# Map view restoration of Aegean–West Anatolian accretion and extension since the Eocene

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[1] The Aegean region (Greece, western Turkey) is one of the best studied continental extensional provinces. Here, we provide the first detailed kinematic restoration of the Aegean region since 35 Ma. The region consists of stacked upper crustal slices (nappes) that reflect a complex paleogeography. These were decoupled from the subducting African-Adriatic lithospheric slab. Especially since ~25 Ma, extensional detachments cut the nappe stack and exhumed its metamorphosed portions in metamorphic core complexes. We reconstruct up to 400 km of trench-perpendicular (NE-SW) extension in two stages. From 25 to 15 Ma, the Aegean forearc rotated clockwise relative to the Moesian platform around Euler poles in northern Greece, accommodated by extensional detachments in the north and an inferred transfer fault SE of the Menderes massif. The majority of extension occurred after 15 Ma (up to 290 km) by opposite rotations of the western and eastern parts of the region. Simultaneously, the Aegean region underwent up to 650 km of post-25 Ma trench-parallel extension leading to dramatic crustal thinning on Crete. We restore a detachment configuration with the Mid-Cycladic Lineament representing a detachment that accommodated trench-parallel extension in the central Aegean region. Finally, we demonstrate that the Sakarya zone and Cretaceous ophiolites of Turkey cannot be traced far into the Aegean region and are likely bounded by a pre-35 Ma N-S fault zone. This fault became reactivated since 25 Ma as an extensional detachment located west of Lesbos Island. The paleogeographic units south of the İzmir-Ankara-Sava suture, however, can be correlated from Greece to Turkey.

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## 1. Introduction

[2] One of the main challenges of solid Earth science is to decipher the physics and dynamics that underlie the complex geological evolution of the crust. The Aegean extensional province located in Greece and western Turkey has been instrumental in linking subduction-related mountain building, continental extension, and the resulting formation and exhumation of metamorphic rocks to plate motions and mantle dynamics [*van Hinsbergen et al.*, 2005a, 2010c; *Jolivet et al.*, 2009; *Faccenna and Becker*, 2010; *Royden and Papanikolaou*, 2011]. In addition, since the recognition of Cordilleran-type low-angle extensional detachments that exhumed metamorphic core complexes in the Cyclades [*Lister et al.*, 1984], the Aegean region has played a prominent role in the analysis of the dynamics of continental extension [e.g., *Tirel et al.*, 2009; *Huet et al.*, 2011a, 2011b].

[3] The Aegean orogen is built up by a series of stacked, ~5–10 km thick (composite) nappes consisting of upper crust that decoupled from now subducted continental and oceanic lithosphere of the Adriatic-African plate [*van Hinsbergen et al.*, 2005a, 2010c; *Jolivet and Brun*, 2010]. These nappes (or "megaunits") were thrusted and stacked from north to south since the Cretaceous [*Faccenna et al.*, 2003; *van Hinsbergen et al.*, 2005a; *Jolivet and Brun*, 2010] and form a strongly shortened representation of a paleogeographical distribution of continental ribbons and deep, sometimes oceanic, basins that existed in the western Neotethys [e.g., *Dercourt et al.*, 1986; *Barrier and Vrielynck*, 2008; *Stampfli and Hochard*, 2009]. Since late Eocene times, while accretion of nappes at the subduction zone continued, previously stacked nappes became stretched in an overall NNE-SSW direction [*Gautier et al.*, 1999; *Forster and Lister*, 2009; *Tirel et al.*, 2009; *Jolivet and Brun*, 2010]. Aegean extension is normally interpreted to result from retreat of the subducting Adriatic-African slab below the Aegean region with respect to Eurasia [*Le Pichon and Angelier*, 1979; *Meulenkamp et al.*, 1988]. This process led to the present-day geometry of a curved, southward convex,

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northward dipping ("amphitheater"-shaped) slab [*Spakman et al.*, 1988, 1993; *Biryol et al.*, 2011].

[4] Extension was largely accommodated by the formation of several large extensional windows: the Rhodope, the eastern mainland Greece-Cycladic, the Menderes and the South Aegean extensional metamorphic complexes (Figure 1). These complexes expose parts of the nappe stack that were metamorphosed during underthrusting [*Jolivet and Brun*, 2010; *Ring et al.*, 2010].

[5] Geological-petrological studies have a tendency to focus on the rocks that were exposed to the most extreme metamorphic conditions and most intense deformation. Therefore, studies in the Aegean region tend to focus on the N-S transect from the Rhodopes, over the Cyclades, to Crete, where the amount of trench-normal extension is presumably at its maximum [e.g., *Jolivet and Brun*, 2010]. However, a large portion of this transect is submerged and therefore inaccessible for field studies, or for direct estimates of the total amount of extension. The amount of Aegean trench-normal extension, however, decreases eastward and (north) westward [e.g., *Brun and Sokoutis*, 2007; *Jolivet et al.*, 2010c; *van Hinsbergen et al.*, 2010a]. These lateral changes in the finite amount of extension are reflected in large-scale vertical axis rotations of the fore-arc domains, clockwise in western Greece and Albania and counterclockwise in south-western Turkey [*Kissel and Laj*, 1988; *van Hinsbergen et al.*, 2005b, 2008, 2010a; *Denèle et al.*, 2011] (Figure 2).

