Paleomagnetism of the Miocene Soma basin and its structural implications on the central sector of a crustal-scale transfer zone in western Anatolia (Turkey)

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A B S T R A C T

The İzmir–Balıkesir Transfer Zone (İBTZ) is a crustal-scale major tectonic feature in western Anatolia accommodating differential extension between the Menderes (MCC) and Cycladic (CCC) core complexes. The kinematics and evolution of the southern part of the İBTZ are well constrained, but its northern continuation remains unstudied. This part is crucial in understanding the complete evolution of western Anatolian tectonics, as well as a possible link between the İBTZ and North Anatolian Fault Zone (NAFZ). In this study, new and extensive paleomagnetic, structural, and stratigraphic data were collected from the Soma basin. These results show that the basin evolved as a part of the İBTZ, including two separate rotational phases. During the first (middle Miocene) phase, deformation was characterized by transcurrent tectonics and associated block rotations as much as ~30° clockwise, during which the İBTZ evolved as a wide dextral shear zone. During the second (late Miocene–recent) phase, the mode of deformation in the Aegean region switched from localized to distributed extension, related to the acceleration of the tearing-off of the African Oceanic slab below the Aegean region. This led to the narrowing of the İBTZ as a discrete brittle shear around the Soma basin, manifested by decoupling of dextral strike-slip and normal faulting. In this period, the Soma basin underwent around −21° counter-clockwise rotation. Based on our new results, the İBTZ extends further to the north and possibly interacts with the southern branch of the NAFZ since the late Miocene.

1. Introduction

The unique position of western Anatolia (Fig. 1a) in the African-European convergent tectonic setting has resulted in a complex deformation history with several large-scale tectonic features. The region is dominated by NNE directed extensional deformation since the late Eocene (Forster and Lister, 2009; Gautier et al., 1999; Jolivet and Brun, 2010; Trief et al., 2009; van Hinsbergen and Schmid, 2012). There are two major hypotheses to explain the interplay between this extension and the observed tectonic features. One hypothesis is the westward escape of Anatolia (Dewey et al., 1986; Şengör, 1979; Şengör and Yilmaz, 1981; Şengör et al., 1985). This process could have led to the formation of the dextral North Anatolian Fault Zone (NAFZ), as well as of the sinistral East Anatolian Fault Zone along which the Anatolian Block moves westwards. The second hypothesis is related to the rollback of the north-eastward subducting Aegean slab and resulting back-arc extension (Birgul et al., 2011; Le Pichon and Angelier, 1979; Meulenkamp et al., 1988; van Hinsbergen et al., 2005; van Hinsbergen et al., 2010b). Several studies have combined these two hypotheses, suggesting that rollback of the Aegean slab could have been a cause for the westward escape of Anatolia (Bozkurt, 2001; Brun et al., 2016).

In any case, Cenozoic extension in western Anatolia has resulted in two regional extensional features, namely the Cycladic (CCC) and Menderes (MCC) metamorphic core complexes. Moreover, another tectonic feature, the İzmir-Balıkesir Transfer Zone (İBTZ), initiated in between the MCC and CCC, and south of the NAFZ. The İBTZ was first identified by Kaya (1981), who divided the region between İzmir and Balıkesir into several NNE-trending Neogene depressions bounded by oblique-slip faults with a considerable strike-slip component. Şengör (1987) proposed that one of these major NE trending strike-slip cross-faults, located at the western boundary of the Kocaçay basin, offset the main Gediz detachment fault near its western end, and delineated it as a transfer fault zone. After these contributions, Okay and Siyako (1993) have suggested that this whole NE-SW trending zone was the...
Fig. 1. (a) Large scale tectonic setting of Anatolia (after Kaymakci et al., 2007; Kocyigit and Ozcag, 2003; Taymaz et al., 2007; Uzel et al., 2013); (b) Miocene paleomagnetic results in western and southwestern Anatolia from previous studies. Numbers refer to (1) Kissel et al. (1987), (2) Sen and Seyitoğlu (2009), (3) van Hinsbergen et al. (2010b), (4) Kaymakci et al. (2007), (5) Uzel et al. (2015), (6) Uzel et al. (2017), (7) Kaymakci et al. (2018). The main structures of western Anatolia are indicated as well. The Soma basin (study area of this research) is highlighted (details in Fig. 2). Abbreviations: GD = Gediz Detachment, AFZ = Akşehir fault zone, BMG = Büyük Menderes Graben, BMD = Büyük Menderes Detachment, CAZF = Central Anatolian fault zone, DB = Demirci Basin, EPF = Ezinepazarı Fault, GG = Gediz Graben, GB = Gordes Basin, IAS = İzmir–Ankara Suture, IBTZ = İzmir-Balikesir Transfer Zone, IEFZ = İnönü-Eskişehir fault zone, KMG = Küçük Menderes Graben, KP = Kozak Pluton, MCC = Menderes Core Complex, MCL = Mid-Cycladic Lineament, NAFZ = North Anatolian fault zone, SB = Selendi Basin, SD = Simav Detachment, SG = Simav Graben, TFZ = Thrace fault zone, TGZ = Tuz Gölü Fault Zone, UGB = Uşak-Güre Basin.

Fig. 2. Somabasin

Quaternary Kula basalts
middle-late Miocene volcanics
middle-late Miocene sediments
early Miocene volcanics
early Miocene sediments
Miocene granites

detachment fault

detachment fault
depositional loci of the Bornova Flysch Zone, a regional olistostrome–mélange belt, during the Late Cretaceous (Erdoğan, 1990; Okay et al., 2012; Sarı, 2012). Recent studies suggest that it has been reactivated as a NE-SW trending shear zone since the Neogene (Erkul et al., 2005; Eros et al., 2012; Morris and Anderson, 1996; Pe-Piper et al., 2002; Philippin et al., 2012, 2014; Ring et al., 1999; Uzel et al., 2013, 2015; Uzel and Sözbilir, 2008; Walcott and White, 1998; Sözbilir et al., 2003). Uzel et al. (2013, 2015) argued that transpressional deformation within this shear zone accommodates differential extensional strain between the MCC and MHC, whereas the Miocene, along the Mid-Cycladic Lineament (MCL). The Miocene volcano-sedimentary units and Quaternary continental units exposed in western Anatolia were all deposited and deformed within the IBTZ (Kaya, 1981; Sözbilir et al., 2011; Uzel and Sözbilir, 2008; Uzel et al., 2012, 2013). In addition, Uzel et al. (2015) and Brun et al. (2016) suggested that the IBTZ is a surface expression of a subduction transform edge propagator fault (STEP-fault; Govers and Wortel, 2005), related to a slab tearing of the Aegean slab due to rollback (Erkul et al., 2005, 2013; Uzel et al., 2015). There is a proposed connection between this slab rollback, the IBTZ, and the NAFZ (Gessner et al., 2013; Özokaymak et al., 2013; Uzel et al., 2013). Understanding the northern continuation of the IBTZ and its relationship with the NAFZ is required for a complete knowledge of western Anatolian tectonics, a seismically active region with frequent earthquakes.

