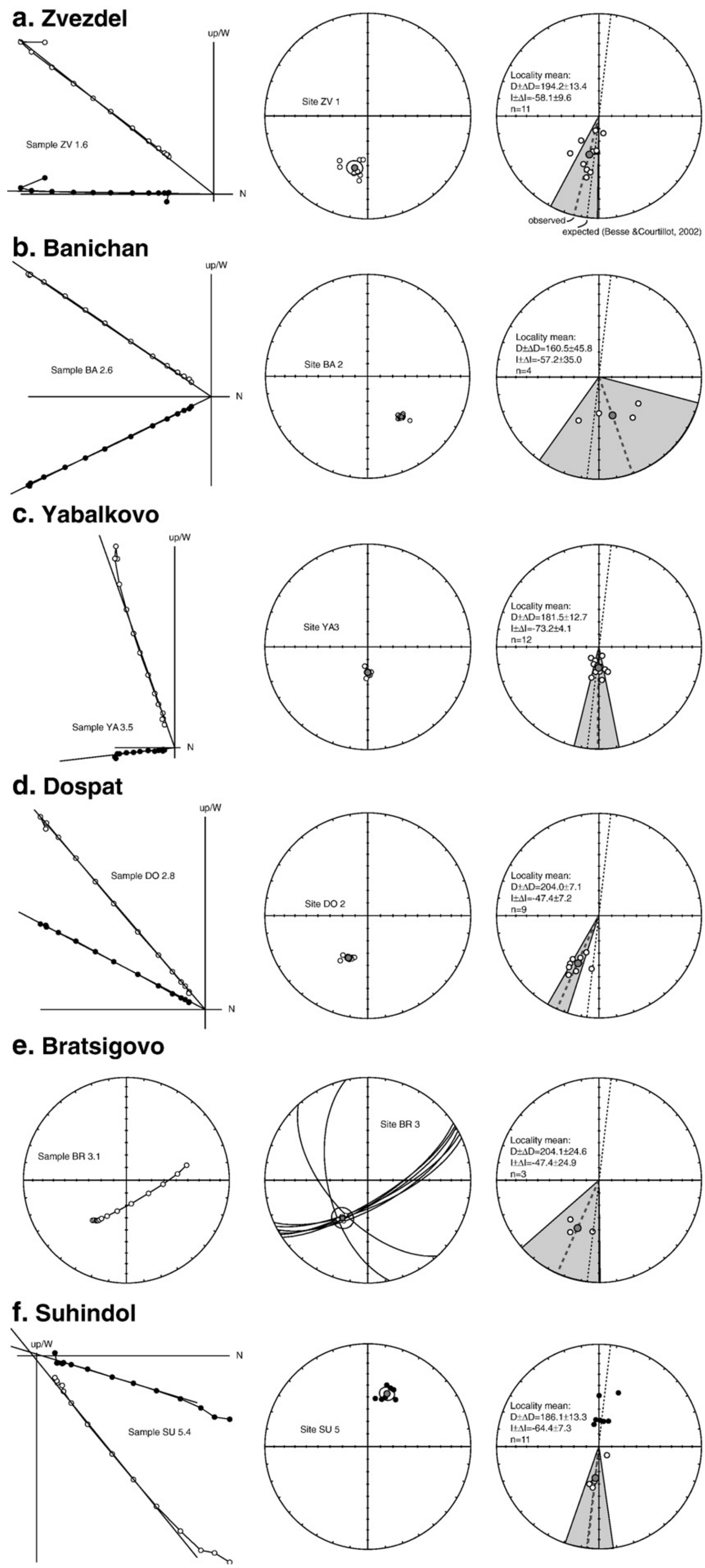


Fig. 2. Structural sketch-map of Bulgaria with shown post-Cretaceous rock. Light (grey)—Neogene–Quaternary sediments (except some locally occurring basalts in central North Bulgaria—sites Suhindol (SU)); Dark (orange)—Paleogene rocks, locally including lower Miocene (Thracia graben and western Bulgaria); v—volcanics; Thick lines—middle Miocene to present extensional faults, forming grabens; dashed line—boundaries of tectonic units; MZ, LU, PV, BO, SU, KA, VA, YA, BR, DO, ZV, BA—sampled localities, fore precise coordinates and names see Table 1 and Appendix A; Filled star—successful site; unfilled—unsuccessful site.


a. Zvezdel
up/W
Sample ZV 1.6
N
Site ZV 1
Locality mean:
D±ΔD=194.2±13.4
I±ΔI=-58.1±9.6
n=11
observed
expected (Besse &Courtillot, 2002)
b. Banichan
up/W
Sample BA 2.6
N
Site BA 2
Locality mean:
D±ΔD=160.5±45.8
I±ΔI=-57.2±35.0
n=4
c. Yabalkovo
up/W
Sample YA 3.5
N
Site YA3
Locality mean:
D±ΔD=181.5±12.7
I±ΔI=-73.2±4.1
n=12
d. Dospat
up/W
Sample DO 2.8
N
Site DO 2
Locality mean:
D±ΔD=204.0±7.1
I±ΔI=-47.4±7.2
n=9
e. Bratsigovo
Sample BR 3.1
Site BR 3
Locality mean:
D±ΔD=204.1±24.6
I±ΔI=-47.4±24.9
n=3
f. Suhindol
up/W
N
Sample SU 5.4
Site SU 5
Locality mean:
D±ΔD=186.1±13.3
I±ΔI=-64.4±7.3
n=11

3. Paleomagnetic sampling

To test whether the Moesian Platform can indeed be considered as stable in the Cenozoic, we sampled upper Cretaceous and younger sediments in the central region of the Moesian Platform (away from deformation at the periphery) where they are found mostly undeformed with subhorizontal bedding orientations (Tari et al., 1997). Because most of these sediments are platform carbonates and marls, which may poorly record the paleomagnetic field, we also sampled lower Miocene volcanic rocks exposed on the central Moesian Platform as well as Eocene to Oligocene volcanoes on the Bulgarian Rhodope, to ensure successful results. As outlined above, the thrusts emplacing the Rhodope and underlying nappes over the Moesian Platform predate the flat-lying volcanic field of the Rhodope. In addition, post-Eocene extensional exhumation phases and associated vertical-axis rotations occur only south of the Bulgarian volcanic fields, e.g. between the Chalkidiki peninsula and the Rhodope, on either side of the Strimon detachment (Kondopoulou and Westphal, 1986; Dimitriadis et al., 1998; Brun and Sokoutis, 2007). Therefore, there are no reasons to expect regional rotation differences between the Rhodope and the Moesian Platform after the Eocene.

We collected approximately 800 samples from 12 localities on the Moesian Platform and Bulgarian Rhodope (Fig. 2). Paleomagnetic samples were collected with a hand-held gasoline-powered drill with water-cooled diamond-coated drill bits. The orientation of all samples was measured with a magnetic compass, and corrected for local declination (4°). Six localities were sampled in upper Cretaceous–Paleocene (Mezdra, Bojouritsa), Eocene (Varna, Pleven) and Miocene (Lukovit, Kavarna) sedimentary rocks of the Moesian Platform. Additionally, we collected (in most cases eight) samples from a total of 58 lava sites in six localities. Twelve sites were sampled in three lower Miocene volcanic plugs (24.0–19.4 Ma K/Ar ages: Yanev et al., 1993) at a locality north of Suhindol (Butovo, Dragomirovo, Chervena; Fig. 2). Although the host sedimentary rocks in which these plugs intruded are not well exposed here, their position in the heart of the Moesian Platform (where observed post-Eocene tilts nowhere exceed 5°), leads us to assume that they did not undergo significant post-emplacement tilting. Furthermore, we collected flat-lying volcanics (lavas and ignimbrites) from volcanoes in the Bulgarian Rhodope. A wealth of K/Ar and Ar/Ar ages have been published for these volcanic rocks, generally clustering between ~31 and 37 Ma, with younger dike swarms down to 25 Ma (Lilov et al., 1987; Yanev and Pecskay, 1997; Yanev et al., 1998; Pecskay et al., 2000). Volcanism in the nearby Greek Rhodope has yielded comparable ages, ranging from ~25 to 31 Ma (Innocenti et al., 1984). Three of the volcanic centers that we have sampled have published ages: Zvezdel (twelve lava sites), with ages of 31–33 Ma (Lilov et al., 1987), and our locality Yabalkovo (twelve lava sites), with ages of 31.5–35 Ma (Lilov et al., 1987). Three other localities were sampled in felsic volcanics of Bratsigovo (four ignimbrite sites), Dospat (ten ignimbrite sites) and the lavas of Banichan (eight lava sites with an age of 28–29 Ma: Pecskay et al., 2000) in the western part of the Rhodope Mountains.