[6] Although most workers realize this strong lateral variation in extension and associated vertical axis rotations, existing restorations of the Aegean region to its pre-extensional state remain conceptual and schematic [*Walcott and White*, 1998; *van Hinsbergen et al.*, 2008; *Brun and Sokoutis*, 2010; *Jolivet and Brun*, 2010; *Royden and Papanikolaou*, 2011]. Also the exact timing of the different stages of extension needs a more careful analysis. In this paper, we provide the first detailed kinematic map view restoration of the Aegean region for the last ~35 Ma, embedded in the Africa-North America-Eurasia plate circuit of *Torsvik et al.* [2012]. We approach this by stepwise restoring the extensional complexes in western and northern Greece, as well as in western Turkey, where their boundaries and temporal history are well constrained. In the case of western Turkey we incorporate and update a recent map view restoration of *van Hinsbergen* [2010]. We then explore the consequences of the restoration for the location and evolution of the main detachment systems of the central and southern Aegean region, calculate the total amount of NNE-SSW trench-normal extension, and the amount of trench-parallel extension. The megaunits that are classically defined in Turkey [*Şengör and Yilmaz*, 1981] have different names than those in Greece [*Aubouin*, 1957; *Jacobshagen*, 1986]. Although Greece indeed contains paleogeographic units that do not exist in Turkey and vice versa, part of the differences in terminology reflect definitions of the same unit changing across political boundaries [see, e.g., *Schmid et al.*, 2008]. In this paper, we follow separate terminologies for both countries in a first step, and discuss to what extent Greek tectonic and paleogeographic units continue into Turkey and vice versa after presenting our reconstruction. Our restoration is mainly based on structural geological criteria, dated via thermochronological data from metamorphic complexes and stratigraphic data from sedimentary basins. We use paleomagnetic constraints on timing and amount of regional vertical axis rotations to independently test the validity of the restoration.

## 2. Approach

[7] The reconstruction is made using the freely available software package GPlates (http://www.gplates.org) [*Boyden et al.*, 2011]. Classic plate reconstructions assume rigid plates and localized deformation at plate boundaries [*Cox and Hart*, 1986]. Rigid plates are modeled as nondeformable polygons. For intensely deformed regions such as the Aegean region, however, this approach is not everywhere useful. In regions of continental extension or shortening, geological units change shape and area over time and cannot be modeled as rigid polygons. For time intervals associated with deformation, we therefore define "topologies" (i.e., deforming polygons) by reconstructing the positions of their boundaries relative to each other through time, leading to deforming-plate reconstructions. As a result, the reconstruction contains internally deforming topologies that at the same time may undergo displacements with respect to each other (Figures 3a and 3b) [*van Hinsbergen*, 2010; *van Hinsbergen et al.*, 2011a].

[8] Metamorphic rocks exhumed in extensional provinces of the Aegean region are generally interpreted to correspond to exhumed portions of nappes that are exposed behind the Hellenic trench. To test whether these inferences are correct, we make a first-order attempt to restore these rocks in their presubduction configuration. We achieve this by "unexhuming" the exhumed metamorphic rocks back into their buried position below their hanging wall (Figures 3c and 3d) and by "unsubducting" them by modeling them as rigidly attached to the African plate from the estimated moment of their peak pressure conditions back in time. This way we eventually provide a crude estimate to test whether the amount of syn-prograde metamorphic convergence is sufficient to bring these rocks to their current relative position in

**Figure 1.** Geological map of the Aegean–West Anatolian region after *Schmid et al.* [2011], modified after *Bornovas and Rontogianni-Tsiabaou* [1983] and *van Hinsbergen et al.* [2010c], following the color correlation scheme to Balkan geology of *Schmid et al.* [2008]. Abbreviations: AAM = Anaximander and Anaxigoras Seam Mounts; AD = Alaşehir Detachment; Ae = Aegina; ASG = Amvrakikos-Sperchios Graben; AT = Aksu Thrust; BMD = Büyük Menderes Detachment; BMG = Büyük Menderes Graben; BS = Botsara syncline (and Klematia-Paramythia Basin); CW = Chelmos Window; GOC = Gulf of Corinth; GG = Gediz Graben; GW = Göçek window; IAD = Itea-Amfissa Detachment; IC = Isthmus of Corinth; KB-LB = Karditsa Basin–Larissa Basin; KM = Kazdağ; MG = Mesta Graben; MKG = Melimala- Kalamata Graben; Mt. Ol. = Mount Olympos; Mt. Os. = Mount Ossa; Pa = Paleros; PGG = Pyrgos-Gythian Graben; PO = Pindos Ophiolite; SD = Simav Detachment; SG = Saronic Gulf; TSZ = Thesprotiko Shear Zone. Numbers refer to the kinematic constraints used in this reconstruction, listed in Table 1.

