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**Propagation of a strike-slip plate boundary
within an extensional environment :
the westward propagation of the North Anatolian Fault.**

***Xavier le Pichon¹, A. M. Celâl Şengör^{2,3}, Julia Kende⁴, Caner İmren⁵,
Pierre Henry⁴, Céline Grall⁶ and Hayrullah Karabulut⁷***

¹ Professeur Honoraire au Collège de France, Europôle de l'Arbois, Bât. Trocadéro, BP 80
13 545 Aix-en-Provence cedex 04 FRANCE

² İstanbul Teknik Üniversitesi, Maden Fakültesi, Jeoloji Bölümü, Ayazağa 34469 İstanbul
TURKEY

³ İstanbul Teknik Üniversitesi, Avrasya Yerbilimleri Enstitüsü, Ayazağa 34469 İstanbul,
TURKEY

⁴ CEREGE, CNRS-Aix Marseille Université, Marseille, FRANCE

⁵ İstanbul Teknik Üniversitesi, Maden Fakültesi, Jeofizik Bölümü, Ayazağa 34469
İstanbul, TURKEY

⁶ Lamont - Doherty Earth Observatory Columbia University, Marine Geology and
Geophysics, New York City, UNITED STATES

⁷ Kandilli Rasathanesi, Boğaziçi Üniversitesi, Kandilli, İstanbul, TURKEY

Abstract

The Sea of Marmara marks a key point in the propagation of the North Anatolian Fault (NAF) toward the northern extremity of the Aegean subduction during the last 12-11 Myr. There is no indication that a localized plate boundary existed to the west of it, north of the Aegean portion of the Anatolia plate, before 2 Ma. Prior to 2 Ma, the shear produced by the motion of Anatolia-Aegea with respect to Eurasia was distributed over the whole width of the Aegean-West Anatolian western portion. This was most probably related to the presence there of a N/S component of extension produced by the southward migration of the Aegean-West Anatolian subduction. Fast subduction of the oceanic Ionian lithosphere had been initiated 15 Ma ago, a few million years before the Anatolia westward motion began 11 Ma ago. These two processes are obviously tightly linked. We document the establishment of the Aegea-Anatolia/Eurasia plate boundary in Plio-Pleistocene time in this geodynamic context.

We show that the beginning of the formation of a localized plate boundary occurred between 4.5 and 3.5 Ma ago at the location of the present Sea of Marmara by the initiation of a shear zone comparable to the Gulf of Corinth one in Central Greece. Thus the first part of the formation of the Sea of Marmara was purely extensional. We discuss the transition from the initial extensional basins to the present strike-slip system, that today cuts across the whole Sea of Marmara and that is called Main Marmara Fault (MMF). We show that the beginning of its development is not earlier than 2.5 Ma ago. Shortly after, the plate boundary migrated west of the Sea of Marmara along the northern border of Aegea from the North Aegean Trough, to the Gulf of Corinth area and

47 to the Kefalonia fault. There, it finally linked with the northern tip of the Aegean
48 subduction zone, completing the system of plate boundaries delimiting the Anatolia-
49 Aegea plate. We have related the remarkable change in the distribution of shear over the
50 whole Aegea from Miocene to Pliocene to the formation of a relatively undeforming
51 block in Pliocene that forced the shear to be distributed over a narrow plate boundary to
52 the north of it. We attribute the formation of this relatively undeforming Aegean block to
53 the northeastward progression of the cold oceanic Ionian slab. We propose that the slab
54 cuts the overlying lithosphere from asthenospheric sources and induces a shortening
55 environment over it.

56

57 ***Keywords***

58 North Anatolian Fault, Propagation of a fault, Subduction, Extension, Aegea, Anatolia

59

60 ***Introduction***

61

62 McKenzie (1970, 1972) first described the westward extrusion of Anatolia-Aegea in
63 terms of plate tectonics, following Ketin's pre-plate tectonics suggestion in 1948. GPS
64 results have confirmed that one could describe the Anatolian extrusion to a first
65 approximation in terms of rigid rotations (Le Pichon et al. 1993, 1995; McClusky et al.
66 2000, 2003; Reilinger et al. 1997; Reilinger et al. 2006; Le Pichon and Kreemer 2010).
67 The northern boundary of this plate separates the Anatolia-Aegea plate to the south
68 from the Eurasia plate to the north and is presently characterized along its whole 2100
69 km length by dextral strike-slip at a rate of about 24 mm/yr. To the east, the portion of
70 the boundary north of Anatolia corresponds to the 1200 km North Anatolian Fault
71 (NAF) that propagated westward since 13-11 Ma (Şengör et al. 2005). The NAF reached
72 the Gulf of İzmit in the Sea of Marmara about 4 Ma ago (Le Pichon et al. 2014) although
73 shear related deformation in the broader Marmara region had already commenced in
74 the late Miocene (Şengör et al. 2005). (See figure 1 for the tectonic map of Aegea-
75 Western Anatolia in figure 1 and figure 2.5 as a key to geographic and geologic features
76 in the Marmara area.) To the west, (see fig.1), the additional 900 km of plate boundary
77 are complex (Şengör 1979; Şengör et al. 1982; Goldsworthy et al. 2002; Armijo et al.
78 2002; Le Pichon and Kreemer 2010; Shaw and Jackson 2010; Floyd et al. 2010; Royden
79 and Papanikolaou 2011; Pérouse et al. 2012). The plate boundary follows successively
80 from east to west the NAF in the Sea of Marmara, called Main Marmara Fault by Le
81 Pichon et al. (2001), the Ganos fault, the North Aegean Trough (NAT) formed by the
82 Saros Basin to the east and the Sporades Basin to the west, followed by a broad zone of
83 shear called the Grecian Shear Zone (GSZ) by Şengör (1979) to finally join through the

dextral Kefalonia Fault the Aegean subduction zone, where the 200 Ma Ionian oceanic slab is subducted at a rate of 34 mm/yr (Pérouse et al. 2012).

The central theme of this paper concerns the mode of establishment of such a strike-slip plate boundary since 13-11 Ma when the Aegea-Anatolia plate was first initiated. We attempt to understand better why a localized pure strike-slip fault has progressively propagated westward as emphasized by Şengör et al. (2005). An essential character of both Anatolia and Aegea is that their geological history suggests very thin or even absent lithospheric mantle (e.g. Şengör 1982). Thus the Aegea-Anatolia plate corresponds to a huge ductile mass. Further the western Aegean portion is subject to N/S extension because of the roll-back of the peri-Aegean subduction (Le Pichon and Angelier 1981; Le Pichon 1982). The ductility of this lithospheric mass and the presence of an important N/S extensional component on its western Aegean portion will occupy a prominent place in our investigation of the formation of this plate boundary. The transition from the well-defined NAF to the east to the more complex Aegean boundary to the west occurs within the Sea of Marmara. We will devote the principal part of this paper to a discussion of the genesis of the Sea of Marmara, as it enlightens the mode of propagation of the NAF through it. We will use new data and interpretations that have been partly published by Grall et al. (2012; 2013), Le Pichon et al. (2014) and Şengör et al. (2014) or are still unpublished (Kende, in preparation).

The present tectonic regime of the North Anatolian Fault and its mechanical behavior

The portion of the NAF east of the Sea of Marmara

We first examine the behavior of the NAF east of the Sea of Marmara. The NAF there, throughout its length, is a stable localized pure dextral strike-slip fault with only local pull-apart and push-up segments (Şengör et al. 2005). The rate of geological motion has been established at about 18 mm/yr (18 over 10 kyr by Hubert-Ferrari et al. 2002, 18.6 over 3 kyr by Kozacı et al. 2009 and 15 to 20 mm/yr in a compilation made by Kozacı et al. 2009). This rate is about 75% of the geodetic slip rate which suggests that 25% of the displacement is taken outside of the main fault plane, presumably along minor faults on both sides of it.

We wish to emphasize an important point concerning the temperature structure of the upper lithosphere along the NAF. This is illustrated by the variation in the Curie Point (580°C for magnetite) depth as mapped by Aydın et al. (2005) in their figure 4. The NAF follows the 20 km depth contour of the Curie Point. Immediately to the south, the Curie Point depth rises to less than 10 km. **Thus, the lower crust and mantle are highly ductile to the south of the fault and much less to the north where the NAF keeps its course.** The mapping of Aydın et al. (2005) demonstrates that the NAF keeps north of the boundary of the ductile lower crust and mantle zone. *This observation leads us to propose that the North Anatolian Fault (NAF) would not stabilize unless the whole crust is in the brittle domain. Otherwise, it has no roots and rapidly loses its "memory".* Geologists have long indicated that the geological history of Anatolia suggests that no lithospheric mantle is present there (e.g. Şengör 1982). One can thus understand why the attempts to enter the bulk of Anatolia, by the Ovacık and Sungurlu faults, were unsuccessful and why the NAF has followed the northern limit of this ductile mass.

The NAF in the Sea of Marmara

Le Pichon et al. (2001) showed that the prolongation of the NAF within the Sea of Marmara, which they called the MMF, is a single throughgoing dextral strike-slip fault system that enters the Sea through the Gulf of İzmit and joins the Ganos fault at its western extremity. The geological slip rate along the eastern Çınarcık portion of this fault has been established to be 18.5 (12.6–29.4, 18.5 preferred rate) mm/yr over 500 kyr by Kurt et al. (2013). Grall et al. (2013) found a rate of 15.1 to 19.7 mm/yr over 405–490 kyr to the west on the Western High (see fig.2.5). On the Ganos fault Aksoy (2009) found 15.1 to 19.7 mm/yr over 20 kyr and Meghraoui et al. (2012) 17 ± 5 mm/yr over 1000 yr. Thus the geological slip rate across the Sea of Marmara is about 18 mm/yr as on the main eastern portion of the NAF. The NAF motion is consequently integrally transferred from the North Anatolian Fault to the Ganos Fault through the MMF. As for the NAF to the east of the Sea of Marmara, the rate along the MMF is also about 75% of the geodetic slip rate which suggests again that 25% of the displacement is taken outside of the main fault plane.

Many authors have assumed the continuity of the rate of motion at the entrance of the Sea of Marmara. They consequently consider that the motion along the NAF there is distributed between a Marmara branch and a Gemlik branch to the south (see fig. 2.5). For example, Gasperini et al. (2011) have measured a strike-slip rate of 4 mm/yr on the southern branch. Should these 4 mm/yr be deducted from the 18 mm/yr on the eastern NAF to obtain the rate of the MMF? Le Pichon et al. (1999) pointed out that this would be true only if the block between the two branches, which they called the Marmara block, had the same motion as the Anatolia block, which would be most unlikely. If the Marmara block has an independent motion, then the motion of this block can be adjusted in such a way that the rate is the same on the eastern NAF and the MMF to

avoid any velocity discontinuity which would be otherwise mechanically very difficult to sustain. This is indeed what was found by Meade et al. (2002), Le Pichon et al. (2003), Reilinger et al. (2006), Le Pichon and Kreemer (2010) and other authors since, who have introduced in their GPS solution an independent Marmara block.

Note however that Gasperini et al. (2011) have measured average rates of about 9 mm/yr on the NAF in the Gulf of İzmit by measuring offsets of a river channel that was drowned about 10 Kyr ago. This is about half the geological slip rate found to the east on the NAF and to the west in the Çınarcık Basin. This discrepancy is puzzling.

It is remarkable that the MMF follows the same 20 km Curie Point limit along the northern edge of the Çınarcık Basin as along the eastern NAF. The portion of the MMF that follows the northern border of the Çınarcık Basin within the eastern Marmara Sea corresponds to a stable localized strike-slip fault (Karabulut et al. 2002; 2011; Bulut et al. 2009; Örgülü 2011; Kurt et al. 2013). While the MMF is entirely contained within the thickly sedimented deep Marmara Basin, another fault system called South Marmara Fault by Le Pichon et al. (2014) was active in Pliocene time over the whole less sedimented southern shelf of the Sea of Marmara. Some activity is still related to this southern fault system. On the southern edge of the İmralı Basin, (Armijo et al. 2002) have described an active extensional system connected to the NAF in the Gulf of İzmit by a right-lateral fault along the shore of Armutlu Peninsula. Current fault activity does not seem continuous along the SW, transtensional, edge of İmralı Basin (see fig.2.5) but a number of active normal faults appear to distribute the extensional component of strain within the basin (Şengör et al. 2014). Moreover, subsidence at the depocenters proceeds at a steady rate since 0.4 Ma at least (Sorlien et al. 2012). Although little seismic activity

is recorded in this area, it is possible that the growth of the İmralı Basin may still today accommodate some significant strain.

The MMF crosses the 20 km Curie Point limit west of Çınarcık, as it changes direction to the WSW toward the Ganos Fault, to enter a hot lower crust and upper mantle area (Aydın et al. 2005). By doing so, it also enters into the serpentinite and mélange-dominated Intra-Pontide suture, which may also affect the mechanical behavior of the fault (see esp. Le Pichon et al. 2014, fig. 1b). There, in the western and central Sea of Marmara, this localized strike-slip fault appears to coexist with remnants of an extensional system, as demonstrated by significant distributed extensional seismic activity to the south of it (Altınok and Alpar 2006; Karabulut, unpublished data, 2015). This western portion of the MMF is thus best understood as an intermediate type of boundary, where a localized strike-slip fault has just been established within a previously existing broad zone of shear affecting the ductile lower crust and upper mantle.

The NAF west of the Sea of Marmara

From the Ganos fault westward into the Gulf of Saros, the Curie Point depth increases again to more than 20 km and the Ganos-Saros Gulf fault is indeed a stable pure strike-slip portion which corresponds to the westernmost extent of the stable pure strike-slip NAF (Karabulut et al. 2006). Recently the Mw 7.0 May 24 2014 earthquake confirmed the pure dextral strike-slip nature of the fault along the Saros Basin, all the way to 25°E (Görgün and Görgün 2015). Gasperini et al. (2011) using the same approach as in the Gulf of İzmit confirmed the existence of dextral strike-slip along the Saros fault. However they evaluated the rate as only about 10 mm/yr over the last 10 kyr post-glacial period.

This discrepancy with the geodetic slip rate of 22 to 24 mm/yr (Le Pichon and Kreemer 2010; Pérouse et al. 2012; Müller et al. 2013) is similar to the discrepancy in the Gulf of İzmit and suggests that there might be some systematic bias in their evaluation. Moreover, we pointed out earlier that trenching studies estimate slip rate on Ganos fault at 17 ± 5 mm/yr (Meghraoui et al., 2012; Aksoy et al., 2009) over similar periods of time.

We conclude that the rate of dextral strike-slip motion is indeed constant from the NAF to the western extremity of the Gulf of Saros fault along the prolongation of the NAF at a geodetic slip rate of about 24 mm/yr. However, as discussed above, although the geological slip rate also appears to be constant at about 18 mm/yr, it is only about 75% of the geodetic slip rate. We proposed earlier that this discrepancy between geodetic and geologic rates may indicate that about 25% of the displacement is taken outside of the main fault plane.

The western portion of the North Anatolian Fault: the undeforming Aegea block, the Ionian slab and the two shear zones that limit it

We now move westward within the Aegean domain. A remarkable result established by the geodetic surveys of Aegea is the existence of an essentially undeforming Aegea block limited by two shear zones (Reilinger et al. 2006; Floyd et al. 2010; Le Pichon and Kreemer 2010; Royden and Papanikolaou 2011; see fig.1). The northern shear zone, as mentioned above, has been called Grecian Shear Zone (GSZ) by Şengör (1979, 1982). It absorbs about 20 - 25 mm/yr of dextral slip. To the east, the Western Anatolian Shear Zone (WASZ) absorbs about 20 mm/yr of sinistral slip. Both zones are characterized by distributed extension across several extensional asymmetric rifts with decollement,

accompanied by rapid rotation at about $5^{\circ}/\text{Myr}$, clockwise for the dextral GSZ and counterclockwise for the sinistral WASZ (see detailed references below).

The rates of rotation mentioned above of about $5^{\circ}/\text{Myr}$ should be detected by paleomagnetic measurements if they persevered over several million years. Such rotations have indeed been detected (Kissel et al. 2003; Van Hinsbergen et al. 2005; Van Hinsbergen et al. 2010). They are clockwise in the western Aegean region and may have reached as much as 40° between 15-13 and 8 Ma with an additional 10° after 4 Ma (Van Hinsbergen et al. 2005). They are counterclockwise within the eastern Aegean region in the WASZ and have been measured as about 20° between 16 and 5 Ma (Van Hinsbergen et al. 2010). These finite rotations thus qualitatively agree with the present geodetic rotations. Of course, paleomagnetic techniques can only detect rotations of relatively short radius. The absence of detectable paleomagnetic rotations thus does not imply the absence of important motions of large rotation radius. In addition, as the paleomagnetic results are obtained by an averaging process between different sites, the distinction between local block rotations and regional rotations is difficult to make, especially in Aegea where numerous local block rotations have been demonstrated to be present. However, an important point is that the rapid phase of rotation detected by paleomagnetism began about 15-13 Ma to the west and 16 Ma to the east confirming that middle Miocene was an important time for the tectonics of Aegea.

