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Paleomagnetic analysis of neotectonic deformation in the Anatolian accretionary collage, Turkey

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ABSTRACT

Closure of the Neo-Tethyan Ocean in the Turkish sector of the Alpine-Himalayan orogen by ca. 12 Ma was succeeded by deformation of a domain between the Eurasia plate, presently bounded by the North Anatolian fault, and the Arabian indenter. Facets of this deformation comprise the crustal thickening and uplift that produced the Anatolian plateau, the establishment of transform faults, and tectonic escape as Arabia has continued to impinge into the collage of Anatolian terranes accreted by closure of the Neo-Tethys. We have compiled a database of neotectonic paleomagnetic results from Anatolia to analyze this deformation. Large rotations (up to 5°/10,000 yr) of small fault blocks along the intracontinental transform faults but do not extend away from these zones and show that seismogenic upper crust is decoupled from lower continental lithosphere undergoing continuum deformation. Between the transforms, large fault blocks exhibit slower rotation rates (mostly <1°/100,000 yr), varying systematically across Anatolia. Large counterclockwise rotations near the Arabian indenter diminish westward, becoming zero and then move clockwise near the limit of tectonic escape. The view that the collage has rotated counterclockwise as a single plate, either uniformly or episodically, during the Neotectonic era is refuted. Instead, deformation has been distributed and differential as the collage adapted to changing tectonic regimes. Crustal extrusion to the west and south has expanded the curvature of the Tauride arc and combined with back-roll on the Hellenic arc to produce the extensional horst and graben province in western Turkey. The latitudinal motions are close to confidence limits but consistent with ~800 km of northward motion of Anatolian terranes over 40 m.y., a figure including up to a few hundred kilometers of closure linked to crustal thickening since the demise of the Neo-Tethys.

Keywords: neotectonics, Turkey, Anatolia, Aegean, paleomagnetism, tectonic rotation, tectonic escape, inclination anomaly

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INTRODUCTION

Present-day Anatolia is part of the Alpine-Himalayan orogenic belt and has formed by the collision of terranes with the Eurasian margin accompanying closure of the Tethys Ocean and by ongoing convergence between the Eurasia and Afro-Arabia plates. The accretionary zone comprises a collage of continental fragments that moved across the ocean from former sites at the northern margin of Gondwana (Robertson et al., 1996). Consumption of the Tethys Ocean occurred in two phases collectively referred to as the Paleotectonic era. Closure of the Paleo-Tethys Ocean was complete by Middle Jurassic times and accompanied deformation of the Pontide orogen now sited in northern Turkey. This region, assigned to the "Cimmerian" continent (Şengör and Yılmaz, 1981), was already located close to Eurasia by Jurassic times (Channell et al., 1996). New ocean basins, collectively comprising the Neo-Tethys, opened in the wake of the accretion of crust to the Pontide orogenic margin (Fig. 1, inset) during Late Permian, Triassic, and Early Cretaceous times (Şengör and Yılmaz, 1981; Robertson et al., 1996). Later in the Mesozoic, two microcontinents, the Sakarya and Kırşehir, were separated by a branch of the Neo-Tethys and drifted northward from equatorial latitudes to amalgamate along the İzmir-Ankara-Erzincan suture zone (IAESZ, Fig. 1) in Late Cretaceous and Early Tertiary times. Ocean closure commenced during the Late Cretaceous, with subduction-related volcanism particularly important in Eocene and Oligocene times as the Kırşehir block impinged into the Sakarya microcontinent (Yılmaz, 1990a). The Cappadoccian ignimbrite succession records the last pulses of subduction-related magmatism, and by middle Miocene times the major tectonic elements of present-day Anatolia were in place following the collision of Arabia with this terrane collage along the Bitlis suture zone.

The succeeding postcollisional regime of crustal thickening and tectonic escape is referred to as the Neotectonic era and has been driven by continued northward motion of Afro-Arabia and differential slippage between the African and Arabian sectors along the Dead Sea fault zone (DSFZ, Fig. 1). A feature of this regime has been the concentration of strike-slip motions along intracontinental transforms including the right lateral North Anatolian fault zone (NAF, Fig. 1) and the left lateral East Anatolian fault zone (EAF, Fig. 1). These fault systems now comprise the main tectonic boundaries of a broad zone of distributed deformation within Anatolia.

Methods for quantifying neotectonic crustal deformation apply to three contrasting timescales: for the short term (10–10^2 yr), kinematics may be resolved from GPS and ground surveying (Barka and Reilinger, 1997). On Holocene timescales ($\sim10^3$–10^6 yr), deformation can be evaluated by geomorphic study of features such as stream offsets and changing elevations of erosion and depositional surfaces. For the longer term ($>10^5$–10^6 yr), recovery of paleomagnetic directions is usually required to quantify the deformation. This latter application has already been effectively used over much of Anatolia due to the widespread preservation of young volcanic rocks, which tend to be good recorders of the paleofield direction (Piper et al., 2002). The paleomagnetic method aims to resolve cumulative motions by comparing ancient directions of magnetization with reference directions of the same age predicted from the paleomagnetic records from the adjoining major continental plates acting as orogenic forelands, in this case Eurasia and Africa. In this article, we compile and update the database of neotectonic paleomagnetic results from the Anatolian region and evaluate the implications of these data for the neotectonic deformation in this sector of the orogen over approximately the last 12 m.y. of its history.

GEOTECTONIC AND GEOLOGICAL FRAMEWORK

Consumption of the Neo-Tethys Ocean along the İzmir-Ankara-Erzincan suture zone and collision along the Bitlis suture in southeastern Anatolia (Fig. 1) produced the major Cenozoic environmental change recorded throughout Anatolia at ca. 11.8 Ma during late Serravallian–earliest Tortonian times. Shallow-water Seravallian sediments are the last marine record and were succeeded by terrestrial sediments as central Anatolia became an upland region characterized by horsts and fault-bounded alluvial depressions ("ovas"). Quasi-continuous volcanism has recorded the waning stages of subduction and subsequent neotectonic fault movements permitting access to asthenosphere melts. It falls into three (western, central, and eastern) provinces, each exhibiting a changing balance of calc-alkaline to alkaline lava compositions that are presumably related to the transition from collisional to strike-slip tectonic regimes (Yılmaz 1990b).

A facet of neotectonic deformation that is poorly understood in this region is orogenic collapse and associated lateral spreading of the crust. Proponents of the mechanism (Dewey et al., 1986; Seyitoğlu and Scott 1996) have predicted that the onset of horst-graben formation in western Anatolia commenced during the late Oligocene–early Miocene following cessation of Paleocene–Eocene collision along the İzmir-Ankara-Erzincan suture zone and associated shortening and overthickening of Aegean-Anatolian crust. Although a genetic link between orogenic collapse and graben formation would imply that the latter has been continuous since the late Oligocene–early Miocene, ground studies show that graben evolution in Anatolia has had a complex history in which back-arc spreading, orogenic collapse, tectonic escape, and roll-back processes have all operated successively or in tandem to form the horst-graben systems (Koçyiğit et al., 1999, 2000). Rift infills in western Anatolia were deposited in extensional regimes mostly during early–middle Miocene and Pliocene–Recent times and were interrupted by an episode of late Miocene–early Pliocene compression. The earlier phase of graben infill reflects deposition temporally related to orogenic collapse along the İzmir-Ankara-Erzincan suture zone (Dewey et al., 1986; Seyitoğlu and Scott, 1996). The earlier extensional phase is separated from

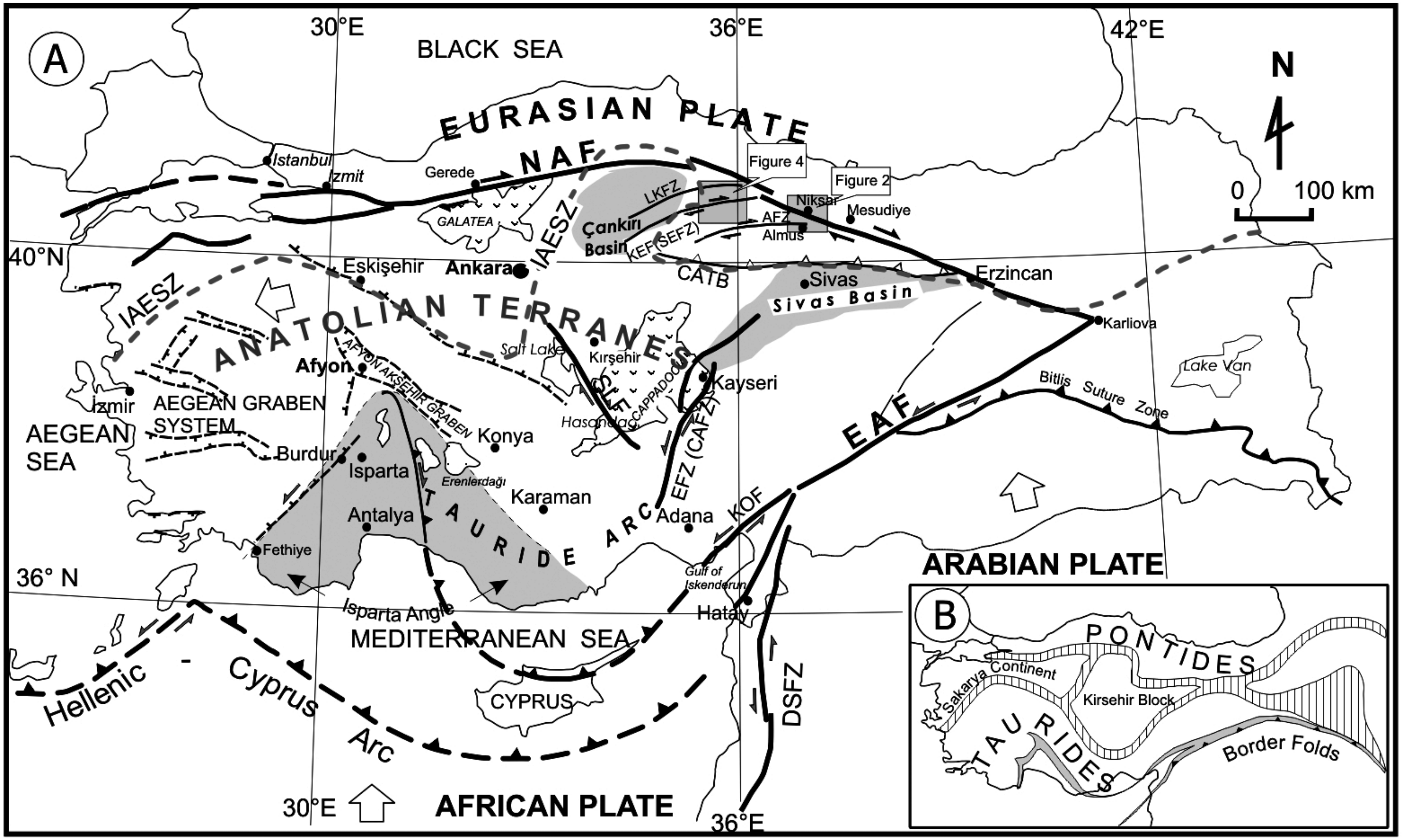


Figure 1. (A) Tectonic framework and location map of the Turkish sector of the Alpine-Himalayan orogenic belt. The large open arrows show the directions of current relative plate motions, and the smaller half arrows show the directions of movement on major strike-slip faults. Tectonic lineaments are abbreviated: AFZ—Almus fault zone; CATB—Central Anatolian thrust belt; DSFZ—Dead Sea fault zone; EAF—East Anatolian fault zone; EFZ (CAFZ)—Ecemiş fault zone (Central Anatolian fault zone); KEF (SEFZ)—Kırıkkale-Erbaa fault zone (becoming the Sungurlu-Ezinepazarı fault zone, or SEFZ, to the we st); KOF—Karatos-Osmaniye fault zone; IAESZ—İzmir-Ankara-Erzincan suture zone; LKFZ—Laçin-Kızılırmak fault zone; NAF—North Anatolian fault zone; SLF—Salt Lake fault zone. The shaded area comprises deformed terranes within the Isparta angle and was formed mainly during the Paleotectonic era by the interference of verging allochthonous units during the final stages of Tethyan convergence. (B) The orogenic framework of Turkey and adjoining regions.

later extension by an ENE-WSW-directed compression that prevailed during late Miocene times (Koçyiğit et al., 1999) contemporaneous with collision along the Bitlis suture zone between the Arabia plate and the Anatolian collage (Fig. 1). This collision initiated the crustal thickening and uplift responsible for producing the Anatolian plateau. By early Pliocene times, seafloor spreading had commenced along the axis of the Red Sea and the differential northward movement of Arabia had initiated the westward extrusion of the terrane collage (Hempton 1987); these events were contemporaneous with the last phase of extension in western Anatolia.

In addition to motions on the transforms delineating the present plate boundaries, deformation within the terrane collage is taken up along second-order faults. These cut the terrane collage and have unclear lateral extents due to Pleistocene cover and limited seismic evidence, although they are probably linked together as an anastomosing network within the Anatolian upper crust. They comprise two sets. The first includes northeast-southwest fractures that splay off from the North Anatolian fault zone into the Anatolian collage; the largest examples are the Kırıkkale-Erbaa and Almus fault zones shown in Figure 1 (KEF and AFZ, respectively). Members of this system can display both right and left lateral strike-slip movements plus normal or reverse-slip components, where there are variations from this general trend (Koçyiğit and Beyhan, 1998) but left lateral strike-slip motions are dominant in eastern Anatolia. A second set of faults has northwest-southeast trends and oblique slip motions, usually with a dominant right lateral motion. Members of this latter set divide the western half of central Anatolia into horsts and grabens bounded by active faults. They are concentrated within a transitional tectonic zone between the strike-slip regime of eastern Anatolia and the extensional regime in the Aegean and western Anatolia (Yılmaz et al., 1987; Fig. 1).

Paleomagnetic studies in the eastern part of Anatolia have demonstrated that block rotations are common throughout Anatolia (Gürsoy et al., 1997, 1998, 1999, 2003b; Kissel et al., 2003; Piper et al., 1996, 1997, 2002; Platzman et al. 1994, 1998; Tatar et al. 1996, 2000, 2002). Two contrasting interpretations have been given to these data. Platzman et al. (1998) and Kissel et al. (2003) interpret them in terms of uniform rotation of a coherent Anatolia plate delimited by the intracontinental transforms in northern, eastern, and southern Anatolia and by the extensional province in western Turkey. We refer to this as the "Anatolia plate model," and it is supported for the short term by GPS evidence (McClusky et al., 2000; Reilinger and McClusky, 2001). In contrast, Gürsoy et al. (1999), Piper et al. (2002), and Tatar et al. (2002) identify a differential character of paleomagnetic rotations and conclude that an Anatolia plate, sensu stricto, does not exist. We refer to this as the "distributed deformation model." Specifically, we note a regional variation of declinations illustrated by a general westward decrease in the amount of counterclockwise rotation across Anatolia toward the horst and graben province of western Turkey. West of the longitude of Ankara, the rotations are variable, but the declinations recovered to date tend to be clockwise from the Galatean region north of Ankara to the Isparta angle close to Afyon in the south (Gürsoy et al., 1999, 2003b; Tatar et al., 2002). The Isparta angle (Fig. 1) is a tectonic bend formed by the interference of verging allochthonous units emplaced during culminating stages of Tethyan convergence (Kissel et al., 1993; Robertson et al., 1996).

A resolution between the Anatolia plate model and the differential deformation model depends in part on the time frame of the rotations. If rotations of similar age are divergent from block to block, the latter model is supported. If older rotations are systematically larger than the younger ones, there is support for rotation of a coherent plate. Gürsoy et al. (1997), Piper et al. (2002), and Tatar et al. (2002) compare rotations in rocks of different age to argue that most, if not all, of this deformation is a late facet of the neotectonic history imparted after crustal thickening. Platzman et al. (1994, 1998) have dated numerous individual lava flows in order to help evaluate this question, but interpret their data in the context of episodic rotation of an integral Anatolia plate.