4. Paleomagnetic analysis

4.1. Rock magnetism and sample treatment

For all samples, the natural remanent magnetisation (NRM) of the specimens was measured on a 2G Enterprises horizontal DC SQUID cryogenic magnetometer (noise level 3×10^{-12} Am^2). The samples were demagnetized stepwise using either thermal (TH) demagnetization for sediments or alternating field (AF) demagnetization for volcanic rocks. Heating took place in a magnetically shielded, laboratory-built furnace using small temperature increments of 30–80 °C. Unfortunately, none of the demagnetizations from the sedimentary sites yielded interpretable paleomagnetic directions, which is mainly the result of very low initial intensities. In Appendix A, we give the GPS coordinates of the sedimentary sites. For the volcanic rocks, AF demagnetization was carried out with 5–20 mT increments up to 120 mT with a degausser interfaced with the magnetometer by a laboratory-built automated measuring device. Identification of the characteristic remanent magnetization (ChRM) was done upon inspection of decay curves, equal-area projections and vector end-point diagrams (Zijderveld, 1967; Fig. 3). Initial intensities range typically from 0.5 to 2.0 A/m. For the volcanic rocks, results reveal fairly simple demagnetization behavior. Within a few sites, one or several samples indicate an overlapping overprint of abnormally high magnetic intensity with random directions likely related to lightning strikes (Fig. 3e). In most samples, however, univectorial decay towards the origin of 90% of the NRM occurs between 15 and 70 mT and is thus defined as the ChRM. This range of unblocking applied AF is typical for titanomagnetite or titanomaghaemite coercivities (Dunlop and Özdemir, 1997). To assess the magnetic mineralogy of the ChRM, the variation of low-field magnetic susceptibility of volcanic rocks from 40 °C to 700 °C has been studied on some representative samples using a KLY3 Kappabridge susceptibility meter with attached CS-3 furnace (Fig. 4). Crushed powder material (~200 mg) was taken from one characteristic specimen from each locality, and was heated and cooled in air in two successive cycles: 40–400–40 °C and 40–700–40 °C. During heating, a moderate increase of susceptibility between 40 and ~300 °C, is followed by a decreases between ~300 and ~400 °C and a sharp decrease between ~525 and 580 °C. During cooling, the variations observed during heating in the lower temperature range (40–400 °C) are not reversible. In contrast, variations in the higher temperature range are reversible. The reversible decrease above 525 °C clearly corresponds to Ti-poor titanomagnetite—probably magnetite—with a Curie temperature close to 580 °C. The lower temperature irreversible variations can be related to titanomaghaemite (rather than Ti-rich titanomagnetite which would result in reversibility at these temperatures (e.g. Biggin et al., 2007)). Titanomaghaemite can result from secondary alteration such as often found in basaltic rocks (e.g. Zhao et al., 2006). The fact that the total intensity is lower after cycling suggests low-Ti titanomaghaemite or maghaemite which inverts to haematite upon heating (rather than Ti-rich titanomaghaemite that would invert to magnetite (Dunlop and Özdemir, 1997)). The lack of secondary directions in the demagnetization diagrams, strongly suggest that maghaemitization did not alter the original paleomagnetic signal carried by Ti-poor titanomagnetite.

4.2. ChRM direction analysis

The few lava sites affected by lightning have within-site randomness of NRM directions and demagnetization paths along great circles that could be identified on equal-area projections and interpreted using the great-circle analysis of McFadden and McElhinny (1988). Otherwise, the most of ChRM directions were calculated by principal component analysis (Kirschvink, 1980). ChRM directions with maximum angular deviation exceeding 15° were rejected from further

Fig. 3. Paleomagnetic results for each locality. Left column shows typical AF demagnetisations (steps in mT: 0, 5, 10, 15, 20, 25, 30, 40, 50, 60, 70, 80, 90, 100) on vector end-point diagrams with solid (open) symbols for projection on the horizontal (vertical) plane. Lines indicates least square line fit (Kirschvink, 1980), except for e. showing equal-area stereographic projection of demagnetisation path along great circle. Central column shows typical site equal-area projection of ChRM directions in upper (open symbols) and lower (solid symbols) hemisphere. Right column shows site-mean directions (all transformed to normal polarity orientation) with associated locality averages (D=declination, I=inclination, ΔD=95% confidence in declination, ΔI=95% confidence in inclination, *n*=number sites). All results are given in tilt-corrected coordinates. Data and statistical parameters are given in Table 1.

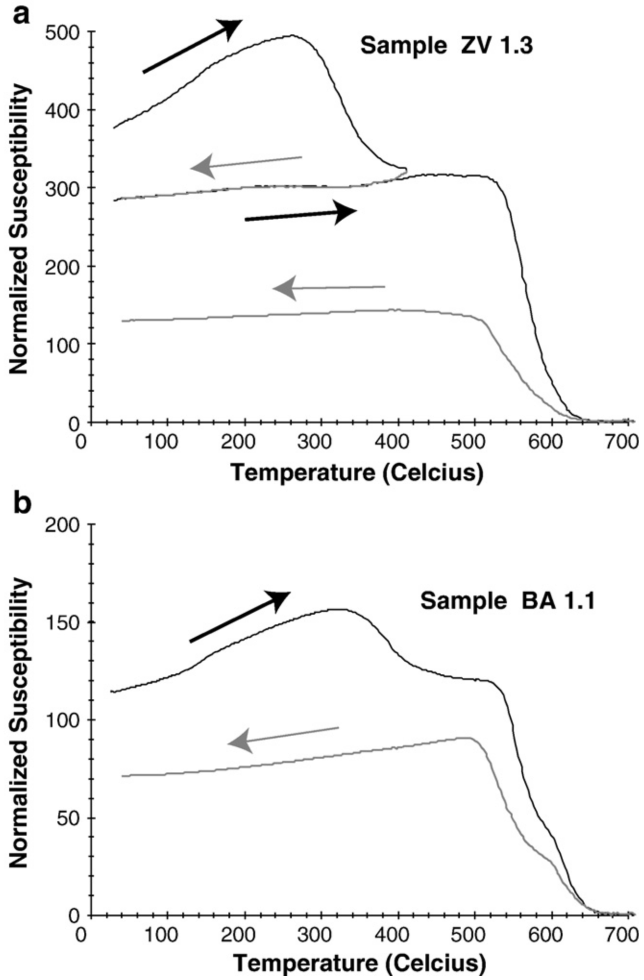


Fig. 4. Typical weak field thermomagnetic curves of normalised susceptibility for representative samples during cycles of heating and subsequent cooling (indicated by arrows). (a) two cycles at 40–400–40 °C and then 40–700–40 °C; (b) one cycle at 40–700–40 °C, see text.