The existence of this undeforming Aegea block is not integrated in recent discussions of the geodynamics of the area (e.g. Royden and Papanikolaou 2010; Vassilikis et al. 2011; Tirel et al. 2013; Pearce 2015). These recent discussions assume rapid lithosphere extension related to the subduction of the oceanic Mediterranean lithosphere since

Middle Miocene. But actually, the episode of rapid extension within the present Aegea block peaked between Serravallian, about 12 Ma ago, and Messinian (7 to 5 Ma) and ceased to exist within this Aegea block after Messinian time (5 Ma) while it continued within the WASZ and was initiated within the GSZ between 2 and 1 Ma ago (Armijo et al. 1996; Sorel 2000; Flotté 2003). We attribute this cessation of extension within the present Aegea undeforming block to the shielding effect of the cold oceanic Ionian slab. The subduction of the Ionian oceanic slab started about 10 Ma ago and this shielding effect thus progressively extended over the whole southern Aegean mostly after Miocene time. Moreover, the slab now dips some 10° to 15° up to a distance of some 300 km from the Hellenic "Trench". Despite the prodigious amounts of Messinian salt probably going down into the subduction channel, this would induce a considerable frictional stress onto the upper plate that would counteract the extension farther north in a situation not too different from the south-central Andes in Chile (see Oncken et al. 2006, especially fig. 16-12 a-f). This would explain why the "dead" block does not extend far into the north in the Aegean Sea. Another explanation proposed by Sonder and England (1989) is that the thinning of the southern Aegean Sea that was initiated about 20 Ma ago may have reached a critical level 3–5 Ma ago because of the cooling of the extending lithosphere, and that consequently the strain associated with the pull toward the arc was transferred further outward, including to western Anatolia. However, their proposal does not explain why the extension continued within northern Aegea and Western Anatolia although the extension had started at about the same time there. We will come back to this important point later on.

We now consider more carefully these two shear zones. *We will show that within the West Anatolian-Aegean domain, outside of the presently undeforming Aegea block, the*

283 *normal mode of shearing there, in this ductile lower crust-upper mantle environment, is*
284 *through rotation of en-echelon normal faults over a 100-200 km wide shear zone*
285 *(McKenzie and Jackson 1983, 1986, 1989; Jackson and McKenzie 1989). Further, the brittle*
286 *distension roots in a shallow decollement above the ductile lower crust. This is because, in*
287 *addition to the ductility of the lower crust and mantle below 10-15 km as in Eastern*
288 *Anatolia, the thinned lithosphere is also affected by strong N/S extension due to the roll-*
289 *back of the peri-Aegean subduction.*

290

291 The GSZ consists of several extensional rifts oblique to the direction of motion of Aegea
292 with respect to Eurasia, the largest of these being the Corinth and Evia grabens (see
293 fig.1). These are asymmetric rifts with active decollements (Rigo et al. 1996; Jolivet et al.
294 2010). In this area, distributed extension is accompanied by rapid clockwise motion at
295 about 5°/Myr (Le Pichon and Kreemer 2010; Floyd et al. 2010; Chousianitis et al. 2010).
296 The width of this portion of the shear zone in a direction perpendicular to the motion of
297 Aegea with respect to Eurasia is about 100-200 km. There is no localized strike-slip
298 boundary there, as evidenced by the absence of strike-slip earthquakes. Yet, Durand et
299 al. (2014) have suggested that there is a localized strike-slip boundary at depth, within
300 the ductile portion of the lithosphere. The existence of shallow dipping decollements
301 flattening at a depth of about 10 km suggests that the lower crust is ductile (Jolivet et al.
302 2010). Further, Tiberi et al (2001) have shown that the Moho topography is unrelated to
303 the surface tectonics as the lower crust appears to have been affected by boudinage. This
304 confirms that the lower crust is most probably ductile there. Thus, there is no reason to
305 assume the existence of a hidden extension of the NAF at depth below the CHSZ as
306 proposed by Durand et al. (2014). Most authors had previously recognized the absence

of a localized strike-slip boundary at depth, notably Şengör (1979) and Goldsworthy et al. (2002).

The WASZ also consists of actively extending asymmetric rifts rotating counterclockwise at about 5°/Myr (Floyd et al. 2010). Karabulut et al. (2013) have put into evidence “small-amplitude and long-wavelength lateral variations of the Moho topography” which suggests to them that “viscous flow in a hot lower crust has smoothed out the lateral variations of crustal thickness.” Aydın et al. (2005) have shown that the Curie Point is at a depth of 5 - 10 km in this area, which confirms that the lower crust is indeed in the ductile regime. The width of the WASZ is similar to the width of the GSZ and equal to about 100-200 km. This width, of the order of the thickness of the average lithosphere, suggests that it is related to block shearing of the whole lithosphere. The deformation in the upper brittle crust is an adjustment to this underlying distributed shear. In conclusion, both the WASZ and the extensional portion of the GSZ have a similar width of about 100-200 km and have a hot and ductile lower crust. The underlying mantle is of course also hot and ductile. We can conclude that this type of shearing zone with rotating distributed extension in the upper crust is preferred in this extensional hot environment where the lower crust and upper mantle are ductile.

We then move to the two portions of the northern Aegean boundary that have not yet been discussed, on both sides of the GSZ: the Sporades Basin, at the western extremity of the North Aegean Trough to the east of the GSZ, and the Kefalonia-Patras fault system to the west. We show that these are portions of the North Aegean Boundary intermediate in character between stable pure strike-slip localized faults and highly distributed extension through rotating faults on top of a sheared lower crust and upper mantle. The

Sporades Basin appears to possess a localized strike-slip fault as well as distributed extension to the north of it (the Thermaikos and Chalkidiki extensional systems). The Kefalonia-Patras fault system has a major strike-slip fault at its northern extremity, the Kefalonia fault, and a minor one at its southern extremity, the Patras fault, with distributed shear in-between. Both portions of boundary appear to be sheared over a width of 100 to 200 km at depth in the lower crust and upper mantle.

A new insight on the behavior of these two portions of the boundary is given by the seismic sequence that followed the January 6 2008 Mw 6.2 earthquake (37.15°N, 22.95°E). This earthquake ruptured the Ionian slab at a depth of 80 km (see Durand et al. 2014) and resulted first in roll-back of the subduction with southwestward motion of the top block toward the trench and then in dextral shear of Aegea with respect to the Eurasia plate. On June 8 1968, an Mw 6.4 dextral strike-slip earthquake activated the Patras-Achaia strike-slip fault over a length of 25 km and then during the second part of 2008, the GSZ (Corinth, Sperchios and Evia rifts) was activated in extension. This extensional activity extended northward into the NAT area within the Thermaikos and Chalkidiki extensional systems.

We now consider the velocity field across the GSZ. Floyd et al. (2010) show that both across the extensional part of the GSZ and across the Kefalonia and Patras-Achaia fault system, there is a continuous linear increase in GPS velocity with respect to Eurasia from about 5-10 mm/yr north of the Aegea block to 30-35 mm/yr within the block. This increase occurs over about 100-200 km, the gradient being somewhat steeper in the purely extensional area than in the Kefalonia-Patras strike-slip area (see also Shaw and Jackson 2010). The increase in velocity over more than 100 km cannot be explained only

by elastic effect because this width is too large. The only reasonable explanation is that it reflects ductile shear in the underlying mantle below, as concluded by Shaw and Jackson (2010) and Floyd et al. (2010).

Considering more carefully the Kefalonia-Patras shear system, the most important fault is the Kefalonia fault that absorbs about 15 mm/yr of dextral slip (Vassilakis et al. 2011) or up to 23 mm/yr (Pérouse et al. 2012). It links the northern extremity of the Aegean subduction to the Aegean northern boundary. As a consequence, the whole Kefalonia-Patras shear area lies on top of the shallow African slab that is subducted at a velocity of about 34 mm/yr, much faster than the 4 mm/yr convergence velocity of the Adriatic crust with the northern Greece crust, north of the Kefalonia fault (Pérouse et al. 2012). The lithosphere being sheared is everywhere thinner than about 50 to 40 km (Pearce 2015). This is a very peculiar situation. We will come back to this point later when discussing the possible role of the slab. Cocard et al. (1999) had shown that this domain had an important clockwise rotation rate centered on the islands of Kefalonia and Zakintos. The rate of rotation is estimated at $6-8^{\circ}/\text{Myr}$ by Chousianitis et al. (2014) and $8^{\circ}/\text{Myr}$ by Vernant et al. (2015). This Patras-Achaia area is limited to the northeast by a zone of left-lateral shear between the Gulf of Amavrakikos and the eastern tip of the Gulf of Corinth. The rate of motion along this zone of shear is estimated to be about 10 mm/yr (Vassilakis et al. 2011; Vernant et al. 2015).

The June 8 2008 earthquake activated the Patras-Achaia fault at the southern extremity of the Kefalonia-Patras shear system as Durand et al. (2014) have demonstrated. This fault is identified by Shaw and Jackson (2010) through its seismicity but only appears on GPS as a change from a gradient in velocity to the north versus a uniform velocity to the

382 south. Vassilakis et al. (2011) tried to evaluate its motion by identifying small more or
383 less rigid "fragments" and conclude that the motion is 7 mm/yr with about 5 mm/yr in
384 extension and 5 mm/yr in strike-slip. This is not in agreement with the pure strike-slip
385 motion revealed by the June 8 earthquake. In any case, their method maximizes the
386 motion on the boundaries between blocks as it ignores any internal deformation. Thus
387 one can only state with certainty that the motion along this fault is too small to be
388 properly characterized by GPS. Indeed, the fault cannot be identified on the field (Shaw
389 and Jackson 2010). Cases of faults with strong seismogenic potential but too small slip
390 rate to be detected by geodetic methods have indeed been described elsewhere (e.g.
391 Friedrich et al. 2004). Thus the Kefalonia-Patras shear area is a zone of continuous shear
392 deformation over a width of at least 100 km bounded by two localized strike-slip faults.
393 The most important is the Kefalonia fault that absorbs about 20 mm/yr of dextral slip
394 (15 mm/yr for Vassilakis et al. 2011, 20 mm/yr for Vernant et al. 2015 and 23 mm/yr
395 for Pérouse et al. 2012). The Patras-Achaïa fault absorbs a maximum of about 5 mm/yr.
396 The June 2008 earthquake activated it, as Durand et al. (2014) have demonstrated.

397
398 We now go back to the NAT. Although Reilinger et al. (2006), Le Pichon and Kreemer
399 (2010), Pérouse et al. (2012) and Müller et al. (2013) all agree that dextral strike-slip at
400 a rate of 22 to 26 mm/yr occurs along the NAT, the presence of simultaneous active
401 extension immediately to the north of the Sporades Basin in the Thermaikos and
402 Chalkidiki extensional systems indicates that the western NAT does not correspond to
403 pure strike-slip localized on a single fault. Pérouse et al. (2012) and Müller et al. (2013)
404 emphasized this point. The activation of the Sporades and Chalkidiki extension, but not
405 of the Sporades dextral strike-slip, during the 2008 seismic sequence confirms that this
406 is an intermediate type of boundary, where a localized strike-slip fault coexists with

remnants of an extensional system, on top of a broad zone of shear affecting the ductile lower crust and upper mantle. We have seen above that this intermediate type of environment also characterizes the central and western Marmara Sea with simultaneous presence both of a localized strike-slip fault, the MMF, and of active remnants of the previous extensional system.

To conclude, west of the Saros Basin, the northern boundary of the Aegean block absorbs 22 to 24 mm/yr of dextral strike-slip over a broad 100-200 km wide zone of shear of the lithospheric mantle, with various expressions within the upper crust. From east to west, there are first both localized strike-slip along the Sporades Basin and simultaneous extension to the north in the Thermaikos and Chalkidiki area. Then there is distributed clockwise rotating extension within the GSZ. Finally, there is distributed shear between two strike-slip faults, the Kefalonia and Patras-Achaïa faults in the westernmost portion. This last portion joins the northern boundary of the Anatolia-Aegea plate to the Aegean subduction plate boundary. It crosses an area of relatively thin crust with no well identified Moho, on top of the shallow subducting slab (Pearce 2015). The crust there probably consists of geologically recent stacks of cold nappes cut from the asthenosphere by the underlying slab as discussed by Pearce (2015). This relatively cold environment accounts for the presence of localized strike-slip faults.

It is important to note that the effect of the deformation in the GSZ and WASZ is to allow the relatively undeforming Aegea block to rotate clockwise away from Anatolia around a pivot situated near the northern extremity of the subduction zone (see for example figure 8 of Le Pichon and Kreemer 2010, in which the GPS vectors are with respect to Anatolia).

This rotation has two effects on the subduction zone. First, it increases the subduction

rate to a value of about 35 mm/yr along the whole NW/SE segment of the Hellenic subduction zone as this rotation compensates for the southeastward decrease of motion of the Anatolia/Eurasia rotation. Second, it maintains the subduction direction strictly perpendicular to the subduction zone (Le Pichon and Angelier 1979). This rotation thus appears to be governed by the roll-back of the subduction zone.

Consequences of the subduction of the Ionian slab

Coming back to the drastic consequences of the subduction of the cold and heavy oceanic Ionian slab, *we argue that the presence of a low dip cold slab in rapid subduction has drastic consequences for the overlying mantle and crust. This is because it cuts the overlying lithosphere from asthenospheric sources and induces a shortening environment over it. This, in our opinion, explains why major extension is absent within the present Aegea block while it was very active between 13 and 5 Ma.*

The cold 200 Ma old oceanic Ionian slab (Truffert et al. 1993) is presently subducted below Aegea at a velocity of about 34 mm/yr, much faster than the 4 mm/yr convergence velocity of the continental Adriatic crust north of the Kefalonia fault with the northern Greece crust (Pérouse et al. 2012). The length of the seismically active slab is about 350 km below Peloponnesus (e.g. Gudmundsson and Sambridge 1988), which indicates that this rapid subduction is 10 Myr old or somewhat older if the average rate of subduction since the initiation of subduction was equal or somewhat smaller than the present one. The volcanic arc related to this subduction was initiated near 2.5 Ma (Le Pichon and Angelier 1979). The slab below the volcanic arc is about 150 km deep and the length of slab from the volcanic arc to the subduction zone is about 300 km. Thus the age indicated for the initiation of subduction is also 10 Ma or somewhat older.

457

458 This age is in fair agreement with the age of formation of the accretionary wedge, which
459 occurred after 15 Ma as established by deep sea drilling (Kastens et al. 1991). Located
460 within a trough near the backstop leading edge, DSDP 377 borehole penetrated early to
461 middle Miocene (between 19 and 15 Ma) turbidites of probable African origin. At the
462 time the turbidites were deposited, the present accretionary wedge could not have
463 existed if the turbidites indeed came from Africa. These turbidites were covered with 15
464 to 14 Ma pelagic marls whereas turbidites continued to be deposited to the south. Thus
465 the change of sedimentation to pelagic marls at 15 Ma is compatible with an initiation of
466 the present accretionary wedge at about this time. The probable limestone backstop is
467 covered by late Miocene basins confirming that it reached its present configuration also
468 in the late Middle Miocene, near 15 Ma (Le Pichon et al. 2002). Thus this margin was
469 formed approximately 15 Ma ago. Le Pichon et al. (2002) proposed that this was a time
470 of reinitialization of the subduction system after slab break. We adopt their proposal.
471 This late initiation of the Ionian oceanic plate would also explain why the Kefalonia
472 transform is geologically quite recent (post-Messinian, 5-7 Ma according to Royden and
473 Papanikolaou 2011). We have noted above that the finite clockwise and
474 counterclockwise rotations detected by paleomagnetism within Aegea were initiated 16
475 to 15 Ma.

476

477 The alternative is that the slab stops being seismically active 10 Myr after being
478 subducted but continues at greater depth as an aseismic slab. This alternative is
479 proposed by authors who postulate continuity of the slab to great depth on the basis of
480 tomographic studies (Bijwaard et al. 1998; Piromallo and Morelli 2003). In this line of
481 thought, Royden and Papanikolaou (2011) and Tirel et al. (2013) agree that rapid

subduction of the cold Ionian lithosphere would only have started at 6-8 Ma, following a slower subduction episode of external carbonate platform on transitional lithosphere. However, in their hypothesis, it is hard to understand how the slab was able to maintain its continuity as it went from oceanic in Eocene time, to semi-continental in Oligocene and lower to middle Miocene, then back to oceanic again in upper Miocene to Recent time. Models that have been proposed to investigate the change due to the subduction of a continent (Martinod et al. 2000; Royden and Husson 2006) suggest that continental subduction induces an increase in the slab dip angle and a decrease in the subduction velocity due to a change in slab buoyancy. They also predict that the slab should break at the former location of the continental block. These models then are difficult to reconcile with the interpretations, which assume continuity of the slab and no break, such as those proposed by Tirel et al. (2013) and Pearce (2015).