Results from the Global Positioning System (GPS) show that surface motions in Anatolia on a decade timescale can be approximated by uniform counterclockwise rotation about a Eulerian pole in northern Egypt (Barka and Reilinger, 1997; McClusky et al., 2000). A long-term extension of this solution would support the Anatolia plate model of Platzman et al. (1998) and Kissel et al. (2003). It is compatible with the westward extrusion of the crust by tectonic escape and is expressed by dominant strike-slip motions within a broad sector of central Anatolia between the North Anatolian fault zone and the East Anatolian fault zone, extending westward from the triple junction at Karlıova where these transforms meet (Fig. 1). Currently central Anatolia is moving westward relative to Eurasia at a rate of ~15 mm/yr, while Anatolia west of the Isparta angle is moving southwest at ~30 mm/yr. The nearly static behavior of the region close to the Isparta angle (Barka et al., 1999) separates an eastern sector dominated by a regime of Pliocene and later compression and strike-slip from a dominantly extensional regime within the eastern part of the Aegean province (Fig. 1). The extensional regime in western Turkey is probably driven predominantly by roll-back on the Hellenic arc (Le Pichon et al., 1995) that has progressively expanded the curvature of this arc during the last 5 m.y. (Kissel et al., 1989). Neotectonic deformation in the Isparta angle is controlled by the interaction of back-arc processes associated with outward motion of the Hellenic arc in the west and extrusion of Anatolian crust by tectonic escape from the east.

PALEOMAGNETIC ANALYSIS OF NEOTECTONIC DEFORMATION

The bulk of the paleomagnetic database from Anatolia summarized in Table 1 is based on studies of volcanic rocks that are typically high-quality recorders of the magnetic field direction at the time of eruption. Rapid cooling and weak fabrics will usually mean that a geomagnetic field direction is accurately recorded, although this is instantaneous in geological terms and therefore a component of the secular variation; sampling of a number of flows is always required to yield a mean direction of paleomagnetic rather than virtual geomagnetic significance. While the orientation of the flows may usually be inferred from pseudo-bedding in lavas, layering in pyroclastic units, or bedding within intralava sediments, the application of tilt adjustment to these data is often unclear because the volcanic units have flowed down primary slopes. When this is the case, most or all of the measured tilt is primary and tilt adjustment is not relevant. Fortunately, this has not proved to be an important problem with volcanic rocks emplaced during the neotectonic era because the topography is young and little affected by later deformation. It is a more serious difficulty in the paleotectonic domains where outcrops are poorer, more degraded, and highly faulted; these factors no doubt contribute to the dispersion of paleomagnetic directions within these older units.

Treatment of all rock collections considered here has followed standard procedures of alternating field or thermal demagnetization. This has normally been accompanied by projection of the demagnetized vector onto orthogonal projections, resolution of component structures, and calculation of equivalent component directions by principal component analysis. Thermal demagnetization is generally considered preferable for resolving component structures in old rocks because this procedure is able to completely unravel the magnetization and the secondary components are often acquired by a thermal mechanism. However, in young volcanic rocks emplaced in the Neotectonic era, secondary components are usually small, and no significant differences have typically been noted between components resolved from application of the two methods. The Paleomagnetic results summarized in Table 1 are derived from Late Cretaceous and younger rocks, and the older units embrace the latter part of the paleotectonic as well as the neotectonic history. The paleotectonic units may have a multiple deformational history incorporating both pre- and postcollisional events.

Some of the sampling is distributed over a large area. The study of Sarıbudak (1989, result 7), for example, covers sampling sites between 31.5°E–37°E and 41°N–42°N. In these cases, the mean site location is listed in Table 1 and the mean direction plotted at this point in Figure 5.

THE PALEOMAGNETIC RECORD OF CRUSTAL DEFORMATION IN ANATOLIA

There are two scales of block rotation resolved from paleomagnetic studies in Turkey. In the vicinity of the intracontinental transform faults, large rotations, sometimes >180° over a time interval of 0.5–2.0 m.y., are observed. This has been recognized with the most important ramifications in the vicinity of the North Anatolian fault zone by the studies of Sarıbudak et al. (1990), Michel et al. (1995), Tatar et al. (1995), and Piper et al. (1997). In the western sector, Sarıbudak et al. (1990) considered that a ~15 km–wide block (Almacık) had undergone a large clockwise rotation as a consequence of ongoing dextral slip between master faults. Michel et al. (1995) were unable to replicate this result and concluded that no such large coherent fault block is present here, although they did observe some large rotations recorded in Palaeogene rocks close to major faults and assigned these rotations to much smaller fault blocks. These large local rotations about both horizontal and vertical axes are superimposed on a general counterclockwise rotation of 10°–20° across the fault zone. Michel et al. (1995) were unable to relate the observed rotations to changes in the stretching direction in these relatively old rocks and concluded that they were the cumulative result of a complex deformational history. Farther east in the North Anatolian fault zone, large clockwise rotations between dextral master faults, locally up to 250°, are recorded by lavas emplaced in the Niksar pull-apart and are only ca. 500 Ka in age (Piper et al., 1997; Fig. 2). These results show that the rotations are more recent than the initiation of the contemporary strike-slip regime along the North Anatolian fault zone and imply that small blocks are rotating in ball bearing fashion between the dextral master faults (Beck, 1976).

Recent road construction south of the town of Niksar has exposed new sections through these lavas, where they have been sampled at three additional locations (in Fig. 2) for which we note data here (Table 2). Demagnetization defines single-component behaviors (Fig. 3) and magnetization directions that are either unrotated or rotated by small amounts clockwise from the present field direction. These data contrast with much larger rotations recorded immediately to the south and east in the same volcanic episode (Fig. 2) and imply that the rotating blocks are a few kilometers or less in size. Three high-precision K-Ar age determinations of 461 ± 7, 523 ± 4, and 568 ± 3 Ka have been obtained from these lavas (Tatar et al., 2004b), and highly variable rotations at rates of up to 5°/10,000 yr during the Brunhes chron are indicated. Comparable rotations recorded in both young (Piper et al., 1997) and much older (Michel et al., 1995) Paleogene volcanic rocks appear to reflect the short life span (~late Pliocene–recent) of the North Anatolian fault zone (e.g., Barka and Hancock,1984; Barka and Gülen, 1989) as the southern boundary of the Eurasia plate in the Anatolian sector.

Fault block deformation across the North Anatolian fault zone has been studied in smaller numbers of lava units elsewhere. Variable and sometime large clockwise rotations are recorded by sporadic Pliocene–Quaternary volcanic units in the Ilgaz and Gerede regions within the northern convex extension of the North Anatolian fault zone between the longitudes 32°E and 34°E (Fig. 1; Piper et al., 1997). In each case, the magnitude of rotation diminishes rapidly away from the fault system and may be entirely confined to small fault blocks between the intracontinental master faults. Thus, the most important geotectonic conclusion to emerge from these results is that intense deformation within a narrow zone of the seismogenic upper crust is decoupled from viscous sheet continuum deformation in

TABLE 1. SUMMARY OF PALEOTECTONIC AND NEOTECTONIC (LATE CRETACEOUS– PRESENT) PALEOMAGNETIC RESULTS FROM CENTRAL ANATOLIA

Location/unit	Age		Location					
Block/region	Stratigraphic	Ma	°E	°N	*N*	D/I(°)	α_{95}(°)*	References
Results from rock units emplaced during the Paleotectonic regime North of the NAF								
1 Gümüşhane[1]	E (M)		39.5	40.4	9	165.3/–34.6	6.3	Kissel et al. (2003)
2 Mesudiye (Group 1)	E (N)		37.7	40.6	7	165.9/–50.4	6.7	Baydemir (1990)
3 Mesudiye (Group 2)	E (R)		37.7	40.6	8	185.6/–45.1	15.1	Baydemir (1990)
4 North of Niksar[2]	E (R)	41.8–43.5	37.0	40.7	17	150.4/–44.9	9.4	Tatar et al. (1995); Platzman et al. (1994)[2]
5 Erbaa	E (R)		36.3	40.8	7	194.6/–48.8	15.3	Piper et al. (1996)
6 Mesudiye	E (N)		37.7	40.5	7	166.8/–49.9	9.6	Orbay and Bayburdi (1979)
7 Central Pontides[3]	Cu (M)		34.0	41.8	8	172.1/–38.6	10.8	Sarıbudak (1989)
8 Kastamonu	E? (R)		33.7	41.2	3	164.2/–31.7	10.6	Piper et al. (1996)
South of the North Anatolian fault zone								
9 Imranlı	E (R)		38.4	39.8	10	146.0/–34.2	6.3	Baydemir (1990)
10 Almus[2]	E (R)	45.3	36.9	40.4	11	146.0/–46.8	6.9	Tatar et al. (1995); Platzman et al. (1994)[2]
11 SW Amasya[4]	E (M)		35.2	40.6	12	168.8/–45.4	12.4	This article[4]
12 Çankırı basin[5]	O (M)		34.1	40.2	6	168.6/–30.7	17.4	Kaymakçı et al. (2003)
13 Akdagmadeni	E (N)		35.8	39.7	3	113.9/–43.5	5.2	Tatar et al. (1996)
14 Yozgat[6]	E (R)		34.7	39.8	5	159.0/–43.3	14.2	Tatar et al. (1996); Kissel et al (2003)[6]
15 Kaman (i)	Pa (R)		33.9	39.2	10	148.0/–61.0	8.7	Sanver and Ponat (1980)
16 Kaman (ii)	Pa (N)		33.9	39.2	3	132.9/–48.4	16.0	Kissel et al. (2003)
17 Bayat	E (R)		34.5	40.8	7	172.1/–47.1	7.6	Piper et al. (1997)
18 Akseki thrust belt	Pa-E (M)		31.8	37.3	9	219.9/–34.7	11.6	Kissel et al. (1993)
19 Sarız	E (R)		36.5	38.5	3	140.3/–38.0	10.0	Kissel et al. (2003)
20 Central Anatolian thrust belt[7]	C–E (M)		36.5	39.9	7	166.9/–42.6	13.5	Gürsoy et al. (1997)
21 Sivas basin[8]	C–E (M)		36.9	39.8	8	155.3/–56.0	14.1	Gürsoy et al. (1997)
22 Tunceli	E(R)		39.6	39.4	4	152.5/–48.0	16.0	Van der Voo (1968)
Results from rock units emplaced during the Neotectonic regime								
23 Galatean province	M (M)		31.8	40.6	15	196.8/–56.5	6.9	Gürsoy et al. (1999)
24 Çankırı basin[9]	$M_{M–U}$ (R)		34.3	40.3	3	185.8/–42.3	33.0	Kaymakçı et al. (2003)
25 Sivas basin[10]	M–Pl (R)	13.8–13.1	36.2	39.4	5	148.3/–41.8	18.8	Gürsoy et al. (1997)[10]
26 Şarkışla block	M–Pl (M)		36.8	39.4	16	150.2/–56.6	12.9	Gürsoy et al. (1997)[11]
27 Kormac Dag block[11]	M–Pl (M)	5.8–5.2	36.9	39.2	10	156.0/–47.6	19.0	Gürsoy et al. (1997)[12]
28 Gürün block[12]	M–Pl (M)	24.0–4.0	37.2	38.7	14	163.8/–51.8	12.2	Gürsoy et al. (1997)
29 Gemerek, R	M (R)	15.2–14.8	36.0	39.2	35	111.9/–60.8	9.6	Krijgsman et al. (1996)
30 Gemerek, N	M (N)	15.2–14.8	36.0	39.2	41	127.9/–32.2	6.7	Krijgsman et al. (1996)
31 Inkonak, R	M (R)	25.8–25.1	37.0	39.4	30	129.8/–37.3	7.7	Krijgsman et al. (1996)
32 Inkonak, N	M (N)	25.8–25.1	37.0	39.4	30	127.1/–49.6	10.0	Krijgsman et al. (1996)
33 Yeniköy, R	M (R)	30.4–28.4	36.4	39.1	49	126.3/–33.3	8.8	Krijgsman et al. (1996)
34 Yeniköy, N	M (N)	30.4–28.4	36.4	39.1	32	149.6/–47.7	7.2	Krijgsman et al. (1996)
35 Haramiköy, R	M (R)	23.0–21.8	31.8	38.5	37	190.1/–40.9	5.3	Krijgsman et al. (1996)
36 Haramiköy, N	M (N)	23.0–21.8	31.8	38.5	45	182.4/–37.2	4.3	Krijgsman et al. (1996)
37 Keseköy	M (N)	21.3–18.3	32.9	40.7	20	189.9/–47.4	5.8	Krijgsman et al. (1996)
38 Afyon (north)[13]	M (M)	21.0–9.3	30.7	39.2	42	201.1/–46.9	5.7	Gürsoy et al. (2003b)
39 Afyon (south)[13]	M (M)	14.8–8.0	29.9	38.7	20	206.3/–49.7	9.7	Gürsoy et al. (2003b)
40 Sille volcanics	M (M)	11.6	32.4	37.8	4	149.5/–51.8	11.3	Tatar et al. (2002)
41 Erenlerdağ (i)[14]	M (M)	9.9	32.1	37.6	12	181.3/–46.4	8.6	Tatar et al. (2002)[14]
42 Kütahya (i)[§]	M (R)		29.9	39.6	5	205.4/–54.6	13.0	Authors in preparation
43 Kütahya (ii)[§]	M (R)		29.9	39.6	6	205.3/–39.8	11.8	Authors in preparation
44 Erenlerdağ (ii)[§]	M–P (M)	6.9–3.5	32.0	37.5	5	179.0/–50.8	13.4	Tatar et al. (2002)
45 Hasandağ	M (M)	13.1	34.2	38.0	7	170.1/–57.6	10.9	Gürsoy et al. (1998)

TABLE 1. *Continued*

Location/unit	Age		Location					
Block/region	Stratigraphic	Ma	°E	°N	*N*	D/I(°)	α_{95}(°)*	References
46 Cappadocian ignimbrites	M–P (M)	11.0–1.1	34.8	38.6	10	174.0/–51.1	6.6	Piper et al. (2002)
47 Ermenek basin	M_L (N)		33.0	36.7	3	179.0/53.0	8.4	Kissel et al. (1993)
48 Antakya region marls	M (M)		36.2	36.1	13	168.0/–43.9	6.9	Kissel et al. (2003)
49 Karacadağ[15]	M_U	6.9–4.7	33.8	37.7	18	175.5/–57.0	5.7	Gürsoy et al. (1998)[15]
50 Karaman (Karadag)[16]	M–P	1.8–2.1	33.2	37.4	15	177.3/–51.3	7.9	Gürsoy et al. (1998)[16]
51 Isparta	P (M)	4.7–4.0	30.8	37.8	15	185.8/–52.6	6.1	Tatar et al. (2002)
52 Kütahya (iii)	P (N)		29.9	39.6	9	205.4/–53.1	6.1	Authors in preparation
53 Kütahya (iv)	P (N)		29.9	39.6	6	161.6/–41.5	9.1	Authors in preparation
54 Karapınar volcanic field	Pl (N)		33.6	37.6	5	156.9/–42.2	14.9	Gürsoy et al. (1998)
55 Nevsehir volcanic field[17]	Pl (M)	9.0–0.0	34.7	38.8	11	182.7/–52.3	10.2	Tatar et al. (2000)[17]
56 Erciyes volcano[18]	Pl (M)	<1.1–0.0	35.8	38.9	24	178.4/–53.0	5.6	Tatar et al. (2000)
57 Osmaniye volcanics	Pl (N)	0.0–0.8	36.0	37.0	8	191.3/–44.6	12.6	Gürsoy et al. (2003a)
58 Karasu rift volcanics	Pl (M)	0.6–0.3	36.3	36.8	51	188.8/–54.7	4.0	Tatar et al. (2004a)

Notes: Directions given are reversed paleofield directions, and locations are the center of the study area. *N* is the number of separate units (sites are not flow units or dikes but individual samples in the study of Krijgsman et al. (1996) included in the calculated mean), and α_{95} is the radius of the cone of 95% confidence about the mean direction. Rock ages are C—Cretaceous; Pa—Paleocene; E—Eocene; O—Oligocene; M—Miocene; P—Pliocene and Pl—Pleistocene. CATB—Central Anatolian thrust belt.

*The radius of the cone of confidence about the mean direction.

§The small roman numerals refer to successive volcanic formations.

[1]The mean of results for nine sites between TK148 and TK164 of Kissel et al. (2003).

[2]The mean direction from north of the North Anatolian fault zone at Niksar is calculated from sites TV6, TV8–10, TV13, TV14, and TV16 of Platzman et al. (1994) and sites 33–42 of Tatar et al. (1995). The mean direction south of the North Anatolian fault zone is calculated from sites TV01, 02, and 11 of Platzman et al. (1994) and sites 50–61 of Tatar et al. (1995).

[3]The mean for Pontide Cretaceous and Eocene sites G001, GI001, BA01, ER01, DE01, S101, S102, and BF01.

[4]The mean for Eocene volcanic succession SW of Amasya (the east continuation of the Çankırı basin) calculated from data in Table 2 plus sites 11–14 and 18 of Piper et al. (1996).