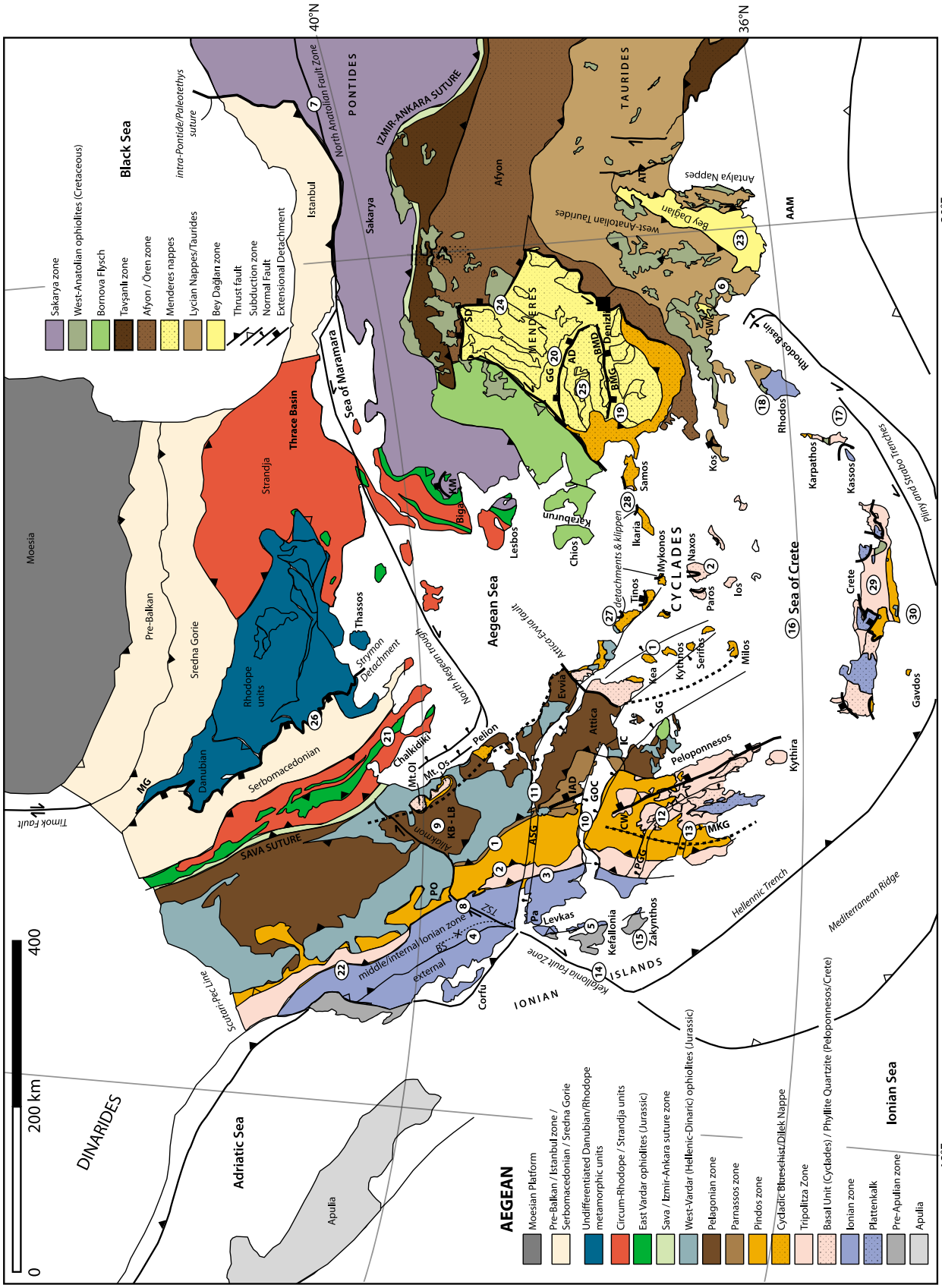

AEGEAN
Moesian Platform
Pre-Balkan / Istanbul zone / Serbomacedonian / Sredna Gorie
Undifferentiated Danubian/Rhodope metamorphic units
Circum-Rhodope / Strandja units
East Vardar ophiolites (Jurassic)
Sava / Izmir-Ankara suture zone
West-Vardar (Hellenic-Dinaric) ophiolites (Jurassic)
Pelagonian zone
Parnassos zone
Pindos zone
Cycladic Blueschist/Dilek Nappe
Tripolitza Zone
Basal Unit (Cyclades) / Phyllite Quartzite (Peloponnesos/Crete)
Ionian zone
Plattenkalk
Pre-Apulian zone
Apulia
Sakarya zone
West-Anatolian ophiolites (Cretaceous)
Bornova Flysch
Tavşanlı zone
Afyon / Ören zone
Menderes nappes
Lycian Nappes/Taurides
Bey Dağları zone
Thrust fault
Subduction zone
Normal Fault
Extensional Detachment
Black Sea
Adriatic Sea
Ionian Sea
Aegean Sea
Sea of Crete
Sea of Maramara
Moesia
Pre-Balkan
Sredna Gorie
Strandja
Thrace Basin
Rhodope units
Danubian
Serbomacedonian
Thassos
Strymon Detachment
Chalkidiki
Mt.Ol
Mt. Os
Pelion
Allakmon
SAVA SUTURE
Timok Fault
Scutari-Pec Line
DINARIDES
Apulia
Corfu
Levkas
Kefallonia
Zakynthos
Kefallonia Fault Zone
IONIAN ISLANDS
Hellennic Trench
Mediterranean Ridge
Peloponnesos
Kythira
Attica
Evvia
Attica-Evvia fault
North Aegean trough
Lesbos
Chios
Karaburun
Samos
Ikaria
Mykonos
Naxos
Paros
Ios
Tinos
Kythnos
Serifos
Milos
Kea
Kos
Rhodos
Karpathos
Kassos
Crete
Gavdos
Rhodos Basin
Pliny and Strabo Trenches
CYCLADES
detachments & klippen
Sakarya
Istanbul
PONTIDES
North Anatolian Fault Zone
IZMIR-ANKARA SUTURE
Afyon
TAURIDES
Antalya Nappes
West-Anatolian Taurides
Bey Dağları
MENDERES
AAM
Denizli
Biga
KM
MG
KB - LB
PO
ASG
IAD
GOC
CW
PGG
MKG
IC
Ae
SG
SD
GG
AD
BMD
BMG
GW
AT
TSZ
BS
Pa
40°N
36°N
30°E
18°E
0
200 km
400