The relationship between simple idealized 2-D models of subduction and actual 3-D complex Aegean subduction is not obvious. For example, the continental subduction north of the Kefalonia transform dips at the same 17° shallow angle as the oceanic subduction to the south (Pearce 2015). However the assumption of a 200-300 km narrow Ionian slab extending uninterrupted over more than 1000 km in spite of these drastic changes in subduction type and subduction direction is based on tomography. But is the assumption justified? The presumed continuous slab depicted by tomography shows significant heterogeneity (see fig. 7c of Bijwaard et al. 1998, and section Ff of figure 9 of Piromallo and Morelli 2003). More importantly, the continuity is undermined by the proven existence of slab breaks in its upper portion. A major slab break, long suspected on the basis of tomographic studies, has been demonstrated to exist between Aegea and Anatolia (Paul et al. 2014). Pearce (2015) concludes his extensive

investigation of seismic imaging of the western Hellenic subduction by stating that “there may be two overlapping slabs beneath northern Greece, one dipping at 17° toward $N60^\circ E$ from 30 to 70 km depth and the other sitting horizontally at 200 km depth”, as shown by Soboudi et al. (2006). He also indicates that he cannot exclude a tear of the slab between northern and southern Greece as proposed by Spakman et al. (1993) and concludes that the slab may be locally segmented in some way at least near its top. The existence of these ruptures comforts the conclusion of Le Pichon and Kreemer (2010) who pointed out that the continuity of the conical seismic slab with a deeper slab could not be maintained without significant rupture.

Furthermore, the extent of the Ionian slab below Aegea, as defined by the intermediate seismicity distribution (Gudmundsson and Sambridge 1988; Shaw and Jackson 2010), is broadly confirmed by the high-resolution surface tomography of Salaün et al. (2012), which shows that the low-velocity anomaly in the upper mantle, between 80 and 160 km depth, present in Anatolia and north central Aegea, is limited to the south by the seismically active Ionian slab. Paul et al. (2014) show that the zone of low velocity anomaly, in Anatolia and north central Aegea, has NE/SW fast-wave polarization axes of shear-wave anisotropy, whereas in southern Aegea, where the seismically active slab is present, the orientation is quite different and significantly smaller.

The mapping of intermediate seismicity depth in the slab by Gudmundsson and Sambridge (1988) and Shaw and Jackson (2010) confirms that the slab is not present below the Corinth and Evia grabens, although Suckale et al. (2009) assume that the slab extends below the eastern extremity of the Corinth Gulf beyond a depth of 100 km. However, this assumption of Suckale et al. (2009) does not appear to be substantiated.

This is because their composite images of the slab show that the well-defined image of the slab disappears at a depth of 100 km before reaching the area below the eastern Gulf of Corinth. The mapping of Gudmunsson and Sambridge also indicates that the slab is present below the Kefalonia-Patras shear zone, as discussed above. And the detailed seismic imaging of the western Hellenic subduction by Pearce (2015) confirms the presence of the slab below the Kefalonia-Patras shear zone, as noted earlier. We can conclude that the slab is absent below the GSZ and present below the Kefalonia-Patras zone, which would explain the tectonic difference below the two zones.

Note that the continuity-of-slab hypothesis also leads to another difficulty as it does not explain why the extensional collapse of the Aegean landmass, which resulted in the formation of the Sea of Crete, could have occurred above a shallow subducting slab that should have created a compressional environment. Drooger and Meulenkamp (1973) have shown that the Southern Aegean landmass fragmented and most of it came below sea level at the end of Serravallian (about 13 Ma, see also Angelier 1979) while it is now part of the undeforming Aegean block since Messinian, at about 5 Ma (Angelier et al. 1982). We submit that the apparent heterogeneity of the tomographic slab may be due to the coalescence of pieces of slab following various tear episodes in Late Eocene and Middle Miocene.

If this tear of the slab did indeed occur prior to the initiation of subduction of the oceanic Ionian slab between 15 and 10 Ma, one can then distinguish between the presently subducting Ionian slab at a rate of 34 mm/yr and with a dip of 17° toward N35°E reaching a maximum depth of 170 km and the older coalesced slabs now dipping at about 45° toward the north between 170 km and about 1200 km (Bijwaard et al. 1998).

This older system, which extends over 1000 km and penetrates into the lower mantle, is most probably anchored within the mantle. Consequently its kinematics are independent of what happens at the surface. Its vertical velocity must be approximately equal to the vertical component of velocity of the subducted Ionian slab, so that contact is maintained between the two subducted pieces. Thus the vertical velocity of the Ionian slab is about 10 mm/yr. This vertical velocity implies an age of the order of 100 Ma for the initiation of subduction of this older slab, which is quite reasonable.

We consequently conclude that the subducted Ionian slab now prevents the asthenosphere from reaching the base of the lithosphere and produces above it a tectonic environment that is cooler and affected by shortening, or at least much less extension. We argue that this is the reason why the essentially undeforming Aegea block coincides with the part of Aegea that is presently underlain by the Ionian slab. This then explains why the Sea of Crete, that resulted from the extensional collapse of the Aegean landmass during Mid to Late Miocene, when a slab was absent below it, following the post-Oligocene break, has been part of the undeforming Aegean block since Messinian, about 5 Ma ago (Angelier et al. 1982). At this time, the Ionian slab began to propagate northeastward below the Sea of Crete and consequently cut off the asthenospheric source on which this extension thrived.

The penetration of the NAF within the western-Anatolian-Aegean landmass 4 Ma ago

Figure 2 shows our proposed reconstruction of the NAF 4 Ma ago. Le Pichon et al. (2014) have shown that at this time the NAF had not yet entered the area of what is now the Sea

of Marmara but had just reached its eastern extremity, near the Gulf of İzmit. We will show that this is the time at which asymmetric extensional rifts similar to those of the present Gulf of Corinth were formed in the southern part of what is now the Sea of Marmara. These rifts were oblique to the direction of motion of Anatolia with respect to Eurasia, with active decollements to the south of a large anticline on the present southern shelf of the Sea of Marmara. This distributed extension, which we consider to have been accompanied by rapid clockwise motion, accommodated the deformation on the western tip of the 4 Ma NAF (Fig. 2, 3a and 3b).

Let us first briefly discuss the geodynamics of the western-Anatolian-Aegean landmass 4 Ma ago. We have seen that the present geodynamics of Aegea and the prolongation of the NAF along its northern boundary are now greatly influenced by the presence of the rapidly subducting Ionian slab. This is because the subduction of the slab resulted in the formation of an undeforming Aegean block on top of the advancing slab. This subduction was initiated about 15 Ma ago, 3 to 4 Myr before the NAF initiated its westward propagation (12 to 11 Ma ago according to Şengör et al. 2005). The fact that subduction of the oceanic Ionian plate immediately preceded the formation of the NAF suggests that slab pull was essential to trigger the extrusion of Anatolia. Before the slab had sufficiently advanced to result in the formation of the relatively undeforming Aegean block, between 15 and 4 Ma ago, the Western Anatolia-Aegea environment was dominated by ductile lower crust and upper mantle and was affected by strong N/S extension due to the roll-back of the peri-Aegean subduction distributed over the whole Aegean. **In this tectonic environment, the shear between the westward motion of Anatolia and Eurasia was distributed over the whole width of Aegea. As a**

consequence, there was no localized northern boundary fault north of Aegea (Fig.2).

This does not mean that extension was not present in the northern Aegean. For example, the North Aegean Trough has deposits of Messinian (7-5 Ma) evaporites (Mascle and Martin 1990) and Messinian carbonates and marls are present along the coasts of Saros Gulf and Çanakkale close to North Aegean Trough (Çağatay et al. 1998, 2006; Melinte-Dobrinescu et al. 2009). But there was no localized zone of shear along it. However as the Ionian slab progressed northeastward below Aegea, distension ceased within the overlying lithosphere and a relatively undeforming Aegean block progressively increased its size. Consequently, the shear could not be distributed over the whole of Aegea. Accordingly, the kinematics had to adjust and a localized zone of shear began to form along the northern boundary of Aegea. This is confirmed by the fact that the age of the GSZ is quite young, less than 2 Ma. Armijo et al. (1996) have shown that the phase of rapid extension in the Gulf of Corinth corresponds to the last million years and Sorel (2000) and Flotté (2003) have dated the initiation of the Gulf of Corinth between 1.7 and 1.1 Ma. Flotté (2003) estimated the total extension to be 15 km. This finite extension should be compared to the present rate of extension of 15 mm/yr. It is equivalent to 1 Myr of extension at the present rate. Nielsen (2003) estimated the dextral offset along the Kefalonia fault at 37 km using the offset of two portions of a seamount that appeared to have been cut in two pieces. This offset would have been produced in 1.6 Myr using the strike-slip rate of 23 mm/yr given by Pérouse et al. (2012). Similarly, Tüysüz et al. (1998) estimated the age of the initiation of dextral motion in the Gulf of Saros as between 2 and 1 Ma, Kurt et al. (2000) as Plio-Quaternary and Ustaömer et al. (2008) as Quaternary. This strike-slip phase followed a compressional phase in the Saros Gulf

region (4-3 Ma) based on the age of the Conkbayırı Formation, an alluvial fan formed on the uplifted southern shoulder of Saros Gulf (northern Gelibolu Peninsula) (see e.g., Sümengen et al. 1987; Çağatay et al. 1998). *Thus the present northern plate boundary of Aegea is less than about 2 Ma from the Gulf of Saros to the Kefalonia fault.*

We should remember however that the 200 km WASZ had been a zone of active extension since Late Miocene and remained in this situation to this day. Consequently the prolongation of the NAF within the Sea of Marmara, which is situated exactly to the north of the WASZ, can be considered to be a leaky transform. The Sea of Marmara thus marks a major transition in the propagation of the NAF toward the Aegean subduction zone.

The Pliocene Sea of Marmara

We now return to the area that will become the Sea of Marmara about 4 Ma ago. Le Pichon et al. (2014) described a continuous gently curvilinear, south-concave zone of deformation about 10 km wide that extended over the whole southern shelf of the Sea of Marmara, from the Gulf of Gemlik to the Dardanelles Strait, in Lower Pliocene time, about 4 Ma (Fig. 3a and 3b). They called this zone of deformation the South Marmara Fault (SMF) system and proposed that the SMF was then a branch of the dextral North Anatolian Fault. They noted that this zone of deformation was associated with a large anticline within the central part of the section, between Marmara and İmralı islands (Fig. 3a, 3b and 3c). However the anticline was not found west of Marmara Island nor east of İmralı Island and the reason for the association of the anticline with the central part of the SMF was not elucidated in their paper.

The basement in the anticline area has been uplifted by more than 1 km, forming a 6 to 10-km-wide anticlinal fold. The consequent uplift led to the erosion of the upper sedimentary units near the summit of the anticline. This folding occurred as the uppermost SE dipping foreset beds of the Alçitepe formation (uppermost Miocene–Lower Pliocene, 5.3 to 3.7 - 3.4 Ma, Melinte-Dobrinescu et al. 2009) were still being deposited to the south of the present southern shelf of the Sea of Marmara. The source of the sediments during their deposition was to the NW and not to the south, which indicates that the Sea of Marmara deep basins did not exist then. Both the upper Miocene (Kirazlı, 11-10 to 5.3 Ma) and Uppermost Miocene - Lower Pliocene (Alçitepe) sedimentary sequences were conformably deposited, then folded and strongly eroded (Fig. 3c and 4). Le Pichon et al. (2014) concluded that the uplift occurred near the end of Lower Pliocene, between approximately 4.5 and 3.5 Ma. The uplift of the basement increases westward toward Marmara Island and the maximum uplift is about 1.5-2 km over Marmara Island that was uplifted and broadly warped at this time (see Le Pichon et al. 2014 for a short discussion of the tectonics of the island). An examination of the topography of the island confirms the broad EW anticlinal upwarp of the island. We reexamined the termination of the anticline west of the island of Marmara for this paper (fig.4). The anticline and the uplift terminate abruptly along a NNW/SSE normal fault west of the island of Marmara.

Le Pichon et al. (2014) noted that the SMF occupies the apex of the anticline. But the SMF continues westward of Marmara island to the Biga Peninsula, just east of the Dardanelles Straits, whereas the anticline does not. If the anticline and the SMF were cogenetic, the disappearance of the anticline west of Marmara Island would be puzzling as one would expect the shortening component to increase as the SMF bends its

direction southwestward west of Marmara. Indeed the SMF is clearly compressional west of the island of Marmara and, during the Pliocene, a compressional regime prevailed in the north of the Gelibolu Peninsula, with the formation of the Anafartalar thrust and the deposition of the Conkbayırı alluvial fan related to it (e.g., Çağatay et al. 1998). However, the anticline along the SMF disappears west of the Island of Marmara, thus this folding appears to be not contractional. This suggests that the formation of the anticline is not related to the SMF. We can conclude further that the SMF occupied the apex of the anticline immediately after the anticline was deactivated, probably because the apex was the weakest portion of the crust as Steckler and ten Brink (1986) predicted that, within a zone of continental thinning, a strength minimum exists just seaward of the hinge zone. But the activity of the SMF was geologically brief as it was terminated when Lower Pliocene fresh water calcareous mudstones were deposited, probably corresponding with the Conkbayırı Formation.

We propose that the anticline is related to a phase of extension comparable with the one now occurring within the GSZ in the Gulf of Corinth and we relate it to the detachments mapped by Bécél et al. (2009) in the southern Sea of Marmara between the sea floor and 13 km depth (see figures 3a, 3b and 3c). We also show in figures 3a and 3b the crests of tilted blocks on top of these detachments as mapped by Bécél et al. (2009). To illustrate the similarity between this system and the Gulf of Corinth extensional system, we show in figure 5 an interpreted lithospheric SW-NE profile (see location on figure 6). The fault locations are from Le Pichon et al. (2014) and Laigle et al. (2008) and the supra-crustal model is based on Bayrakci et al. (2013) while the detachment geometry comes from Bécél et al. (2010). The Moho depth is the Kende et al. (in prep.) model and will be discussed later when presenting figure 6. Interpretation of seismic profiles of Turkish

Petroleum (TPAO) were also consulted to constrain the fault and sediment deposition geometries on the southern shelf. The locations of microearthquakes are from Kandilli Observatory data. We have added in figure 5, below the profile, the sketch of Jolivet et al. (2010) that illustrates the possible future evolution of the Gulf of Corinth to illustrate the similarity between this structure and the Gulf of Corinth one. The possible evolution of the Gulf of Corinth described by Jolivet et al. (2010) is that of a typical Cordilleran-style metamorphic core complex. The origin of the antiform has been attributed to the footwall isostatic rebound by many authors since Spencer (1984) and Wernicke et al. (1985). We consider that figure 5 demonstrates that the Southern Shelf anticline and the decollement of B  cel et al. (2009) are cogenetic. Because we know the age of formation of the anticline, about 4 Ma, we conclude that the area corresponding to the southern portion of the present Sea of Marmara was similar to the present Gulf of Corinth about 4 Ma.

In figures 3a, 3b and 3c, the anticline has two distinct portions, a Marmara portion to the west and an   mralı one to the east. They both have the same orientation and the same width. But they appear to be left-laterally offset. We have sketched in dotted line along the direction of this offset a possible shear. A similar shear with the same direction might limit the   mralı anticline portion and separate it from the Armutlu uplift. Finally, another one might limit the western extremity of the   znik Basin. In the tectonic context of the time, these shears might correspond to left-lateral x shears and be related to the clockwise rotation that must have affected the whole area in front of the western tip of the NAF, rotation which is attested by the clockwise rotation of 16   detected paleomagnetically by Av  sar and    seven (2009) on volcanic Eocene rocks of the Armutlu Peninsula. It is interesting to note in this context that the orientation of the large   mralı

canyon is compatible with the direction of these probable x-shears. But we do not have enough structural data to identify this other possible İmralı canyon x-shear. Note however that, as noted above, the fault along the western end of Armutlu Peninsula has been interpreted by Armijo et al. (2002) as being presently an active dextral fault limiting the presently extending İmralı Basin.

The presence of the İmralı portion of the anticline led us to investigate whether it could also be associated with a detachment, similar to the Marmara one, to the west of the Armutlu Peninsula. We noted on the SEISMARMARA seismic profile in the İmralı Basin, shown in fig 2.b of Laigle et al. (2008) (see location of the profile in figures 3a and 3b), that the basement dipping northward could be interpreted as a detachment and the basement between İmralı and Çınarcık as a huge tilted block. As we had no other information, we gave to this detachment the same orientation as the Bécel et al. (2009) detachment. We converted it into depth following the conversion used by Bécel et al. (2009). This double basin interpretation, i.e. a western one (paleo-Central basin) and an eastern one (paleo-İmralı Basin), leads us to propose that the whole area corresponding to the present Sea of Marmara was occupied about 3.5 to 4.5 Ma by two asymmetric extensional basins, with detachments dipping NNE, within a broad zone that rotated clockwise in front of the tip of the advancing NAF, as depicted schematically in figure 2. This area then should correspond to the oldest portion of the Sea of Marmara and should have a purely extensional origin.