[5]The mean of sites KAL, DAW, GUV, HAM3, INC3, and HAM4 of Kaymakçı et al. (2003).

[6]The mean result for the Yozgat volcanic-sedimentary succession calculated from data for sites 156–160 of Tatar et al. (1996) and sites TK109, TK115, and TK117 of Kissel et al. (2003).

[7]The result for rock units above the Central Anatolian thrust belt includes sites C1–3 and C199–201 of Gürsoy et al. (1997) and sites TK100 and TK104 of Kissel et al. (2003).

[8]The result for units between the Central Anatolian thrust belt and the Sivas thrust calculated from sites C240 and E72–75, E127–131, and E241 of Gürsoy et al (1997).

[9]The mean of sites KUC, HAL1, and UR of Kaymakçı et al. (2003).

[10]The neotectonic units between the Central Anatolian thrust belt and Sivas thrust include sites Q14–18 of Gürsoy et al. (1997) and sites S10 and S11 of Platzman et al. (1998).

[11]The units used to calculate a mean for the Kormac Dağ Block are Q6–8, Q21, Q22, Q69, Q70, PL23, and M76–78 of Gürsoy et al. (1997) and sites S2 and S4 of Platzman et al. (1998).

[12]The Gürün block includes sites Q65–68 and PL25 and PL26 of Gürsoy et al. (1997) and sites S3, S5, S7, S8, and S9 of Platzman et al. (1998).

[13]The Afyon South result is calculated from sites 2, 3, 6–11, 13–19, 22, 76, 77 80, and 81, and the Afyon North result is based on sites 25–29, 31, 34–51, 53, 55–64, 67–75, 78, and 79 of Gürsoy et al. (2003a). These groups are separated by the Afyon-Aksehir graben (Fig. 1).

[14]The Erenlerdag group mean includes sites K1 and K2 of Platzman et al. (1998).

[15]The mean for the Karacadağ volcanic complex includes sites K3, K4, K5, K9, and K10 of Platzman et al. (1998).

[16]The mean for the Karadağ volcano includes sites K6 and K7 of Platzman et al. (1998).

[17]The mean calculated for the Nevsehir volcanic field comprises sites 20 and 22–26 of Tatar et al. (2000) and sites N1–3, N5, and N6 of Platzman et al. (1998).

[18] To derive an overall mean for the Erciyes volcano and the surrounding volcanic field sited within the Erciyes pull-apart basin, we have meaned sites 1–4, 7, 9–13, 15–19, 28, 29, and 31–37 from the study of Tatar et al. (2000).

lower crust and continental lithosphere (Piper et al., 1997; c.f. England and McKenzie, 1982). Gürsoy et al. (2003a) also identify large rotations in young lavas close to the Karataş-Osmaniye fault (KOF, Fig. 1), a continuation of the East Anatolian fault zone into the Gulf of İskenderun in southern Turkey; these latter data imply block rotations at rates of ~1°/10,000 yr within this left lateral system.

The second scale of block rotation resolved from paleomagnetism is typically an order of magnitude less than these rotations determined near the transforms. It appears to apply to the weak crust in Anatolia between the North Anatolian fault zone and the East Anatolian fault zone, but prior to the Pliocene establishment of the North Anatolian fault zone it probably also included terranes within the Pontides north of the North Anato-

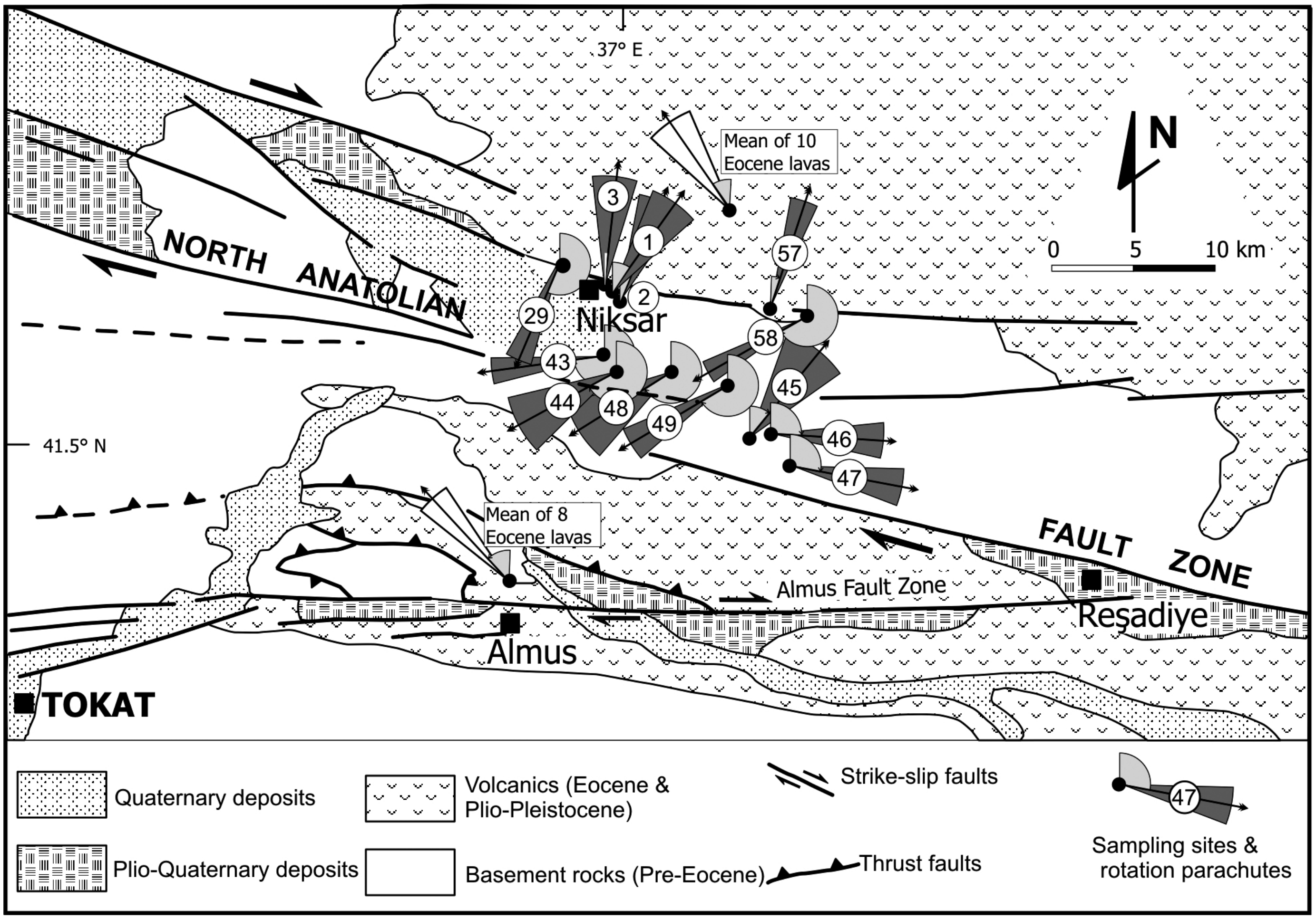


Figure 2. Magnetic declinations (normal polarity) derived from Eocene- and Pleistocene-age lava flows in the Niksar sector of the North Anatolian fault zone after Piper et al. (1997), supplemented by three additional sites (Table 2) in the Brunhes-age lavas emplaced in the Niksar pull-apart basin. Sites 1–3 refer to data included in this article (Table 2), and sites 29, 43–49, 57, and 58 refer to the study of Piper et al. (1997).

lian fault zone (Sarıbudak, 1989; Piper et al., 1996; Kissel et al., 2003). In the example shown in Figure 2, comparable counterclockwise rotations are recorded in the Eocene volcanic successions north and south of the North Anatolian fault zone investigated by Platzman et al. (1994) and Tatar et al. (1995) following an early study by Van der Voo (1968). In Table 1, mean results are calculated for Eocene volcanic sequences north and south of the North Anatolian fault zone by combining data from these two studies. On the north side of the fault zone, mean directions of D/I = 147/–48° (7 sites, α_{95} = 19°) and D/I = 152/–43° (10 sites, α_{95} = 11°) are yielded by the data of Platzman et al. (1994) and Tatar et al. (1995), respectively (D/I—declination [°E] and inclination [positive down, negative up]; α_{95}—95% cone of confidence). There is probably some duplication between these sites, although it is not obvious from the site mean data, probably because of differing field assessments of lava orientation. Data from eight of nine Eocene flows studied by Tatar et al. (1995) supplement data from three sites of Platzman et al. (1994) on the south side of the fault zone to yield a mean of D/I = 146/–47° (11 sites, α_{95} = 7°, Fig. 2). Age determinations from the Eocene volcanic rocks on each side of the North Anatolian fault zone are marginally different (Platzman et al., 1994), although these differences are within error limits and, because polarities are uniformly reversed, it seems likely that they record a volcanic episode confined to a single chron. Polarity chrons ranged from 1.01 to 2.48 million years in duration, with the longest reversed episode at 46.3–43.8 Ma (Cande and Kent, 1995) during the interval embracing age determinations on these volcanics (Platzman et al., 1994)

The revised group means calculated by incorporating the collective data identify 30° of counterclockwise rotation from the present field on the north side of the fault zone and 34° on the south side in the Niksar sector since the Eocene. Farther to the east, Baydemir (1990, results 2 and 9 in Table 1) used vol-

TABLE 2. SITE AND GROUP MEAN DIRECTIONS OF MAGNETIZATION FROM ADDITIONAL PALEOMAGNETIC STUDY OF VOLCANIC UNITS IN CENTRAL–NORTHERN ANATOLIA

Site	D_{is}	I_{is}	D_{tc}	I_{tc}	*N*/*n*	*R*	α_{95}	*k*
Brunhes chron lavas, Niksar pull-apart basin								
1	47.3	60.1	33.5	55.1	3/3	2.96	18.3	46.4
2	22.3	37.2	22.3	37.2	7/7	6.93	6.5	88.2
3	3.5	48.0	3.5	48.0	7/7	6.93	6.7	83.2
Eocene volcanics, SW Asmaya region								
4	196.9	–56.9	137.2	–58.4	7/3	2.96	16.6	15.0
6	168.7	–72.9	187.8	–43.6	5/4	3.98	8.0	138.4
7	177.4	–78.6	184.4	–54.7	7/5	4.54	27.2	8.7
8	205.2	–63.9	220.1	–39.2	8/7	7.73	11.0	26.0
9	172.5	–63.7	184.7	–39.3	6/5	4.97	6.2	154.3
10	3.7	61.1	10.4	35.7	3/3	3.00	5.9	442.3
11*	67.6	6.0	69.3	0.0	8/7	6.89	8.3	54.2
12	153.8	–48.2	151.4	–23.4	7/4	3.92	15.0	38.5
Mean**			168.8	–45.4	12	11.17	12.4	13.2

Notes: D_{is}, I_{is}, $D_{tc,}$ and I_{tc} are the declination and inclination of the mean remanence direction—in situ and tilt-adjusted, respectively—derived from *n* samples out of a site population of *N*. *R* is the resultant vector, α95 the radius of the cone of confidence about the mean direction, and *k* the Fisher precision parameter.

*Site excluded from mean calculation.

**Includes sites 11–14 and 18 of Piper et al. (1996).

canic rocks, also of Eocene age, to demonstrate counterclockwise rotations, in this case 44° on the north side and 53° on the south side of the fault zone. This differential counterclockwise rotation is possibly an expression of rotation on the second-order faults splaying westward from the North Anatolian fault zone into the Anatolian collage in this sector (Fig. 1). Although the differences are too small to establish with confidence by paleomagnetism, they are consistent with the GPS evidence if the North Anatolian fault zone is a major locus of differential rotation of Anatolian terranes in this region: the contemporary rotation rate of 1.3° ± 0.1° per million years deduced from the latter evidence (Reilinger and McClusky, 2001) would be equivalent to ~5° of rotation during the life span of the North Anatolian fault zone and compatible with the differences in magnetic declination.

The southern margin of the North Anatolian fault zone has also been studied farther west in the Gerede region, where the fault zone changes to an ENE-WSW trend (Fig. 1), by Platzman et al. (1994) and Piper et al. (1997), and in the Ilgaz-Bayat region by the latter authors. In the former area, these investigations show declinations with large amounts of both clockwise and counterclockwise rotation, which is probably a result of dissection of this zone into numerous small fault blocks; no representative regional direction can be calculated here. In the latter area (result 17 in Table 1), a small amount of coherent counterclockwise rotation is present farther away from the North Anatolian fault zone.

Kaymakçı et al. (2003) use paleomagnetic data to develop a model for the indentation of the Çankırı basin at the northern perimeter of the Kırşehir block accompanying collisional impingement into the Sakarya microcontinent. Oroclinal bending appears to have been in progress here during the Eocene and probably into the Oligocene (contributing to the distribution of declinations in result 12 in Table 1) but was complete prior to middle Miocene times (result 24).

Second-order strike-slip faults with east-west to ENE-SSW trend splaying westward from the North Anatolian fault zone distribute deformation from right lateral strike slip on the transform fault zone into the Anatolian collage and, as noted in the Niksar-Almus and Mesudiye examples mentioned earlier, are an important facet of the late neotectonic framework (Fig. 1). A member of this system, the Sungurlu-Ezinepazarı fault zone (SEFZ, Figs. 1 and 4), cuts the southern margin of the Çankırı basin (Fig. 4). Piper et al. (1996) report results from five sites in an Eocene volcano-sedimentary succession on the southeast side of this fault, yielding a group mean rotated counterclockwise. To improve the definition of this result, we have sampled a further eight volcanic units in this region; the site locations are shown in Figure 4 together with the sites of the previous study. Following progressive thermal demagnetization, some samples at all sites yielded stable vectors with significant group means of which six are of reversed, one is of normal polarity, and one has an intermediate direction (Table 2); a typical vector plot is illustrated in Figure 3B. When seven additional sites (1N, 6R) are included with the five sites of reversed magnetization of Piper et al. (1996), a mean for the block south of the Sungurlu-Ezinepazarı fault zone of D/I = 169/–45° yields result 11 in Table 1. The Eocene declination here is close to the Oligocene

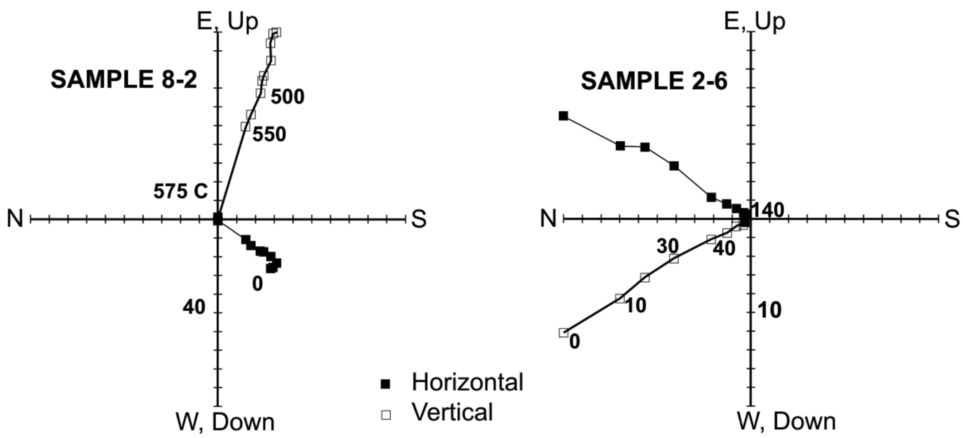


Figure 3. Orthogonal projections of the magnetizations in a lava sample from the Niksar pull-apart basin (2–6) and in Eocene lava (8–2) from the eastern sector of the Çankırı basin (see Figures 1 and 4 for locations). Projections onto the horizontal plane are shown as solid symbols, and those onto the vertical plane are shown as open symbols. The figures against the axes refer to demagnetizing alternating fields in mT or °C. The figures on the axes are intensities in 10^{-5} Am^2/kg.

declination derived from data of Kaymakçı et al. (2003) in the Çankırı basin to the west (result 12 in Table 1).

Both Tatar et al. (1996) and Kissel et al. (2003) have studied Eocene volcanic-sedimentary successions west of the town of Yozgat, where the Almus-Tokat block is overthrust to the north by the Central Anatolian thrust belt (CATB, Fig. 1). Uniform reversed polarity with a S-SSE declination is identified in each case. A mean result for five sites from the former study and three sites from the latter is summarized as result 14 in Table 2; four sites with anomalously shallow inclination (<10°) from the latter study may represent imperfect tilt adjustments and are excluded. A similar Eocene volcanic-sedimentary succession 100 km to the east in the vicinity of Akdağmadeni is of normal polarity and rotated more strongly counterclockwise (Tatar et al., 1996, result 13 in Table 1).