Figure 1

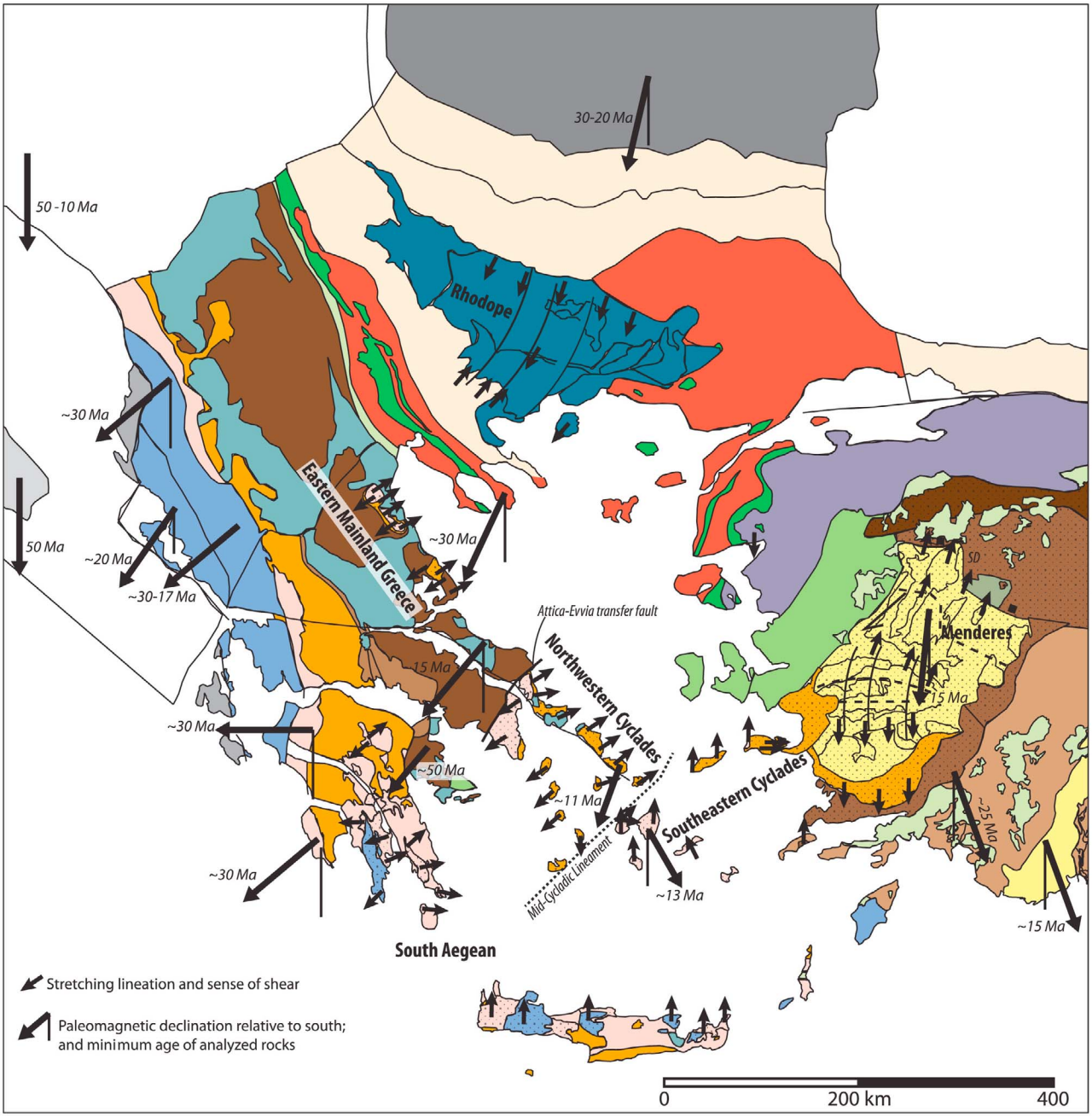



**Figure 2.** Map showing the regional pattern of stretching lineations in the metamorphic complexes of the Aegean–West Anatolian region as well as the locations of the breakaways (dotted lines) of the major extensional metamorphic complexes. Lineation directions and senses of shear compiled from *Walcott and White* [1998], *Brun and Sokoutis* [2007], *Rosenbaum et al.* [2007], *Marsellos and Kidd* [2008], *Diamantopoulos et al.* [2009], *Tirel et al.* [2009], *van Hinsbergen and Boekhout* [2009], *Jolivet et al.* [2010c] and *van Hinsbergen et al.* [2010a]. Paleomagnetic declinations compiled from *Horner and Freeman* [1982, 1983], *Kissel et al.* [1984, 1985, 1989, 1995], *Kissel and Poisson* [1987], *Kissel and Laj* [1988], *Tozzi et al.* [1988], *Márton et al.* [1990], *Scheepers* [1992], *Speranza et al.* [1992, 1995], *Morris and Robertson* [1993], *Speranza and Kissel* [1993], *Mauritsch et al.* [1995, 1996], *Morris* [1995], *Morris and Anderson* [1996], *Beck et al.* [2001], *van Hinsbergen et al.* [2005b, 2008, 2010a, 2010b], *Broadley et al.* [2006] and *de Leeuw et al.* [2012].