In addition, there is another crucial uncertainty in constraining the tectonic history of western Anatolia. Several studies suggested that the Cenozoic extension took place in a single deformation phase (Gloody and Hetzel, 2007; Kaymakci, 2006; Seyitoğlu et al., 2000, 2002, 2004), while others proposed that it occurred during multiple deformation phases, separated by periods of inversion and tectonic quiescence (Beccalotto and Steiner, 2005; Bozkurt, 2001; Bozkurt and Sözbilir, 2004, 2006; Bozkurt and Wittewed, 2005; Emre and Sözbilir 2007; Kaya et al., 2004, 2007; Kaymakci 2006; Köyüşitik et al., 1999; Purvis and Robertson 2004, 2005; Sözbilir, 2001; Yilmaz et al., 2000). The role and evolution of the IBTZ within this tectonic setting were not well understood until recently, as paleomagnetic studies in the Aegean region were not focussed on the IBTZ. Only recently, Uzel et al. (2013, 2015, 2017) conducted structural and paleomagnetic research within the southern part of the IBTZ and adjacent regions (Fig. 1b). They argued that at least three strike-slip dominated deformation phases occurred throughout the IBTZ during the Neogene. Their paleomagnetic data constrained at least two distinct rotational phases separated by a middle Miocene angular unconformity (Uzel et al., 2015, 2017). Deformational and rotational trends within the IBTZ were different from the adjacent CCC and MCC regions. This indicates that the IBTZ evolved as a system of rigid-body rotations related to orthogonal extension (Jolivet and Brun, 2010; Uzel et al., 2015). However, these observations do not completely agree with the conclusions of van Hinsbergen et al. (2010b), who did not recognize the occurrence of rigid-block rotations within parts of the IBTZ and only distinguished one Miocene rotational phase in western Anatolia, related to asymmetrical exhumation of the MCC. Similarly, the data provided by Uzel et al. (2015) also contradict Kondopoulo et al. (2011), who argued that the coastal region of western Anatolia is characterized by a “chaotic pattern” of coexisting clockwise (CW) and counterclockwise (CCW) rotations.

In this context, the next logical step is to investigate the northern continuation of the IBTZ system, since this will provide crucial insight in understanding the connection between the IBTZ and the NAFZ. In this regard, new data will elucidate the relationship between the multiple-phase formation of the IBTZ of Uzel et al. (2013, 2015, 2017) and the single-phase asymmetrical exhumation of the MCC of van Hinsbergen et al. (2010b). Therefore, the goal of this research is to constrain the Neogene deformation history of the Soma basin, western Anatolia, and adjacent regions (Fig. 1). This area is located at a cross-point directly north of the area studied by Uzel et al. (2013, 2015, 2017), directly west of the area studied by van Hinsbergen et al. (2010b) and proximal to the southern strands of the NAFZ. In this contribution, we will present the results of newly acquired paleomagnetic data, as well as new structural and stratigraphic and anisotropy of magnetic susceptibility (AMS) results.

2. Geological setting

Western Anatolia is a region with a complex deformation history, where NNE trending extension and exhumation has taken place since the late Eocene (Şengör et al., 1985). The main geological features of the region are the MCC, the CCC, and the NE-SW trending IBTZ. The Soma basin is one of the basins located within the central part of the IBTZ.

2.1. Menderes core complex (MCC)

The geology of the region directly east of the IBTZ is characterized by the MCC and several E-W to NE-SW trending Neogene basins. The stratigraphy of the MCC consists of two sequences; a high-grade metamorphic core which consists of augen gneisses, metagranites, schists, paragneisses, and metagabbros, and a lower-grade metamorphic cover comprising schists, quartzites, amphibolites, phyllites, and marbles. Its metamorphism history can be traced back to late Proterozoic to early Paleozoic Pan-African events as well as to Mesozoic to Cenozoic Alpine orogenic events (Akkök, 1983; Bozkurt and Park, 1997; Candan et al., 1997, 2001; Okay, 2001).

The NE-SW trending basins in the northern part of the MCC are often called the Northern Menderes Massif basins. The largest of these are from west to east the Gördes, Demirci, Selendi, and Uşak-Güre basins (Şengör, 1987; Yilmaz et al., 2000). The basins are bounded on their northern and southern margins by E-W trending detachment faults as well as high angle normal faults. On their eastern and western margins, the basins are bounded by dextral or sinistral strike-slip faults, which were interpreted by Şengör (1987) as cross-faults related to the differential extension. Therefore, most of these basins have been interpreted as upper crustal supradetachment basins related to rapid, NNE-SSW trending extension of young, hot crust, resulting in domal uplift and exhumation of the crustal-scale MCC during the Oligocene-Miocene (Bozkurt, 2000; Bozkurt and Sözbilir, 2006; Çiftçi and Bozkurt, 2009; Köyüşitik et al., 1999; Sözbilir, 2001, 2002). According to van Hinsbergen et al. (2010b), this exhumation occurred asymmetrically with the Büyükb Menderes and Gediz detachments defining a pivot point, separating CW rotation in the north and CCW rotation in the south.

2.2. Cycladic core complex (CCC)

The CCC and related detachment faulting are less exposed in western Anatolia compared to the Menderes region. It is mainly exposed in the southern margin of the Kocaçay Basin (Sözbilir et al., 2011) in a crescent-shape belt that extends into adjacent Greek islands such as Samos and Ikaria. The main event that influenced the CCC was the Aegean subduction during the Eocene, which resulted in high-pressure metamorphism (Jolivet and Brun, 2010). In some places, high-temperature metamorphism superimposed high-pressure metamorphism (Brun et al., 2016; Philippin et al. 2012). The main rocks that comprise the CCC are mica- and calc-schists, marbles, meta-cherts, serpentinites, and meta-volcanic rocks in western Anatolia (Okay, 2001; Sözbilir et al., 2011). Like the MCC, the exhumation of the CCC took place during the Oligocene-Miocene (Vandenberg and Lister, 1996). It is cross-cut by basins with a progressively more continental sedimentary infill of similar age separated by large-scale detachments, comparable to the MCC (Brun et al., 2016). The CCC has experienced a greater degree of extension since the Eocene compared to the MCC (Gessner et al., 2013; Ring et al., 1999; Uzel et al., 2013).
2.3. İzmir-Balıkesir Transfer Zone (İBTZ)