The Sivas basin to the east and south of this latter region (Fig. 1) comprises a thick sequence of sediments and volcanics laid down on ophiolite sheets emplaced during later stages of the basin's paleotectonic history and on the Akdağmadeni metamorphic complex. It is now considered essentially a postcollisional tectonic element (Yılmaz, 1994). Deformation has been dominated by north-south to northwest-southeast compressive regimes. The infilling succession is undeformed in the central part of the basin near the town of Sivas but becomes progressively incorporated into southward-migrating folds and thrusts toward the north (including the Central Anatolian thrust belt, Fig. 1) and into northward-migrating thrusts toward the south (Guezou et al., 1996). Subsequently, strike-slip deformation on ENE-WSW faults has dissected the region into a number of major blocks accommodating the tectonic escape of Anatolia. Extensive paleomagnetic study of this region by Gürsoy et al. (1997) has been supplemented by study of nine sites (with the latter including K-Ar age data) by Platzman et al. (1998) and of two sites by Kissel et al. (2003). The directions summarized by results 21 and 22 in Table 1 separate the results from above and below the Central Anatolian thrust belt from the results from Late Cretaceous to Eocene units. Both data sets show counterclockwise rotation, but because some of these magnetizations appear to be overprints (Gürsoy et al., 1997), they may be representative of the neotectonic deformation only.

Results from south of the Central Anatolian thrust belt are separated into data sets from fault blocks delineated by the major strike-slip faults. Result 25 incorporates the central part of the Sivas basin between the Central Anatolian thrust belt and the northward-directed Sivas thrust. Results 26–28 embrace neotectonic units in blocks emplaced successively to the southeast between the Sivas thrust, the Deliler fault, the Sarız-Kangal fault, and an area south of this latter fault system, respectively (see Fig. 3 of Gürsoy et al., 1997). The rotations identified by results 25–28 are uniformly counterclockwise and diminish in magnitude from north to south (Fig. 5). This latter observation suggests that neotectonic deformation resulting from ongoing impingement of Arabia into the Anatolian collage has been concentrated in the vicinity of the Sivas basin and the southwest continuation of this region into the Ecemiş or Central Anatolian fault zone (EFZ or CAF, Fig. 1). The Central Anatolian fault zone currently exhibits lower levels of seismic activity than the East Anatolian fault zone (Demirtaş and Yılmaz, 1996) but is interpreted by Koçyiğit and Beyhan (1998) to be a major left lateral intracontinental transform. In support of this we note that cumulative rotations on the west side of the Central Anatolian fault zone from Cappadocia to Erenlerdağı are lower in magnitude than on the east side (Fig. 5). However Jaffey and Robertson (2001) found that Plio-Quaternary motion across the Central Anatolian fault zone has been mainly extensional, and they speculate that counterclockwise rotation in this sector has rotated the fault zone away from an optimal orientation for accommodating strike-slip motions during tectonic escape. This interpretation predicts that in this part of Anatolia left lateral motion has now largely transferred to the southeast to become focused along the East Anatolian fault zone.

Group mean paleomagnetic directions from the region incorporating the Central Anatolian thrust belt and the Sivas basin embrace rock units with a wide age range, and because the mean

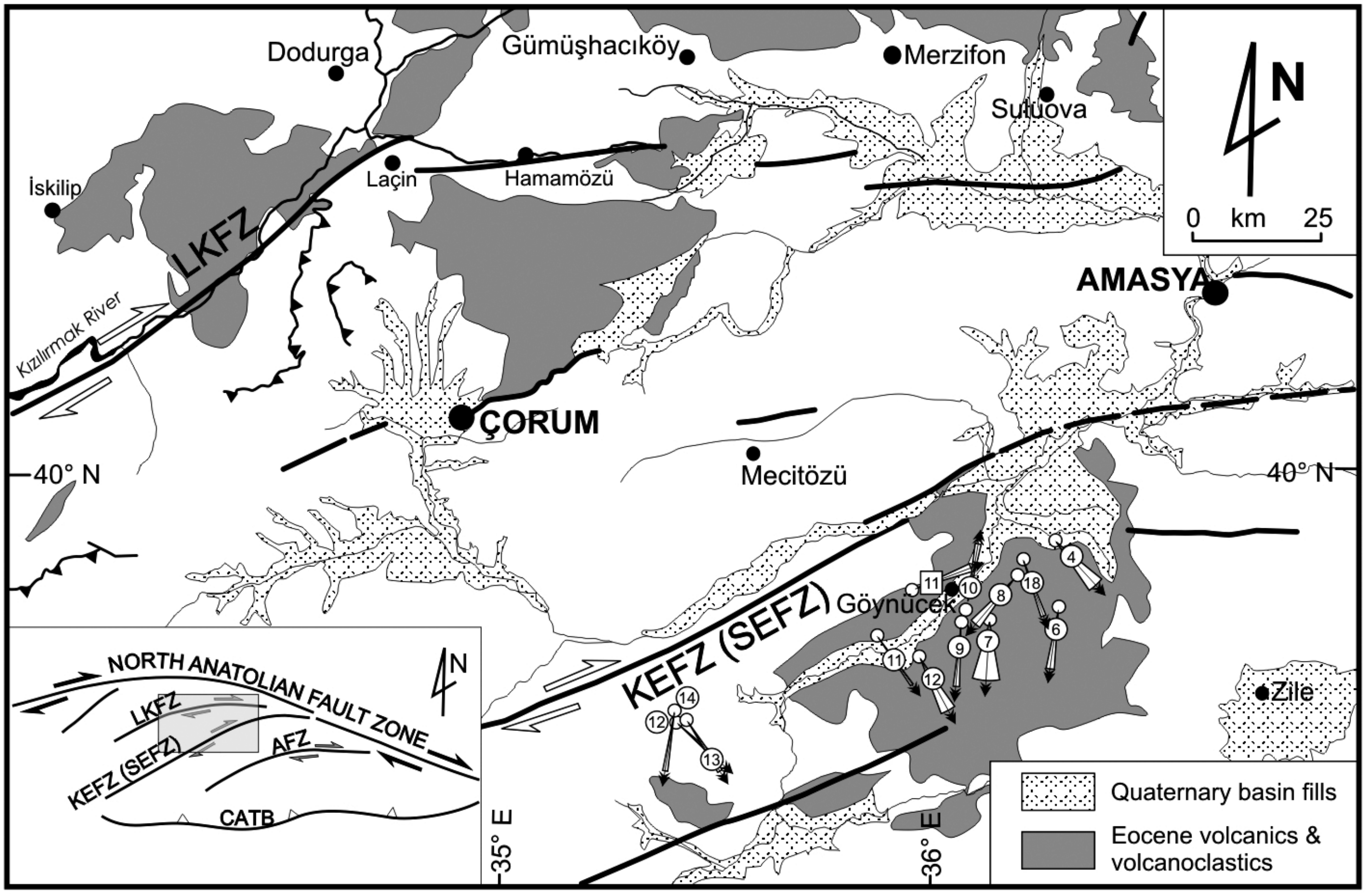


Figure 4. Outline geological map of the eastern part of the Çankırı basin (see Fig. 1), west-southwest of the town of Amasya, showing sampling sites and magnetic declinations derived from paleomagnetic study of Eocene-age volcanic units. The boxed site 11 is included in the results of this study (Table 2), and the circled site 11 and sites 12–14 and 18 are the results of Piper et al. (1996). The inset map shows the relationship of this region to the second-order strike-slip faults side splaying from the North Anatolian fault zone. KEF (SEFZ)—Kırıkkale-Erbaa fault zone (becoming the Sungurlu-Ezinepazarı fault zone, or SEFZ, to the west); LKFZ—Laçin-Kızılırmak fault zone.

rotations appear to be rather uniform across this zone (results 25–28 in Table 2), they provide an opportunity to evaluate the temporal development of the regional rotation. Most age assignments are based on a broad allocation to the Miocene, Pliocene, or Pleistocene periods based on stratigraphic and morphological criteria; just six of the studied units are directly linked to age dates (Platzman et al., 1998). The units assigned to the Pliocene are too few to yield a result significantly removed from the present dipole field ($N = 5$, D/I = 164/–52°, $\alpha_{95} = 26°$), but the Miocene and Pleistocene units, of obviously different age, yield similar directions of magnetization of D/I = 151/–54° ($N = 27$, $\alpha_{95} = 9.6°$) and D/I = 156/–47° ($N = 12$, $\alpha_{95} = 13.0°$). These are not significantly different; together with a comparable mean derived from Paleotectonic units covering the same region (D/I = 166/–53°, $\alpha_{95} = 7°$), they indicate that most, possibly all, of the block rotations in this region are young and confined to the last 2–3 m.y. (Fig. 6).

Results 36 and 37 summarize data from a large paleomagnetic collection (Gürsoy et al., 2003b) on an ~120 km north-south transect spanning the Afyon-Akşehir graben, a northwest-southeast-trending neotectonic feature showing strong contemporary seismic activity near the southwest limit of the zone of strike-slip terranes (Fig. 1). Rotation of blocks defined by faults paralleling this system certainly influence these data but are evidently minor, because a small amount of uniform clockwise rotation is observed on both sides of the fault zone. In Table 1, two mean directions are calculated, one for sites south of the Afyon-Akşehir graben and the other for sites to the north; the northern result is rotated, possibly around the arcuate form of the graben, marginally less than the result to the south. Current age determinations from this middle–late Miocene volcanic province are fairly dispersed, and it is also possible that the volcanic events in the south are more recent than those in the north (Table 1).

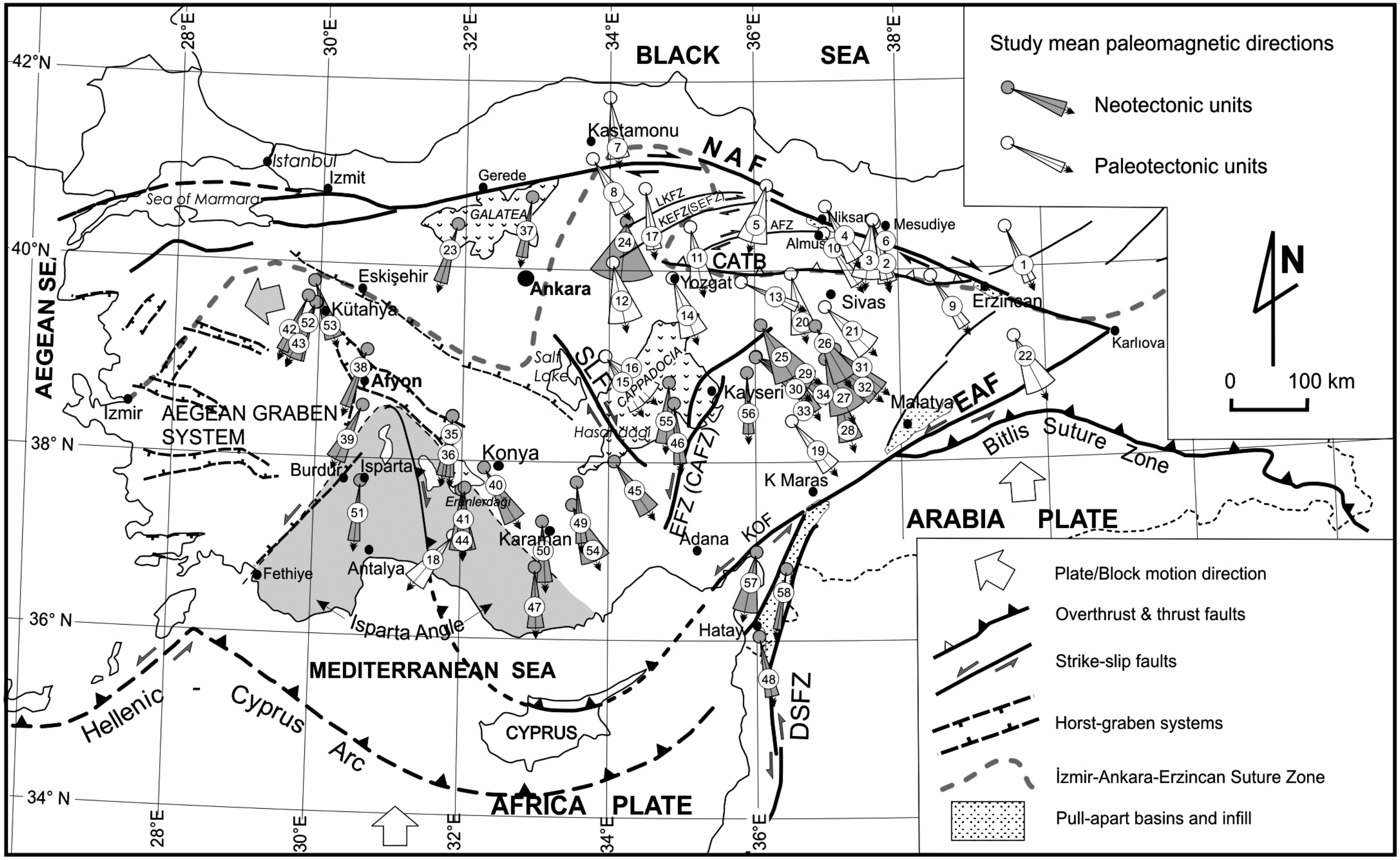
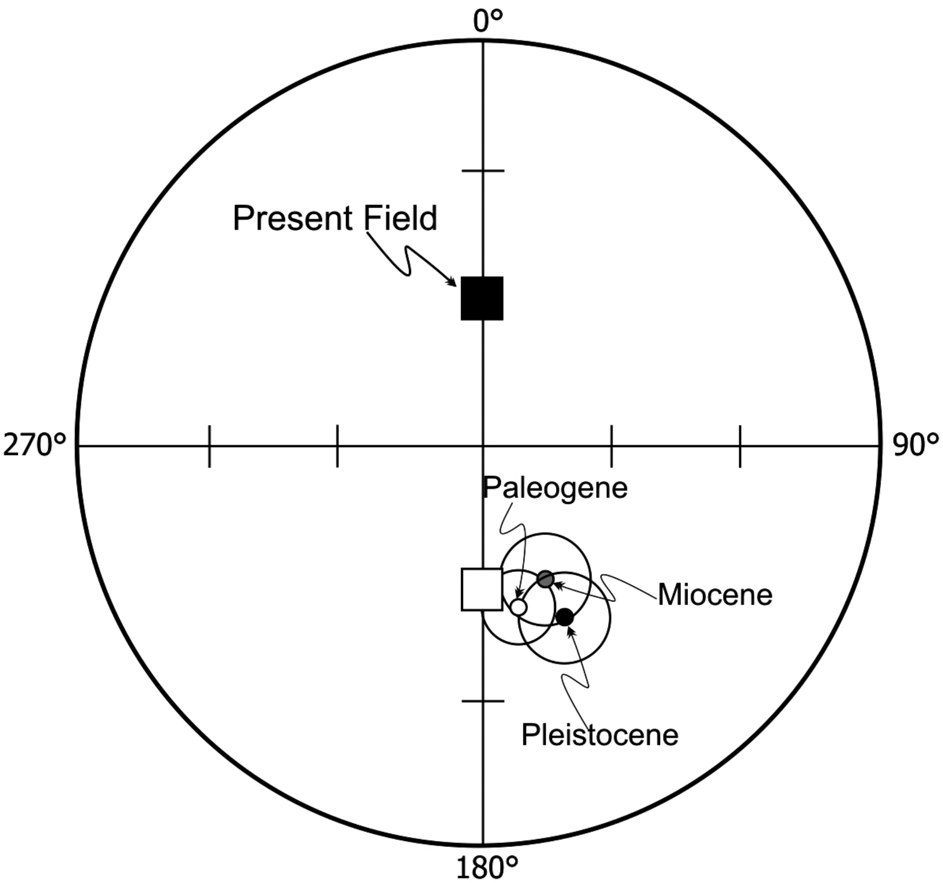


Figure 5. Schematic tectonic map of Turkey and adjoining region showing the distribution of major faults active during the Neotectonic regime and declinations of mean paleomagnetic vectors with arcs embracing 95% confidence from Late Cretaceous and younger rock units (all data with common reversed polarity; see Table 1). The thick dashed line (the İzmir-Ankara-Erzincan suture zone) is the site of Neo-Tethyan convergence during the paleotectonic era. AFZ—Almus fault zone; CATB—Central Anatolian thrust belt; DSFZ—Dead Sea fault zone; EAF—East Anatolian fault zone; EFZ (CAFZ)—Ecemiş fault zone (Central Anatolian fault zone); KEFZ (SEFZ)—Kırıkkale-Erbaa fault zone (becoming the Sungurlu-Ezinepazarı fault zone, or SEFZ, to the west); KOF—Karatos-Osmaniye fault zone; LKFZ—Laçin-Kızılırmak fault zone; NAF—North Anatolian fault zone; SLF—Salt Lake fault zone.