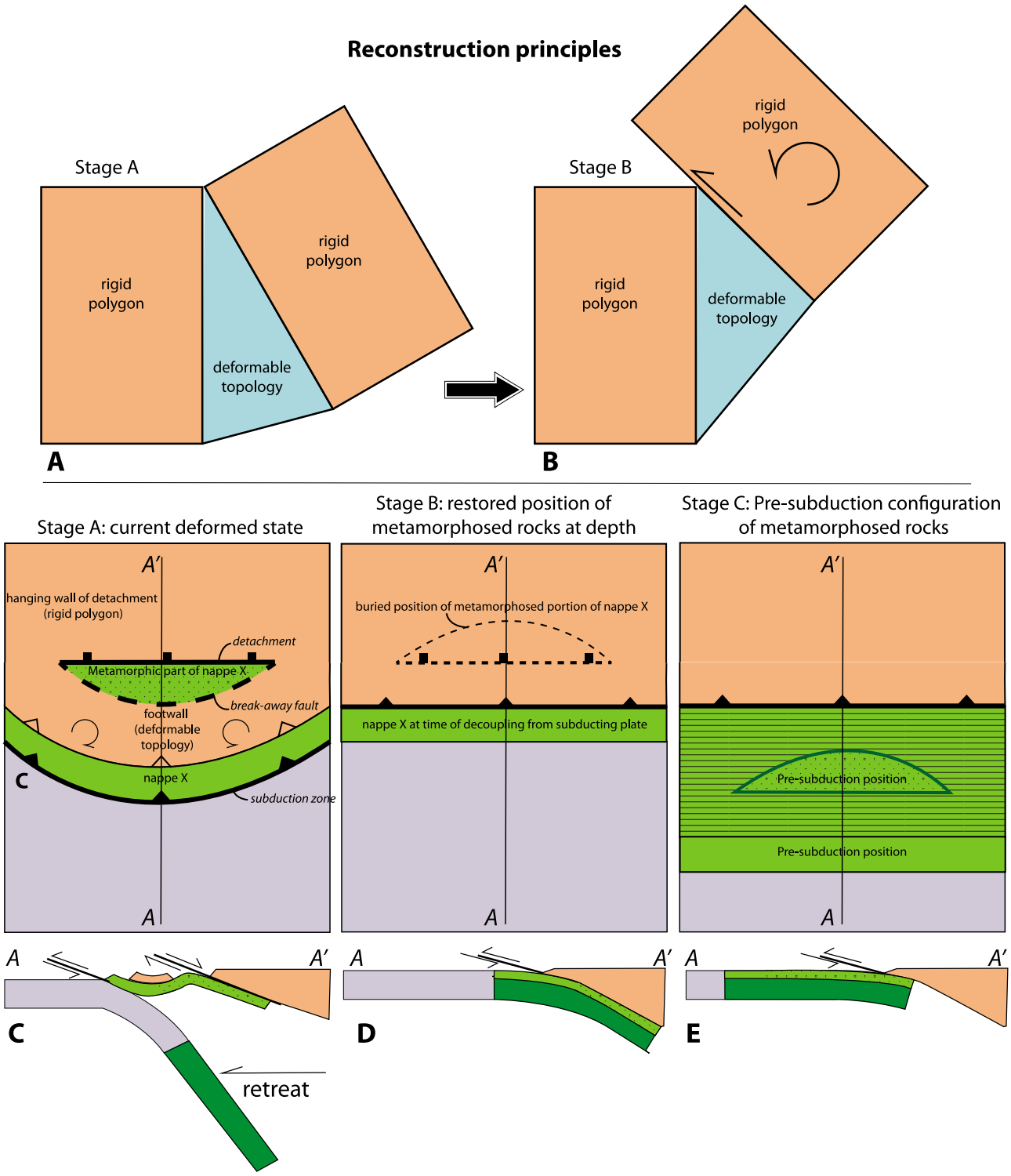



**Figure 3.** (a and b) Illustration of the principle of using deforming polygons ("topologies"): the boundaries of the topologies move relative to each other over time, leading to changing shapes and areas, amid other topologies or rigid polygons that move relative to each other. For example, extending basins or shortening fold-thrust belts can be modeled as topologies; (c–e) Restoration of exhumed metamorphic rocks: first, the position of the exhumed metamorphic rocks below the surface is restored by restoring the extension that exhumed them (Figures 3c and 3d), followed