The IBTZ was recognized in many studies as a NE-SW trending transtensional shear zone, accommodating differential extension between the MCC and CCC (Erkül et al., 2005; Ring et al., 1999; Uzel and Sözbilir, 2008; Sözbilir et al., 2011; Uzel et al., 2013, 2015). It is dominated by strike-slip deformation from the Miocene onwards, as evidenced by GPS data and earthquake focal mechanisms (Aktüğ and Kılıçoğlu, 2006; Inan et al., 2012; Uzel et al., 2013; Zhu et al., 2006). It has been suggested that the IBTZ is a surface expression of a slab-tear induced by the rollback of the Aegean slab (Uzel et al., 2015). This slab-tear forms a lateral boundary of the Hellenic trench system. Mantle windows often coincide with the location of the slab edge developed during the formation of a slab-tear (Biryol et al., 2011; Govers and Wortel, 2005; J. Westerweel et al., 2013; van Hinsbergen et al., 2010a), according to results from tomography (Biryol et al., 2011; Paul et al., 2014; van Hinsbergen et al., 2010a), surface observations (Gessner et al., 2013; Ring et al., 1999; Erkül et al., 2005, 2013; Uzel et al., 2013, 2015) and geochemistry (Aldanmaz et al., 2000; Altunkaynak et al., 2014; van Hinsbergen et al., 2010a; Ersoy et al., 2012a; Pe-Piper et al., 2002; Karacik et al., 2013). There is indeed evidence for such a mantle window, adding further credibility to the hypothesis that the IBTZ is a surface expression of a slab-tear. The formation of the IBTZ is also manifested by corresponding NE-aligned volcanism (Genc et al., 2001; Uzel and Sözbilir, 2008; Sözbilir et al., 2011). Previous studies constrained at least three deformation phases (Uzel et al., 2013) and two rotational phases (Uzel et al., 2015, 2017) within the IBTZ (Fig. 1b). The first deformation phase was dominated by NE-SW trending transtension during the early to late Miocene. This led to the formation of NE-SW trending basins within the zone. It has been suggested that these basins follow inherited structural trends (Kaya, 1981). This deformation phase was followed by Pliocene overall pure strike-slip deformation, which coincided with the final exhumation phase of the MCC, the formation of the NAFZ, and the cessation of displacement along the MCL. During the third deformation phase in the late Pliocene-Quaternary, the IBTZ evolved from a wide shear zone into a relatively narrow fault zone. During this phase, extensional and strike-slip deformation were completely decoupled from each other with NW–SE trending sinistral and NE–SW trending dextral strike-slip faults occurring alongside E-W trending normal faults. Extension was oriented NNW–SSE to NNE–SSW during this last phase within the IBTZ (Uzel et al., 2013), while the regional extension was N–S to NE-SW oriented since the Oligocene (Bözkurt, 2003; Bözkurt and Sözbilir, 2004; Lips et al., 2001; Seyitoğlu and Scott, 1996; Sözbilir, 2001).

According to paleomagnetic data by Uzel et al. (2015), the first rotational phase is expressed in early Miocene volcano-sedimentary rocks as an average net 23 ± 6° CW rotation. This was followed by a reorganization in the rotation pattern during the second phase, which is characterized by an average net rotation of 22 ± 11° CCW (Uzel et al., 2015). This suggests that the IBTZ region underwent a significant 45° CW rotation in the middle Miocene, followed by a 22° CCW rotation after the late Miocene. In addition, the narrowing of the IBTZ occurred progressively following the first rotational phase.

2.4. Soma basin

The Soma basin is one of the Neogene basins within the IBTZ, developed on the Eocene to Palaeozoic basement (Figs. 1 and 2). Because early studies on the geology of the basin were mainly focused on its abundant coal-bearing deposits, it was thought to be an intramontane basin, which developed in the topographic depressions related to Alpine deformation of the pre-Neogene basement (Incî, 1998, 2002). However, more recent studies suggest that faulting played an important role in the formation of the Soma basin, as it is bounded on all sides by high angle faults and horsts (Arpaliyigit, 2004). The most recent studies (Uzel and Sözbilir, 2008; Uzel et al., 2012, 2013; Uzel, 2017) provided evidence for the occurrence of both major strike-slip and normal faulting in the IBTZ basins, and therefore they interpret these basins as transtensional strike-slip basins. Overall, the deposits in the Soma basin can be divided into three stratigraphic sequences separated by regional unconformities. These sequences include, from older to younger, (i) pre-Neogene rocks, (ii) Neogene volcano-sedimentary units, and (iii) Plio-Quaternary units (Arpaliyigit, 2004; Incî, 2002; Kaya et al., 2004; Özkaynak et al., 2013). The focus of this research will be on the Miocene volcano-sedimentary units (Fig. 3) since they were deposited and deformed during the main tectonic events that formed the IBTZ.

2.4.1. Neogene volcano-sedimentary units

The base of the Miocene volcano-sedimentary sequence is characterized by alluvial fan conglomerates deposited along the basin margins, derived from the adjacent topographic highs that consist of basement lithologies (Incî, 2002). Uzel et al. (2017) concluded that an erosional period took place during the middle Miocene, and this hiatus is hence called the Middle Miocene Unconformity (MMU). It divides the Miocene volcano-sedimentary rocks into a lower and upper sequence (Fig. 3). In the Soma basin, the lower sequence is called the Soma Formation, while the upper sequence is named as DeniŞ Formation (Nebert, 1978; Incî, 2002). Our observations indicate that the Soma Formation consists mainly of limestones alternating with sandstones, siltstones, and marls. These deposits are intercalated with organic-rich, coal-bearing layers as well as pyroclastic, biotite- and plagioclase-rich tuff and tuffite deposits. Grey to white marl is the most abundant lithology in both the lower and the upper sequence in the basin. Occasional gastropod-rich marlstone deposits are present as well. Organic-rich layers contain fossil leaves, while common sedimentary structures are mud cracks and load casts. The dominant environment was fluviolacustrine throughout the deposition of the Soma and DeniŞ formations (Incî, 2002).

According to our field observations and revised stratigraphy of the area, these two formations are separated from each other by the Yunudağ volcanics (Fig. 3), which formed as a NE-SW trending volcanic ridge within the IBTZ during the early Miocene (21.0–15.9 Ma, Borsi et al., 1972; Ercan et al., 1996; Ersoy et al., 2012b; Uzel et al., 2020). We subdivided the volcanic rocks of the Soma basin into two categories: (i) pinkish to greyish, porphyritic, biotite- and plagioclase-rich andesites and rhyolites and (ii) grey to black, aphanitic olivine bearing basalts and trachybasalts, which sometimes form columnar joints. Both categories can contain empty vesicles, suggesting a subaerial environment during extrusion. The first extrusion identifiable as the Yunudağ volcanics cover the entire succession of the lower sequence (Fig. 3). In the second phase of volcanism, the Dededağ basalts (Fig. 3), was synchronous with the deposition of the upper sedimentary sequence. According to our field observations, basalt and basaltic andesite lava flows and their pyroclastic deposits are often intercalated with the DeniŞ Formation. Here, the contact between extrusion and sediment is characterized by baked contacts at the bottom of the lava flow, while no baking took place at the top, ruling out the possibility of a magmatic sill. This second phase of volcanism also covers the upper sequence successions.