An intriguing point is that the western anticline segment increases in amplitude westward toward Marmara Island, whereas the eastern one increases in amplitude eastward toward İmralı Island. This opposite polarity would be expected if these basins

formed a relay between the Eastern NAF and a just initiated paleo-Ganos-Gulf of Saros Fault. This situation then would be similar to the present situation of the GSZ as a relay between the NAT and the Kefalonia-Patras Fault system (see figure 16 of Shaw and Jackson 2010).

Figure 6 shows the present distribution of the marine Alçıtepe Formation and of the continental Pliocene after the map of the General Directorate of Mineral Research and Exploration (2002), Görür et al. (1997) and Sakıncı et al. (1999). This is the first time when marine Mediterranean and the Marine Paratethyan sequences began mixing in the later history of the Marmara Sea and the well-controlled shape of the marine ingression indicates that it was the southern fault systems that were governing the route of the invading seas (see especially Figs. 9 and 10 in Sakıncı et al. 1999). The distribution of the Alçıtepe neritic formation in the west corroborates that the basins, formed between approximately 4.5 and 3.5 Ma, were situated in the southern portion of the present Sea of Marmara. The distribution of the formations to the west of the present Sea of Marmara appears to be compatible with the initiation of a paleo-Ganos-Gulf of Saros Fault during Uppermost Miocene – Lower Pliocene.

We conclude that sometime between 4.5 and 3.5 Ma, two left-laterally offset basins were formed in a purely extensional context and now correspond to the oldest part of the present Sea of Marmara. At the end of this tectonic episode, the NAF entered for the first time in the Marmara area, through the present Gemlik Bay, within what would later become the shelf of the Sea of Marmara, following the axis of the İmralı and Marmara anticlines. It then presumably joined a paleo-Ganos-Gulf of Saros Fault. Le Pichon et al. (2014) named this branch of the NAF the South Marmara Fault. Note that, in our

interpretation, the present Gemlik fault was part of the early SMF. Thus, we propose that the SMF was the primary structure in Pliocene and that prior to Pliocene, the shear was distributed over the whole western Anatolia. There was no primary structure then.

The formation of the SMF marked the end of the initial extensional phase and the beginning of the second strike-slip phase, although the transition between the two stages was most probably gradual because extension is still present today in the southern part of the Sea of Marmara. During this second phase, the strike-slip motion migrated northward and established itself along the present MMF. It was responsible for the architecture of the northern part of the Sea of Marmara, which was submitted to a strikingly different strike-slip context than the one prevailing during the first purely extensional phase.

Crustal thinning below the Sea of Marmara

Before getting into the discussion of this second mostly strike-slip phase of formation of the Sea of Marmara, we need to examine crustal thinning there in the context of the overall thinning prevalent in Aegea and Western Anatolia since Middle Miocene. Crustal thinning has been investigated by different means with consistent results over the Whole Aegean -Western Anatolian area (e.g. Tiberi et al. 2001; Tirel et al. 2004; Karabulut et al. 2013; Pearce 2015). The crust has been systematically thinned by more than 10 km below the Aegean Sea (Tirel et al. 2004; Pearce 2015) as well as below the Sea of Marmara (Bécel et al. 2009; Karabulut et al. 2013), the GSZ (Tiberi et al. 2001) and the WASZ (Karabulut et al. 2013). The shallowest Moho is found at a depth of 24 km below the NAT, 23-24 km beneath the Sea of Crete (Tirel et al. 2004) and 23-24 km below the GSZ (Tiberi et al. 2001). In the latter case, Tiberi et al. (2001) pointed out that

the highest Moho point is offset with respect to the Gulf of Corinth. This offset suggested to them that the lower crust is affected by boudinage. Below the WASZ, Karabulut et al. (2013) observed long-wavelength variations of Moho depth, from about 25 km below the Sea of Marmara (in agreement with Bécel et al. 2009) to about 32 km beneath the İzmir–Ankara suture and to about 25 km beneath the Menderes Massif. This long-wavelength variation suggested to these authors that viscous flow in a hot lower crust has smoothed out the lateral variations of crustal thickness, as mentioned earlier.

This brief review indicates that the overall crustal thinning in Aegea –Western Anatolia has proceeded locally to maxima that lifts the Moho to a depth of about 25 km, below the Gulf of Corinth, northern Aegea, the Sea of Crete, the Sea of Marmara and the WASZ. This value of 25 km is too prevalent to be purely coincidental. It affects regions where thinning started at widely diverse ages between 20 and 2 Ma. We suggest that this depth of about 25 km of Moho corresponds to the stage of thinning at which the gravitational potential energy of the Aegean continental lithosphere, with respect of the potential energy of the adjacent Ionian oceanic lithosphere, reaches its minimum value (see discussion for example in Le Pichon 1982). Thus, from that stage on, the thinning can only propagate laterally outward. These results suggest furthermore that the lower crust has a long-term viscous mechanical behavior, with boudinage deformation, during the thinning process, as mentioned earlier.

These important remarks will guide us in the following discussion of the thinning process below the Sea of Marmara in which we will use a new Moho model developed by Kende et al. (in prep, see Annex). Kende et al. (in prep.) selected 30 km for the depth of Moho corresponding to an absence of thinning. This value of 30 km is the one given by

831 Karabulut et al. (2013). We preferred it to the 35 km given by B  cel et al. (2009) as they
832 have few constraints away from the offshore domain and as their 35 km value does not
833 seem compatible with the model of Karabulut et al. (2013) although we note that their
834 N-S 2D model spreads the thinning on a larger area than in our model. Our resulting
835 Moho depth model shown in figure 7 maps the Moho depth below the whole Sea of
836 Marmara. The Moho is shallowest at two locations, one north of Marmara Island and the
837 other southwest of the   ınarcık Basin, with a smooth variation in between. The largest
838 amount of crustal thinning there is about 10 km. The western zone of highest thinning is
839 narrower and impinges on the southern shelf whereas the eastern one is wider and
840 occupies the whole width of the   ınarcık Basin. We note further that the thinning under
841 the Sea of Marmara is asymmetric, steeper to the north and more gradual to the south.
842 Below the MMF, the Moho depth is remarkably constant near 26 km (see figure 10f that
843 will be discussed later). It deepens abruptly to more than 29 km as the fault leaves the
844 Sea of Marmara, both to the east and the west. The maximum thinning is situated about
845 10 km to the south of the MMF in the western Marmara Moho high and about 15 km to
846 the south of the MMF in the eastern   ınarcık Moho high. In between, near the profile
847 shown in figure 5, the maximum thinning coincides with the MMF. The asymmetry is
848 clearly seen on the profile of figure 5 and appears to reflect the asymmetry of the
849 decollement mapped by B  cel et al. (2009). This suggests that the development of the
850 deep crustal structure of the Sea of Marmara has been controlled by the two initial
851 Al  tepe southern rifts. This control can also be seen in the dual nature of the zone of
852 maximum thinning, which reflects the emplacements of the two Al  tepe rifts. The
853 eastern   ınarcık Moho high corresponds to the paleo-  mralı Basin whereas the western
854 Marmara Moho high corresponds to the paleo-Central Basin.

855

However, it is also striking that there is no one to one correlation between the upper crust structure and the lower crust and mantle one. For example, the zone of maximum thinning to the west extends below the island of Marmara. In addition, although one should be aware of the imperfect resolution of the Moho mapping that was used, the Moho topography is quite smooth, which suggests that the lower crust is indeed governed by a ductile rheology that allows it to flow on a geological time scale. We noted earlier that Karabulut et al. (2013) had used the smoothness of the Moho topography in Western Anatolia as an argument in favor of the flow of the lower crust. This change of rheology between upper and lower crust is also suggested in figure 5 by the confinement of the seismicity of the MMF to the upper crust. Thus the upper crust there appears to coincide with the brittle crust. We conclude that our investigation agrees with our previous discussion on the ductile nature of the lower crust and upper mantle over the Aegean-West-Anatolian area. Although the structure of the Moho topography below the Sea of Marmara appears to have been in great part controlled by the formation of the two Alçitepe extensional grabens, its evolution appears to be partly decoupled from the evolution of the upper crust. In addition, it must have been influenced by the later geological strike-slip phase, with the migration of the strike-slip fault northward to the present emplacement of the MMF and the formation of the more recent Tekirdağ and Çınarcık basins.

In figure 8 we compare the “Basement” topography after Bayrakci et al. (2013) to the topography of the Moho shown in figure 7. The “Basement” of Bayrakci et al. (2013) does not coincide everywhere with the crystalline basement. Basement in figure 8 is assumed to have a seismic velocity larger than 4.2 km/s. This is because they find a velocity step between 3.9 km/s and 4.5 km/s. It can be considered to be the layer on

which occurred the onset of sedimentation in the newly formed North Marmara Trough. The difference between the topography of the basement and the topography of the Moho is striking. Actually, at first sight, one does not see any significant correlation between the two topographies. However, a closer examination shows that the two E-W elongated deep basement troughs that are the main characteristic of the topography of the basement appear to be controlled by the two Alçitepe grabens. The eastern Çınarcık basement trough corresponds to the paleo İmralı-Basin whereas the western Tekirdağ-Central Basin basement trough is closely correlated to the paleo-Central Basin. But at the difference of the two Moho highs, the linear structure of the troughs follows the direction of the present MMF. The MMF lies at the axis of the western trough, whereas the MMF, which follows the base of the continental slope in the Çınarcık Basin, is parallel but offset 10 km to the north of the eastern trough.

It is remarkable that it would be impossible to predict the present topography characterized by three main basins, Çınarcık, Central and Tekirdağ, separated by two highs, Central and Western, on the basis of the topography of the basement! That one cannot predict the topography in such an area from the topography of the Moho is not surprising, in view of the decoupling effect of the ductile lower crust. But nobody, to our knowledge, expected the absence of correlation between the topography of the basement and the sea floor topography. This indicates that the two highs are structures that have no roots in the basement. Both highs thus are the products of sedimentary cover tectonics decoupled from the basement, most probably triggered by the strike-slip motion along the MMF. However, this remark needs to be qualified by the fact that the Central High is a zone where the crust is somewhat thicker than either east and west.

The evolution of the NAF in the Sea of Marmara from Lower Pliocene to Present

Age of onset of sedimentation in the newly formed North Marmara Trough

To better understand the propagation of the NAF in the Sea of Marmara after the formation the Alçıtepe dual-basins structure, we now try to evaluate our present knowledge of the age of the onset of sedimentation within the different parts of the Sea of Marmara. A synthetic map is presented in figure 9. This map reflects our still considerable uncertainties. We have chosen to classify the ages in four categories.

The oldest basins, between 4.5 and 3.5 Ma are those that were formed during the Alçıtepe extensional stage discussed above. To the west, an onset age of 5 to 3.5 Ma in the South Central area was proposed on the basis of a heat flow model by Grall et al. (2012). Although this evaluation is fragile, it agrees with the 4.5-3.5 Ma that we proposed above for the basin produced during Alçıtepe by extension along the Marmara decollement mapped by Bécel et al. (2009). We thus adopt an age of 4.5 to 3.5 Ma. This zone most probably extends further west to the western extremity of the Marmara anticline in the southern Tekirdağ Basin. To the east, in the North İmralı Basin, Sorlien et al. (2012) show a profile along a line north of the present depocenter in their figure 13. If one applies a constant rate of sedimentation and takes into account the effect of compaction, the basin in this zone should be at least 2 Ma. It is definitely older than the Çınarcık Basin to the north or the Tekirdağ Basin to the west. Having no stratigraphic constraints on the deeper horizons, we consider reasonable to assume that the age is the one predicted by the formation of the İmralı decollement and anticline between 4.5 and 3.5 Ma.

930 *The intermediate 2.5 to 1.5 basins* might correspond to the initiation of shearing along
931 the MMF. A maximum age of 1.5 to 2.5 Ma has been proposed by Grall et al. (2012) for
932 the Central Basinette. They show that if one assumes that the subsidence rates have
933 been constant since the beginning of the formation of this basin, the age cannot exceed
934 2.5 Ma. But although the rate of subsidence has been constant since 250-300 ka, a more
935 rapid subsidence may have been present earlier, which would lead to a younger age. The
936 heat flow model of Grall et al. (2012) confirms that this basin is younger than the basin
937 in the South Central area. We assume for the Western High a similar age in view of the
938 great thickness of sediment (Şengör et al. 2014; Bayrakci et al. 2013). This zone might
939 extend all the way to westernmost Çınarcık (see discussion below)

940 *The youngest 1.5 to 1 Ma basins* correspond the Tekirdağ and Çınarcık basins. Seeber et
941 al. (2004) proposed that the Tekirdağ Basin was created by oblique slip. They obtain a
942 maximum age of 1.4 Ma with the parameters used in their model. We believe that if one
943 takes into account the compaction, the maximum age may not exceed 1 Ma. Sorlien et al.
944 (2012) and Kurt et al. (2013) used the same model to predict an age increasing from
945 east to west in the Çınarcık Basin with less than 1 Ma for the easternmost Çınarcık and
946 up to 1.4 in the central portion. We extrapolate an age of 1.5 to 2.5 Ma for westernmost
947 part of Çınarcık. Note that the young age of the Çınarcık Basin suggests that the İzmit-
948 Sakarya segment of the NAF that now enters the İzmit Bay is definitely younger than the
949 İzmit fault that enters the Gemlik Bay to the south.

950 *Figure 9 then is compatible with the existence of a southern Marmara basin formed*
951 *between 4.5 and 3.5 Ma and a more complex composite younger northern Marmara basin*
952 *that propagated outward to the east and to the west sometime after 2.5 Ma along the*
953 *newly formed SMF.*

Subsidence along the MMF since 2.5 Ma

In the following, we explore the possibility that since 2.5 Ma the northern Marmara basins along the present MMF developed through extension accommodated mostly by oblique slip on the transform fault itself (Seeber et al. 2004, 2006; Sorlien et al. 2012; Kurt et al. 2013). Such a model was originally proposed by Seeber et al. (2004) to explain the asymmetry and time-transgressive behavior characteristic of basins, such as the Tekirdağ basin, that form on the releasing sides of bends on strike-slip faults. The authors cited above have shown that this type of model is quite successful to account for the formation of the basins of the Sea of Marmara that are directly linked to the MMF, in particular the Tekirdağ at its western extremity and the Çınarcık at its eastern extremity. We apply the model through its simplest form. We assume that the side of the fault that is not subsiding is fixed and that the basement dip is parallel to the slip vector. Then the plunge (angle of slip motion with respect to the horizontal) of the slip vector along the main fault is the arc tangent of the subsidence velocity divided by the horizontal velocity. It is thus directly related to the subsidence rate through the rate of horizontal slip. This rather crude model is only used to test the general viability of a process of formation of the basins through extension mostly accommodated by oblique slip as discussed above.

To derive the plunge values presented in figure 10c we assumed a horizontal slip along the MMF of 18 ± 2 mm/yr (Kurt et al. 2013; Grall et al. 2013) in agreement with the general discussion we had earlier on the geological slip rates along the NAF. The subsidence rates (figure 10b) were derived from various data discussed below.

Çınarcık: Kurt et al. (2013) presented a stratigraphic interpretation of the TAMAM dense grid of multichannel seismic profiles. The subsidence rates were derived from the difference between the depths of the green-H5 horizon, a seismic reflector described by Sorlien et al (2012) which age is estimated to be 540 - 630 ka [Sorlien et al. 2012; Grall et al. 2013], on each side of the fault. From the east, the subsidence rate decreases, starting at 5.5 - 4.7 mm/yr (Figures 3 and 4 in Kurt et al. 2013) to about 3.7 - 3.2 mm/yr in the middle part of the basin and 1.8 - 1.5 mm/yr to the west (Figure 3 in Kurt et al. 2013), giving the corresponding plunge values of 17.1°, 11.6° and 5.9°. The 17.1° plunge value also corresponds to the dip of the basement along the fault at the eastern corner of the basin.

Kumburgaz: From the same TAMAM survey, Sorlien et al. (2012) presents a profile cutting the Kumburgaz Basin on its western part. Following the same method, a subsidence rate of 1.5 - 1.3 mm/yr and a 4.7° plunge is derived from the green-6 (another seismic reflector described by Sorlien et al. 2012) depth variation of 0.8 km across the fault (Figure 11 in Sorlien et al. 2012).