by restoring the presubduction position by assuming a fixed position relative to the subducting plate prior to the moment of peak pressure (Figures 3d and 3e).

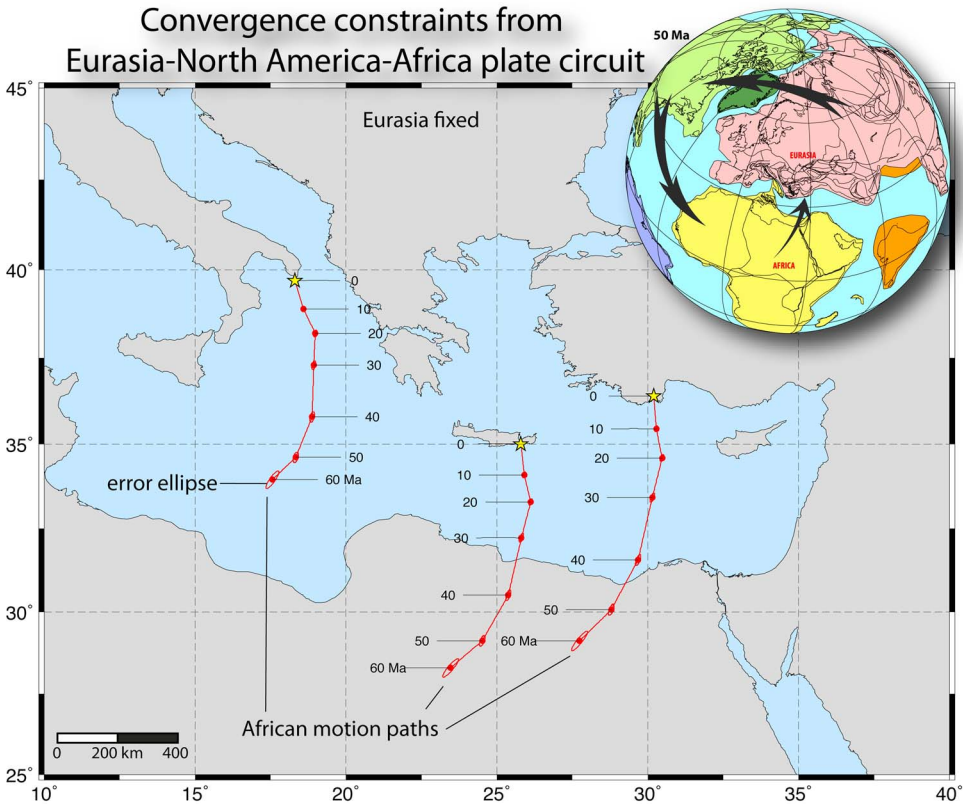



**Figure 4.** Motion path of Africa relative to the fixed Eurasia for since 60 Ma, at two coordinates: Apulia, southern Italy (39.7°N, 18.3°E) and the İzmir-Ankara suture zone, Turkey (39.5°N, 29°E), based on the plate circuit model A of *van Hinsbergen et al.* [2011b].

the nappe stack; we provide a crude estimate of their original palinspastic position on the African plate prior to subduction (Figures 3d and 3e).