2.4.2. Plio-Quaternary units

According to stratigraphic field observations, it appears that the Pliocene Kumköy Formation rests unconformably on the Soma and DeniŞ Formations (Fig. 3). The Kumköy Formation is hence considered as the sedimentary succession of post-Miocene sedimentation that corresponded to the non-volcanic edifice of the IBTZ in the Soma basin. This formation consists of well-sorted, surrounded to rounded ellipsoidal pebble conglomerates with imbrications, sandstones and mudstones with cross-bedding as well as soft-sediment deformation structures such as load casts and flame structures, and finally pisolithic-rich lacustrine carbonates. Collectively, these features indicate that the depositional environment of the Kumköy Formation was an association of
Fig. 2. (a) Geological map of Turkey with the study area of this research and paleomagnetic sampling locations indicated (after Inci, 1998; GDMRE, 2002; and this study). Triangles (squares) denote volcanic (sedimentary) sites, respectively; red stars indicate sites for paleostress measurements. Site abbreviations: AD = Arpadere, BG = Bağalan, BO = Beyoba, BY = Bayat, DK = Dereköy, EV = Evcler, GB = Göçbeyli, GL = Gilembe, HM = Hamidiye, IL = Ilyaslar, KD = Karadere, KG = Küçükçan, KN = Kınık, KP = Kapaklı, SV = Selvili. Structural abbreviations: AF = Akhisar Fault, BF = Bakur Fault, GFZ = Gilembe Fault Zone, KP = Kerkaç Fault, KFZ = Kaleköy Fault Zone. (b) rose diagrams prepared from the orientations of folds. Note the orientation of F1 and F2 are almost perpendicular to each other. F1 folds developed as buckle folds while F2 folds are developed as forced folds above normal faults.
Fig. 3. Integrated stratigraphic column of the Soma basin based on our observations and previous studies (Arpalyığiç, 2004; İnci, 2002). Rotation data of the IBTZ are from Uzel et al. (2015). The abbreviations of sampling localities are the same as in Fig. 2.
fluvial and lacustrine environments.

The most recent deposits of the Soma basin include Quaternary al-
luvial deposits developed along present-day streams. They rest un-
conformably on the older units (Fig. 3). Our observations show that
they consist mainly of typical alluvial fan deposits characterized by
reddish-brown colluvial fan-apron deposits along with slope breaks and
proximal matrix-supported cross-bedded conglomerates and sand-
stones. The rest of the Quaternary infill of the Soma basin consists of
cross-bedded channel conglomerates and sandstones, as well as mud-
stones. These are typical meandering river deposits, as indicated by
their morphology that includes point-bars, a high sinuosity of the river
channel, and epsilon cross-bedding in the coarse clastics wherever they
are exposed. These sediments contain clasts derived from both the pre-
Neogene basement and Neogene volcano-sedimentary units.

2.5. Structural geology

We have updated the 1/25000 scale geological map of the Soma
basin in the light of data from İnci (1998) and MTA (2002). The re-
sulting geological map of the study area and the revised stratigraphy
are shown in Figs. 2 and 3, respectively. In the sections below, the
newly acquired structural and kinematic data from the Soma basin is
presented according to the revised stratigraphy of the area (Figs. 4 and
5).

2.5.1. Faults

The faults observed within the Soma basin can be classified into
three general groups: NE-SW trending dextral strike-slip faults, NW-SE
trending sinistral strike-slip faults, and E-W trending normal faults.
These three fault sets are henceforth called D1, S1, and N1 (Fig. 5).

D1 faults are characterized by NE-SW striking fault planes with high
dip angles (> 60°) and well-constrained near-horizontal slickenside
pitches (Figs. 4a-b and 5). They generally displaced lower sequence
units and the basement. The same is true for the NW-SE trending S1
faults. However, S1 faults generally have a larger normal component, as
indicated by steeper slickenside pitches. They exhibit less constrained
slip directions compared to D1 (Figs. 4c and 5). N1 faults are oriented
approximately E-W and have higher dip angles and slickenside pitches
compared to D1 and S1 (Fig. 4g). The most notable exceptions on D1
faults are the Bakır and Kirkağaç faults: unlike other E-W trending
normal faults, they have N-S to NW-SE strikes and have normal fault
characteristics. Most of the N1 faults are cross-cutting the whole sedi-
mentary successions, including Quaternary alluvium, indicating recent
activity (Fig. 4f-i). In addition, the N1 faults cut and displace D1 and S1
faults (Fig. 2) except for the Gelembe Fault Zone (GFZ in Fig. 2), a
member of D1 group, which displaces all other structures and is still
active at present (Emre et al., 2016).

The slip directions and constructed paleostress con-
figurations in-
dicate that the D1 faults were developed under NNW-SSE directed ex-
tension and WSW-ESE directed compression (Fig. 4c), while inter-
mediate stress was subvertical (σ1: 258°N/15°, σ2: 041°N/72°, σ3:
165°N/10°). Subvertical intermediate and horizontal major and minor
stress orientations indicate that these faults were developed in a
transcurrent (strike-slip) tectonic setting along the IBTZ.

A reliable paleostress analysis could not be performed for the S1
faults because the number of fault slip data were not sufficient to

Fig. 4. Representative field photographs of deformational structures of D1 (a, b), S1 (c), F1 (d), F2 (e) and N1 (f-i).
constrain the tectonic setting. Nevertheless, the available data suggest tri-axial strain conditions for S1, as indicated by subhorizontal and subvertical stress orientations (Fig. 4f). Field observations and relative age relationships (cross-cutting with D1 faults) suggest that S1 faulting is most likely a conjugate fault system along the IBTZ.

The paleostress configurations constructed for the N1 fault set indicate that they were formed under approximately N-S extension. Computed kinematic data show that the principal stress axes are oriented as $\sigma_1$: 047°N/73°, $\sigma_2$: 277°N/11°, $\sigma_3$: 184°N/12° (Fig. 4i). As seen in Fig. 4h, the slickensides trend almost radially, indicating uniaxial stress conditions that develop when the magnitudes of two of the stress components ($\sigma_2$ and $\sigma_3$) are almost equal. The resultant geometry implies a multi-directional extension. If such conditions develop when $\sigma_1$ is vertical, such stress conditions tend not to be associated with relative block rotations.

2.5.2. Folds

Two distinct fold axis orientations are present in the Soma basin (Fig. 2). The first set of folds has an approximately NE-SW trend (Fig. 2b), indicating WNW-ESE directed shortening. This fold set is henceforth called F1, and it can be observed in the Neogene volcano-sedimentary units (Fig. 4d). These folds are occasionally cut and displaced by N1 faults across the study area (Fig. 2a).