analysis. For each site, the Vandamme (1994) cut-off was applied to discard widely outlying ChRM directions. Averages and cones of confidence on lava site-mean directions were determined using Fisher (1953) statistics assuming random within-site errors (Table 1). We applied a stringent cut-off, rejecting lava site-means with $k<50$, since expected within-site scatter in a lava recording a spot-reading of the Earth's magnetic field should be minimal. In total, 393 samples from 51 sites from six volcanic localities provided conclusive directions, or well-constrained great circles (Table 1). Two sites from the ignimbritic Dospat locality sharing a common true mean direction at the A level of McFadden and Lowes (1981) were combined into one site on the basis that they probably represent the same spot-reading of the magnetic field. This may result from the same ignimbrite being inadvertently sampled twice or two ignimbrites formed within years or tens of years. Application of the fold test is precluded by the absence of sufficient bedding-tilt within and between localities. The outcome of the parametric reversals test of Tauxe (1998) is positive with 95% confidence for the Suhindal locality recording dual polarity (all other localities have exclusively reverse polarity), supporting a primary origin for the magnetisation. For each site-mean direction, the associated Virtual Geomagnetic Pole (VGP) position was calculated. For each locality and for all sites combined, the average VGP with associated 95% confidence circle (A95) was calculated using Fisher statistics (Fisher, 1953); Table 1). Within each locality, the relatively low amounts of VGPs (3 to 12) do not provide sufficient sampling of the geomagnetic field to confidently average out secular variation. In contrast, a meaningful number of sites to average out secular variation is provided by combining our 50 site-VGPs to construct a single VGP for the Moesian Platform and the Bulgarian Rhodope. After exclusion of two outlying VGPs using the Vandamme (1994) cut-off, the remaining 48 site-VGPs yield a late Eocene–early Oligocene pole for the Moesian Platform and Bulgarian Rhodope ($\lambda=80.2°$N; $\varphi=96.9°$E; $A_{95}=4.8°$; $K=19.3$; Scatter$=18.8$). This VGP is statistically indistinguishable from the Eurasian Apparent Polar Wander Path (APWP) of this time of Besse and Courtillot (2002) ($\lambda=82.8°$N; $\varphi=158.1°$E; $A_{95}=3.8°$), Schettino and Scotese (2005) ($\lambda=84.1°$N; $\varphi=175.8°$E; $A_{95}=8.1°$) and Torsvik et al. (in press) ($\lambda=82.7°$N; $\varphi=152.5°$E; $A_{95}=2.8°$). In addition, the obtained VGP scatter is comparable to geomagnetic field model predictions for scatter values at this latitude (McFadden et al., 1991). Together with the positive reversals test and the rock magnetic analysis, these observations strongly suggest our results record primary magnetisations providing suitable directions for rotational analysis.

4.3. Rotation analysis

To estimate the possible rotation with respect to a stable reference, we compare VGPs to existing Eocene to Oligocene APWPs of Eurasia at 20 and 30 Ma (Figure Plots). Observed inclination and declination with associated ΔD and ΔI (the error bars on the declination and inclination, respectively, calculated through Eqs. A.60 and A.57 in Bulter (1992)) are compared to expected declination and inclination calculated from the APWPs at an average sampling location (41.56°N, 25.33°E). Based on our dataset, we have no ground to suspect significant rotation differences between our localities. Locality-mean declinations are within 95% confidence intervals of each other and statistically indistinguishable from declinations expected by APWP of stable Eurasia except for a slight offset of the Dospat ($D\pm\Delta D=203.9\pm6.2°$) locality with respect to Yabalkovo locality ($D\pm\Delta D=181.5\pm12.7°$; see Fig. 3). However, comparison between locality-mean directions can be considered meaningless in terms or rotation or absence of rotation, because any directional difference can easily be the result of secular variation not averaged out by the insufficient number of sites within each locality. Consequently, the slight declination offset from the locality Dospat may result from insufficient sampling of secular variation (as suggested by the lower scatter for this locality), and should not be taken as hard evidence for a local rotation with respect to other localities. Moreover, significant rotations between our localities would result in higher than expected scatter values for the overall VGP of the Moesian Platform and Bulgarian Rhodope that averages all our site-VGPs. As noticed above, the scatter associated with this VGP is not higher than the expected scatter values, and thus suggests no significant rotations between localities. This is further supported by the lack of important regional deformation after emplacement of these flat-lying rocks. Given that no significant internal rotations can be identified, we further investigate the possibility of a wholesale regional rotation by comparing the overall VGP of the Moesian Platform and Bulgarian Rhodope to 20 Ma and 30 Ma poles from three existing APWPs of Eurasia (Fig. 5a). In all cases the rotation is not statistically significant at the 95% confidence level, thus precluding important wholesale regional rotation after the late Eocene–early Oligocene formation of these rocks.

An opportunity to expand our dataset is provided by the existence of published data from 15 paleomagnetic site-mean directions from upper Eocene–lower Oligocene volcanics located directly south of the Bulgarian Rhodope (nearby Thrace, across the border in northern Greece (Kissel et al., 1986a)). 10 out of these 15 sites are selected after applying our cut-off of $k<50$. These provide similar results to the one obtained in the present study (Fig. 5b). The 10 VGPs calculated are combined to our original 50-VGP dataset. After 4 VGPs are removed by the Vandamme (1994) cut-off, we derived an overall VGP

Table 1
New and published paleomagnetic results of Eocene–Oligocene volcanic sites of the Moesian Platform and Rhodope