## 3. Plate Kinematic Constraints

[9] The deformation history of the Aegean–West Anatolian region occurred within the context of convergence between Africa and Eurasia. The reconstruction presented here combines finite rotations describing the Cenozoic motion between Africa, North America and Eurasia, using the most recent kinematic constraints from the central and northern Atlantic Ocean [*Müller et al.*, 1999; *Gaina et al.*, 2002]. This kinematic model is similar to the Africa-Eurasia plate circuit of *Torsvik et al.* [2012], with minor modifications as in model A of *van Hinsbergen et al.* [2011b]. Reconstruction uncertainties were propagated through the Africa–North America–Eurasia plate circuit following the procedure of [*Doubrovine and Tarduno*, 2008]. Because the kinematic models of *Müller et al.* [1999] and *Gaina et al.* [2002] do not use the same time steps, we have compiled a common set of ages, comprising all reconstruction ages in the two individual reconstructions, and estimated "missing" Africa–North America and North America–Eurasia rotations, and their errors, using the interpolation technique described by *Doubrovine and Tarduno* [2008]. For illustrative purposes (Figure 4), we also interpolated individual plate pair rotations at 10 Ma increments. Coeval rotations (original and/or interpolated) were then combined, and their uncertainties were calculated using established formulations [e.g., *Stock et al.*, 1990].

[10] Figure 4 shows the relative plate motion path of Africa relative to (fixed) Europe for three coordinates: 39.7°N, 18.3°E (corresponding to southernmost Apulia in southern Italy), 35.0°N, 25.8°E (corresponding to Ierapetra on Crete) and 36.4°N, 30.2°E (corresponding to the southern Bey Dağları platform in western Turkey). These motion paths (Figure 4) track the amount of Africa-Europe convergence that must have been accommodated at and north of these points through time. The 95% confidence ellipses for the reference points are typically smaller than 40 km along the major semiaxis for the last 40 Ma.

[11] We restore Aegean deformation relative to the Moesian Platform (or "Moesia"), which was essentially part of stable Eurasia throughout the Cenozoic (see section 4) and we restrict our reconstruction to an area south of the Scutari-Pec line in northern Albania (Figure 1).

## 4. Tectonic Architecture of Greece and Western Turkey

### 4.1. Greece

[12] Figure 1 shows the geological map of Greece and western Turkey. We now briefly summarize the main tectonic units of this area, forming the pattern that we reconstructed into its preextensional configuration. The Moesian Platform borders the Aegean fold-thrust belt in the north. Moesia contains a Precambrian basement, overlain by Upper

Cambrian to Cenozoic clastic sediments and platform carbonates [*Tari et al.*, 1997; *Matenco et al.*, 2003]. Since mid-Cretaceous times only very minor faulting occurred between Moesia and Eurasia (e.g., along the IntraMoesian Fault [*Hippolyte*, 2002; *Matenco et al.*, 2003; *Ozclon et al.*, 2007; *Schmid et al.*, 2008]). To the west, Moesia is bordered by the curved, dextral Timok Fault that accommodated north directed motion of west Aegean, Albanian and Macedonian megaunits relative to the Moesian platform in Oligo-Miocene time [*Fügenschuh and Schmid*, 2005]. In this paper, however, we do not reconstruct the area to the west of the Moesian platform in detail.

[13] The originally WNW-ESE trending nappe stack of the Aegean region has been curved into an orocline [*van Hinsbergen et al.*, 2005b, 2008], with nonmetamorphic parts of the nappe stack covering most of the area, surrounding the extensional metamorphic complexes (Figures 1 and 2). The nappes are least affected by extension in mainland Greece where they represent typically 5–10 km thick, internally folded and thrusted rock units, mapped based on tracing bounding thrusts and recognizing similar sedimentary and/or metamorphic facies and age [*Aubouin*, 1957; *Jacobshagen*, 1986]. The nappe stack is SW to south facing. It overrides the Apulian platform to the SW and the subducting African plate in the south, respectively. The most internal units (Circum-Rhodope/Strandja, Serbomacedonian, Sredna Gora megaunits, Figure 1) thrust northward toward the Moesian platform, mostly in Jurassic to early Cretaceous times followed by a reversal in subduction polarity in the Late Cretaceous. To a minor extent, northward thrusting of the Sredna Gora and Pre-Balkan units over Moesia resumed in Cenozoic times [*Sinclair et al.*, 1997; *Burchfiel et al.*, 2008], but this is neglected in our reconstruction for simplicity. Since late Cretaceous time, first oceanic crust of the Vardar-Axios (or Sava) ocean subducted northward below the most internal units, followed by the arrival of the previously mentioned continental ribbons and deep, sometimes oceanic, intervening basins that resulted in formation of the southward growing nappe stack [*van Hinsbergen et al.*, 2005a; *Schmid et al.*, 2008; *Jolivet and Brun*, 2010; *Bonev and Stampfli*, 2011].