The second fold set (F2) affected Neogene volcano-sedimentary units, but only within the hanging wall blocks of (N1) normal faults (Fig. 4e). These folds are generally open to gently plunging folds, and their axes are mostly parallel to the nearby E-W trending normal faults (Fig. 2b). These orientations suggest that the region deformed either in the NE-SW directed contraction (Bozkurt, 2000; Bozkurt and Sözbilir, 2004; Bozkurt and Rojay, 2005; Özkaymak et al., 2013), provided that they are buckle folds, or they are forced folds developed in response to normal faulting in the basement, which would indicate N-S extension. Similar to F1 folds, the orientations of the F2 folds are not compatible with the stress conditions that created D1 and S1 faulting. Their close proximity and parallel orientations to the N1 normal faults indicate that F2 folds are bending related forced folds (as opposed to buckle folds) developed in response to N-S directed extension, which also controlled normal faulting. Such a mechanism is already proposed for some of the folds in the Gediz graben (Çiftçi and Bozkurt, 2009; Seyitoglu et al., 2000; Sözbilir, 2001, 2002) and the Denizli basin (Kaymakci, 2006). According to field observations and geological mapping, the F2 folding was probably formed during or after the late Miocene and is related to the high angle normal (N1) faulting across the Soma basin, while F1 folds developed due to WNW-ESE directed $\sigma_1$ since the late Miocene.

3. Methods

3.1. Structural analysis

Field-based structural mapping and kinematic analysis were conducted to investigate the nature and order of deformation that occurred in the Soma basin. The orientations of fault planes and fold axes, as well as cross-cutting relationships, provide constraints on different deformation phases. Kinematic indicators, such as slip lineations (slicken-olths, slickensides, slickenfibres, etc.), stratigraphical offset features, and Riedel shear geometries, are used to identify shear sense. The collected fault slip data are used to reconstruct paleostress configurations for each deformation phase. The computations were performed using Win_TENSOR (Delvaux and Sperner, 2003) software. The ranking scheme from Delvaux and Sperner (2003) was applied for checking the quality of the fault slip data. Finally, field observations and geological cross-sections were utilized to determine the relative ages of different structures and lithologies.

3.2. Paleomagnetism

Paleomagnetic analysis of rocks is an effective method for determining the deformation history of strike-slip fault zones with prevailing simple shear conditions, like the IBTZ, because it can constrain vertical-axis rotations with respect to the present-day geographic north (Christie-Blick, 1985; Tauxe, 2010). In addition, analysis of the AMS in sediments provides constraints on paleostress directions for comparison with the structural data (Hrouda, 1982; Tauxe, 2010).

In total, we distinguished 25 early Miocene volcanic sites distributed over four localities: Arpadere (AD), Göçbeyli (GB), and Karadere (KD); and 15 middle-late Miocene volcanic sites from the localities of Başalan (BG), Gelembe (GL) and Bayat (BY). Furthermore, we distinguished 24 early Miocene sedimentary sites across seven localities: Selvili (SV), Haylasar (HL), Knuk (KN), Evcler (EV), Dereköy (DK), Beyoba (BO) and Kapaklı (KP); and 17 middle-late Miocene sedimentary sites from the localities of Hamidiye (HM) and Küçüküney (KG). The distribution of all localities is shown in the geological map of Fig. 2. From these localities, 664 conventional paleomagnetic core plug samples (Ø 25 mm) were collected in stratigraphic order using a gasoline-powered drilling machine, consisting of 338 volcanic and 326 sedimentary samples (Fig. 6).

Volcanic rocks cool relatively fast and therefore retain spot readings of the geomagnetic field, while sedimentary rocks average out paleo-secular variation due to relatively slow sedimentation rates. The magnetic signal of sedimentary rocks is overall weaker compared to volcanic rocks. The availability of suitable outcrops determines the sampling distribution for paleomagnetic study. As a result, the sizes and
boundaries of tectonic blocks are sometimes poorly constrained, which is further complicated by the occurrence of several deformation phases in the study area. For these reasons, several sites from the same volcanic locality were sampled. In addition, all sampled localities are distributed as equally as possible over the study area and over different lithologies. The exception to this is the Soma open coal pit mine, which contained no suitable outcrops due to intense deformation. For measurement purposes, all samples were oriented using a magnetic compass and, if possible, a sun compass for volcanic rocks. Furthermore, the bedding orientation was measured at every site to check whether a geographic or tectonic coordinate system yields more consistent directions. For the volcanic rocks, bedding planes are checked with the closest

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**Fig. 6.** Geological cross-sections of representative paleomagnetic sampling sites from localities Bağalan (BG), Hamidiye (HM) and Küçükgüney (KG), including lithologies, sampling distribution and field pictures.
sedimentary bedding for the reliability of their paleo-horizontal position whenever possible, although these paleo-horizontal measurements remain prone to errors related to unclear bedding surfaces or the effect of paleo-topography causing non-horizontal deposition of lava flows. Core orientations and bedding strikes were corrected for the declination of the present-day field of ~5° (Thébault et al., 2015). In the laboratory, the cores were cut into 22 mm specimens for sedimentary cores and 11 mm for volcanic cores, because of their overall stronger magnetic signal.

3.2.1. Thermal variation of magnetic susceptibility

The dominant magnetic carriers and chemical alterations of a selection of samples from different localities were determined by measuring their mass-normalized bulk magnetic susceptibility at increasing temperature steps. This was done using an AGICO KLY-3 (noise level $3.2 \times 10^{-13}$ A m$^2$). The samples were powdered, weighted, and put into a

Fig. 6. (continued)
quartz-glass sample holder. The mass-normalized bulk magnetic susceptibility was measured during several heating and cooling cycles at steps of 60 °C, starting at 180 °C up to a maximum of 580 °C.

3.2.2. Anisotropy of magnetic susceptibility (AMS)

The AMS of a selection of sedimentary samples from all localities was measured to determine whether their magnetic fabrics have a mainly sedimentary or tectonic origin. In both cases, initial depositional compaction will impart an oblate magnetic fabric with the minimum axis $K_3$ perpendicular to the bedding due to compaction. In the case of a purely sedimentary fabric, the $K_1$ and $K_2$ axes are indistinguishable. Upon deformation, the maximum axis $K_1$ of the AMS tensor will gradually be directed parallel to the orientation of maximum extension or equivalently perpendicular to the orientation of maximum contraction resulting in distinctly separated $K_1$ and $K_2$ distributions in a tectonic fabric (Hrouda, 1982). In such cases, the geological context is crucial for determining whether the results indicate a pure extensional or compressional setting. When the statistical measurement errors are too high, all three axes will be indistinguishable; such results need to be discarded from further analysis and interpretation.

In this study, the AMS tensor was measured and calculated using an AGICO Kappabridge MFK1-FA (noise level $2.1 \times 10^{-13}$ A m$^2$). Jelinek statistics (Jelinek, 1978, 1981) were used for the calculations of the AMS tensor. The resulting data were viewed and interpreted using AGICO Anisoft 4.2.