Central Basin: Grall et al. (2012) presented two models of a profile cutting the basin: a steady-state model based on subsidence rate extrapolation and a model based on seismic line interpretation. We used the depth variation across the fault of the 2 Myr level in the first model and of the 1.5 Myr level on the second model as they are presented in their supplementary material [Grall et al. 2012]. Both models yield the same subsidence rate of 2 mm/yr and a plunge of 6.3°.

Western High: Grall et al. (2013) indicate a sedimentation rate over the last 39 kyr that reaches a maximum of 1.5 mm/yr in the eastern and western small basins pinching in the Western High (Figure 6 in Grall et al. 2013). Correcting for the compaction of shallow sediment of 65-70% porosity, subsidence at crustal level at these sites should account for about one third of the seafloor subsidence rate. This gives a subsidence rate of 0.5 mm/yr corresponding to a plunge of less than 2° over the Western High.

Tekirdağ: The plunge values of 8° and 14° for, respectively, the Tekirdağ Basin and the Ganos small basin off Ganos are inferred by Seeber et al. (2004) assuming that the Tekirdağ Basin geometry is due to a stationary onset point, relatively to the south side of the fault. Thus, the 8° plunge of the basin floor along the fault must correspond to the fault plunge. For the Ganos small basin, Seeber et al. (2004) observed that the absence of horizontal offset where the fault enters the basin could be explained by a displacement plunge equal to the basin rim slope of 14°. Note that because the Tekirdağ Basin is forming northward of the fault, the plunge values are considered as negative, as for the Central Basin.

Moreover, we considered a relative subsidence equal to zero on average across the Western High and on the Central High on either side of the Kumburgaz Basin. The uncertainties were derived from values given in the cited studies, when available. Consequently, most of them were calculated based on uncertainties on age models and on horizontal slip rate (15 to 20 mm/yr) but do not take into account the uncertainty in the relative vertical displacement.

The main interest of figures 10a and 10b is to illustrate the clearly anomalous nature of the western Tekirdağ Basin and of the eastern Çınarcık Basin with their fast outward propagation rate. There, the model proposed by Seeber et al. (2004) appears to account in an elegant way for the evolution of these basins in a pure strike-slip mode. In between the rates are relatively modest and probably can be accounted for by simple additional N/S distension which is known to be present because of its seismic signature, especially to the south of the western and central part of the Sea of Marmara (Altınok and Alpar 2006; Karabulut personal communication, 2015). We proposed earlier that in the western and central Sea of Marmara, the MMF coexists with remnants of an extensional system. Finally, as noted earlier, to the east, the active faulting and subsidence in the İmralı Basin area may also involve continued activity on early structures formed under extension.

Figure 10e shows the sediment thickness along the MMF after Bayrakci et al. (2013) and confirms the very large thickness of sediment below the Western High, up to more than 6 km. We discussed this point earlier. Figure 10f gives the depth of the Moho from the inversion of Kende et al. (in preparation) shown in figure 7. In figure 10d the crustal thickness is obtained by subtraction of the bathymetry and sediment thickness from the Moho depth. The constancy of the Moho depth along the MMF at about 26.5 km is in contrast with the relatively large variation in the crustal thickness. This contrast again testifies that the lower crust acts as a decoupling level in the structuration of the Marmara area. We can also note that the overall thinning of the crust is relatively modest.

To conclude, the evidence we have discussed is compatible with a two-stage formation

of the Sea of Marmara. First, two left-laterally offset basins, a paleo-Central basin to the west and a paleo-İmralı Basin to the east, were formed between 4.5 and 3.5 Ma, in a purely extensional context, in front of the advancing NAF. They now correspond to the oldest part of the present Sea of Marmara. The formation of the SMF on the southern shelf marked the end of the initial extensional phase and the beginning of the second strike-slip phase. During this second phase, the strike-slip motion migrated northward and established itself along the present SMF. Sometime after 2.5 Ma, the northern basins were formed along the present MMF through extension accommodated mostly by oblique slip on the transform fault itself (Seeber et al. 2004, 2006; Sorlien et al. 2012; Kurt et al. 2013). However the transition between the two stages has most probably been gradual because extension is still present today, especially in the southern part of the western and central Sea of Marmara as well as within the İmralı Basin.

From extension to strike-slip

The evidence we have presented demonstrates, in our opinion, that the formation of the Marmara deep basin began about 4.5 Ma during a 1 Myr purely extensional phase, similar to the one producing today the Corinth rift. Furthermore, the evidence we discussed above indicates that the proposition made by Le Pichon, Taymaz and Şengör (2000) that the present MMF is a continuous dextral strike-slip fault which transfers the slip of the eastern NAF from the Gulf of İzmit to the Ganos fault with a geological slip rate of 18 mm/yr is correct. The question then is when did the tectonic change from extension to strike-slip occur and how much finite motion corresponded to each of these tectonic phases. This question does not have simple definitive answers today.

To approach this problem we may start from the total slip on the eastern branch of the

NAF, which is estimated today at about 75 ± 10 km (Şengör et al. 2005). This is equivalent to 4.1 Myr of slip at the present geological slip rate of 18 mm/yr, much smaller than the age of initiation of about 11 Ma. This led Şengör et al. (2005) to assume a slip rate progressively increasing from zero 11 Ma to the present rate during the last Myr. With their model, the total finite offset at the beginning of the purely extensional phase in the Marmara area between 4.5 and 3.5 Ma is slightly more than 20 km and the geological rate of slip at that time about 12 mm/yr. The modelization of Provost et al. (2003) has demonstrated that the present NAF implies an extremely low friction coefficient of less than 0.05 on the fault plane, fifteen times smaller than normal. Such a low friction coefficient indicates that the principal displacement zone is well developed and has reached its maturity through the effect of regularization by many successive earthquakes, which would have produced a total offset of at least 20 km (e.g. Chester et al. 1993). 20 km is probably the minimum amount of offset one may expect at this time for the NAF to produce on its tip a well developed extensional shear zone. We accept this 20 km estimate of Şengör et al. (2005) for the offset at 4 Ma and the corresponding probable rate of slip of 12 mm/yr at this time.

Note that this conclusion has an important consequence for the tectonics of Aegea from Upper Miocene to Lower Pliocene. This is because, as a consequence, during this period, the subduction rate must have been significantly less than today, probably not more than 20 mm/yr as an average, with a progressive increase between 15-12 Ma and 4 Ma, compared to the present subduction rate of 34 mm/yr.

We now move to the recent period of strike-slip. The formation of the Tekirdağ and Çınarcık basins by the present MMF occurred at approximately the present 18 mm/yr

geological slip rate since about 1.5 Ma. This indicates that about 27 km of total slip occurred since 1.5 Ma on the MMF. We are left with approximately 30 km of total slip for the period between 3.5 and 1.5 Ma. This gives an average rate of about 15 mm/yr, which appears reasonable in view of the many uncertainties. The question then is how this approximate amount of 30 km of total slip was distributed? What was the part of the emerging MMF between 2.5 and 1.5 Ma, the part of possible intermediate systems of strike-slip between the SMF and the MMF and the part absorbed in the still existing extensional system? We do not have the information necessary to answer this. We can only state with certainty that the present strike-slip system along the MMF was fully functional at the present rate since 1.5 Ma.

Le Pichon et al. (2001) and Le Pichon et al. (2003) had been impressed by the observation that the MMF cuts across older existing structures in the Sea of Marmara. They concluded that the MMF appeared in a late stage of evolution of the Sea of Marmara. Noting that back-slipping the MMF by 4 km appears to restore the Central Bassinette to a symmetric shape, Le Pichon et al. (2001) concluded the MMF appeared only about 200 ka. This estimate appeared to be reinforced by the discovery by Armijo et al. (2002) of an offset of 3.5 km of a N/S fold on the Central High. This conclusion was further elaborated by Rangin et al. (2004) and Şengör et al. (2005). It is clear now that this interpretation was not correct and that the event that affected the MMF about 200 ka was probably a readjustment of the MMF (Grall et al. 2012). But this does not change the observation that the MMF cut across the Marmara basin more than 2 Ma after its initial extensional formation and that the structures related to it, such as the two elongated basement troughs mapped by Bayrakci et al. (2013), are superimposed on an older principally extensional framework.

1127

1128 ***Additional discussion of the geodynamics***

1129

1130 We pointed out throughout this paper that the lithospheric mantle of the Aegea-Anatolia
1131 plate appears to be very thin or even absent and thus that the plate corresponds to a
1132 huge ductile mass. This is because the Curie Point isotherm (580°C) south of the NAF is
1133 nearly everywhere at a depth of less than 20km (Aydın 2005), which puts the
1134 temperature at more than 900°C at the Moho. This Curie Point isotherm rises to less
1135 than 10 km in southwestern Anatolia! McKenzie et al. (2005) have shown that, in
1136 continental as well as oceanic lithosphere, the brittle seismogenic zone is limited at its
1137 base by the 600°C isotherm. And numerous studies have confirmed that the part of the
1138 mantle that is above 600°C does not contribute significantly to the elasticity of the plate.
1139 As a consequence, the mantle in most of Anatolia and Aegea, south of the NAF, has
1140 negligible mechanical resistance. Further, because the temperature within the lower
1141 crust exceeds 750°C, many models predict that channel flow is possible if the
1142 composition of the crust is not granulitic (e.g. Beaumont et al. 2004). In addition,
1143 because partial fusion is prevalent in many places within the lower crust, transient
1144 channel flow would appear as soon as the rate of partial fusion is sufficiently high
1145 (McKenzie and Jackson 2002). We can conclude that the mechanical resistance of the
1146 Aegean-Anatolian lithosphere lies essentially within its upper brittle crust and that
1147 everything happens as if the lower crust acts as a decoupling level. The existence of a
1148 decoupling level, somewhere within the lower crust or at the Moho, has been an
1149 important acquisition of modern tectonics (e.g. Wernicke et al. 2008). The presence of a
1150 decoupling level below the upper crust explains the observation we made earlier that
1151 the NAF appears not to be able to stabilize unless the whole crust is in the brittle

domain. If there is possible flow in the lower crust, the upper part of the fault is disconnected from its roots and rapidly loses its "memory".

If the Aegean-Anatolian lithosphere is mechanically so weak, then how is it possible that traction at its southwestern extremity can be transmitted over 1500 km to its eastern extremity? The answer lies in the role played by the surface relief. The sometime exclusive interest given to the role of the detachments of the slabs in the last few years has led to a neglect of the very important role played by the evolution of topography, with the notable exception of Floyd et al. (2010) for the Aegean-West Anatolian area.

McKenzie (1972) and Le Pichon (1982) had both proposed that the westward movement of the Anatolian lithosphere responded to a gradient in gravitational energy. Currently the elevation of this plateau is about 2 km and the elevation regularly decreases westward to below sea level within the Aegean Sea (see figure 11). The existence of this westward slope of 1.5/1000 results in a gradient of potential gravitational energy that enables the transmission of boundary forces to occur smoothly over the whole length of the Anatolia-Aegea plate (see Le Pichon and Kreemer 2010 for a more complete discussion of these dynamics). And this gradient explains how the southwestward trench pull upon the Aegean lithosphere can lead to a westward traction of Anatolia. But when and how was this gradient established?

Let us first consider when this gradient was established? There is a paradox there. The last marine sediments deposited above the present East Anatolian Plateau date from Serravallian, 13-11 Ma (Şengör et al. 2003). Thus the uplift started at this time, as indicated by the volcanic activity, that was initiated 11 Ma and became plateau-wide at 8 Ma. Şengör et al. (2003) reasonably attributed part of the uplift of the East Anatolian

1177 Plateau to slab break-off, arguing that the asthenosphere appeared to lie immediately
1178 below the crust, as evidenced by its anomalously low seismic velocity. The existence of a
1179 slab break-off was also proposed by Faccenna et al. (2006) and by many others since, as
1180 tomographic studies have demonstrated that the plunging African slab below Turkey is
1181 detached from the African plate. But, during the same Serravallian time, most of what
1182 now forms the central and southern Aegean Sea was rapidly subsiding. Drooger and
1183 Meulenkamp (1973) have described the spectacular fragmentation of the southern
1184 Aegean landmass that passed below sea level about 13 Ma. Yet, the Aegean Sea crust is
1185 also underlain by hot asthenosphere (Salaün et al. 2012, Paul et al. 2014), as below the
1186 East Anatolian Plateau. As discussed earlier, the subsidence there, in spite of the
1187 presence of hot mantle, results from the fact that, during middle to late Miocene, the
1188 crust of the Aegean Sea was thinned by more than 10 km.

1189
1190 Since the work of Aubouin et al. (1976), it has been demonstrated that after the
1191 Paleogene phase of nappe stacking, the central Aegean had been the site of intense
1192 crustal extension since the beginning of Miocene. This extension had formed core
1193 complexes through uplift of the lower middle continental crust to the surface during the
1194 middle to late Miocene (Lister et al. 1984). It seems reasonable to assume that the
1195 present location of the core complexes were in late Oligocene to early Miocene the sites
1196 of maximum crustal thickness and highest topography, and that they became the sites of
1197 maximum crustal extension (Le Pichon et al. 2002). We have argued earlier that this
1198 occurred because the Aegean portion of the slab broke, sometime in late Oligocene,
1199 about 10 Myr earlier than below the East Anatolia Plateau. Consequently, at this time,
1200 the asthenosphere rose to come in contact with the lower crust at the Moho. As its
1201 southern boundary was a free boundary, able to migrate southward, the crust reacted by

1202 southward extension, whereas, 10 Myr later, below the East Anatolia Plateau, no
1203 extension was possible because of the collision to the south at the Bitlis suture: the only
1204 possibility was uplift. We thus had a situation in early Miocene in which the Aegean and
1205 westernmost Anatolia were relatively high whereas the Eastern Anatolia Plateau was
1206 close to sea level, the exact reverse of what it is now.

1207
1208 We now try to consider how the gradient was established. The topographic and crustal
1209 thickness profiles in figure 11 indicate that there is a linear relationship between
1210 topographic elevation h and crustal thickness T : $h = (T - 27)/9.5$ in km. The crustal
1211 thickness for sea level is 27 km and for an altitude of 2 km, on the plateau, 46 km. This is
1212 to be compared to the relationship for Tibet, $h = (T - 35)/7$ (Le Pichon et al. 1992). The
1213 linear relationship demonstrates that a simple Airy type isostasy prevails there. The fact
1214 that, to produce 1 km of additional elevation, one has 9.5 km of additional crustal
1215 thickness in Anatolia-Aegea instead of 7 km in Tibet indicates a lower density contrast at
1216 the Moho for Anatolia-Aegea than for Tibet. For example, using the values of upper crust
1217 density of 2.67 and of lower crust density of 2.95, as adopted by Kende et al. (in
1218 preparation), the density of the mantle would be 3.26 for Anatolia-Aegea and 3.39 for
1219 Tibet. Of course, this model is very crude but it does confirm that the mantle below the
1220 Moho is unusually light, and consequently hot, below Anatolia-Aegea.

1221
1222 If the thickening to 45 km of the crust below the East Anatolia Plateau can be explained
1223 by crustal thickening north of the Arabian indenter, no such effect can be invoked to
1224 explain the variation in thickness of the crust in central and western Anatolia. The only
1225 process that we are able to invoke is lower crust flow from the East Anatolia Plateau
1226 toward the west. Such a flow should actually be expected in view of the large amount

of magmatism since 8 Ma. This magmatism would have greatly decreased the viscosity of the middle-lower crust (McKenzie and Jackson 2002). Actually, this tectonic situation can be compared with what is now happening on the eastern border of the Tibetan Plateau. This eastern border has been uplifted, without substantial shortening of the upper crust, since 8 Ma, because of the occurrence of east-southeastward flow of the middle-lower crust (e.g. Royden et al. 2008). We conclude that an eastward flow of the lower crust of the East Anatolia Plateau has led to the progressive uplift of Anatolia from east to west.

But was the mantle involved in this westward flow? We have argued that a level of decoupling exists in Anatolia-Aegea within the lower crust. Consequently, the mantle may not necessarily take part in this flow. However, the asthenosphere that had risen up below the East Anatolia Plateau had to escape outward, being pushed by the advancing Arabian indenter and the only possibility of escape for the asthenosphere was to the southwest, toward the Mediterranean, between the Arabian indenter to the east and the Aegean slab to the west. Do we have any indication that such a flow has existed? We can again consider the Tibetan area where the orientation of the fast-wave polarization axes of shear waves aligns itself with directions of principal shear indicating a flow in the mantle, in the same direction as the surface one, with deformation rates exceeding 75% (Holt 2000). On the southeastern border of the Tibetan Plateau, however, Flesch et al. (2005) conclude that the mantle flow, as indicated by these polarization axes, is probably decoupled from the crust. The eastward flow is actually close to the extensional direction, at an angle with the southeastward surface flow, suggesting that deformation rates are smaller there.