Age	D	I	N	R	α_{95}
Plio-Pleistocene	156	-47	12	11.10	13.0
Miocene	151	-54	27	24.21	9.6
Paleogene	166	-53	11	10.72	7.1

Figure 6. Mean reversed directions of magnetization with 95% confidence cones derived from rocks of the Sivas basin and adjoining blocks to the south-southeast. The paleomagnetic results from the region are grouped into site mean data of Late Cretaceous–Oligocene, Miocene, and Pleistocene age, and the squares indicate the direction of the present mean dipole field in this region. *R* is the magnitude of the resultant vector derived from *N* studies, and α_{95} is the radius of the cone of 95% confidence about the mean direction, expressed in terms of declination (∞E) and inclination (positive down, negative up).

REGIONAL ROTATIONS DEDUCED FROM PALEOMAGNETIC STUDY

In deformed terranes, tectonic rotation (R′) is determined by comparison of the mean directions with the predicted paleofield axis at the study location derived from the adjoining stable plate. The Anatolian terranes were accreted by closure of the Paleo-Tethys and the Neo-Tethys against the margin of Eurasia, and the latter continent is the appropriate comparator. Both observed and reference directions have confidence limits (ΔD and ΔD_{ref}, respectively). Beck (1980) proposed a confidence limit on R′ defined by

$$\Delta R' = \sqrt{(\Delta D^2 + \Delta D_{ref}^2)}.$$

Demarest (1983) concluded that this value overestimates the errors and showed that, provided α_{95} is small and preferably less than 10°, a standard correction factor is applicable. For $n \geq 6$, the correction factor lies between 0.78 and 0.80. ΔR′ is then derived from the equation

$$\Delta R' = 0.8 \geq (\Delta D^2 + \Delta D_{ref}^2).$$

Only if R′ > ΔR′ can a rotation be regarded as significant at a 95% confidence level. The paleofield directions predicted at the location of each data compilation in Table 1 have been calculated from the updated Eurasian apparent polar wander path (APWP) of Besse and Courtillot (2002) to determine the cumulative rotation at each location since the time of magnetization. Rotations with errors determined from the previous equation are summarized in Table 3.

In principle, contrasts in magnetic inclination (following adjustment for any tectonic tilt) are the signature of relative latitudinal movement. In practice, these latter differences have

TABLE 3. TECTONIC ROTATIONS OF PALEOMAGNETIC RESULTS FROM ANATOLIA DETERMINED BY COMPARISON WITH THE EURASIAN APWP OF BESSE AND COURTILLOT (2002)

Result	Age	Rotation	Result	Age	Rotation	Result	Age	Rotation
1	E	–23.7 ± 5.7	21	C–E	–33.0 ± 11.6	41	M	–3. 0 ± 7.1
2	E	–22.9 ± 6.0	22	E	–36.0 ± 13.1	42	M	20.1 ± 10.5
3	E	–3.2 ± 12.4	23	M	10.5 ± 5.6	43	M–P	20 ± 9.6
4	E	–38.3 ± 8.0	24	M	1.2 ± 26.5	44	M	–4.5 ± 10.8
5	E	6 ± 12.5	25	M–Pl	–36.3 ± 15.1	45	M–P	–14.3 ± 8.9
6	E	–22 ± 8.1	26	M–Pl	–34.4 ± 10.4	46	M	–19.4 ± 5.5
7	Cu	–16.3 ± 8.9	27	M–Pl	–28.6 ± 15.3	47	M	–5.3 ± 6.9
8	E	–24.2 ± 8.9	28	M–Pl	–20.8 ± 9.9	48	M	–16.4 ± 5.7
9	E	–42.8 ± 5.7	29	M	–72.7 ± 8.0	49	M–P	–8.9 ± 4.8
10	E	–42.7 ± 6.1	30	M	–56.7 ± 5.8	50	P	–7.1 ± 6.6
11	E	–19.7 ± 10.3	31	M	–56.8 ± 6.5	51	P	2.1 ± 5.2
12	O	–18.8 ± 14.2	32	M	–59.5 ± 8.3	52	P	21.8 ± 5.2
13	E	–74.6 ± 4.6	33	M	–61.2 ± 7.4	53	Pl	–22.0 ± 7.5
14	E	–29.3 ± 11.7	34	M	–37.9 ± 6.2	54	Pl	–26.8 ± 12.1
15	Pa	–40.2 ± 7.5	35	M	3.9 ± 4.8	55	Pl	–1.4 ± 8.4
16	Pa	–55.3 ± 13.1	36	M	–3.8 ± 4.1	56	Pl	–5.5 ± 4.8
17	E	–16.3 ± 6.6	37	M	3.5 ± 5.1	57	Pl	7.2 ± 10.2
18	Pa–E	32.1 ± 9.6	38	M	15 ± 4.8	58	Pl	5.7 ± 3.7
19	E	–48.2 ± 8.4	39	M	22.1 ± 7.9			
20	C–E	–21.6 ± 11.1	40	M	–34.8 ± 9.2			

Notes: Rock ages are C—Cretaceous; Cu—Upper Cretaceous; Pa—Paleocene; E—Eocene; O—Oligocene; M—Miocene; P—Pliocene; and Pl—Pleistocene.

proved difficult to apply on the neotectonic timescale, because in the majority of cases geomagnetic inclination along the Alpine-Himalayan orogen has been found to be too shallow to be accommodated by actual northward latitudinal movement (Cogné et al., 1999; Beck et al., 2001; Piper et al., 2002).

Platzman et al. (1998) and Kissel et al. (2003) conclude that the Anatolian region is rotating either uniformly (Kissel et al., 2003) or episodically (Platzman et al., 1998) as a coherent plate. Unfortunately, the former analysis excludes or dismisses a large fraction of the data, and the latter analysis is based on age dates from individual volcanic units that record stages in the secular variation and are unlikely to represent satisfactory time averages of the geomagnetic field. In conflict with these interpretations, the distribution of group mean declinations is found to change systematically across Anatolia (Fig. 5) in a way that shows that this region is not a coherent plate but is instead subject to distributed deformation. In central northern Anatolia, magnetic declinations in Eocene units within the Pontides are rotated counterclockwise on both sides of the right lateral North Anatolian fault zone (Fig. 2 and section headed "Crustal Deformation in Anatolia"), and most regional rotation in this area must therefore have predated initiation of the transform in late Pliocene times. Although the bulk of these rotations in northern Anatolia probably occurred before initiation of the North Anatolian fault zone (cf. Fig. 2), farther to the south comparable rotations are found in both Paleotectonic and Neotectonic units and support the observation from the Sivas basin and adjoining blocks (Fig. 6) that most regional rotations have occurred during the last few million years. In the Cappadocian ignimbrite succession, for example, a low rate of counterclockwise rotation averaging ~1°/m.y. is identified between 8.2 and 1.1 Ma, which subsequently accelerated after the last ignimbrite eruption at ca. 1.1 Ma (Piper et al., 2002). The cumulative rotation diminishes progressively to the west from the Sivas basin (Gürsoy et al., 1997) to Cappadocia (Piper et al., 2002) and on to the Karaman, Hasandağı (Gürsoy et al., 1998), and Erenlerdağı regions, where no significant rotation is evident (Tatar et al., 2002; see Fig. 1 for regional locations). Farther to the west, in the Isparta and Afyon areas near the western perimeter of extruded terranes, the declinations become slightly clockwise (Gürsoy et al., 2003b). Thus, the declinations south of latitude 40°N (all but one in neotectonic units younger than ca. 12 Ma) radiate around the Tauride arc, implying that crustal extrusion has had the net effect of expanding the perimeter of this arc (Fig. 7).

The distribution of tectonic rotations as a function of age (Fig. 8) confirms that the older Paleotectonic units are not rotated substantially more than the Neotectonic units. Hence rotations, at least within the Anatolian collage (the studied units in the north are mostly of Paleotectonic age), were concentrated within the latter part of the Neotectonic era. Also, the lower, and sometimes the clockwise, rotations are found within the western half of the Tauride arc (i.e., west of longitude ~33°E; see Fig. 8). Thus, the general picture to emerge from rotation of the paleomagnetic vectors from the present geomagnetic field di-

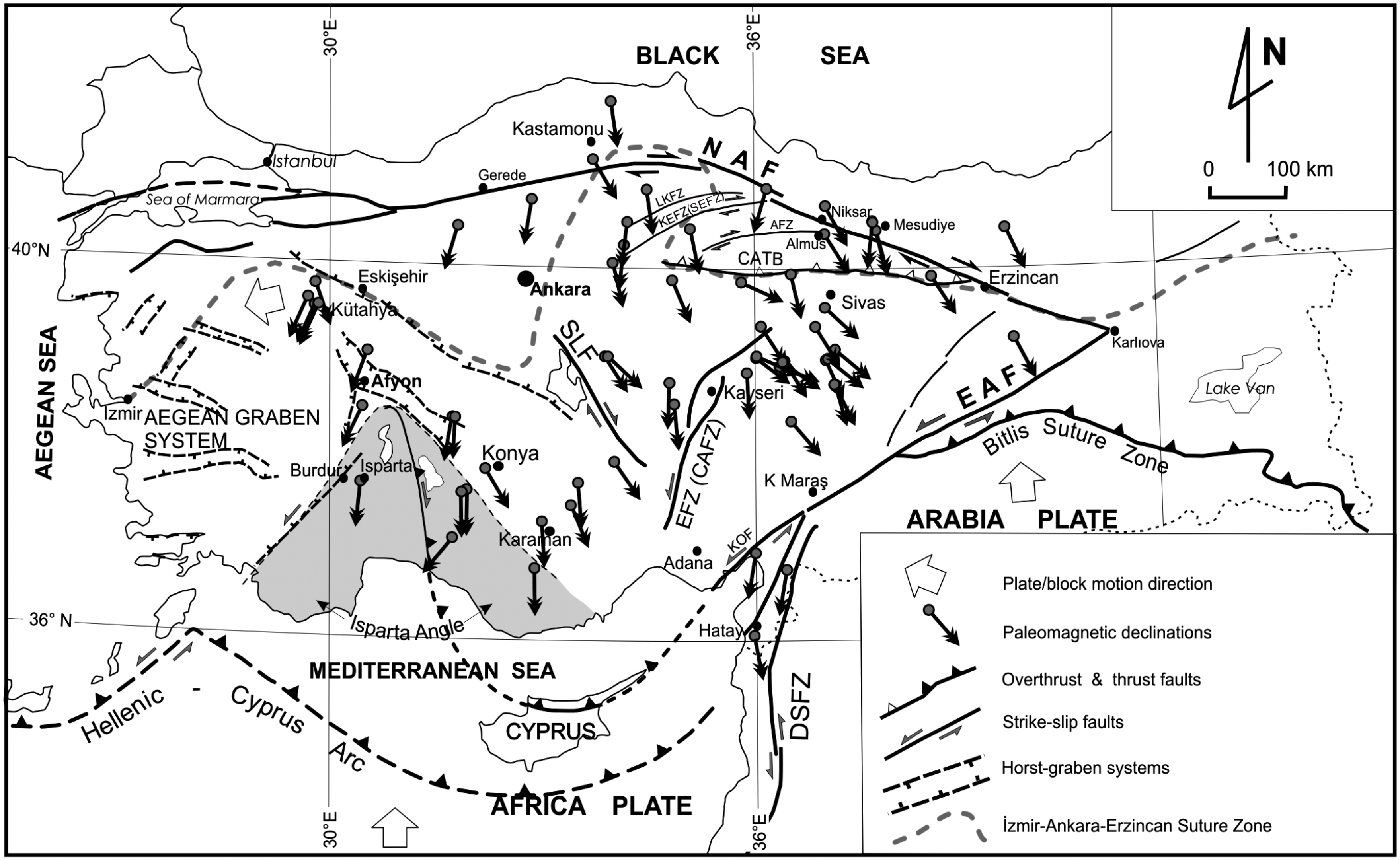


Figure 7. Regional trends of paleomagnetic declinations across Anatolia with an outline of the tectonic framework. AFZ—Almus fault zone; CATB—Central Anatolian thrust belt; DSFZ—Dead Sea fault zone; EAF—East Anatolian fault zone; EFZ (CAFZ)—Ecemiş fault zone (Central Anatolian fault zone); KEF (SEFZ)—Kırıkkale-Erbaa fault zone (becoming the Sungurlu-Ezinepazarı fault zone, or SEFZ, to the west); KOF—Karatos-Osmaniye fault zone; LKFZ—Laçin-Kızılırmak fault zone; NAF—North Anatolian fault zone; SLF—Salt Lake fault zone.

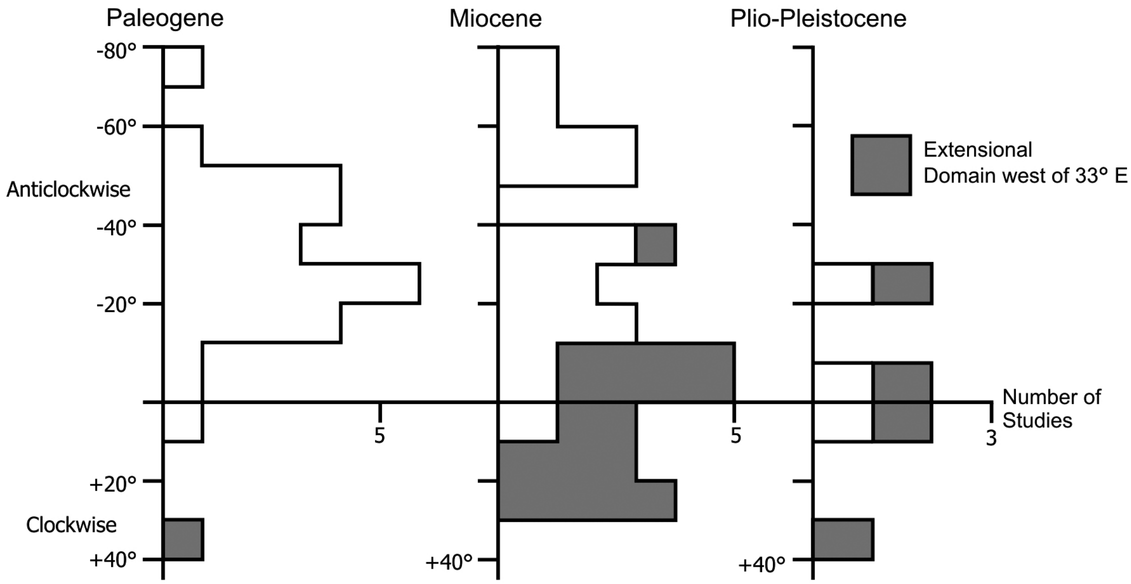


Figure 8. Frequency distribution of tectonic rotations derived from paleomagnetic studies in Late Cretaceous–Oligocene, Miocene, and Pleistocene rock units. The shaded areas are results from rock units sited in the western part of the zone of Anatolian extruded terranes (west of the longitude at 33°E, a line bisecting the Tauride arc; see Fig. 1).

rection as shown in Figure 7 is supported by calculated rotations (Table 3) and temporal variations (Fig. 8).

In summary, the case refuting the Anatolian plate model (Kissel et al. 2003) is based on five observations:

1. Although the paleomagnetic declinations come from rock suites with a range of ages, there is no apparent tendency for the rocks of older age to be systematically rotated more than younger rocks (Figs. 6 and 8).
2. Where it is possible to compare rotations in rocks of different ages from the same or adjoining tectonic blocks, as in Figure 6, they show little or no tendency to be larger in older rocks than in younger rocks prior to the Pliocene. Specifically, Paleogene rotations are typically no larger than Neogene rotations (Fig. 8)
3. In other regions such as Cappadoccia (Piper et al., 2002), where there is evidence for small amounts of uniform rotation since Miocene times, the rotation rate differs from rotation rates identified to the east (Fig. 6) and the west (Fig. 6 and Table 3).
4. The cumulative rotations determined from paleomagnetism are in any case unlikely to have resulted from a uniform stress regime of long continuity, because progressive clockwise rotation of the regional stress field appears to have occurred across Anatolia since Pliocene times. This is evident, for example, in the Ankara area, where regional compression moved from northwest-southeast before late Pliocene times to north-south during Pleistocene times (Koçyiğit, 1991), and in western Anatolia, where regional extension moved from north-south to northeast-southwest over broadly the same period (Glover and Robertson, 1998).
5. Magnetic directions have rotated across the area of Anatolia in the way predicted from westward extrusion of an upper brittle crust to the west and south by the indentation of Arabia in the east (Fig.7). While the limitations of modeling mean that the character of this distributed deformation is not well reproduced in small-scale experiments (Hubert-Ferrari et al., 2003), it is expected from the temporal changes in the tectonic regimes (Koçyiğit, 1991; Westaway, 2003) and from the presence of numerous crustal dislocations, such as terrane boundaries inherited from the paleotectonic collisional history.