3.2.3. Vertical-axis rotations

Our samples were demagnetized stepwise in order to obtain their characteristic remanent magnetization (ChRM) vectors. For most samples, this was done using thermal demagnetization. Volcanic samples were heated stepwise with increasing temperature increments of 20–50 °C in an oven until they reached a temperature of 580 °C, while most sedimentary samples were heated until a temperature of about 420 °C. At these maximum temperatures, the samples were sufficiently demagnetized to determine the ChRM, while thermal alteration is minimized. In addition, some samples were demagnetized using...
alternating field (AF) demagnetization. These samples were placed manually in a Helmholtz coil to provide a non-magnetic environment. Stepwise increasing alternating fields of 10 mT were applied in three directions for all samples until a field of 60 mT was reached. After each demagnetization step, the natural remanence magnetization (NRM) of the sample was measured on a 2G Enterprise horizontal cryogenic magnetometer equipped with three DC SQUIDs (noise level 3.0 × 10^{-12} Am^2) or an AGICO JR6 spinner magnetometer (noise level 2.5 × 10^{-11} Am^2).

For all paleomagnetic interpretations and subsequent statistics, the open-source and platform-independent portal “Paleomagnetism.org” was used (Koymans et al., 2016). The demagnetizations of all measured samples were plotted in orthogonal demagnetization plots (Zijderveld, 1967) for interpretation. Characteristic components were derived from this using principal component analysis (Kirschvink, 1980) or the great circle best-fit analysis (McFadden and McElhinny, 1988), while standard Fisher statistics (Fisher 1953) were used to calculate the means and corresponding dispersions (k, K) and cones of confidence (a95, A95) of the ChRM and virtual geomagnetic pole (VGP) distributions. Subsequently, errors in declination (ΔD) and inclination (ΔI) were calculated after Butler (1992). A 45° cut-off is applied to the ChRM/VGP distribution to determine which sites were used for constraining locality mean directions. Samples from tilted beds were corrected for their paleoenvironment. The formation of magnetite does not disturb the determination of the ChRM/ VGP from Deenen et al. (2011, 2014) was used to determine whether a certain distribution represents paleosecular variation (PSV) or a spot reading. In the latter case, A95 < A95_max. In case A95 > A95_max additional sources of scatter are present, such as small-scale rotational differences, chemical alteration, or measurement errors. The cartesian coordinate bootstrap test from Tauxe (2010) was used for determining whether two distributions from the same locality share a common true mean direction (CTMD). The fold test of Tauxe and Watson (1994) was utilized to check whether the magnetizations in a certain locality were acquired before or after tilting. Both tests were applied whenever possible.

4. Results
4.1. Paleomagnetism
4.1.1. Bulk magnetic susceptibility
The mass-normalized bulk magnetic susceptibility for four representative volcanic samples and two sedimentary samples from different localities is shown in Fig. 7 after seven consecutive heating and cooling cycles. Plots of intensity decay during thermal demagnetization are shown as well for comparison.

Andesitic sample AD06.3 shows a gradually increasing susceptibility up to 360 °C with only minor susceptibility changes during the first five heating cycles. These minor changes are likely related to chemical alteration due to oxidation. Between 360 °C and 420 °C, the susceptibility reaches a maximum. This maximum can be interpreted as a Hopkinson peak, which marks the transition from blocking temperatures to Curie temperatures for magnetic minerals in the sample (Hopkinson, 1989). The sharp susceptibility drop associated with the Curie temperature is observed at higher temperatures, and the susceptibility reaches a minimum at 580 °C, indicating that magnetite (Fe3O4) is the dominant magnetic carrier. However, the inflection of the curve towards a slighter decrease in susceptibility around 550 °C in-and K2 measurements form well-defined clusters (Fig. 8a and 8b). All three axes have a similar general orientation in both localities,

4.1.2. Anisotropy of magnetic susceptibility (AMS)
The AMS of selected samples from all sedimentary localities was measured. The measurement errors in samples from most localities were too high, resulting in undistinguishable fabrics for those localities. The large errors may be attributed to low overall intensities of the samples. Therefore, these localities were discarded. Only three localities (Evcler, Kapakli, and Selvili) yielded results with low measurement errors, with a largely sedimentary fabric after bedding correction. The resulting AMS ellipsoids of these three localities are shown in Fig. 8, alongside their inferred extension directions. In all three cases, K1 is oriented almost vertically, perpendicular to the bedding plane within its error margin. Furthermore, the K1 and K2 axes form a girdle perpendicular to the K3-axis, resulting in an oblate AMS ellipsoid, as can also be seen from the shape parameter T (Fig. 8f).

AMS results of Evcler and Selvili show a tectonic fabric where K1 and K3 measurements form well-defined clusters (Fig. 8a and 8b). All three axes have a similar general orientation in both localities,
indicating approximately N-S extension parallel to \( K_1 \) or, equivalently, perpendicular E-W trending contraction parallel to \( K_2 \). The same orientations can be inferred from Kapaklı (Fig. 8c) even though the AMS axes show relatively poor clustering, although they still agree with the AMS results from Evciler and Selvili. Therefore, all results together (Fig. 8d) consistently indicate an approximately N-S extension and/or E-W contraction in the Soma basin during the deposition of the upper sequence. When comparing these results with the kinematic analyses (Fig. 5) and mapped structures (Figs. 2 and 4), we conclude that approximately N-S oriented extension took place along the İBTZ during the early Miocene.

4.1.3. Vertical-axis rotations

In total, 432 samples were demagnetized to obtain their ChRM directions. In many samples, a small viscous component is removed during the first few demagnetization steps. The mean paleomagnetic data for every measured site is listed per locality in Supplementary Table S1 in both a geographic and tectonic reference frame. The average declination \( D \) and its error \( \Delta D_x \) from each locality were used to describe the vertical-axis rotations. Fig. 9 shows representative orthogonal demagnetization plots (Zijderveld, 1967) for both volcanic (Fig. 9a) and sedimentary samples (Fig. 9b). Example plots of white limestones from Beyoba (BO) and Kapaklı (KP) are indicated as well.
Unfortunately, these two lower sequence localities gave no results due to the low intensities of the samples; they were subsequently discarded. Magnetite is the dominant magnetic carrier in the majority of samples based on unblocking temperatures, but the demagnetization of sample GL03.2 suggests that a minority of samples could contain additional small occurrences of higher-coercivity components, such as hematite. Fig. 10 shows the equal-area projections of the ChRM directions and their means for all localities. We divide the localities into three separate blocks, according to their age and position in the basin (Fig. 2).