In figure 11, we show the NE/SW orientations of the fast-wave polarization axes of shear waves detected by Paul et al. (2014) over most of Anatolia and the northern Aegean Sea. These are remarkably similar to the extensional directions from GPS (e.g. fig. 7 of Le Pichon et al. 2010) or from earthquake fault plane solutions (e.g. Floyd et al. 2010), including the same counterclockwise rotation from eastern Anatolia to the northern Aegean Sea noted by Paul et al. (2014). Thus the upper mantle appears to be decoupled from the crust, flowing along the lithospheric extensional direction, toward the easternmost Mediterranean. This direction of flow is not surprising, as the new Aegean slab has blocked the possible outflow of the asthenosphere in the Aegean-West Anatolian area in the last few million years. Finally, this discussion helps us to understand why, although the westward extrusion of the Anatolia-Aegean plate was initiated 11 Ma, the formation and westward propagation of the NAF took a few million more years to develop, as the westward gravitational gradient was being established.

Conclusions

The Sea of Marmara appears to be a key point in the history of the propagation of the NAF toward the northern extremity of the Aegean subduction during the last 10 Myr. There is no indication that a localized plate boundary existed north of the Aegean portion of the Anatolia plate before 2 Ma. Prior to this age, the shear produced by the motion of Anatolia with respect to Eurasia was distributed over the whole width of the its Aegean-West Anatolian western portion. This was most probably related to the presence there of a N/S component of extension produced by the southward migration of the Aegean subduction. It is no coincidence that fast subduction of the oceanic Ionian

lithosphere was initiated a few million years before the Anatolia westward motion. These two processes are obviously tightly linked.

The beginning of the formation of a localized plate boundary occurred between 4.5 and 3.5 Ma at the location of the present Sea of Marmara by the initiation of a shear zone comparable to the present Grecian Shear Zone in Central Greece. Thus, the first part of the formation of the Sea of Marmara was purely extensional. The present strike-slip system that today cuts across the whole Sea of Marmara and that is called MMF, began to develop after 2.5 Ma. Shortly after, the plate boundary migrated westward along the northern border of Aegea from the NAT, to the GSZ and to the Kefalonia fault. There, it finally linked with the northern tip of the Aegean subduction zone, completing the system of plate boundaries delimiting the Anatolia-Aegea plate. We have related the remarkable change in the distribution of shear over Aegea from Miocene to Pliocene to the formation of a relatively undeforming block in Pliocene that forced the shear to be distributed over a narrow plate boundary to the north of it. We attribute the formation of this relatively undeforming Aegean block to the northeastward progression of the cold oceanic Ionian slab. This is because the slab cuts the overlying lithosphere from asthenospheric sources and induces a shortening environment over it. We noted however that another possible contribution was the lower subduction rate during the Miocene.

Finally, we wish to make three remarks that may have significant implications for the evaluation of the seismic risk in the Sea of Marmara:

1) The present structure of the Sea of Marmara results from the superposition of a purely strike-slip system on top of an initial purely en-echelon extensional system. This superposition explains the occurrence of large strike-slip earthquakes together with significant extensional earthquakes that reactivate earlier extensional structures.

2) The MMF is a continuous dextral strike-slip fault through the whole basin with a present seismic activity strikingly different from the portions of land fault broken by the 1999 Kocaeli and the 1912 Ganos earthquake. Its continuity and relative homogeneity suggests that one cannot discard the possibility that the whole MMF will break in a single earthquake as proposed by Le Pichon et al. (1999).

3) The MMF closely follows the base of the northern Marmara margin east of 28°E. As a result there is a 4-6 km offset between the basement on both sides of the fault. Jim Rice (personal communication to XLP, 2002) pointed out that this would induce a significant asymmetry of elastic deformation. This asymmetry has very important implications that have essentially been ignored up to this day.

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1326 for many insightful suggestions.

1327

1328 **Annex**

1329 Moho model

1330 The model of Kende et al (in preparation) was obtained through a gravity inversion in
1331 the Fourier domain of the measured gravity corrected from the upper crust geology. The
1332 gravity and topography data used are satellite data [Smith et al. 1997] retrieved from
1333 the Scripps Institution of Oceanography website. Gravity data were refined, when
1334 possible, with the records from the Bodenseewerk KSS 31 gravimeter during the Marsite
1335 Cruise on board the NO “Pourquoi Pas?” in October and November 2014. To avoid
1336 superficial crustal effects, the topography was filtered with a 20 km window Gaussian
1337 filter before being used to apply a Bouguer correction onshore. Then, the offshore
1338 gravity was corrected from the known basin geology. The two sediment layers basin
1339 model from Bayrakci et al. (2013) was used, completed on the southern shelf with TPAO
1340 multichannel seismic line interpretations. The correction was made in the Fourier
1341 domain using the Parker method [Parker 1972]. A first Moho depth variation was
1342 computed by inverting the residual gravity anomaly and a Lower Crust/Upper Crust
1343 boundary (LC/UC) was defined 9 km above this depth model. An iterative process was
1344 then launched. It used the sediment basin model and the newly defined LC/UC to correct
1345 the measured gravity with the Parker method, obtaining a new gravity residual. The
1346 inversion of this new gravity residual led to a new Moho depth model. This process,
1347 repeated twice, enhanced the overall model significantly as it lowered the final residual
1348 anomaly.

1349

The method developed for the study of Kende et al. (in preparation) is close to the Parker-Oldenburg method [Oldenburg 1974] and is based on a least-square linear inversion with preconditioning [Tarantola 2005]. This method proved to be fast and effective for large scale 2D models. It also avoided part of the human bias associated with forward modeling [Aitken et al. 2013]. However, because this model relies on a simplified basin structure, the choice of the densities of the two sedimentary layers has a strong impact on the Moho depth variation amplitude. Data well constrain the Moho geometry only at spatial wavelengths longer than crustal thickness. To damp Moho oscillations at shorter wavelengths, Moho deviation from the mean regional value is preconditioned to zero and a third parameter (Moho variance scaling) entered by the user defines the weight of this preconditioning. After defining a range of possible mean densities for each of these parameters, the final values used for the results presented here were those that best agreed with B  cel et al. (2009) and Karabulut et al. (2013) Moho depth models.

B  cel et al. (2009) presented a Moho model based on a dense grid of Multi-channel Marine Seismic profiles with a high penetration depth retrieved during the SEISMARMARA Leg 1, completed with the data from on-land temporary seismometers and Ocean Bottom Seismometers (OBSs). Their study was designed so that the profiles crossed each other at OBS positions and had land stations in their continuity. Crossing profiles permitted, for instance, to identify side echoes that could appear as flat interface events on some profiles, while OBSs first arrival time of refracted waves on the basement were used to build a basement velocity and geometry model. Finally, on land station records of refraction on the upper mantle (Pn waves) were integrated, which was critical for the 2D model under the Northern Marmara Trough as only 2 OBSs

1375 clearly recorded refraction on the Moho. Bécel et al. (2009) concluded that there had
1376 been a 5 km crustal thinning below the Northern Marmara Trough and under the İmralı
1377 Basin as the Moho height was essentially constant at a depth of 26 km. Their study
1378 however put no constraint on the possible existence of small scale thinning variations.
1379 Because the North-South Moho depth variation obtained was close to the data
1380 resolution, they could not resolve conclusively the existence of possible N-S variations.

1381

1382 Karabulut et al. (2013) on the other hand published a N-S 650 km long profile of
1383 lithospheric structures. This profile crosses the Sea of Marmara in the eastern part of the
1384 Tekirdağ Basin. They relied on a set of 17 permanent broadband seismic stations and 23
1385 temporary stations installed as part of the SIMBAD project to fill the gaps and create a
1386 spacing between stations of about 15 km along the transect. After a year of recording,
1387 they selected 60 events with a magnitude superior or equal to 5.5 and a tolerable noise
1388 level. They then performed a depth migration of the P-to-S converted phases and
1389 interpreted the Moho as the level of sharp increase of amplitude undulating between 25
1390 and 32 km. Their station coverage is not tight enough at the level of the Marmara Sea to
1391 determine the detailed Moho depth variation but Karabulut et al. (2013) showed an
1392 uplift of 6 to 7 km centered under the sea with a width of less than 250 km. The highest
1393 Moho under the Sea of Marmara obtained in the two studies is the same, about 25 km.

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Draft

1928

1929 **Captions**

1930

1931 *Figure 1*

1932 Tectonic map of Aegea-Western Anatolia. Violet indicate shortening structures Yellow,
1933 normal faults Green, left-slip faults White, North-Anatolian dextral fault family including
1934 the normal faults of the Grecian Shear Zone (Şengör 1979). All faults are Quaternary,
1935 because they cut Quaternary deposits established by sequence correlations in the Sea of
1936 Marmara. Following Clarence Allen's rule (Allen 1975), we assume all are active or
1937 potentially so. The light colored presently undeforming Aegean block revealed by
1938 geodesy is limited by dashed lines. The dashed red contours of the Aegean slab are after
1939 Gudmundsson and Sambridge (1988). The two yellowish shear zones that limit it, the
1940 GSZ and NAT to the north with clockwise rotation and the WASZ to the east with
1941 counterclockwise rotation, absorb about 20-25 mm/yr of shear, dextral for the northern
1942 zone and sinistral for the eastern one. The figure is modified from Şengör and Zabcı (in
1943 press).

1944

1945 *Figure 2*

1946 Proposed tectonic map of Aegea-Western Anatolia about 4 Ma. White line, the westward
1947 propagating North-Anatolian dextral system has just reached the area of the present Sea
1948 of Marmara that is occupied by a system of clockwise rotating normal-fault bounded
1949 blocks similar to the GSZ, with extensional decollements in front the tip of the NAF.
1950 Modified from Şengör and Zabcı (in press)

1951

1952 *Figure 2.5*

1953 Index map of geographic and geologic features within the Sea of Marmara area
1954 discussed in the text.

1955

1956 *Figure 3a*

1957 Map of the southern shelf anticlinal system. Faults on the shelf after Le Pichon et al.
1958 (2014). Contours every 100 m. (light lines) and 500 m. (dark lines) are estimates of
1959 amount of eroded material based on the seismic profiles. See Le Pichon et al. (2014) for
1960 network of seismic profiles used. The 3.5 Ma South Marmara Fault (SMF) is shown in
1961 red. Possible x-shears related to the clockwise rotation of the blocks between 4 and 3.5
1962 Ma are shown as dotted lines, in purple. Red barbed lines are direction of propagation of
1963 clastics. Contours between the sea floor and 13 km depth of detachments north of the
1964 Marmara anticline after Bécel et al. (2009). Black lines with tick marks show crests of
1965 tilted blocks on top of these detachments after Bécel et al. (2009). The detachment north
1966 of the İmralı anticline is drawn on the basis of the SEISMARARA seismic profile in the
1967 İmralı Basin, shown in fig 2b of Laigle et al. 2008. The location of this profile is shown by
1968 the red line. See text. M1 is the Marmara 1 industrial well.

1969

1970 *Figure 3b*

1971 Enlargement of figure 3a.

1972

1973 *Figure 3c*

1974 Stacked profiles from East to West with erosional contours on top of it.

1975

1976 *Figure 4*

1977 Detail of contours of anticline around Marmara Island. Red line, SMF. Yellow barbed
1978 lines, direction of propagation of clastics. The top of the Upper Miocene Kirazlı
1979 formation rises abruptly across the normal fault to less than 250 m. near the island. The
1980 log of the Marmara 1 drill hole situated to the east of Marmara Island is given for
1981 reference. See Fig.3a and Le Pichon et al. (2014).

1982

1983 *Figure 5*

1984 Interpreted lithospheric SW-NE profile (see location on figure 7). Fault locations from Le
1985 Pichon et al. (2014) and Laigle et al. (2008), supra-crustal model based on Bayrakci et al.
1986 (2013), detachment geometry based on Bécél et al. (2010) and Moho depth from Kende
1987 et al.(in prep.) model. TPAO line interpretations were also consulted to constrain the
1988 fault and sediment deposition geometries on the Southern shelf. Violet faults represent
1989 the Main Marmara Fault system in the Central Basin. Black dots are micro seismicity
1990 localizations from Kandilli Observatory data. For comparison, we have shown at the
1991 same scale below the profile the sketch of Jolivet et al. (2010) that illustrates the
1992 possible future evolution of the Gulf of Corinth. Green, phyllite, quartzite. Grey, ductile
1993 domain.

1994

1995 *Figure 6*

1996 Distribution of the marine Alçıtepe Formation (uppermost Miocene–lower Pliocene, 5.3-
1997 3.7 to 3.4 Ma) in blue and the Pliocene undifferentiated continental deposits in yellow
1998 after the 1/500 000 map of the General Directorate of Mineral Research and Exploration
1999 (2002), Görür et al. (1997) and Sakıncı et al. (1999). Map of decollements and anticlines
2000 from figure 3a. Red pattern: deformation zone along the SMF.

2001

2002 *Figure 7*

2003 Moho topography in meters after Kende et al. (in preparation). See text. The thick black
2004 line is the MMF. Blue line is location of figure 5 profile. Detachments contours and block
2005 crests as in figure 3a.

2006

2007 *Figure 8*

2008 “Basement” topography after Bayrakci et al. (2013) superposed to Moho depth contours
2009 of figure 7 in meters. Basement is defined as having a seismic velocity larger than 4.2
2010 km/s and does not coincide everywhere with crystalline basement. Thick black line:
2011 MMF. Yellow lines: contours of detachments as in figure 7. Submarine fault scarps after
2012 Armijo et al. 2002 are represented in white. Black crosses are inverted nodes and red
2013 ones are fixed nodes in the inversion of Bayrakci et al. (2013).

2014

2015 *Figure 9*

2016 Synthesis of information on age of onset of sedimentation in the newly formed North
2017 Marmara Trough from Grall (2015, in preparation).

2018

2019 *Figure 10*

2020 Compilation of different data along a transect following the MMF. a) Map of the Sea of
2021 Marmara. MMF in red based on Şengör et al. (2014) and contour lines of the Moho depth
2022 from figure 7 b) Subsidence rate along the MMF retrieved from various studies (see
2023 details in this paper). Dotted line indicates a lack of information to determine the
2024 uncertainty. c) Fault plunge along the fault (see details in this paper). d), e) and f)
2025 Crustal thickness, sediment thickness and Moho depth from Bayrakci et al. (2013)
2026 basement model and Kende et al. (in prep) Moho model presented in figure 7.

2027

2028 *Figure 11*

2029 Topography of the Anatolia-Aegea area. Note the 2 000 meters altitude of the East
2030 Anatolia Plateau, opposite the Arabian indenter and the progressive decrease of the
2031 elevation westward to less than sea level below the Aegean Sea. The depth contours of
2032 the Aegean slab are shown in dashed black lines. The fast-wave polarization axes of
2033 shear waves detected by Paul et al. (2014) are also shown (red their measurements,
2034 blue previous measurements). Note the progressive counterclockwise rotation of the
2035 axes as one gets closer to the Aegean slab. The E/W profiles along 39.5°N show the
2036 smoothed topography in blue and the crustal thickness in red after Karabulut et al.
2037 (2015).