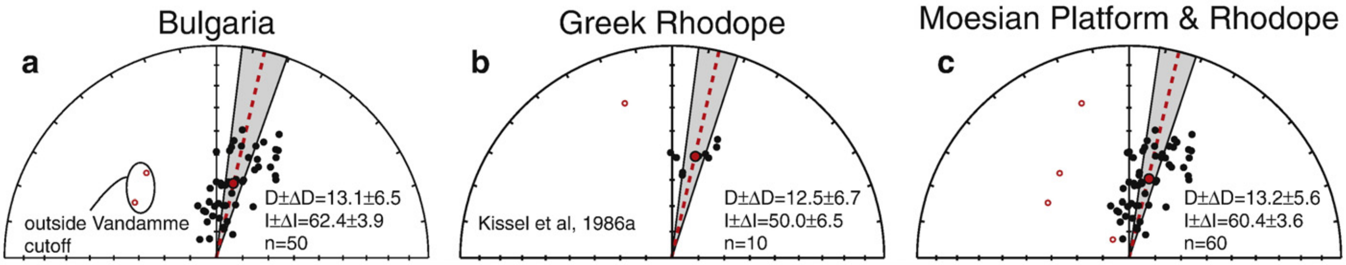
Locality						Tilt	Tilt Corrected					Tilt Corrected							Age
Lava/neck site	Lat	Lon	Type	Na	Nc	strike/dip	λ	ϕ	*k*	A95	Palat	D	ΔD	I	ΔI	*k*	α95	pol	Ma
New data																			
Suhindol				12	11		84.1	62.1	25.2	9.3	45.3	6.1	13.3	64.4	7.3			n/r	19.4–24.0 [43]
SU 1*	43.1681	25.1263	gc	8	6	000/00	83.0	126.9				189.2		−60.4				r	
SU 2*	43.1681	25.1263	gc	8	7	000/00	84.6	113.1				187.4		−61.9				r	
SU 3	43.1681	25.1263	gc	8	7	000/00	78.6	128.1				194.5		−58.9		896.6	2.0	r	
SU 4	43.1681	25.1263	gc	8	7	000/00						188.8		−56.5		44.2	9.2	r	
SU 5	43.2902	25.1736		8	7	000/00	65.9	156.9				19.7		43.5		141.1	5.1	n	
SU 6	43.2902	25.1736	gc	8	8	000/00	76.4	−155.2				0.1		48.7		75.5	6.4	n	
SU 7	43.2902	25.1736	gc	8	8	000/00	55.2	3.6				138.1		−80.6		987.2	1.8	r	
SU 8	43.2285	25.1526		8	8	000/00	80.5	34.3				2.5		69.1		205.2	4.2	n	
SU 9	43.2285	25.1526		8	8	000/00	77.8	55.5				10.3		69.6		193.2	4.0	n	
SU 10	43.2285	25.1526		8	8	000/00	73.9	76.6				20.5		68.4		66.3	6.9	n	
SU 11	43.2285	25.1526		8	7	000/00	80.4	−1.7				353.0		68.4		113.8	5.7	n	
SU 12	43.2285	25.1526	gc	8	6	000/00	74.1	−1.6				347.0		71.9		993.3	2.1	n	
Yabalkovo				12	12		72.5	28.1	45.1	6.5	58.9	181.5	12.7	−73.2	4.1			r	31.5–35.0 [27]
YA 1	42.0415	25.2345	gc	8	8	000/00	57.3	48.1				215.6		−79.1		181.2	4.1	r	
YA 2	42.0415	25.2345	gc	8	8	000/00	62.6	38.1				195.9		−78.6		215.0	3.8	r	
YA 3	42.0415	25.2345		8	8	000/00	77.7	20.0				178.1		−70.2		544.2	2.4	r	
YA 4	42.0415	25.2345	gc	8	7	000/00	67.0	45.7				197.3		−75.5		120.7	5.5	r	
YA 5	42.0446	25.2450	gc	8	8	254/05	71.6	17.2				174.9		−74.0		1975.7	1.2	r	
YA 6	42.0446	25.2450		8	7	254/05	73.0	−4.3				165.0		−71.4		370.2	3.1	r	
YA 7	42.0446	25.2450		8	7	254/05	55.9	16.5				160.7		−82.4		1283.6	1.7	r	
YA 8	42.0446	25.2450		8	7	254/05	73.2	−20.8				160.2		−68.8		313.5	3.1	r	
YA 9	42.0346	25.2773		8	8	000/00	70.3	32.5				185.1		−74.8		243.1	3.6	r	
YA 10	42.0346	25.2773		8	8	000/00	85.4	−25.4				175.0		−63.3		68.8	6.7	r	
YA 11	42.0346	25.2773		8	8	000/00	77.3	46.5				187.7		−69.8		53.9	7.6	r	
YA 12	42.0346	25.2773		8	8	000/00	79.1	86.5				193.9		−64.6		128.8	4.9	r	
Zvezdel				12	11		79.5	112.9	20.2	10.4	38.8	194.2	13.4	−58.1	9.6			r	31.0–33.0 [27]
ZV 1	41.2781	25.2499		9	9	000/00	72.0	165.1				192.9		−45.2		66.2	6.4	r	
ZV 2	41.2745	25.2493	gc	8	8	000/00	82.6	106.0				189.9		−61.0		328.6	3.1	r	
ZV 3	41.2727	25.2488		8	5	000/00	87.1	91.2				183.6		−61.3		153.0	6.2	r	
ZV 4	41.2712	25.2478	gc	8	8	000/00	68.2	175.1				191.6		−38.6		70.5	6.6	r	
ZV 5	41.2656	25.2511		8	8	000/00	86.2	78.3				184.2		−62.2		124.2	5.0	r	
ZV 6	41.2651	25.2509		8	8	000/00	72.6	178.1				188.7		−43.5		253.9	3.5	r	
ZV 7	41.2542	25.2561		8	–	000/00						–		–		–	–		
ZV 8	41.2539	25.2561		8	7	000/00	63.7	88.3				216.1		−65.7		219.9	4.1	r	
ZV 9	41.2471	25.2567		8	8	000/00	71.4	147.5				198.1		−48.9		82.7	6.1	r	
ZV 10	41.2469	25.2480	gc	8	7	000/00	58.8	119.4				217.7		−51.8		122.9	5.5	r	
ZV 11	41.2447	25.2473		9	9	324/15	63.0	37.1				193.9		−78.1		218.8	3.5	r	
ZV 12	41.2444	25.2466		8	8	324/15	66.4	9.3				165.8		−75.9		93.7	5.8	r	
Bratsigovo				4	3		66.4	140.9	35.4	21.0	30.7	204.1	24.6	−47.4	24.9			r	late Eo-Oligocene
BR 1	42.0175	24.2601	gc	8	8	224/20	75.7	177.4				187.3		−48.0		122.0	5.0	r	
BR 2	42.0173	24.2603		8	8	244/12	60.0	142.3				208.9		−41.5		201.1	3.9	r	
BR 3	42.0173	24.2603	gc	8	8	244/12	59.5	122.7				216.0		−50.7		3401.8	0.9	r	
BR 4	42.0169	24.2604	gc	8	8	244/12						236.8		−54.2		28.6	10.5	r	
Dospat				10	9		67.1	139.1	72.3	6.1	30.7	204.0	7.1	−47.4	7.2			r	late Eo-Oligocene
DO 1	41.4102	24.0940		8	8	000/00	74.9	116.8				199.6		−58.3		264.6	4.1	r	
DO 2	41.4102	24.0940		8	7	000/00	60.1	135.3				211.1		−44.3		220.3	4.1	r	
DO 3–5	41.4103	24.0922		16	16	000/00	69.0	134.6				203.3		−50.6		217.8	2.5	r	
DO 4	41.4104	24.0919		8	8	000/00	62.3	128.9				211.2		−48.8		163.8	4.3	r	
DO 6	41.4111	24.0917		8	8	000/00	63.9	151.0				202.5		−40.8		372.4	2.9	r	
DO 7	41.4131	24.0920		8	8	000/00	75.2	114.1				199.4		−59.0		94.8	5.7	r	
DO 8	41.4130	24.0912		8	7	000/00	74.8	175.7				188.1		−46.4		198.7	4.3	r	

(continued on next page)