The sizes of the blocks involved in these differential rotations are not usually well defined by surface outcrop due to selective erosion and burial beneath Pleistocene to recent infill. Figure 5 suggests that they are on a 10–100 km scale. The Cappadocian region, for example, could be rotating largely as a single ~100 km block bounded by the Central Anatolian and Salt Lake faults (Fig. 5). The Sivas basin on the east side of the Central Anatolian fault zone is dissected by major faults strike-slip spaced on a ~50 km scale that probably define the block boundaries, although their similar northeast-southwest trend provides little scope for differential rotation.

INTERPRETATION OF NEOTECTONIC MAGNETIC INCLINATIONS

Magnetic inclinations are more difficult to use for neotectonic analysis than declinations because they are often, although not exclusively, shallower than inclinations predicted from the geocentric axial dipole (GAD) assumption by amounts usually considered tectonically unreasonable. This is in part because they are recording a global phenomenon that produces far-sided paleomagnetic poles and requires a geomagnetic field during the late Tertiary, with time-averaged properties including nondipole components (Wilson, 1971). However, inclination shallowing by amounts greater than an amount attributable to this geomagnetic cause has been widely recorded within the Tethyan realm, specifically within the terranes of central Asia (e.g., Thomas et al., 1992; Cogné et al., 1999; Gilder et al., 2003). Of the possible causes of this phenomenon cited by Cogné et al. (1999), three are relevant to the discussion here: (1) a magnetic field anomaly, (2) compaction-induced shallowing of magnetizations, and (3) tectonic movements between Eurasia and the orogenic terranes. Although inclination shallowing tends to be slighter in igneous units than in sedimentary ones (Bazhenov and Mikolaichuk, 2002), it is often present. Volcanic formations, which predominate in the collections considered here, are identified as accurate recorders of the field direction during rapid cooling and do not invariably show inclination shallowing (Tatar et al., 2004a). When erroneously applied, tilt adjustments can produce inclination shallowing. Adjustments for tilt are not usually relevant to data from the Neotectonic domains, because any dip in volcanic units here is considered essentially primary; it could be the case with results from Paleotectonic units where tilt adjustment is considered a key part of the analysis. Hence, compaction-induced effects do not appear to be an adequate explanation, at least for the Miocene and younger data.

A departure from the GAD is recorded by an axial quadrupole during the last few million years, and this complexity must contribute to the data used to calculate reference paths for the major plates (Beck et al., 2001). Cogné et al. (1999) suggest that the reference APWP could itself be the cause of the large amount of inclination shallowing observed in central Asia due to internal deformation of the Eurasia plate. It remains possible, then, that residual inclination differences observed within the Tethyan region could be the expression of either a regional anomaly, imperfections in the reference path, or northward transport of terranes or a combination of more than one of these causes.

To evaluate the possible tectonic significance of inclination differences in Anatolia, we have undertaken an inclination-only analysis (Kono, 1980) of the data of Table 1 within latitude windows of 2° over successive steps of 1° across Anatolia. The results are summarized in Figure 9. Palaeolatitudes computed from the Eurasian master curve (Besse and Courtillot, 2002) for the region of central Anatolia (longitude 33° is used here) at 0 Ma lie between 2° and 3° south of the present-day latitude and are probably a reflection of the small long-term axial octopolar contribution to the magnetic field source. The paleolatitudes computed from Eocene–Oligocene, Miocene, and Plio-Pleistocene paleolatitudes plot a farther 2°–6° south of the latitudes predicted from the Eurasian path. The Eocene results (embracing results from Late Cretaceous to Oligocene units) are compatible with 600–800 km of northward transport of these terranes relative to Eurasia during the last ~40 m.y. Specific interest attaches to the Miocene results, because these data are contemporaneous with the final closure of the Neo-Tethys, they include rock suites recording the last pulses of subduction, and they date the initiation of uplift of the Anatolian plateau. Estimates of northward movement of between 200 and 800 km relative to Eurasia are possible here, and the anomaly declines from north to south (Fig. 9). We do not know the width of the crust bordering Eurasia that might have accommodated this convergence, but an ~2 km uplift of the ~600 km–wide Anatolide-Pontide region with compensating crustal thickening of ~10 km at depth would be compatible with the lower estimate. From these data we also favor a conservative estimate of northward transport of the Anatolian terranes, because the Plio-Pleistocene results continue to resolve paleolatitudes placing them 200–400 km south of the Eurasian prediction (Fig. 5). These results include data from Brunhes epoch volcanic fields, which, although not obviously differentiated from older Pliocene data (see Table 1), are unlikely to record northward movement discernible by paleomagnetism. It is therefore possible that internal deformation of the Eurasia plate is responsible for this anomaly (Cogné et al., 1999).

If latitudinal movement of terranes can be identified from the paleomagnetic data through the complexities introduced by the addition of long-term nondipole components (and possibly a regional anomaly), we would expect to see a contrasting picture emerging from the Aegean extensional province in western Turkey and the Aegean Sea. The tectonics of the Aegean region is driven predominantly by roll-back of the Hellenic arc, a process that is expanding the curvature of the arc and producing in its wake one of the most rapidly extending regions on the globe. Considerable uncertainty surrounds the duration of this extensional regime (Meulenkamp et al., 1988; Le Pichon et al., 1995; Kondopoulou, 2000), but we might expect to observe that younger Neotectonic terranes show less northward motion than their counterparts in Anatolia or even show indications of southward motion.

Unfortunately, the database from the Aegean extensional province (considered here to be between 23°E–30°E and 35°N–42°N), although comparable in size to that from Anatolia (Table 4), has poor latitudinal distribution: only results of Miocene age are satisfactory for this test. Paleocene–Oligocene results are concentrated in the Greek and Bulgarian Rhodopes beyond the northern limit of the extensional domain and identify a similar latitudinal difference to units in Anatolia of equivalent age, with ~900 km of northward motion relative to Eurasia implied (Fig.

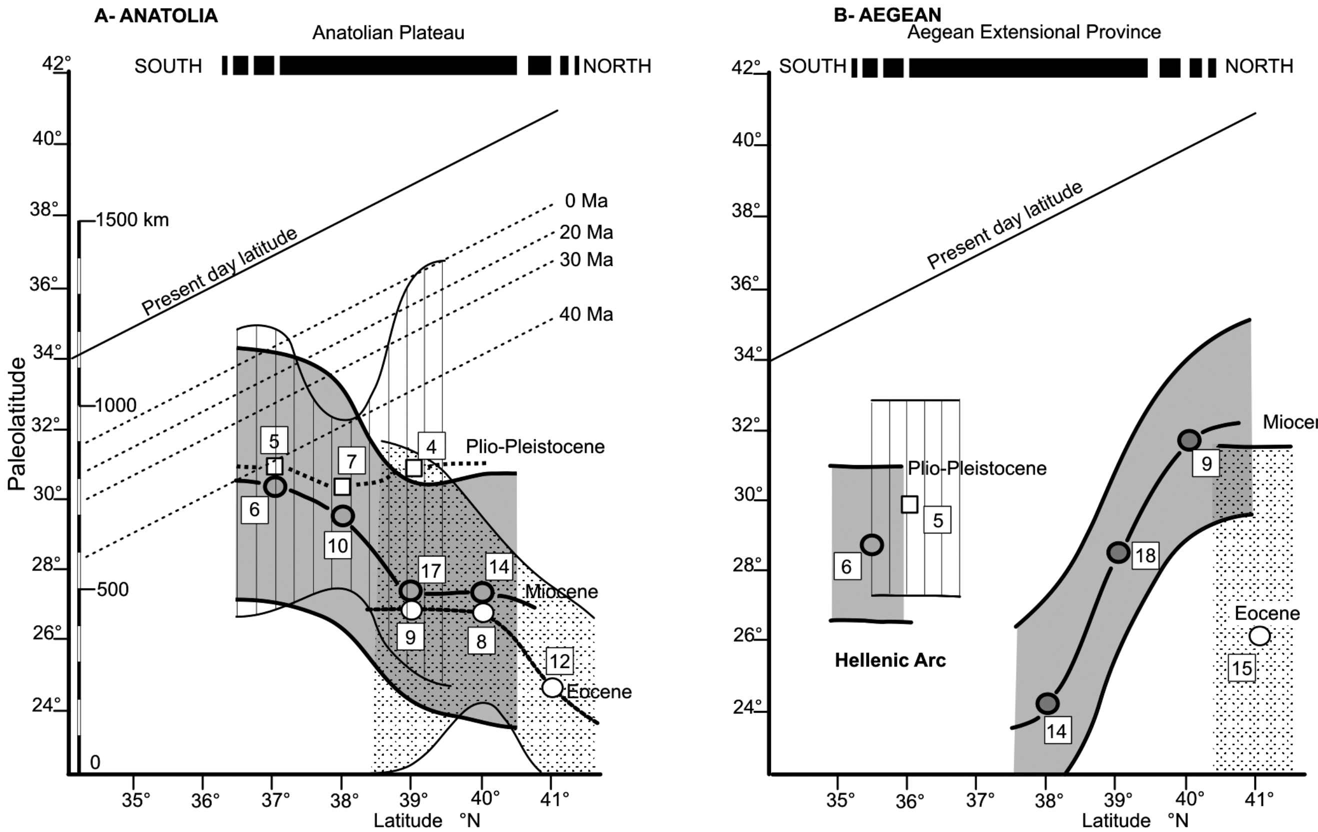


Figure 9. Distributions of mean paleolatitudes averaged in overlapping 2° windows of latitude incremented in steps of 1° as derived from inclination-only analysis of paleomagnetic studies in Late Cretaceous–Oligocene, Miocene, and Pliocene–Pleistocene rock units (Tables 1 and 3). The bands are 95% confidence limits, and the numbers in boxes refer to the number of studies. The sloping lines in (A) are paleolatitudes predicted at the latitude of central Anatolia at 10 Ma intervals from Eocene time to the present as derived from the apparent polar wander path of Eurasia (Besse and Courtillot, 2002).

9). The Miocene paleolatitudes are not systematically different from the Anatolian paleolatitudes. Only data at 39°N and 40°N might indicate southward transport by the crustal extension; the result at 38°N has anomalously shallow inclination. The Plio-Pleistocene mean for the Aegean in Figure 7 excludes an anomalously shallow inclination recorded by Løvlie et al. (1989) and shows ~3° divergence from the predicted field inclination, which is very similar to the Anatolian results of this age. Thus, at present there seems to be no distinction in paleolatitude between the Anatolian strike-slip and the Aegean extensional provinces, although both regions show a regional discrepancy from the field predicted from the Eurasian APWP. We conclude that either the APWP is an inadequate comparator due to a reason such as internal deformation of this major plate or that a regional anomaly is present within the Tethyan realm.

SUMMARY AND CONCLUSIONS

In spite of the obvious limitations imposed by the number of samples and the lack of structural and temporal control of many of the individual studies summarized in Table 1, the overall pattern of regional declinations in remarkably consistent. Away from the immediate vicinity of the intracontinental transforms, the largest rotations are concentrated near the pincer zone, where the northern perimeter of the Arabian indenter has impinged into the Anatolian collage. Rotation diminishes rapidly into the broader lower-strain region of accreted terranes to the west. The radial pattern of neotectonic declinations follows the curvature of the Tauride arc, changing from counterclockwise in Cappadocia and Karaman to near zero in Erenlerdağı, then clockwise into the Isparta angle and the Afyon region near the western extremity of the zone of extruded terranes. The temporal control, albeit weak, implies that this rotation was concentrated in Plio-Pleistocene times. This interval incorporates the time frame within which the intracontinental transforms were established to delineate the main zone of neotectonic deformation (Barka and Hancock, 1984; Barka and Gülen, 1989; Westaway, 2003), and slippage along these transforms has no doubt accommodated much of the westward extrusion of crust.

This pattern of rotations would be consistent with an ex-

TABLE 4. SUMMARY OF PALEOGENE AND NEOGENE PALEOMAGNETIC RESULTS FROM THE AEGEAN EXTENSIONAL DOMAIN BETWEEN 23° AND 17°E AND 35° AND 42°N

	Rock unit	Age: Stratigraphic	Age: Ma	°E	°N	D/I	α_{95}*	References
1	Arnea and Gomati intrusives	E–O	40.5–30.5	23.8	40.2	217.0/–31.0	9.0	Kondopoulou and Westphal (1986)
2	Chalkidik batholith	P–O	50–30	23.8	40.2	217.0/–28.0	9.0	Kondopoulou and Westphal (1986)
3	Thrace volcanics	E–O	45–36.6	26.0	41.0	187.0/–46.5	7.5	Kissel et al. (1986b)
4	Thrace volcanics	E–O	35–25	25.9	41.1	182.0/–52.0	6	Spais (1987)
5	Thrace intrusions	E–O	35–25	25.9	41.1	203.0/–46.0	5	Spais (1987)
6	Biga region volcanics	O		26.5	38.9	188.2/–42.2	9.8	Tapırdamaz and Yaltırak (1995)[1]
7	Elatia volcanics	O		24.3	41.5	194.5/–38.5	12.0	Atzemoglou et al. (1994)
8	Elatia intrusions	E–O	47–39	24.3	41.5	202.0/–39.0	11.0	Atzemoglou et al. (1994)[5]
9	Medousa ignimbrites	O	30	24.9	41.2	197.0/–46.0	5.4	Haubold et al. (1997)
10	Zagradenia volcanics	O		24.4	41.5	199.0/–49.0	4.5	Haubold et al. (1997)
11	Bulgarian volcanics	O–M	34–23	25.8	42.0	180.6/–54.8	16.3	Nozharov et al. (1977)
12	Canakkale volcanics			26.5	40.0	185.9/–54.0	4.3	Kissel et al. (1987)
13	Rhodope massif volcanics	E–M	33–18	24.2	41.3	203.7/–49.1	7.5	Atzemoglou et al. (1994)
14	Strymonikos lavas	M–O	30–20	23.0	41.1	206.0/–47.0	–‡	Westphal et al. (1991)
15	Xanthi intrusions	O	28	24.9	41.1	208.0/–53.5	9.5	Atzemoglou et al. (1994)
16	Kavala intrusions	M	18	24.0	40.9	204.3/–53.0	11	Atzemoglou et al. (1994)
17	Lemnos volcanics	M	21–17	25.2	39.8	214.0/–48.0	15.0	Westphal and Konopoulou (1993)
18	Karaburum volcanics	M	21.3–17.0	26.5	39.5	229.0/–51.0	16.4	Kissel et al. (1987)
19	Biga region volcanics (i)§	M		26.5	39.8	141.2/–56.6	15.3	Tapırdamaz and Yaltırak (1995)[2]
20	Biga region volcanics (ii)§	M		26.5	39.9	207.2/–58.0	15.5	Tapırdamaz and Yaltırak (1995)[3]
21	Bergama volcanics	M	17.5	27.0	38.8	202.0/–39.0	20.2	Kissel et al. (1987)
22	Izmir-Bergana volcanics	M	18.2–7.0	27.0	39.0	147.0/–52.0	8.7	Kissel et al. (1987)
23	Izmir-Foca lavas and sediments	M	18.5–15.0	27.0	38.3	151.0/–49.0	11.2	Kissel et al. (1986b)
24	Chios sediments	M	17–15	26.1	38.3	155.0/–45.0	12.0	Kondopoulou et al. (1993a)
25	Lesbos volcanics	M	18–15	26.4	39.3	192.0/–49.0	15.0	Kondopoulou (2001)
26	Lesbos volcanics	M	22–16	26.2	38.3	184.3/–48.5	6.9	Beck et al. (2001)
27	Lesbos volcanics	M	18.0–15.5	26.2	39.2	186.0/–49.5	6.8	Kissel et al. (1987)
28	Skyros volcanics	M	15	24.5	38.9	206.0/–45.0	7.7	Kissel et al. (1986a)
29	Volos volcanics			21.9	39.0	178.0/–59.0	7.6	Kissel et al. (1986a)
30	Chios lavas	M	15–14	26.1	38.2	155.0/–45.0	16	Kondopoulou et al. (1993b)
31	Kalamos	M	14.8–13.2			206.1/–51.8	4.2	Morris (1995)
32	Naxos granodiorite	M	13–11	25.4	37.0	153.4/–37.6	8.6	Morris and Anderson (1996)
33	Tinos/Island	M	12.3–11			207.8/–41.9	9.9	Avigad et al. (1998)
34	Seres intrusions	M	15–10	23.8	41.2	190.0/–52.0	7.6	Kondopoulou (1986)[4]
35	Evia volcanics	M	14–2	24.0	38.6	228.7/–45.8	6.6	Kissel et al. (1986a)
36	Kimi (Evia) andesites	M	15–13	24.1	38.6	231.0/–40.0	12.5	Morris (1995)
37	Evia marls	M		24.1	38.6	199.0/–42.0	11.7	Morris (1995)
38	Kythira	M				171.9/–39.4	8.1	Duermeijer et al. (2000)
39	Mykonos island	M	13–9	25.3	37.5	218.4/–23.0	7.3	Morris and Anderson (1996)[5]
40	Crete sediments	M	13–7	25.0	35.0	176.5/–45.0	5	Laj et al. (1982)
41	Kastellios sediments, Crete	M	11–10	24.5	35.3	172.0/–52.0	4.5	Şen et al. (1986)
42	Tinos intrusions	M	11.5	25.2	37.6	207.8/–41.9	9.9	Avigad et al. (1998)
43	West Crete	M	9.7–6.7	24.0	35.4	166.5/–45.9	7.4	Duermeijer et al. (1998)
44	Central Crete	M	9.7–6.7	25.0	35.2	145.3/–45.9	8.6	Duermeijer et al. (1998)
45	East Crete	M	9.7–6.7	25.6	35.1	167.9/–48.7	2.9	Duermeijer et al. (1998)
46	Kassos	M		27.0	35.4	157.3/–48.1	7.7	Duermeijer et al. (2000)
47	Kimi (Evia)	Mu–P		24.1	38.6	230.8/–40.3	12.5	Morris (1995)
48	Samos sediments	M	6	26.8	37.8	174.0/–51.0	4.7	Şen and Valet (1986)
49	Milos volcanics	P–Pl	3–1	24.4	36.1	169.0/–50.0	9	Kondopoulou and Pavlides (1990)
50	Crete sediments	P	6–2	25.0	35.0	180.1/–46.0	7	Laj et al. (1982)
51	Rhodes sediments	P–Pl		28.0	36.2	162.5/–46.8	8.3	Duermeijer et al. (2000)
52	Rhodes sediments	Mu–Pl	6–2	28.0	36.2	180.9/–46.9	8.2	Laj et al. (1982)
53	Rhodes sediments	P–Pl	0.5–3.5	28.0	36.2	181.7/–35.6	13.7	Løvlie et al. (1989)
54	Karpathos	P		27.2	35.5	162.0/–55.9	7.5	Duermeijer et al. (2000)