Draft

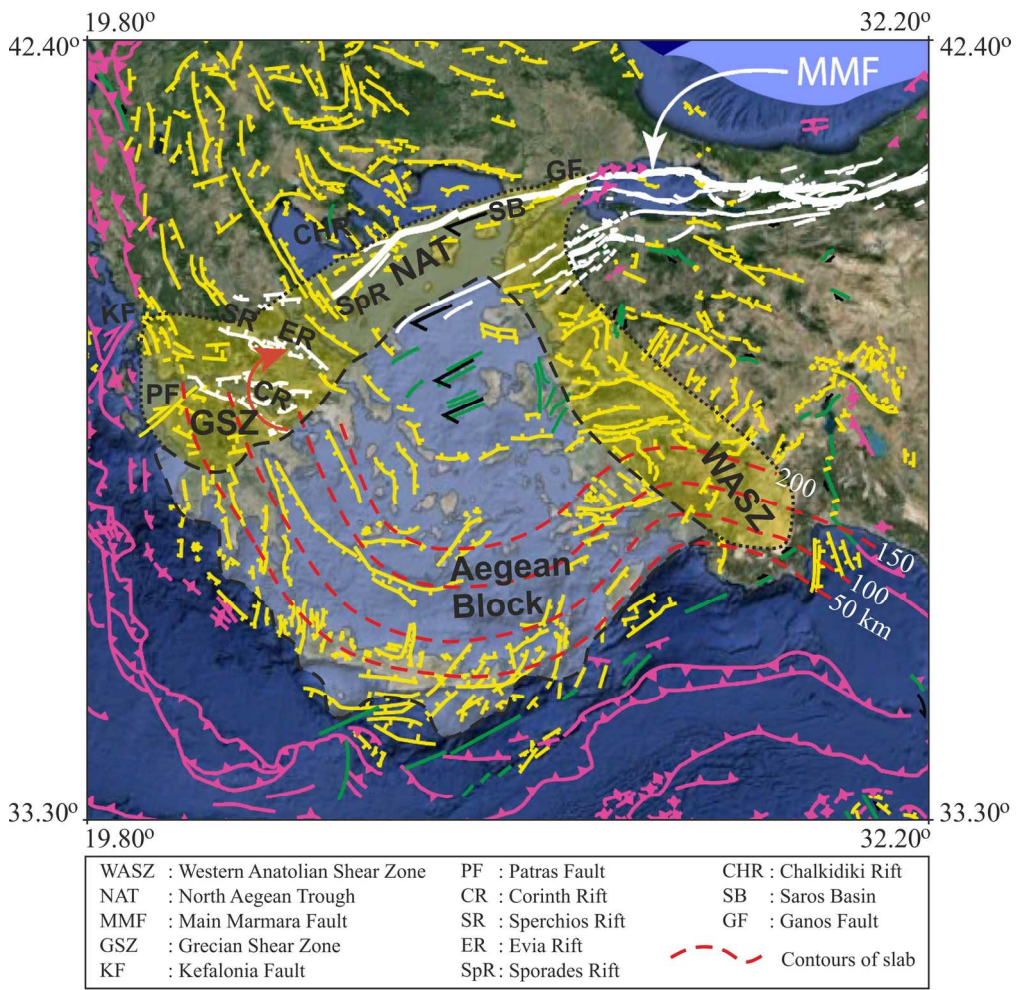
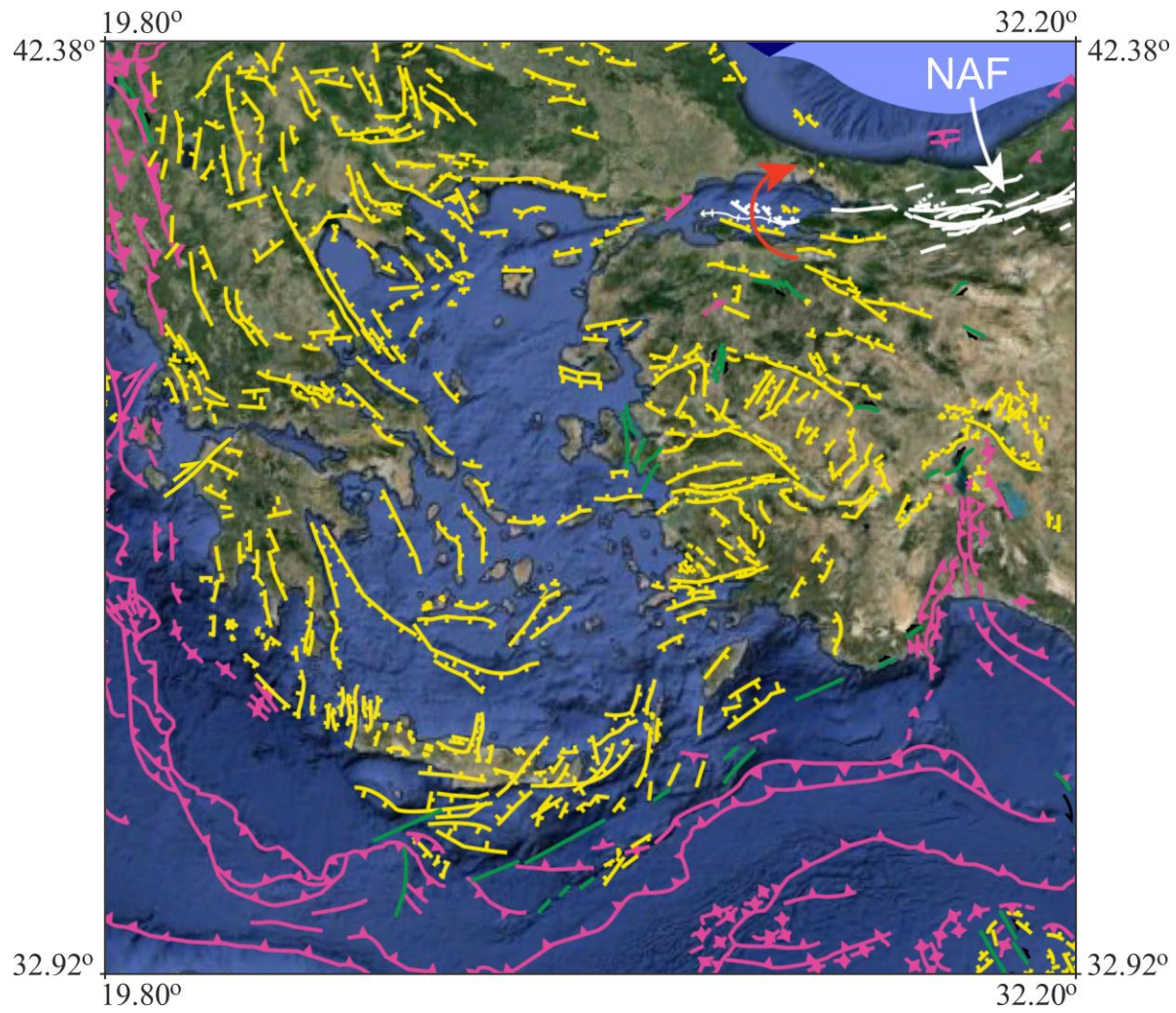
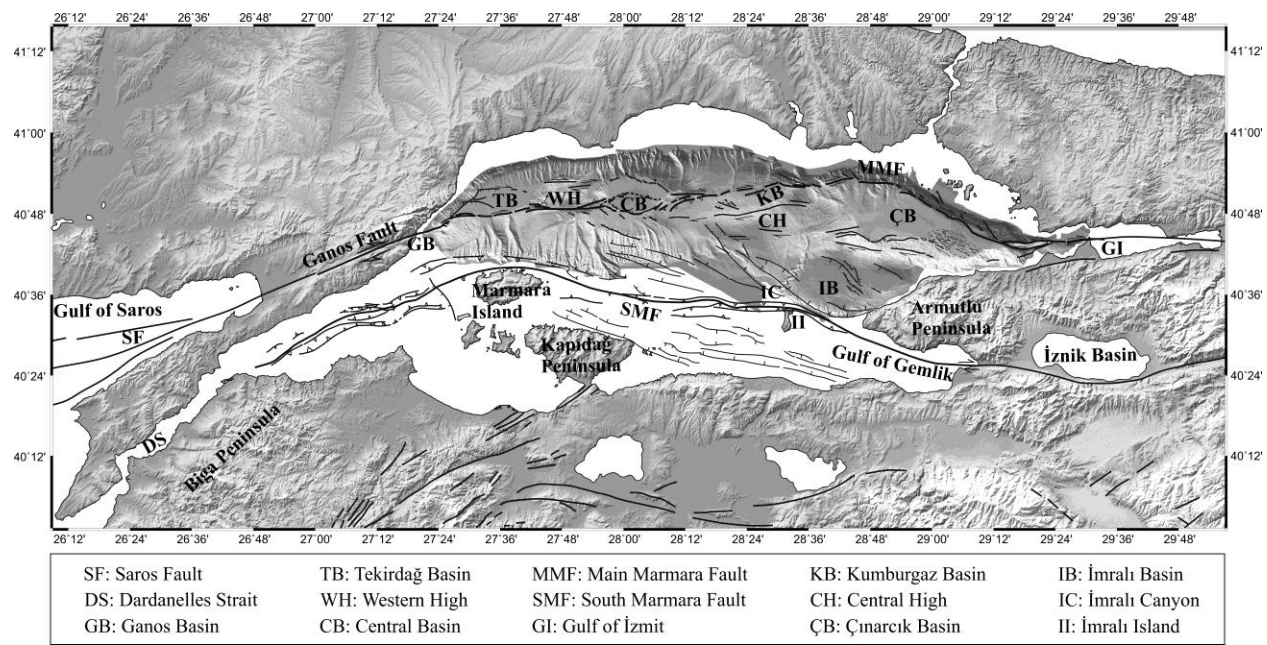
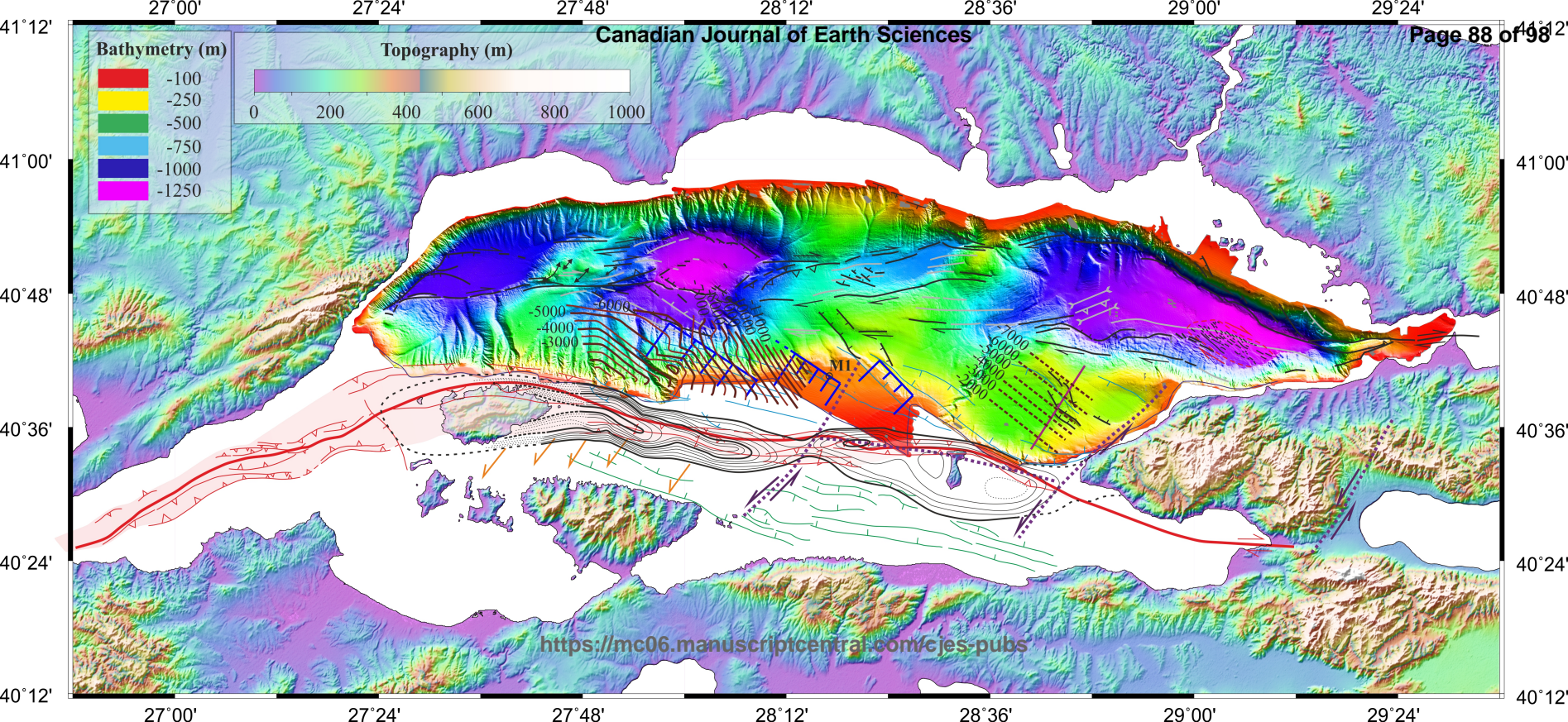
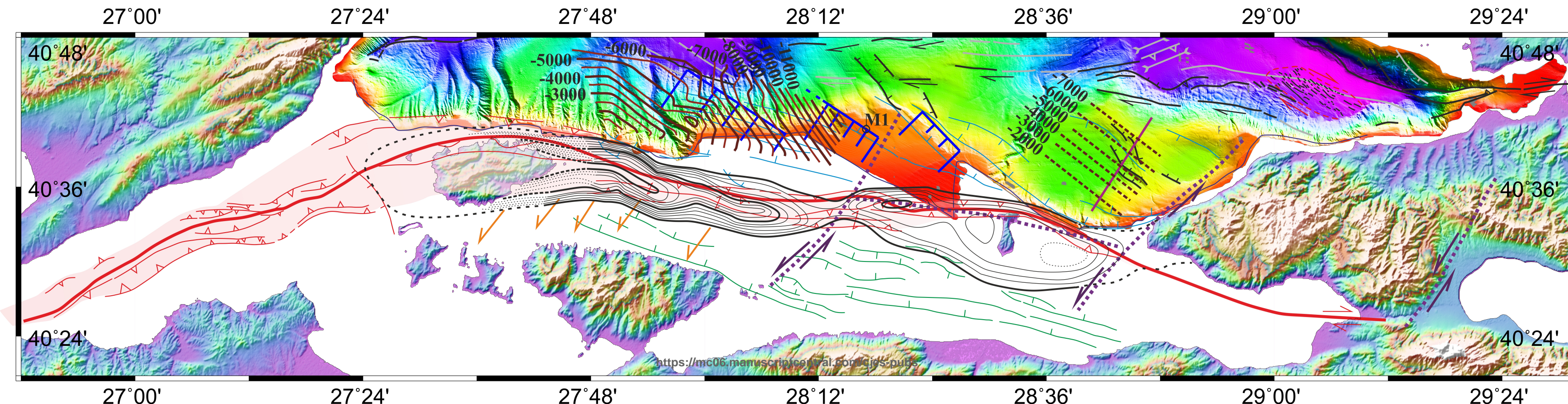
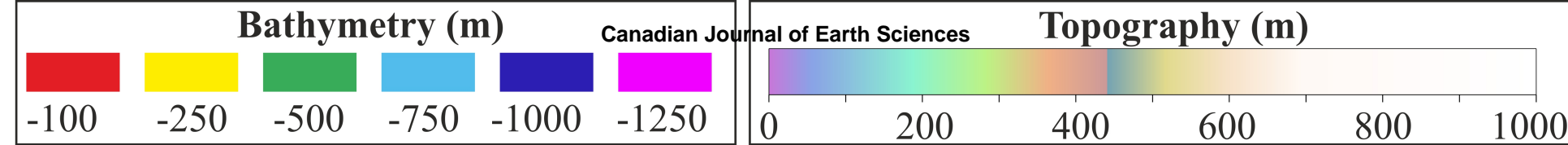


Figure 1
153x149mm (300 x 300 DPI)









27°24'

27°48'

28°12'



28°36'

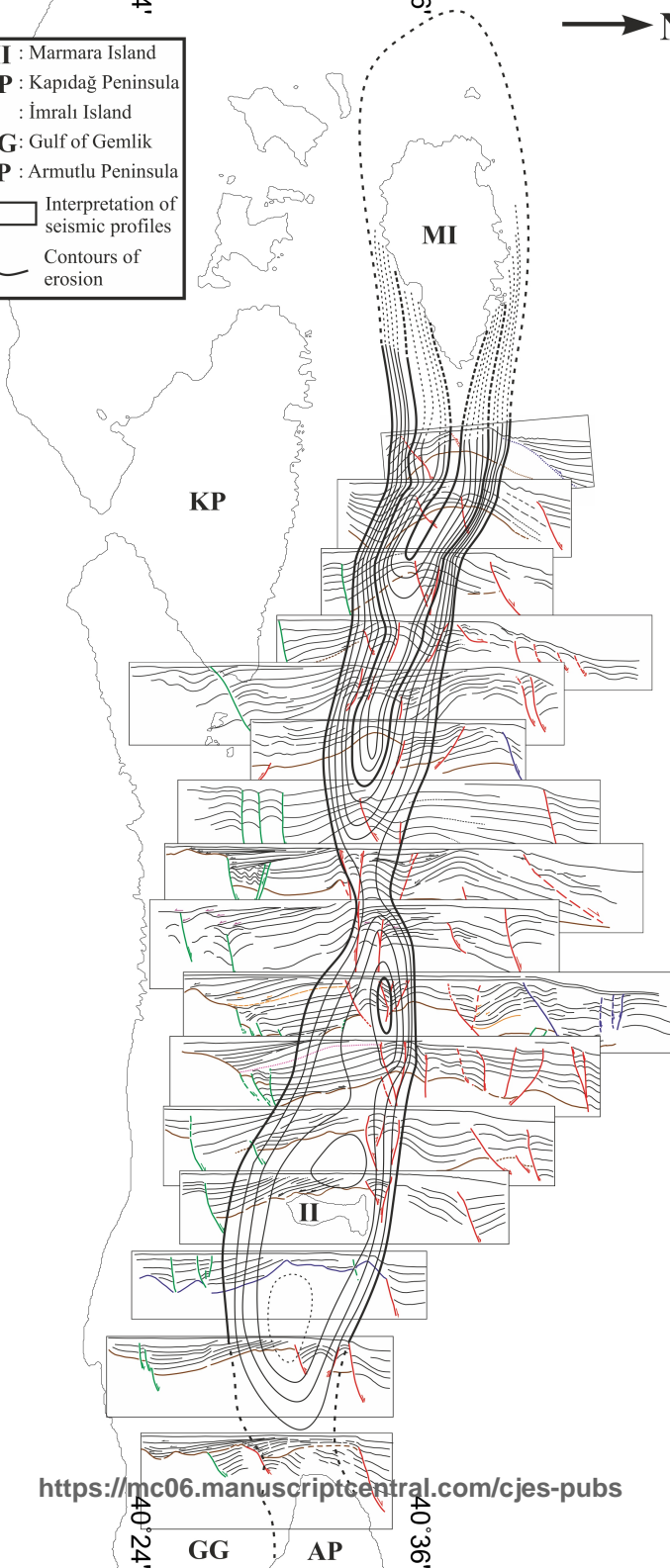
27°24'