**Kink block**: Göçbeyli (GB), Arpadere (AD), Karadere (KD), Kink (KN) and Dereköy (DK) are all sampled localities from the south-western quadrant of the study area (Fig. 2) and together constitute the Kink block (Fig. 10a). All localities are from the lower sequence of early Miocene age. The Göçbeyli location consists of nine volcanic sites (lava flows) from the lower sequence. Sites GB05 and GB06 both produced chaotic directions with low k values (4.2 and 5.8, respectively). Additionally, these sites were discarded by the 45° cut-off. The remaining sites produced a mean with a very shallow inclination (−7°) after tectonic correction. Therefore, the geographic reference frame gives a more consistent result for the Göçbeyli location (Table 1). Arpadere is part of the Yuntdağ volcanics (Fig. 3), and seven (andesitic) extrusions were sampled at this locality. It exhibits a well-determined CW rotation of 24 ± 10° in a geographic reference frame (Table 1). Arpadere is part of the Yuntdağ volcanics (Fig. 3), and seven (andesitic) extrusions were sampled at this locality. It exhibits a well-determined CW rotation of 24 ± 10° in a geographic reference frame (Table 1). Arpadere is part of the Yuntdağ volcanics (Fig. 3), and seven (andesitic) extrusions were sampled at this locality. It exhibits a well-determined CW rotation of 24 ± 10° in a geographic reference frame (Table 1). Arpadere is part of the Yuntdag volcanics (Fig. 3), and seven (andesitic) extrusions were sampled at this locality. It exhibits a well-determined CW rotation of 24 ± 10° in a geographic reference frame (Table 1).
Fig. 9. Representative volcanic (a) and sedimentary (b) orthogonal vector diagrams (Zijderveld, 1967) from different lithologies in the study area. Open/closed circles indicate projections on the vertical/horizontal plane; tectonic (tilt corrected, tc) or geographic (no tc) reference frame is indicated, as well as thermal demagnetization steps. Vertical/horizontal projections of the characteristic components are shown in red/green. Some examples from rejected sites are shown (c) and two thermal demagnetization intensity decay diagrams (d).
are two sedimentary localities in the south-western quadrant of the study area (Fig. 2). KN01 is the only used site from the Kınık locality; the other three sites from this locality have very scattered directions (A95 > A95max) and are therefore not used for the final result (Table S1). The average ΔDx and A95 values from KN01 become slightly lower after tectonic correction. In addition, the tectonic correction resulted in a more consistent fit with the results from the previous three localities, producing a CW rotation of 51 ± 20° (Fig. 10a, Table 1). Dereköy produces a consistent CW rotation of 42 ± 9° after tectonic correction (Fig. 10a, Table 1). We applied the fold test on all localities from the Kınık block, but considering the large 95% confidence interval [3,94%] we consider it as indeterminate and hence inconclusive (Fig. S1a). Because the mean results from the Kınık block localities are similar and consistently do not plot in the present-day field even in a geographic reference frame, the fold test results are probably related to shortcomings of the fold test itself taking not into account the effect of non-horizontal deposition (paleo-topography) or noncoaxial differential tilting of the bedding planes alongside errors in bedding measurements of the thick volcanic flows making up the bulk of reliable results from the Kınık block (AD, KD, GC) rather than to post-folding magnetization. The final mean direction from the early Miocene sites of the Kınık block gives a CW rotation of 33 ± 7° (Fig. 10c, Table 1).

**Bakır block:** Selvili (SV), Ilyaslar (IL), and Gelembe (GL) are the three sampled localities in the south-eastern quadrant of the study area, separated from the south-western localities (Fig. 2), and together make up the Bakır block (Fig. 10b). Ilyaslar consists of lower sequence fluvio-lacustrine and pyroclastic deposits. Selvili is similar to Ilyaslar in terms of stratigraphic position (Figs. 2 and 3). It consists of two sites (SV1 and SV2), which form a NE-trending anticline. Before tilt correction, the two sites give more consistent directions implying that the magnetization has been acquired after tilting. Unfortunately, the results are too scattered to produce a reliable fold test for this locality, and many samples are rejected by the 45° cut-off (Fig. 10b, Table S1). Applying tilt correction results in a shallow inclination (25°) and a very low k value (7.6), indicating a high amount of scatter. In addition, the resulting rotation in geographic coordinates is consistent with Ilyaslar.
Table 1

Mean paleomagnetic results from this study for the Bakır and Kınık blocks, and the combination of the two blocks forming the main results for the Soma basin. The results of locality EV and the late Miocene localities in the northern Soma basin are listed as well. All locality means are displayed in Fig. 10, while detailed paleomagnetic results are listed in Table S1. N = number of samples/sites after a fixed cut-off (45°), out of a total of Ns samples; D, I = mean declination, inclination; k, α95 / K, A95 = dispersion and 95% cone of confidence of the directional distribution / of the VGP distribution; ΔD, ΔI = the error in declination, inclination based on A95 of the VGP distribution (Butler, 1992); A95min, A95max refers to the confidence envelope of Deenen et al. (2011, 2014) for sampling paleoaxial variation.

<table>
<thead>
<tr>
<th>Locality / Site</th>
<th>Formation</th>
<th>Lithology</th>
<th>N</th>
<th>Ns</th>
<th>D</th>
<th>ΔD</th>
<th>I</th>
<th>ΔI</th>
<th>k</th>
<th>α95</th>
<th>K</th>
<th>A95min</th>
<th>A95max</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kınık (K)</td>
<td>Soma Formation</td>
<td>Sandstone, limestone, tuff</td>
<td>7</td>
<td>6</td>
<td>41.9</td>
<td>9.3</td>
<td>52.9</td>
<td>8.4</td>
<td>41.9</td>
<td>7.4</td>
<td>39.4</td>
<td>4.8</td>
<td>7.8</td>
</tr>
</tbody>
</table>
| 29 ± 12° CCW as the youngest rotational phase. This resulted in a CCW rotation of −29 ± 15° in tectonic coordinates (Fig. 10d, Table S1). If we combine the results of all middle-late Miocene sites, we find a rotation of −21 ± 12° CCW as the youngest rotational phase.

5. Discussion

5.1. Spatio-temporal relationships

The IBTZ is interpreted as a large-scale NE-SW trending strike-slip shear zone in an extensional setting accommodating differential strain between the Menderes and Cycladic core complexes (Sözbilir et al., 2003; 2011; Uzel and Sözbilir, 2008; Uzel et al., 2013, 2015). The observed styles of structures within the IBTZ display a Riedel deformation pattern in map view (Fig. 11a and b). Therefore D1 dextral
strike-slip faults in the Soma basin correspond to the synthetic Riedel (R-) shears, while S1 sinistral strike-slip faults are the antithetic Riedel (R’-) shears based on their general orientations and field relationships. Both fault sets have normal oblique components, which is a common feature in strike-slip shear zones. This implies that both fault sets were developed simultaneously under NNW-SSE directed extension and related WSW-ENE trending compression (Figs. 5 and 10). The AMS results agree with this general configuration of the paleostress orientations, where the K1 axis is parallel to $\sigma_3$, and the K2 axis is parallel to the $\sigma_1$ direction (Fig. 11c).