Notes: Rock ages are C—Cretaceous; Cu—Upper Cretaceous; Pa—Paleocene; E—Eocene; O—Oligocene; M—Miocene; Mu—upper Miocene; P—Pliocene; and Pl—Pleistocene.

*Radius of the cone of confidence about the mean direction.

§The small roman numerals refer to successive volcanic formations.

‡Denotes error not quoted in original.

[1]Oligocene result based on sites TR10, TR11, and TR19 of Tapırdamaz and Yaltırak (1995) and sites TH290 and TH300 of Kissel et al. (1986b).

[2]Miocene (i) result based on sites BA5, BA6, B2, and B10 of Orbay et al. (1995) and site TR21 of Tapırdamaz and Yaltırak (1995).

[3]Miocene (ii) result based on sites BA4, BA7, and B4 of Kissel et al. (1989) and sites IZ53, IZ50, TR08, and TR13 of Tapırdamaz and Yaltırak (1995).

[4]Includes data from Atzemoglou et al. (1994). [5]Includes the results of Haubold et al. (1997).

pansion of the extruded terrane perimeter to the west and amplification of curvature in the Tauride arc to the south. It further implies that tectonic escape contributes to producing the extensional regime in the west by expanding this perimeter of the domain. The present velocity of the crust in central Anatolia deduced from GPS is ~2 cm/yr to the WSW relative to Eurasia and suggests counterclockwise motion about a remote pole of rotation in the southeast Mediterranean region (McClusky et al., 2000). This analysis of contemporary motions from GPS clearly contrasts with the long-term distribution of rotations about local axes deduced from paleomagnetism. The deduced rates of rotation are also significantly larger than the rate of contemporary rotation of $1.3 \pm 0.1°$/m.y. deduced from GPS coverage of Anatolia (with the exception of the central part of the Tauride arc, where rotation has been near zero). The discrepancy between the short-term assessment derived from GPS and the long-term assessment deduced from paleomagnetism has several possible explanations. It could reflect long-term temporal changes in the regional stress field such as those highlighted in Section 2, or it could arise because the GPS results do not accommodate longer-term internal deformation of the rotating blocks or embrace sufficiently long periods to be representative of geological time. Furthermore, although the directions of motion deduced from GPS appear to be consistent, the rates of movement are variable, and a longer timespan of observation is probably required to clarify this point.

Although the regional coverage of neotectonic rotations is still incomplete (and is indeed largely constrained by the distribution of young volcanic rocks), the picture emerging is one of distributed crustal deformation with considerable regional variation (Fig. 7). Most rotation probably occurred in the south later than in the north, both because the effects of collision were realized earlier in the north and because the bulk of rotation demonstrably predates initiation of the North Anatolian fault zone here. In the south, most rotation has occurred during the last few m.y. and presumably succeeded the crustal thickening that followed late Miocene collision with Arabia along the Bitlis suture zone.

The conclusions derived from neotectonic magnetic inclinations are less clear, because the differences are within or close to 95% confidence levels and must be separated from the regional effects of inclination shallowing. They are compatible with up to 1000 km of northward movement of Anatolian terranes since ca. 40 Ma and during later stages of NeoTethyan closure, with up to a few hundred kilometers of this northward motion linked to crustal thickening since Miocene times.

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REFERENCES CITED

Atzemoglou, A., Kondopoulou, D., Papamarinopoulos, S., and Dimitriadis, S., 1994, Paleomagnetic evidence for block rotations in the western Greek Rhodope: Geophysical Journal International, v. 118, p. 221–230.

Avigad, D., Baer, G., and Heimann, A., 1998, Block rotations and continental extension in the central Aegean Sea: Paleomagnetic and structural evidence from Tinos and Mykronos, Cyclades, Greece: Earth and Planetary Science Letters, v. 157, p. 23–40, doi: 10.1016/S0012-821X(98)00024-7.

Barka, A.A., and Gülen, L., 1989, Complex evolution of the Erzincan Basin (eastern Turkey): Journal of Structural Geology, v. 11, p. 275–283, doi: 10.1016/0191-8141(89)90067-9.

Barka, A.A., and Hancock, P.L., 1984, Neotectonic deformation patterns in the convex-northwards arc of the North Anatolian Fault, *in* Dixon, J.G., and Robertson, A.H.F., eds., The geological evolution of the Eastern Mediterranean: Geological Society of London Special Publication 105, p. 763–773.

Barka, A., and Reilinger, R., 1997, Active tectonics of the Eastern Mediterranean region: Deduced from GPS, neotectonic and seismicity data: Annali Di Geofisica, v. 40, p. 587–610.

Barka, A., Reilinger, R., Şaroğlu, F., and Şengör, A.M.C., 1999. The Isparta Angle: Its importance in the Neotectonics of the Eastern Mediterranean Region, *in* Pişkin, Ö., et al., eds., Proceedings of the International Earth Science Colloquium on the Aegean Region, 9–14 October 1995, İzmir, Turkey, p. 3–17.

Baydemir, N., 1990, Paleomagnetism of the Eocene volcanic rocks in the eastern Black Sea region: Istanbul Üniversity Müh Fakulti, Yerbilimleri Dergisi, v. 7, p. 167–176 (in Turkish).

Bazhenov, M.L., and Mikolaichuk, A.V., 2002, Paleomagnetism of Palaeogene basalts from the Tian Shan, Kyrgysta: Rigid Eurasia and dipole geomagnetic field: Earth and Planetary Science Letters, v. 195, p. 155–166, doi: 10.1016/S0012-821X(01)00586-6.

Beck, M.E., 1976, Discordant paleomagnetic pole positions as evidence of regional shear in the western Cordillera of North America: American Journal of Science, v. 276, p. 694–712.

Beck, M.E., 1980, Paleomagnetic record of plate margin processes along the western edge of North America: Journal of Geophysical Research, v. 87, p. 7115–7131.

Beck, M.E., Burmester, R.F., Kondopoulou, P., and Atzemoglou, A., 2001, The paleomagnetism of Lesbos, NE Aegean, and the eastern Mediterranean inclination anomaly: Geophysical Journal International, v. 145, p. 233–245, doi: 10.1111/j.1365-246X.2001.00376.x.

Besse, J. and Courtillot, V., 2002. Apparent and true polar wander and the geometry of the geomagnetic field over the last 200 Myr: Journal of Geophysical Research, v. 107, no. B11, article no. 2300, doi:10.1029/20000JB000050.

Cande, S.C., and Kent, D.V., 1995, Revised calibration of the geomagnetic polarity timescale for the Late Cretaceous and Cenozoic: Journal of Geophysical Research, v. 100, p. 6093–6095, doi: 10.1029/94JB03098.

Channell, J.E.T., Tüysüz, O., Bektaş, O., and Şengör, A.M.C., 1996, Jurassic–Cretaceous paleomagnetism and paleogeography of the Pontides (Turkey): Tectonics, v. 15, p. 201–212, doi: 10.1029/95TC02290.

Cogné, J.P., Halim, N., Chen, Y., and Coutrillot, V., 1999, Resolving the problem of shallow magnetisations of Tertiary age in Asia: Insights from paleomagnetic data from the Qiangtang, Kunlun and Qaidam blocks (Tibet, China), and a new hypothesis: Journal of Geophysical Research, v. 104, p. 17,715–17,734, doi: 10.1029/1999JB900153.

Demarest, H.H., 1983, Error analysis for the determination of tectonic rotation from paleomagnetic data: Journal of Geophysical Research, v. 88, p. 4321–4328.

Demirtaş, R., and Yılmaz, R., 1996, Seismotectonics of Turkey: Ministry of Public Works and Settlement, Ankara.

Dewey, J.F., Hempton, M.R., Kidd, W.S., Şaroğlu, F., and Şengör, A.M.C., 1986, Shortening of continental lithosphere: The neotectonics of eastern Anatolia—A young collision zone, *in* Coward, M.P., and Reis, A.C., eds., Collision tectonics: Geological Society of London Special Publication 19, p. 2–36.

Duermeijer, C.E., Krijgsman, W., Langereis, C.G., and Ten Veen, J.H., 1998, Post early Messinian counter-clockwise rotations on Crete: Implications for the Late Miocene to Recent kinematics of the southern Hellenic Arc: Tectonophysics, v. 298, p. 177–189, doi: 10.1016/S0040-1951(98)00183-8.

Duermeijer, C.E., Nyst, M., Meijer, P.Th., and Langereis, C.G., 2000, The Neogene evolution of the Aegean arc: Indications for a Pleistocene tectonic rotation phase: Earth and Planetary Science Letters, v. 176, p. 509–525, doi: 10.1016/S0012-821X(00)00023-6.

England, P.C., and McKenzie, D.P., 1982, A thin viscous sheet model for continental deformation: Geophysical Journal of the Royal Astronomical Society, v. 70, p. 295–321.

Gilder, S., Chen, Y., and Sen, S., 2003, Oligo-Miocene magnetostratigraphy and rock magnetism of the Xishuigou section, Subei (Gansu Province, western China), and implications for shallow inclinations in central Asia: Journal of Geophysical Research, v. 106, p. 30505–30521, doi: 10.1029/2001JB000325.

Glover, C.P., and Robertson, A.H.F., 1998, Neotectonic intersection of the Aegean and Cyprus tectonic arcs: Extensional and strike-slip faulting in the Isparta Angle, SW Turkey: Tectonophysics, v. 298, p. 103–132, doi: 10.1016/S0040-1951(98)00180-2.

Guezou, J.C., Temiz, H., Poisson, A., and Gürsoy, H., 1996, Tectonics of the Sivas Basin: The Neogene record of the Anatolian accretion along the Inner Tauric Suture: International Geology Review, v. 38, p. 901–925.

Gürsoy, H., Piper, J.D.A., Tatar, O., and Temiz, H., 1997, A paleomagnetic study of the Sivas Basin, Central Turkey: Crustal deformation during lateral extrusion of the Anatolian Block: Tectonophysics, v. 271, p. 89–105, doi: 10.1016/S0040-1951(96)00242-9.

Gürsoy, H., Piper, J.D.A., Tatar, O., and Mesci, L., 1998, Paleomagnetic study of the Karaman and Karapınar volcanic complexes, central Turkey: Neotectonic rotation in the south-central sector of the Anatolian Block: Tectonophysics, v. 299, p. 191–211, doi: 10.1016/S0040-1951(98)00205-4.

Gürsoy, H., Piper, J.D.A., and Tatar, O., 1999, Paleomagnetic study of the Galatean Volcanic Province, north-central Turkey: Neogene deformation at the northern border of the Anatolian Block: Geological Journal, v. 34, p. 7–23, doi: 10.1002/(SICI)1099-1034(199901/06)34:1/2<7::AID-GJ812>3.0.CO;2-0.

Gürsoy, H., Tatar, O., Piper, J.D.A., Heimann, A., and Mesci, L., 2003a, Neotectonic deformation in the Gulf of Iskenderun, Southern Turkey, deduced from paleomagnetic study of the Ceyhan-Osmaniye Volcanics: Tectonics, v. 22, p. 1067–1079, doi: 10.1029/2003TC001524.

Gürsoy, H., Piper, J.D.A., and Tatar, O., 2003b, Neotectonic deformation in the western sector of tectonic escape in Anatolia: Paleomagnetic study of the Afyon region, central Turkey: Tectonophysics, v. 374, p. 57–79, doi: 10.1016/S0040-1951(03)00346-9.

Haubold, H., Scholger, R., Kondopoulou, D., and Mauritsch, H.J., 1997, New paleomagnetic results from the Aegean extensional province: Geologie en Mijnbouw, v. 76, p. 45–55, doi: 10.1023/A:1003134610530.

Hempton, M.R., 1987, Constraints on Arabian plate motion and extensional history of the Red Sea: Tectonics, v. 6, p. 687–705.

Hubert-Ferrari, A., King, G., Manighetti, I., Armijo, R., Meyer, B., and Tapponier, P., 2003, Long-term elasticity in the continental lithosphere: Modelling the Aden Ridge propagation and the Anatolian extrusion process: Geophysical Journal International, v. 153, p. 111–132, doi: 10.1046/j.1365-246X.2003.01872.x.

Jaffey, N., and Robertson, A.H.F., 2001, New sedimentological and structural data from the Ecemiş Fault Zone, Southern Turkey: Implications for its timing and offset and the Cenozoic tectonic escape of Anatolia: Journal of the Geological Society of London, v. 158, p. 367–378.

Kaymakçı, N., Duermeijer, C.E., Langeris, C., White, S.H., and Van Dijk, P.M., 2003, Paleomagnetic evolution of the Çankırı Basin (Central Anatolia, Turkey): Implications for oroclinal bending due to indentation: Geological Magazine, v. 140, p. 343–355, doi: 10.1017/S001675680300757X.

Kissel, C., Averbuch, O., Frizon de Lamotte, D., Monod, O., and Allerton, S., 1993, First paleomagnetic evidence for a post-Eocene clockwise rotation of the Western Taurides thrust belt east of the Isparta reentrant (Southwestern Turkey): Earth and Planetary Science Letters, v. 117, p. 1–14, doi: 10.1016/0012-821X(93)90113-N.

Kissel, C., Laj, C., and Mazaud, A., 1986a, First paleomagnetic results from Neogene formations in Evia, Skyros and the Volos region and the deformation of Central Aegea: Geophysical Research Letters, v. 13, p. 1446–1449.