27°48'

28°12'

28°36'

- MI** : Marmara Island
KP : Kapıdağ Peninsula
II : İmralı Island
GG : Gulf of Gemlik
AP : Armutlu Peninsula
-  Interpretation of seismic profiles
 Contours of erosion



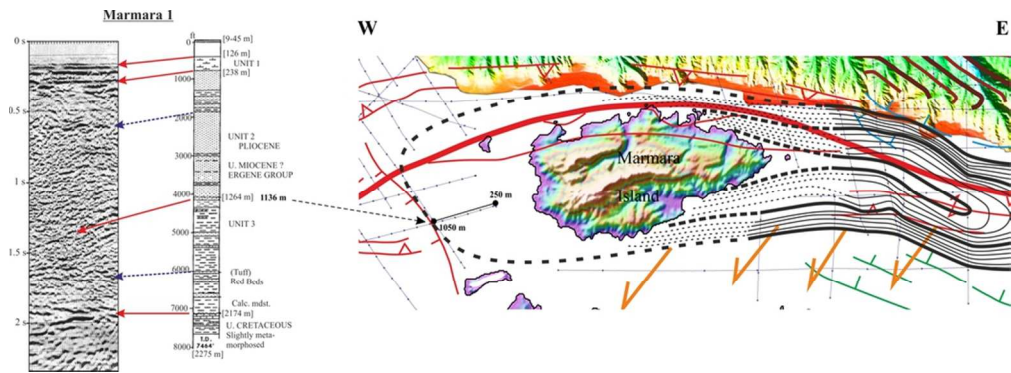


Figure 4
88x32mm (300 x 300 DPI)

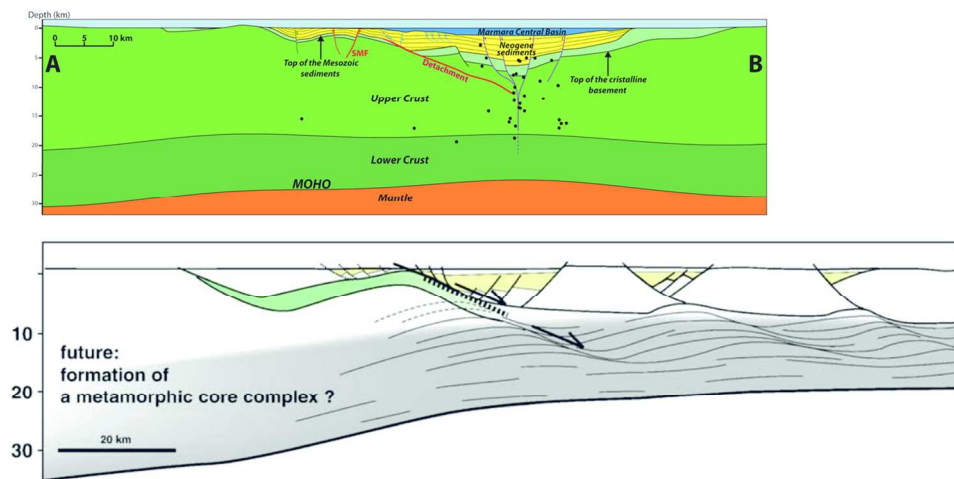


Figure 5
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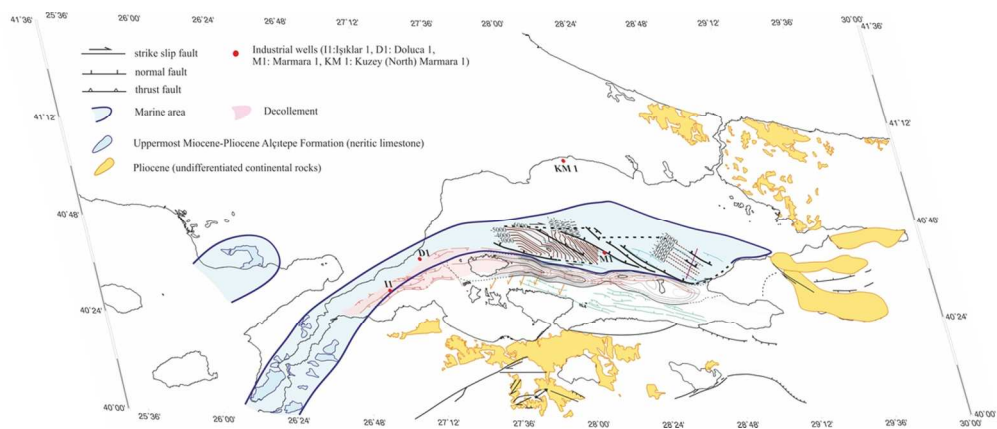


Figure 6
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Draft

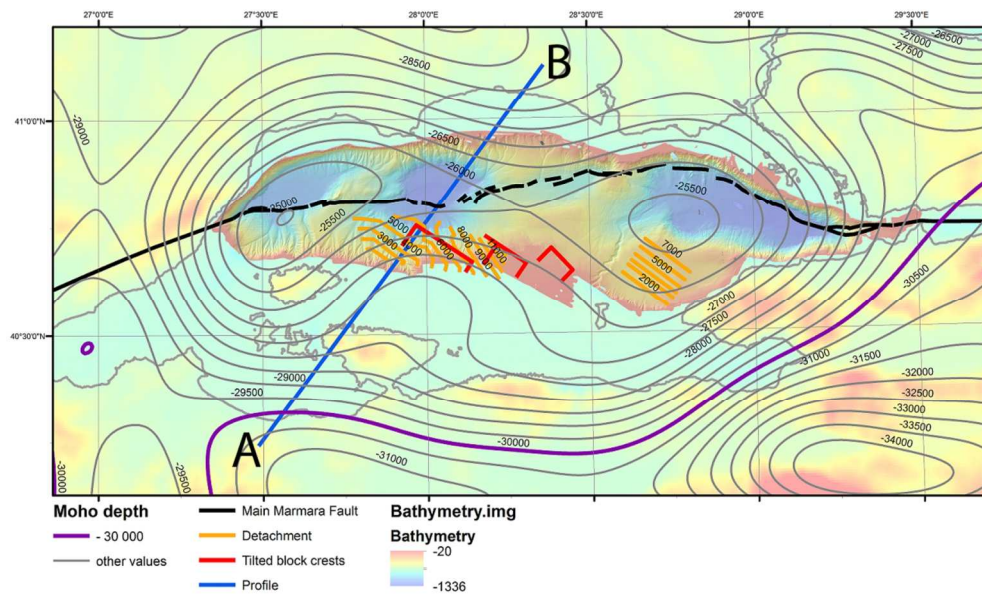


Figure 7
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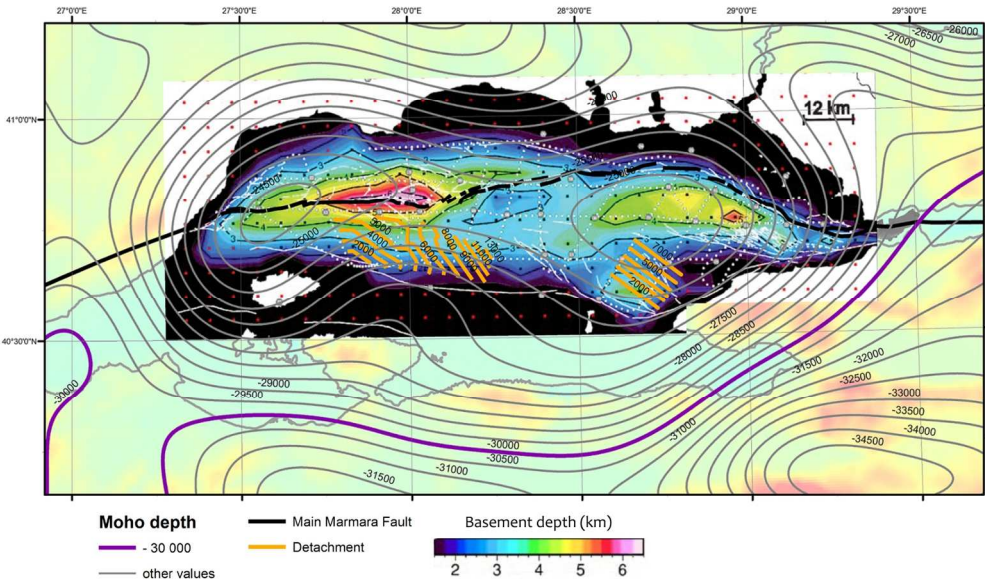


Figure 8
157x93mm (300 x 300 DPI)

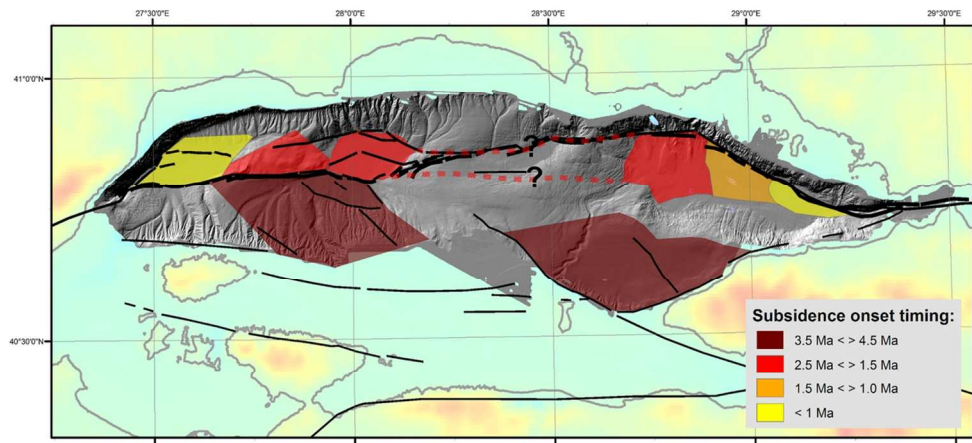


Figure 9
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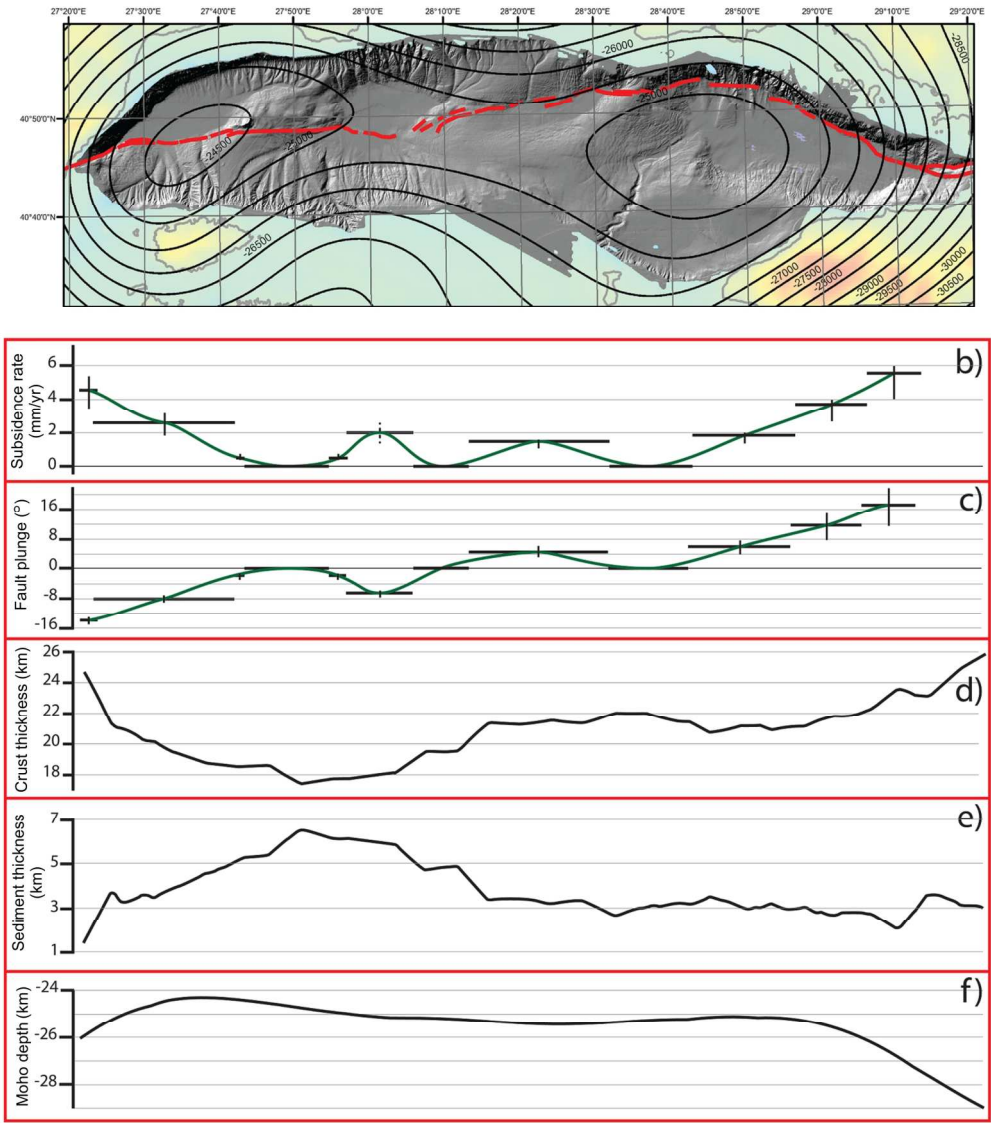


Figure 10
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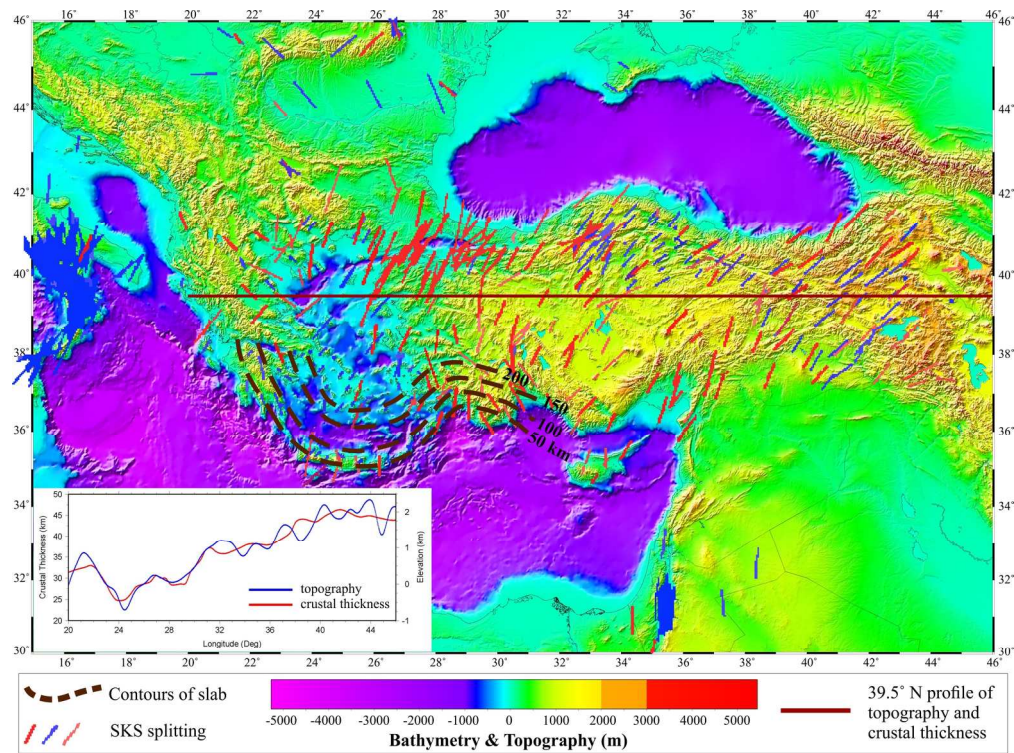


Figure 11
176x130mm (300 x 300 DPI)