Table 1 *(continued)*

Locality						Tilt	Tilt Corrected					Tilt Corrected							Age
Lava/neck site	Lat	Lon	Type	Na	Nc	strike/dip	λ	ϕ	*k*	A95	Palat	D	ΔD	I	ΔI	*k*	α95	pol	Ma
DO 9	41.4135	24.0878		8	8	000/00	59.3	139.8				210.2		−41.3		185.2	4.1	r	
DO 10	41.4138	24.0876		8	8	000/00	58.1	149.6				207.1		−35.1		147.3	4.6	r	
Banichan				8	4		75.3	286.6	7.1	37.2	32.5	160.5	45.8	−57.2	35.0			n/r	28–29 Ma [44]
BA 1	41.3862	23.4203	gc	8	6	000/00	89.4	15.7				179.9		−60.9		94.1	6.9	r	
BA 2	41.3857	23.4211		8	6	000/00	55.4	−75.9				140.4		−47.2		332.1	3.7	r	
BA 3	41.3854	23.4218		9	7	000/00	45.2	−59.4				124.1		−51.7		58.9	7.9	r	
BA 4	41.3854	23.4218		8	8	000/00						133.7		−61.2		31.9	10.0	r	
BA 5	41.3832	23.4240		8	5	000/00						162.2		−42.2		36.2	12.9	r	
BA 6	41.3820	23.4254		8	8	000/00						145.0		−49.3		37.5	9.2	r	
BA 7	41.3817	23.4354		8	5	000/00						141.2		−43.3		26.2	13.3	r	
BA 8	41.3807	23.4254	gc	8	8	054/15	67.8	130.9				25.1		51.1		221.3	3.7	n	
20 Ma pole				12	11		84.1	62.1	25.2	9.3	45.3	6.1	13.3	64.4	7.3			n/r	
30 Ma pole				39	37		79.9	101.7	18.3	5.7	43.2	13.5	7.8	61.9	4.7			n/r	
Combined pole				50	48		80.2	96.9	19.3	4.8	43.9	13.0	6.7	62.6	3.9			n/r	
* site average is a great circle: direction is determined comparing these great circles with the directions from this locality using the method of [50]																			
Previously published data																			
(Kissel et al. 1986a)																			
Thessaly (Greece)				15	9		75.5	157.5	80.6	5.8	30.7	12.5	6.7	50.0	6.5			n/r	late Eo-Oligocene
TH 297	41.1125	25.5353			10		73.7	154.4				194.6		−49.0		133.0	3.8	r	
TH 301	40.9012	25.5384			6		80.6	−167.6				2.5		51.0		52.0	8.0	n	
TH 307	40.9013	25.5402			9		84.2	135.8				7.0		58.0		302.0	2.7	n	
TH 300	40.8887	25.5479			8		79.0	−149.9				359.0		49.0		172.0	3.7	n	
TH 306	41.1061	25.5637			9							13.8		35.2		19.0	10.0	n	
TH 292	41.0998	25.8663			10		83.4	148.0				187.0		−56.8		52.0	6.0	r	
TH 296	41.2417	25.8883			9							189.9		−58.7		19.0	10.2	r	
TH 291	41.1566	25.8883			10		65.7	150.3				22.0		43.2		50.0	6.2	n	
TH 293	41.1250	25.9482			9							335.3		38.3		37.0	7.6	n	
TH 294	41.1755	25.9482			8							229.0		−45.5		10.0	15.4	r	
TH 290	40.9328	26.1752			6		70.2	144.9				200.0		−48.6		107.0	6.0	r	
TH 304	41.1534	26.2161			10							161.5		−27.5		31.0	7.8	r	
TH 289	40.9580	26.2192			10		70.9	151.1				197.8		−47.4		117.0	4.0	r	
TH 303	41.1597	26.2287			11		58.1	−120.7				162.8		−24.8		73.0	5.0	r	
TH 288	40.9832	26.2414			5		64.6	155.8				201.0		−40.0		84.0	6.8	r	
Eo-Oligocene pole																			
Combined pole				60	56		80.1	112.4	21.9	4.2	41.3	13.2	5.6	60.4	3.6			n/r	

Lat=latitude of the site; Lon=longitude of the site; Type indicates ‘gc’ when the site average was constructed using the great circle method of McFadden and McElhinny (1988); *N*a=number of samples analysed; *N*c=number of sampled used to construct the site average; Tilt: strike and dip applying the left hand rule; λ =longitude of the virtual geomagnetic pole (VGP); ϕ =latitude of the VGP; K=Fisher (1953) precision parameter for the average pole; A95=95% confidence limit of the average pole; Palat=paleolatitude; D=declination; ΔD=95% confidence limit of the declination; I=inclination; ΔI=95% confidence limit of the inclination; k=Fisher (1953) precision parameter for the average direction (in case of lava sites); α95=95% confidence limit of the average direction (in case of lava sites); pol=normal (*n*) or reverse (*r*) polarity.



locality	n	*Besse & Courtillot (2002) reference direction (30 Ma)*				*Torsvik et al (2008) reference direction (30 Ma)*				*Schettino & Scotese (2004) reference direction (30 Ma)*			
Moesian platform/Rhodope		rotation	error	flattening	error	rotation	error	flattening	error	rotation	error	flattening	error
20 Ma pole	11	0.1	13.7	-5.9	7.7	0.1	13.9	-5.6	7.8	3.6	14.3	-7.2	8.3
30 Ma pole (Bulg. + Greece)	46	6.9	9.1	-6.0	6.0	6.2	9.1	-5.5	6.0	9.9	12.8	-6.1	9.2
all	56	6.6	7.3	-4.5	5.2	5.9	6.6	-4.0	4.5	9.6	11.5	-4.6	8.7

Fig. 5. Equal-area projections for the Bulgarian data (a), the northern Greek data of Kissel et al. (1986a) (b) and the combination of all available data from the upper Eocene and Oligocene of the Moesian Platform and Rhodope (c). All data have been plotted as normal directions. Data and statistical parameters are given in Table 1. D = declination, I = inclination, ΔD = 95% confidence limit in declination, ΔI = 95% confidence limit in inclination, n = number of lava sites. Data from the Greek Rhodope were previously published by Kissel et al. (1986a). D: Comparison of our new Eocene–Oligocene average direction for the Moesian Platform and Rhodope with reference directions at 30 Ma obtained from the apparent polar wander paths of Besse and Courtillot (2002), Schettino and Scotese (2005) and Torsvik et al. (in press). Poles are translated to directions at the coordinate 41.56°N/25.33°E.

(n = 56; λ = 80.1°N; φ = 112.4°E; A_{95} = 4.2°; K = 21.9; Scatter = 15.6), yielding D ± ΔD/I ± ΔI of 13.2 ± 5.6/60.4 ± 3.6 (Fig. 5c). This result compares well with the 14.5° declination previously published for five lower Oligocene lavas in the Greek Rhodope (Atzemoglou et al., 1994), however, as the latter publication does not report the lava site-means, we cannot incorporate these results in our average. When compared to the Eurasian APWPs, our data expanded with the Greek Rhodope data also indicate no statistically significant regional rotation (Fig. 5). We therefore conclude that the Moesian Platform and Rhodope can be considered to be stable entities when compared in the kinematic reconstructions, to the large rotations in the Aegean and Carpathian domains since ~30 Ma.

5. Regional kinematic implications

5.1. Implications for the Moesian Platform and Rhodope

The majority of the previous paleomagnetic results obtained from the Moesian Platform (Dolapchieva, 1994) and overlying nappes of the Balkanides have been obtained from pre-Cenozoic units. Analyses in middle Triassic and Jurassic rocks from the Balkanides generally show no deviation from the Eurasian directions (Kruczyk et al., 1988; Muttoni et al., 2000). In some places, however, remagnetization has been identified: South of the Moesian Platform, (Kruczyk et al., 1990) sampled Jurassic sediments from the Srednogorie zone which were remagnetized, possibly in the late Cretaceous. Jordanova et al. (2001) concluded that sites in the Triassic to Cretaceous from the Balkanides of western Bulgaria were remagnetized based on negative fold tests in these sediments. Based on the best fit of the remagnetized directions with comparison with the Eurasian APWP, the age of this remagnetization could be Eocene, which would imply a ~15° clockwise rotation of the Balkanides. Jordanova et al. (2001) suggested that these rotations are local and associated with the last compressional phases that affected the Balkanides in the middle Eocene. This is a possible scenario, given that we find neither significant post-Eocene rotation, nor evidence for a post-Eocene remagnetization. Together with our results, the paleomagnetic data from the Moesian platform indicates that some local rotations (~15° clockwise) may have occurred during Eocene compression but that after this time, no significant rotation affected the Moesian Platform and the Rhodope region.