Kissel, C., Kondopoulou, D., Laj, C., and Papadopoulos, P., 1986b, New paleomagnetic data from Oligocene Formations of Northern Aegea: Geophysical Research Letters, v. 13, p. 1039–1042.

Kissel, C., Laj, C., Poisson, A., Savaçın, Y., Simeakis, K., and Mercier, J.L., 1987, Paleomagnetic evidence for Neogene rotational deformations in the Aegean domain: Tectonics, v. 5, p. 783–795.

Kissel, C., Laj, C., Poisson, A., and Simeakis, K., 1989, A pattern of block rotations in Central Aegea, *in* Kissel, C., and Laj, C., edrs., Palaeomagnetic rotations and continental deformation: Dordrecht, Kluwer, p. 115–130.

Kissel, C., Laj, C., Poisson, A., and Görür, N., 2003, Paleomagnetic reconstruction of the Cenozoic evolution of the Eastern Mediterranean: Tectonophysics, v. 362, p. 199–217, doi: 10.1016/S0040-1951(02)00638-8.

Koçyiğit, A., 1991, Changing stress orientation in progressive intracontinental deformation as indicated by the Neotectonics of the Ankara region (NW Central Anatolia): Turkish Association of Petroleum Geologists Bulletin, v. 3, no. 1, p. 48–55.

Koçyiğit, A., and Beyhan, A., 1998, A new intracontinental transcurrent structure: The Central Anatolian Fault Zone, Turkey: Tectonophysics, v. 284, p. 317–336, doi: 10.1016/S0040-1951(97)00176-5.

Koçyiğit, A., Yusufoğlu, H., and Bozkurt, E., 1999, Evidence from the Gediz graben for episodic two-stage extension in western Turkey: Journal of the Geological Society of London, v. 156, p. 605–616.

Koçyiğit, A., Ünay, E., and Saraç, G., 2000. Episodic graben formation and extensional neotectonic regime in west central Anatolia and the Isparta Angle: A case study in the Akşehir-Afyon Graben, Turkey, *in* Bozkurt, E., et al., eds., Tectonics and magmatism in Turkey and the surrounding area, Geological Society of London Special Publication 73, p. 405–421.

Kondopoulou, D., 1986, Tertiary rotational deformation in the Greek Serbomacedonian Massif: Bulgarian Geophysical Journal, v. 12, no. 4, p. 71–80.

Kondopoulou, D., 1994**[AQ1]**, Some constraints on the origin and timing of magnetisation for Mio-Pliocene sediments from N. Greece: Bulletin of the Geological Society of Greece, v. 30, no. 5, p. 53–66.

Kondopoulou, D., 2000, Paleomagnetism in Greece: Cenozoic and Mesozoic components and their geodynamic implications: Tectonophysics, v. 326, p. 131–151, doi: 10.1016/S0040-1951(00)00150-5.

Kondopoulou, D., and Pavlides, S., 1990. Paleomagnetic and neotectonic evidence for different deformation patterns in the South Aegean volcanic arc: The case of Milos Island, *in* Proceedings of the International Earth Science Congress on Aegean Regions, Izmir, p. 210–223.

Kondopoulou, D., and Westphal, M., 1986, Paleomagnetism of the Tertiary intrusives from Chalkidiki (northern Greece): Journal of Geophysics, v. 59, p. 62–66.

Kondopoulou, D., De Bonis, L., Koufos, G., and Şen, Ş., 1993a. Paleomagnetic data and biostratigraphy of the Middle Miocene vertebrate locality of

Thymiana, Chios Island, Greece, *in* Proceedings of the 2nd Congress of the Hellenic Geophysical Union, Florina, 626–636.

Kondopoulou, D., Leci, V., and Symeakis, C., 1993b. Paleomagnetic study of the Tertiary volcanics in Chios Island, Greece, *in* Proceedings of the 2nd Congress of the Hellenic Geophysical Union, Florina, 676–687.

Kono, M., 1980, Statistics of paleomagnetic inclination data: Journal of Geophysical Research, v. 85, p. 3878–3882.

Krijgsman, W., Duermeijer, C.E., Langeris, C.G., Bruijn, H., Saraç, G., and Andriessen, P.A.M., 1996, Magnetic polarity stratigraphy of late Oligocene to middle Miocene mammal-bearing continental deposits in central Anatolia (Turkey): Newsletter of Stratigraphy, v. 34, no. 1, p. 13–29.

Laj, C., Jamet, M., Sorel, D., and Valente, J.P., 1982. First paleomagnetic results from Mio-Pliocene series of the Hellenic sedimentary arc, *in* Le Pichon, X., Augustithis, S.S. and Mascle, J., eds., Geodynamics of the Hellenic Arc and Trench, special volume: Tectonophysics, v. 86, p. 45–67.

Le Pichon, X., Chamot-Rooke, N., Lallemant, S., Noomen, R., and Veis, G., 1995, Geodetic determination of the kinematics of central Greece with respect to Europe: Implications for eastern Mediterranean tectonics: Journal of Geophysical Research, v. 100, B7, p. 12,675–12,690, doi: 10.1029/95JB00317.

Løvlie, R., Stole, G., and Spjeldnaes, N., 1989, Magnetic polarity stratigraphy of Pliocene and Pleistocene marine sediments from Rhodes, eastern Mediterranean: Physics of the Earth and Planetary Interiors, v. 54, p. 340–352, doi: 10.1016/0031-9201(89)90251-3.

McClusky, S., Balassanian, S., Barka, A., Demir, C., Ergintav, S., Georgiev, I., Gürkan, O., Hamburger, M., Hurst, K., Kahle, H., Kastens, K., Kekelidze, G., King, R., Kotzev, V., Lenk, O., Mahmoud, S., Mishin, A., Nadariya, M., Ouzounis, A., Paradissis, D., Peter, Y., Prilepin, M., Reilinger, R., Sanli, I., Seeger, H., Tealeb, A., Toksöz, M.N., and Veis, G., 2000, Global Positioning System constraints on plate kinematics and dynamics in the eastern Mediterranean and Caucasus: Journal of Geophysical Research, v. 105, p. 5695–5719, doi: 10.1029/1999JB900351.

Meulenkamp, J.E., Wortel, W.J.R., Van Wamel, W.A., Apaskman, W., and Hoogerduyn Strating, E., 1988, On the Hellenic subduction zone and geodynamic evolution of Crete in the late middle Miocene: Tectonophysics, v. 146, p. 203–215, doi: 10.1016/0040-1951(88)90091-1.

Michel, G.W., Waldhor, M., Neugebauer, J., and Appel, E., 1995, Sequential rotation of stretching axes, and block rotations: A structural and paleomagnetic study along the North Anatolian Fault: Tectonophysics, v. 243, p. 87–118, doi: 10.1016/0040-1951(94)00193-D.

Morris, A., 1995, Rotational deformation during Paleogene thrusting and basin closure in eastern central Greece: Paleomagnetic evidence from Mesozoic carbonates: Geophysical Journal International, v. 121, p. 827–847.

Morris, A., and Anderson, M., 1996, First paleomagnetic results from the Cycladic Massif Greece and their implications for the Miocene extension directions and tectonic models in the Aegean: Earth and Planetary Science Letters, v. 142, p. 397–408, doi: 10.1016/0012-821X(96)00114-8.

Nozharov, P.B., Veljovic, D., and Petkov, N.I., 1977, Paleomagnetism of some Upper Cretaceous and Tertiary volcanites in Sredna Gora and Eastern Rhodopes: Compte Rendu Acadamie Bulgare Science, v. 30, p. 843–845.

Orbay, N., and Bayburdi, A., 1979, Paleomagnetism of dykes and tuffs from the Mesudiye region and rotation of Turkey: Geophysical Journal of the Royal Astronomical Society, v. 59, p. 437–444.

Orbay, N., Sanver, M., Tapırdamaz, C., Hisarlı, M., and Özçep., F., 1995, Batı Anadolu'nun Mikro Bloklarının Paleomagnetizması ve Genç Tektonik Evrimi: TUBITAK Projesi (YBAG-127).

Piper, J.D.A., Moore, J., Tatar, O., Gürsoy, H., and Park, R.G., 1996, Paleomagnetic study of crustal deformation across an intracontinental transform: The North Anatolian Fault Zone in Northern Turkey, *in* Morris, A., and Tarling, D.H., eds., Paleomagnetism of the Mediterranean Regions: Geological Society of London Special Publication 105, p. 299–310.

Piper, J.D.A., Tatar, O., and Gürsoy, H., 1997, Deformational behaviour of continental lithosphere deduced from block rotations across the North Anatolian Fault Zone in Turkey: Earth and Planetary Science Letters, v. 150, p. 191–203, doi: 10.1016/S0012-821X(97)00103-9.

Piper, J.D.A., Gürsoy, H., and Tatar, O., 2002, Paleomagnetism and magnetic properties of the Cappadocian Ignimbrite Succession, Central Turkey and Neogene tectonics of the Anatolian Collage: Journal of Volcanology and Geothermal Research, v. 117, p. 237–262, doi: 10.1016/S0377-0273(02)00221-4.

Platzman, E.S., Platt, J.P., Tapırdamaz, C., Sanver, M., and Rundle, C.C., 1994, Why are there no clockwise rotations along the North Anatolian Fault Zone?: Journal of Geophysical Research, v. 99, p. 21,705–21,716, doi: 10.1029/94JB01665.

Platzman, E.S., Tapırdamaz, C., and Sanver, M., 1998, Neogene anticlockwise rotation of central Anatolia (Turkey): Preliminary paleomagnetic and geochronological results: Tectonophysics, v. 299, p. 175–189, doi: 10.1016/S0040-1951(98)00204-2.

Reilinger, R., and McClusky, S., 2001, GPS constraints on block motions and deformation in western Turkey and the Aegean: Implications for earthquake hazards, *in* Taymaz, T., ed., Symposia on seismotectonics of the north-western Anatolia-Aegean and recent Turkish earthquakes: Istanbul, Istanbul Technical University, 14–20.

Robertson, A.H.F., Dixon, J.E., and Brown, A., 1996, Alternative tectonic models for the late Palaeozoic–early Tertiary development in the Eastern Mediterranean region, *in* Morris, A., and Tarling, D.H., eds.), Paleomagnetism and tectonics of the Mediterranean region: Geological Society of London Special Publication 105, p. 239–263.

Sanver, M., and Ponat, E., 1980. Paleomagnetism of the magmatic rocks in Kirsehir and surrounding area: Bulletin of the Kandilli Observatory, Istanbul University, p. 1–11.

Sarıbudak, M., 1989, New results and a paleomagnetic overview of the Pontides in northern Turkey: Geophysical Journal International, v. 90, p. 521–531.

Sarıbudak, M., Sanver, M., Şengör, A.M.C., and Görür, N., 1990, Paleomagnetic evidence for substantial rotation of the Almacık flake within the North Anatolian Fault Zone, NW Turkey: Geophysical Journal International, v. 102, p. 563–568.

Şen, Ş., and Valet, J.P., 1986, Magnetostratigraphy of Late Miocene continental deposits in Samos, Greece: Earth and Planetary Science Letters, v. 80, p. 167–174, doi: 10.1016/0012-821X(86)90030-0.

Şen, Ş., Valet, J.P., and Ioakim, C., 1986, Magnetostratigraphy and biostratigraphy of the Neogene deposits of Kastellios hill, Central Crete, Greece: Paleogeography, Paleoclimatology and Paleoecology, v. 53, p. 321–334, doi: 10.1016/0031-0182(86)90066-0.

Şengör, A.M.C., and Yılmaz, Y., 1981, Tethyan evolution of Turkey: A plate tectonic approach: Tectonophysics, v. 75, p. 181–241, doi: 10.1016/0040-1951(81)90275-4.

Seyitoğlu, G., and Scott, B.C., 1996, The cause of N-S extensional tectonics in western Turkey: Tectonic escape vs. back-arc spreading versus orogenic collapse: Journal of Geodynamics, v. 22, nos. 1–2, p. 145–153, doi: 10.1016/0264-3707(96)00004-X.

Spais, C., 1987, Paleomagnetic and magnetic fabric investigations of Tertiary rocks from the Alexandroupolis area, NE Greece [Ph.D. thesis]: University of Southampton, Southampton, England.

Tapırdamaz, C., and Yaltırak, C., 1995, Trakya'da Genc Volkanik Kayaçlar Üzerinde Yapılan Paleomagnetik Çalışma: Jeofizik, v. 9–10, p. 213–218.

Tatar, O., Piper, J.D.A., Park, R.G., and Gürsoy, H., 1995, Paleomagnetic study of block rotations in the Niksar overlap region of the North Anatolian Fault Zone, Central Turkey: Tectonophysics, v. 244, p. 251–266, doi: 10.1016/0040-1951(94)00241-Z.

Tatar, O., Piper, J.D.A., Gürsoy, H., and Temiz, H., 1996, Regional significance of Neotectonic counterclockwise rotation in Central Turkey: International Geology Review, v. 38, p. 692–700.

Tatar, O., Piper, J.D.A., and Gürsoy, H., 2000. Paleomagnetic study of the Erciyes sector of the Ecemiş Fault Zone: Neotectonic deformation in the southeastern part of the Anatolian Block, *in* Bozkurt, E., et al., eds., Tec-

tonics and magmatism in Turkey and the surrounding area: Geological Society of London Special Publication 173, p. 423–440.
Tatar, O., Gürsoy, H., and Piper, J.D.A., 2002, Differential neotectonic rotations in Anatolia and the Tauride Arc: Paleomagnetic investigation of the Erenlerdağı Complex and Isparta volcanic district, south-central Turkey: Journal of the Geological Society of London, v. 159, p. 281–294.
Tatar, O., Piper, J.D.A., Gürsoy, H., Heimann, A., and Koçbulut, F., 2004a, Neotectonic deformation in the transition zone between the Dead Sea Transform and the East Anatolian Fault Zone, Southern Turkey: A paleomagnetic study of the Karasu Rift volcanism: Tectonophysics, v. 385, p. 17–43, doi: 10.1016/j.tecto.2004.04.005.
Tatar, O., Temiz, H., Gürsoy, H., and Guezou, J.C., 2004b, Orta ve Kuzey Anadolu Miyosen-Kuvaterner volkanizması ışığında Avrasya-Anadolu Levhaları çarpışması: Ankara, TUBİTAK Projesi No: YDABÇAG 198Y092, 55 p..
Thomas, J.C., Perroud, H., Cobbold, P.R., Bazhenov, M.L., Burtman, V.S., Chavin, A., and Sadybakasov, E., 1992, A paleomagnetic study of Tertiary formations from the Kyrgyz Tien Shan and its tectonic implications: Journal of Geophysical Research, v. 98, p. 9571–9589.
Van der Voo, R., 1968, Paleomagnetism and the Alpine tectonics of Eurasia, Part 4: Jurassic, Cretaceous and Eocene pole positions form northeastern Turkey: Tectonophysics, v. 6, no. 3, p. 251–269, doi: 10.1016/0040-1951(68)90053-X.
Westaway, R., 2003, Kinematics of the Middle East and Eastern Mediterranean updated: Turkish Journal of the Earth Sciences, v. 12, p. 5–46.
Westphal, M., and Kondopoulou, D., 1993, Paleomagnetism of Miocene volcanics in the vicinity of the north Aegean Trough: Annales Tectonicae, v. 7, no. 2, p. 142–149.
Westphal, M., Kondopoulou, D., Edel, J.B., and Pavlides, S., 1991, Paleomagnetism of late Tertiary and Plio-Pleistocene formations from N. Greece: Bulletin of the Geological Society of Greece, v. 25, no. 3, p. 239–250.
Wilson, R.L., 1971, Dipole offset: The time average paleomagnetic field over the past 25 million years: Geophysical Journal of the Royal Astronomical Society, v. 22, p. 481–504.
Yılmaz, A., 1994, An example of post-collisional trough: Sivas Basin, Turkey: Proceedings, 10th Petroleum Congress of Turkey, Ankara, 21–23.
Yılmaz, Y., 1990a, Allochthonous terranes of Tehyan Middle East: Anatolia and the surrounding regions: Philosophical Transactions of the Royal Society of London, Series A, v. 331, p. 611–624.
Yılmaz, Y., 1990b, Comparison of young volcanic associations of western and eastern Anatolia formed under compressional regime: A review, *in* Le Fort, P., et al., eds., Collision magmatism, special volume: Journal of Volcanology and Geothermal Research, v. 44, p. 69–87.
Yılmaz, Y., Şaroğlu, F., and Güner, Y., 1987, Initiation of the neomagmatism in East Anatolia: Tectonophysics, v. 134, p. 177–199, doi: 10.1016/0040-1951(87)90256-3.

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