We categorized the development of D1 and S1 faults as deformation Phase 1 and propose that it represents a transtensional deformation phase related to the initiation of the IBTZ as a wide dextral shear zone during the Miocene (Fig. 11a-d). This deformation style fits with other studies, such as NNW-SSE directed extension and NE-SW striking dextral strike-slip faulting in the southern part of the IBTZ around İzmir (Kaya, 1979; Erkül et al., 2005; Sözbilir et al., 2011; Uzel et al., 2013).

The E-W trending normal faults from fault set N1 cut and displace Phase 1 structures (Fig. 2), and also deformed the Pliocene Kumköy Formation. Most of the N1 normal faults have oblique-slip components, and they accommodate the development of folds. On the other hand, the NE-SW striking Gelembe Fault Zone, which is still active at present (Emre et al., 2013), is a dextral strike-slip fault (D1) that is contemporaneous with the activities of N1 faults. This suggests that these structures were developed under a new deformation regime, Phase 2, characterized by approximately N-S directed extension and E-W directed $\sigma_2$ related compression, while $\sigma_1$ is subvertical.

The F1 folds are not completely parallel to the expected folding direction of deformation phase 1 (compare “F” and “F1” in Fig. 11). These folds are mainly perpendicular to the $\sigma_2$ directions and parallel to $\sigma_3$ directions of deformation phase 2, implying that they are buckle folds developed in response to approximately WNW-ESE directed compression together with N1 faults (Fig. 11e). On the other hand, the E-W trending F2 folds are developed almost parallel to the nearby normal faults, and they are located mainly on the downthrown blocks of the normal faults (Fig. 11e-f). This relationship implies that these
folds are forced folds developed due to the bending of the cover rocks above normal fault blocks. Unlike F1 buckle folds – with fold axes perpendicular to the compression direction and parallel to the local extension direction (Fig. 11f) – these forced folds develop in extensional settings (cf. Janecke et al., 1998) and their axes are perpendicular to the local extension direction. Çiçti and Bozkurt (2009), Sözbilir (2002), and Uzel et al. (2013) found evidence for extensional folding along the Gediz Graben. The middle Miocene and younger rocks of the Soma basin are affected by F2 folding. Therefore, we propose that this folding must belong to Phase 2, contemporaneous with N1 normal faults. Although both deformation Phases 1 and 2 are of transtensional nature, field and age relationships show that extensional deformation became more dominant during the second phase and is currently still active.

5.2. Rotational phases

The paleomagnetic results of this study acquired in the early Miocene localities show an average net rotation of 33 ± 7° CW for the Kınık block and 26.5 ± 6° CW for the Bakır block. The rotation of the Bakır block is slightly less (~6°) than that of the Kınık block. The directions of the two blocks do not share a CTMD according to the coordinate bootstrap test, but their errors in declination overlap. After combining the results of the two blocks, we arrive at an average rotation of 29 ± 5° for the south-central Soma basin (Table 1). It seems that the ChRM of both blocks have possibly been acquired during strike-slip faulting along the IBTZ, and the difference in rotation (~6°) between them may reflect a differential CW rotation through time. This would also provide an additional explanation for the inconclusive and negative paleomagnetic results from the Kınık and Bakır blocks, alongside the potential influence of paleo-topography and uncertainties in tilt correction on the fold tests, as discussed above (Section 4.1.3), implying that magnetic acquisition was contemporaneous with folding in an actively deforming region. In any case, the results imply that the first rotational stage of the Soma basin took place within a dextral shear zone, which is also evidenced by the coeval development of the Riedel deformation pattern within the IBTZ during the middle Miocene (Phase 1, Fig. 11). Furthermore, the paleomagnetic results from this study are in good agreement with previous paleomagnetic results obtained from the IBTZ. Uzel et al. (2015) reported an average net CW rotation of 23 ± 6° from the Yünstağı region, while the Söke basin in the southern part of the IBTZ also has a rotation of 23 ± 6° CW (Uzel et al., 2017). These rotations are in good agreement with the rotation of the Bakır block in this study and have the same sense (CW) of rotation as those from the Kınık block, although the latter shows a significant larger CW rotation by ~1°, which may be an augmented rotation caused by local rotation of smaller blocks within the large shear zone. Indeed, the rotations within the Kınık block vary between 16° and 51° (Table 1), which indicates internal deformation of the block. We note that the rotations within the Yünstağı region vary as well, between 16° and 33° (Uzel et al., 2015).

The second rotational phase in the IBTZ is shown by a −22 ± 11° CCW rotation from middle-late Miocene rocks (Fig. 12), which took place since the late Miocene (Uzel et al., 2015). In this study, we find a −21 ± 12° CCW rotation from the late Miocene rocks in the Soma basin, identical to the IBTZ rotations (Uzel et al., 2015). Therefore, we suggest the early Miocene rocks of Soma basin experienced a significant CW rotation (~50° in total) in the middle Miocene, which is almost similar in sense and magnitude (~45° CW) as found in the central sector of IBTZ (Uzel et al., 2015). Hence, we conclude that the Soma basin has experienced the same rotational history as the other Miocene basins along the IBTZ, while it was different from the Miocene basins on top of the MCC (van Hinsbergen et al., 2010b; Uzel et al., 2015; Kaymakçı et al., 2018).
subducting African oceanic slab below the western Anatolian - Aegean region (overriding plate) and subsequent acceleration of roll-back. The response of the overriding plate was segmentation into small structural blocks with differential rotation, delimited by approximately NE-SW strike-slip and E-W normal faults. The results also imply a possible link between the İBTZ and North Anatolian Fault Zone (NAFZ), at least since the late Miocene, inviting further investigation of the region.

CRediT authorship contribution statement

Jan Westerweel: Investigation, Data curation, Formal analysis, Writing - original draft. Bora Uzel: Investigation, Conceptualization, Methodology, Writing - review & editing. Cornelis G. Langereis: Investigation, Supervision, Methodology, Software. Nuretdin Kaymakci: Investigation, Supervision, Conceptualization, Writing - review & editing. Hasan Sözbilir: Supervision, Validation.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Fig. 12. Schematic representation of western Anatolian rotational phases since the Miocene (after Özkaymak et al., 2013; Uzel et al., 2013 and references therein). (a) Rotation of different tectonic blocks are drawn using the results from van Hinsbergen et al. (2010b) for the Northern Menderes Massif (NMM) basins, from Uzel et al. (2015) for the İBTZ and the Menderes massif, and the results of this study (Kinik, Bakır and late Miocene); pink (blue) shading refers to early (late) Miocene paleomagnetic results. Note that the Gediz Graben separates domains with CW rotation in the northern Menderes region from CCW rotation in the southern Menderes region. Paleostress directions $\sigma_1$ and $\sigma_3$ are indicated with red and blue arrows, respectively. Black lines with boxes (on the hanging-wall block) are normal faults. The recently active faults are thicker. (b) Simplified tectonic block model of the first deformational phase and its rotation senses.
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Appendix A. Supplementary material

Supplementary data to this article can be found online at https://doi.org/10.1016/j.jseaes.2020.104305.

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