5.2. Implications for the Carpathian arc

Kinematic reconstructions of the evolution of the Carpathian arc are based on quantitative constraints mostly from structural geology data and vertical-axis block rotations inferred from paleomagnetic studies (e.g. Balla, 1987; Linzer, 1996; Zweigel et al., 1998; Csontos and Voros, 2004). Two distinctive tectonic domains separated by the mid-Hungarian deformed belt (Csontos and Nagymarosy, 1998) have been defined: the northwestern Carpatho–Pannonian domain (the Alcapa block) with systematic counterclockwise rotations and the south-eastern Carpatho–Pannonian domain (the Tisza block) with systematic clockwise rotations (Fig. 1). Based on existing kinematic reconstructions, the motion of Tisza was roughly coeval with the rotation and emplacement of the Alcapa block (following Ustaszewski et al. (in press), we infer a minimum amount of ~25° of counter-clockwise rotation of the Alcapa block since the middle Miocene). These reconstructions assume the Moesian Platform to have formed a stable promontory during the motion of the Tisza block. First, northward motion of Tisza along the western edge of the Moesian Platform is evidenced by late Cretaceous nappe stacking followed by arc-parallel extension dated late Eocene–Oligocene by thermochronology (e.g. Schmid et al., 1998; Fügenschuh and Schmid, 2005). Secondly, the Tisza block—driven by eastward roll-back of the Carpathian slab and northward motion of Apulia—rotated clockwise around the northwestern corner of the Moesian Platform and translated eastward into the Carpathian embayment (Balla, 1987; Sperner et al., 2002; Csontos and Voros, 2004). The timing of this second phase is still controversial because of discrepancies between structural geology data and the age of teconic rotations derived from paleomagnetism (Patrascu et al., 1994; Fügenschuh and Schmid, 2005).

We review here briefly the available paleomagnetic data pertaining to the rotation of the Tisza block, adding data from Cretaceous rocks to the existing review of Dupont-Nivet et al. (2005; see Appendix A). In reviewing the data, we have attempted to exclude results superseded by more reliable paleomagnetic data and/or better age control from the same location. In particular, published results from the Apuseni Mountains with an Eocene–Oligocene age (Patrascu et al., 1994) that are commonly used for kinematic reconstructions (Schmid et al., 1998; Fügenschuh and Schmid, 2005) have been found to be of Maastrichian age (Panaiotu, 1998). When compared to the Eurasian APWP, our review shows that the mean post-Cretaceous clockwise rotation (79.1 ± 9.6°) is indistinguishable from the mean Eocene–Oligocene clockwise rotation (72.2 ± 16.6°) suggesting that no significant rotation can be demonstrated between Cretaceous and Oligocene time and that most of the eastward motion and rotation of Tisza occurred after the Oligocene. The mean post-Eocene–Oligocene data can be directly compared to the reference pole of the Moesian Platform derived from the present study to derive a 68.4 ± 16.7° post-

Oligocene clockwise rotation of Tisza with respect to the Moesian Platform. This rotation is statistically indistinguishable—albeit higher—from the structurally constrained 53° clockwise rotation prediction of Schmid et al. (1998). This prediction is required by (1) the accommodation of 140 km of shortening between the Tisza Block and the East European cratonic margin and (2) the reconstruction of a pivot point for the Tisza block rotation from the arcuate shape of the nortwestern edge of the Moesian Platform (i.e. the curved Timok and Cerna–Jiu fault and associated structures (Fig. 1)). The slightly higher paleomagnetic rotation compared to this prediction may result from the 140 km shortening being an underestimation, the defined pivot point being too far from the Tisza block or, more likely, overestimation of the block rotation as a result of additional clockwise rotations due to local dextral shear associated with the emplacement of the Tisza block. For the Miocene rotation results, it is necessary to distinguish between data of the Tisza block proper (mostly from the Apuseni Mountains and the Transylvania Basin, Fig. 1) from the Southern Carpathians data on the margin of the Tisza block. Indeed, middle Miocene data from foreland deposits of the Southern Carpathians revealed systematic post-13 Ma clockwise rotations (average of 31.6± 6.7°, see Appendix A) interpreted as local rotations due to distributed dextral shear during eastward motion of the Tisza block into the Carpathian embayment (Dupont-Nivet et al., 2005). This suggests distributed dextral shear within the Southern Carpathian belt and supports that this mechanism accommodated the dextral motion without the occurrence of major surface strike-slip faults along the Southern Carpathians as previously suggested (Fügenschuh and Schmid, 2005). The absence of significant rotation in rocks younger than ~6 Ma along the Southern Carpathians suggest that the dextral motion of the Tisza block occurred between ~13 and ~6 Ma, in good agreement with basin analysis data (Bertotti et al., 2003; Matenco et al., 2003). A set of data from the Tisza block proper (Apuseni Mountains and Transylvania Basin) provides a remarkable progressive record during the clockwise rotation between roughly 15 and 10 Ma (Panaiotu, 1998, 1999). Allowing that this set of data is representative for the entire Tisza block, it indicates that most if not all of the rotation occurred in this short time frame. In light of our results showing the post-Eocene stability of the Moesian Platform, we confirm that the Tisza rotation can be fully attributed to the emplacement of the Tisza block around the Moesian Platform with associated dextral shear along its western and northern margin. In addition, the fact that the Moesian Platform has not rotated together with the Tisza block at this time, implies that the coeval southward migration and rotation of the Aegean arc did not include the Moesian Platform, and should be accommodated within the Aegean region.

5.3. Implications for the Aegean arc

A 600-km long region stretching from northern Albania over mainland western Greece to the Peloponnesos underwent approximately 50° clockwise rotation (e.g. Horner and Freeman, 1983; Kissel and Laj, 1988; Morris, 1995; Speranza et al., 1995; van Hinsbergen et al., 2005a). This region is bounded to the north by the Scutari–Pec transform fault (Fig. 1), north of which, in the Dinarides, no Cenozoic rotations occurred (Kissel et al., 1995). The youngest rocks which experienced this amount or rotation include the upper Burdigalian of the Klematia–Paramythia basin in western Greece (with an age as young as ~17 Ma (van Hinsbergen et al., 2005c)) and volcanic rocks on Evia of 15 and 13 Ma (Kissel et al., 1986b; Morris, 1995), leading van Hinsbergen et al. (2005a) to argue that the majority of clockwise rotation occurred after 15 Ma. Rocks of 8 Ma and younger in the Florina–Ptolemais–Lava basin in northern Greece, as well as on Corfu experienced no more than 10° clockwise rotation (van Hinsbergen et al., 2005a), suggesting that the bulk of the west Aegean rotation occurred between 15 and 8 Ma.

Several lines of evidence suggest that this region can be considered as a single rotating block. The western Aegean region displays a series of nappes which are internally folded. The strike of these folds is proportional to the amount of rotation (Horner and Freeman, 1983; van Hinsbergen et al., 2005a) and provides a clear structural grain showing little post-folding deformation of the region. Local rotations remain confined along discrete fault zones such as the Gulf of Corinth and the Thesprotiko shear zone (van Hinsbergen et al., 2006). This, in combination with the very consistent amount of rotation all over this region, led Kissel and Laj (1988) and van Hinsbergen et al. (2005a) to conclude that this region can be considered as a single rotating block, with little internal fragmentation.

Our new reference direction for the Moesian Platform and Rhodope allows us to accurately quantify the rotation difference between western Greece and the Rhodope, which is critical to evaluate kinematic models for the Aegean region. However, for western Greece Kissel and Laj (1988) and van Hinsbergen et al. (2005a) mention 'approximately 50°' of rotation, but do not give statistical values for this average. Therefore, we calculated the average declination and inclination from all published sites obtained from the Oligocene of western Greece and Albania. Ninety site-mean directions remain after discarding locally-rotated sites in central Greece and the northern Peloponnesos, following van Hinsbergen et al. (2005a). The average given in Table 2 (D±ΔD/I±ΔI of 51.2±5.6/34.7±6.5) is based on 85 site-mean directions, (five were excluded by the Vandamme (1994) cut-off: see Appendix A). The outcome of the parametric reversals test of Tauxe (1998) on these data is positive with 95% confidence. The inclination error is very large, which can be straightforwardly explained by inclination shallowing due to compaction in these sediments. This result compared to our newly obtained direction for the Moesian Platform yields a 38.0±7.2° clockwise rotation difference between western Greece and the Moesian Platform/Bulgarian Rhodope (Table 2).

Kissel and Laj (1988) and Kissel et al. (2003) argued that the pivot point of the rotation was located at the northwestern tip of the west Aegean domain in northern Albania at the Scutari–Pec transform fault (Fig. 1). However, 38.0±7.2° clockwise rotation difference between the west-Aegean domain and the Moesian Platform around this pivot point would require 265±50 km of N–S to NE–SW extension in the Aegean back-arc at the latitude of the Gulf of Corinth, 400 km away from northern Albania. The total amount of N–S extension in the

Table 2
Oligocene direction for the west-Aegean domain, and comparison of the declination with our new reference direction for the Moesian Platform and Rhodope

Locality	Reference direction					Pole			Direction						Age
	lat	lon	n	n (vD)	λ	ϕ	κ	A95	palat	D	ΔD	I	ΔI	pol	Ma
West Aegean	38.15	21.44	90	85	58.8	197.7	13.6	4.3	19.1	51.2	4.6	34.7	6.5	n/r	23–33
Moesian Platform	41.56	25.33	60	56	80.1	112.4	21.9	4.2	41.3	13.2	5.6	60.4	3.6	n/r	20–37
Rotation difference										38.0	7.2				

The inclination in the west Aegean domain is very low with respect to the expected inclination, which can be straightforwardly explained by inclination shallowing due to sedimentary compaction. Data used to construct the new west-Aegean paleomagnetic direction for the Oligocene are given in online Appendix B. See caption of Table 1 for explanation of abbreviations in the table header. n=number of sites; n (vD): amount of sites used for calculation of poles and directions after the Vandamme (1994) cut-off.

Aegean region since approximately 25 Ma was calculated at approximately 300–400 km (Gautier et al., 1999; Jolivet, 2001). Much of this extension, however, was accommodated by formation of metamorphic core complexes, which occurred for a large part prior to 15 Ma (Jolivet, 2001). Moreover, this amount includes up to 100 km of extension accommodated on Crete and in the Sea of Crete (Ring et al., 2001), which lie south of the rotating domain. Thus, the amount of extension required by a rotation with a pivot point in northern Albania is by no means in line with the Aegean reconstructions. This is not to say that there was no extension at all. The Strimon detachment in the Greek Rhodope was active after 15 Ma (Dinter and Royden, 1993; Sokoutis et al., 1993; Brun and Sokoutis, 2007; see discussion below) and additional exhumation and extension occurred in that time in eastern central Greece (Lacassin et al., 2007). Although precise extension estimates on these structures are still scarce, the total amount of post-15 Ma extension in this region is unlikely to be much more than 150 km. We thus conclude that the pivot point of the western Aegean rotations cannot have been positioned in northern Albania, but was more likely placed within the rotating domain itself. Given the uncertainties in estimating the amount and distribution of post-15 Ma extension in Greece it is not possible to determine the location of the pivot point with certainty, but several lines of argumentation lead us to postulate its location on a meridian line located on the western margin of the Moesian Platform, around the longitude of Mt Olympos, i.e. ~22.50°E (Figs. 1 and 6). Firstly, no important middle to late Miocene NE–SW extension occurred west of this line, even though there is a rotation difference to be accommodated between Albania and the eastern Balkanides/Moesian Platform. Secondly, the geodynamics of the Aegean region make this location likely: east of this line, the African slab has been (and likely still is) rolling back south(west)ward, whereas west of this line, the motion of Apulia was continuously northward together with Africa (Besse and Courtillot, 2002; Govers and Wortel, 2005). Within this geodynamic framework, rotation of the west-Aegean domain about a pivot point located on this line would be a likely consequence.

This is in line with a pivot point position proposed by Brun and Sokoutis (2007) based on the independent study of the Eocene to Miocene Southern Rhodope Core Complex (Fig. 1). This triangular core complex developed in two stages, in a dome-in-a-dome sense, comparable to the Menderes core complex of western Turkey (Gessner et al., 2001). The first dome evolved since ~40 Ma, the second one cooled below ~300 °C since 15 Ma. Lineations within this core complex display a gradual ~20° trend change. A paleomagnetically determined ~20° clockwise rotation difference between the Chalkidiki peninsula and the Bulgarian Rhodope (Kondopoulou and Westphal, 1986) suggests the core complex formed during this rotation (with the new Bulgarian Rhodope reference direction obtained here, the data of for the Chalkidiki Peninsula since late Eocene–early Oligocene times ($D \pm \Delta D$ / $I \pm \Delta I = 31.1 \pm 11.6/37.1 \pm 15.3$; Appendix A) yields a similar clockwise rotation difference across the Southern Rhodope Core Complex of 17.9±12.6° (Table 2)). However, Brun and Sokoutis (2007) suggest that the rotation occurred continuously since ~40 Ma which is at odds with the interpretation of van Hinsbergen et al. (2005b) who suggested that the rotation of the Chalkidiki peninsula occurred as part of the rotation of the west

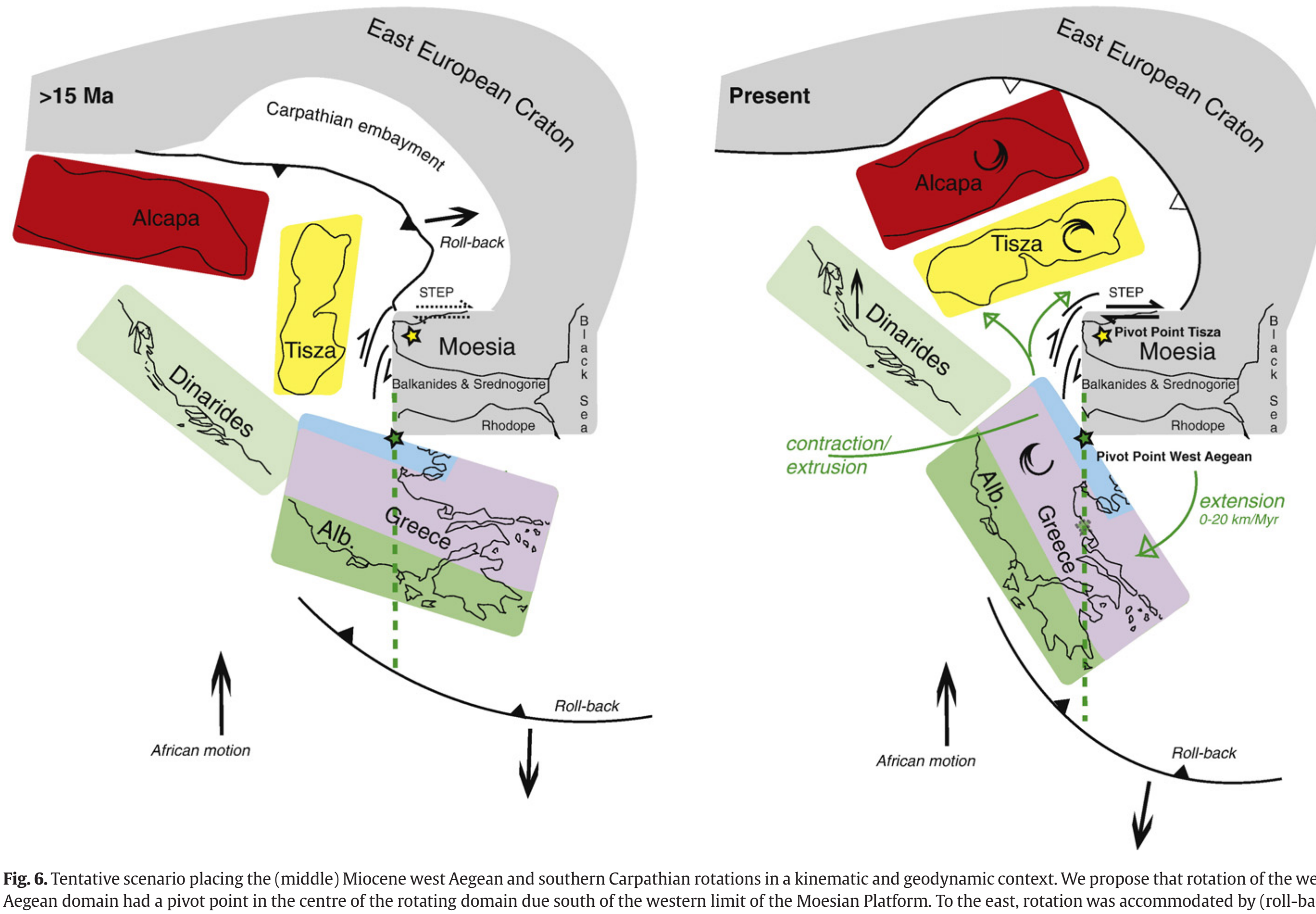


Fig. 6. Tentative scenario placing the (middle) Miocene west Aegean and southern Carpathian rotations in a kinematic and geodynamic context. We propose that rotation of the west-Aegean domain had a pivot point in the centre of the rotating domain due south of the western limit of the Moesian Platform. To the east, rotation was accommodated by (roll-back-related) extension in the Aegean back-arc. To the west, rotation was accommodated by escape of the Tisza block into the Carpathian back-arc, triggered by eastward roll-back of the Carpathian subduction zone, and coeval rotations along the northern edge of the Moesian Platform related to activity of the south Carpathian STEP fault (Dupont-Nivet et al., 2005; Govers and Wortel, 2005). Rotation of the Alcapa Block and the reconstruction of the Carpathian Embayment are taken from Ustaszewski et al. (in press).

Aegean region after 15 Ma. The (sub-)Pelagonian nappe of eastern mainlaind Greece revealed no rotation between Eocene and early Miocene times in the Mesohellenic basin (van Hinsbergen et al., 2005b), whereas this nappe already accreted to the overriding plate—including the Rhodope—before the Eocene (Ricou et al., 1998; van Hinsbergen et al., 2005a). In addition, there is no structural evidence for deformation that may have accommodated a rotation difference between the Pelagonian zone of mainland Greece and the Chalkidiki peninsula before the post-15 Ma rotation of western Greece. A possible scenario satisfying both the timing of rotation and the 20° change in lineation trends is that the first dome of the Southern Rhodope Core Complex formed without rotation of the Chalikidiki peninsula with respect to the northern Rhodope and the second dome formed in response to the ~20° rotation since ~15 Ma.

The finite rotation of the west-Aegean domain is however ~20° more than Chalkidiki suggesting the west Aegean pivot point may have started at the northwestern tip of the southern Rhodope core complex, as proposed by Brun and Sokoutis (2007), but then moved southward through time, either gradually or suddenly.

Extension rates implied by a pivot point halfway the rotating domain as indicated in Fig. 6 would increase from the pivot point eastward from 0 to a maximum of ~20 km/Myr (assuming the Aegean block east of the pivot point is ~250 km long and rotated ~30° clockwise with respect to the Bulgarian Rhodope between 15 and 8 Ma; the final 10° occurring after 8 Ma (van Hinsbergen et al., 2005b)). This is in reasonable agreement with average extension rates estimated for the last ~30 Ma in the Aegean system of Gautier et al. (1999) and Jolivet (2001) of ~10–15 km/Myr.

5.4. Kinematic link between Aegean and Carpathians?

Using this new pivot point has interesting tectonic and kinematic implications. The rotation of the Aegean domain northwest of the pivot point in central Greece postulated above requires well over 100 km NE–SW convergence between Albania and the western Moesian Platform. This convergence may have been accommodated by compressional deformation and/or northward extrusion of a wedge of material in between these regions (Fig. 6). On the one hand, compressional deformation is supported by regional geological features of this region. The Albanian equivalents of the Tripolitza and especially Pindos nappes have been largely overthrusted by the Vardar ophiolites, thus suggesting a larger amount of convergence than in Greece. On the other hand, extrusion may have been associated to the northward extrusion and clockwise rotation of the Tisza block around the northwestern edge of the Moesian Platform (Dupont-Nivet et al., 2005). We favor this scenario supported by the coincidence of the Tisza block emplacement with the Aegean rotations (van Hinsbergen et al., 2005a). However, extrusion requires coeval strike-slip motions on either side of the wedge that remain to be identified in the field, and the larger Albanian shortening remains to be quantified and dated. Clearly, further work is required to test these hypotheses, but the tentative reconstruction we propose provides the best fit for the kinematic and rotation information available for the Aegean and Carpathian regions.

6. Conclusion

Kinematic reconstructions for the Carpathian and Aegean arcs, loci of development of fundamental models for mountain building, orogenic collapse and geodynamics, generally assume the Moesian Platform, located in the core of the Carpathian–Balkan arc and the southern Rhodope Mountains, to be stably attached to the East European craton during the Cenozoic. In this paper we provide a new paleomagnetic reference direction ($D\pm\Delta D/I\pm\Delta I=13.2\pm5.6/60.4\pm3.6$), which confirms this assumption: there is no significant deviation of the Eocene–Oligocene declination of the Moesian platform and Rhodope with respect to published contemporaneous European apparent polar wander paths.

This new reference point has important implications for the regional kinematic evolution of the south-eastern European region. We reviewed paleomagnetic results obtained from the Tisza Block which rotated around the northwestern margin of the Moesian platform during its post-Eocene emplacement into the Carpathian back-arc. The Tisza block underwent 68.4±16.7° of clockwise rotation with respect to the Moesian platform during the middle Miocene (~15–10 Ma). This result is statistically indistinguishable from estimates derived from structural geology (Schmid et al., 1998). In addition, our finding confirms that the eastward emplacement of the Tisza block into the Carpathian back-arc was associated with dextral shear along the southern margin expressed by systematic clockwise rotations between 13 and 9 Ma in the Southern Carpathians foreland (Dupont-Nivet et al., 2005).

The new Eocene–Oligocene reference direction for the Moesian platform and Rhodope allows to accurately quantify the rotation difference with the west Aegean domain. We reviewed and recalculated the post-Oligocene rotation difference (which was largely accommodated between 15 and 8 Ma) at 38.0±7.2°. Classically, the west-Aegean rotation is considered to occur around a pivot point located on the northwestern tip of the rotating domain, in northern Albania (Kissel and Laj, 1988). This would, however, imply that the west-Aegean rotation is entirely accommodated in the Aegean back-arc, requiring 265±50 km of extension between the Moesian platform and the Gulf of Corinth. Aegean extension largely occurred prior to 15 Ma, and it is unlikely that more than 150 km of extension occurred in the Aegean back-arc during the west-Aegean rotation. We therefore propose that the pivot point of the west-Aegean rotation was located approximately in the middle of the rotating domain on the meridian line of the western margin of the Moesian platform, with extension to its east.

This new scenario predicts that material should be contracted or extruded in between northern Albania and the Moesian platform. This prediction is in line with the contemporaneous evolution of the Tisza block, rotating around the Moesian platform and into the Carpathian back-arc. We therefore postulate that the west-Aegean rotation is kinematically balanced by a combination of Aegean back-arc extension and the motion of the Tisza Block around the northwestern edge of the Moesian platform.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.epsl.2008.06.051.

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