[14] The first ribbon continent to arrive and collide with Eurasia after inception of northward subduction was the Pelagonian zone, including the previously obducted West Vardar ophiolites (Figure 1). This resulted in the final closure of a remnant of the Vardar ocean [e.g., *Stampfli and Borel*, 2002] and thereby the formation of the Sava Zone (or Sava suture) [*Jacobshagen*, 1986; *Ricou et al.*, 1998; *Schmid et al.*, 2008; *Ustaszewski et al.*, 2009, 2010]. The exact timing of collision of the Pelagonian platform (including ophiolites that obducted in Jurassic time [*Dimo-Lahitte et al.*, 2001; *Pamić et al.*, 2002; *Liati et al.*, 2004; *Schmid et al.*, 2008]) is still ill constrained but is of late Cretaceous age [*Schermer*, 1990; *Lips et al.*, 1998; *Ricou et al.*, 1998]. This was followed by the decoupling of Pelagonian upper crust from the downgoing Adriatic-African slab, and a jump of frontal accretion to its southwestern edge (in modern coordinates) in Paleocene times [*van Hinsbergen et al.*, 2005a; *Jolivet and Brun*, 2010] that led to the underthrusting of the Pindos-Olonos nappe (hereafter called "Pindos nappe") (Figure 1). The Pindos nappe consists of Triassic to Upper Cretaceous radiolarites and carbonates deposited on unknown basement, regarded as either oceanic [e.g., *Stampfli and Borel*, 2002] or thinned continental crust and lithosphere [e.g., *Schmid et al.*, 2008]. These sediments are overlain by an upper Paleocene to upper Eocene flysch (~60–35 Ma [*Richter et al.*, 1978; *Degnan and Robertson*, 1998]), with the oldest flysch age providing a first-order estimate for the minimum age of inception of subduction of the megaunit and the youngest flysch age allowing inference of the end of its subduction, its accretion to the overriding plate, and the associated uplift [see, e.g., *van Hinsbergen et al.*, 2005c] (Table 1). Windows in the eastern part of the Pelagonian nappe (Mount Olympos, Mount Ossa and Mount Pelion), as well as in the Cycladic area show that it is underlain by the high-pressure/low-temperature (HP-LT) metamorphic Cycladic Blueschist unit, which is widely regarded as the metamorphosed lateral equivalent of the Pindos nappe based on a similar stratigraphic age range and tectonostratigraphic position [*Jolivet et al.*, 2010b]. The Cycladic Blueschist unit was metamorphosed to HP-LT conditions in Eocene–early Oligocene times (up to ~18–22 kbar, 500–600°C; see overviews by *Ring et al.* [2007a, 2011], *Tirel et al.* [2009] and *Jolivet and Brun* [2010]) and locally include Permo-Carboniferous granites intruding an older basement [*Keay and Lister*, 2002; *Tomaschek et al.*, 2008], covered by a metasedimentary carbonate-rich sequence that contains mafic Triassic volcanics [*Bröcker and Pidgeon*, 2007]. Similar volcanics are known from the basal Pindos nappe in western Greece [*Pe-Piper and Piper*, 1991]. The Cycladic Blueschist unit contains detrital zircons in metasandstones as young as ~79 Ma [*Bulle et al.*, 2010]. The Cycladic Blueschists are associated with a consistent ~55–40 Ma age range for high-pressure metamorphism, overprinted by retrograde greenschist facies metamorphic parageneses between ~30–25 Ma and younger [*Kumerics et al.*, 2005; *Putlitz et al.*, 2005; *Bröcker and Franz*, 2006; *Lagos et al.*, 2007; *Ring et al.*, 2007b; *Schneider et al.*, 2011] (see review by *van Hinsbergen et al.* [2005c]). Underthrusting of the Pindos unit continued until the end of deposition of the youngest flysch, i.e., late Eocene, ~35 Ma, by which time most of the Cycladic Blueschist was already exhuming [e.g., *Jolivet and Brun*, 2010]. This indicates that the Pindos and Cycladic Blueschist units merely represent the same paleogeographic realm and cannot be strictly considered as one single semirigid nappe complex.

[15] In the modern fore arc, the Pindos nappe is underlain by the Gavrovo-Tripolitza nappe (hereafter referred to as "Tripolitza nappe") (Figure 1), which consists of platform carbonates of Late Triassic to Eocene age, underlain by a Triassic volcanic series (the Tyros beds) [*Institut de Géologie et Recherches du Sous-sol–Institut Français du Pétrole* (*IGRS-IFP*), 1966; *British Petroleum Company*, 1971; *Fleury et al.*, 1979; *Jacobshagen*, 1986; *Zambetakis-Lekkas et al.*, 1998]. The platform carbonates are overlain by an Oligocene flysch (~33–23 Ma) [*IGRS-IFP*, 1966; *Richter et al.*, 1978; *Peeters et al.*, 1998]. In the Cycladic and west Aegean extensional metamorphic complexes, the Cycladic Blueschist unit is underlain by the so-called "Basal Unit," a metamorphosed platform carbonate succession with up to Eocene sedimentary ages and Oligocene metaflysch, metamorphosed at HP-LT metamorphic conditions (~8–10 kbar, 400°C) with the youngest estimates of peak metamorphic ages clustering around 24–21 Ma [*Ring et al.*, 2001;

**Table 1.** Kinematic Constraints Used in This Reconstruction[a]

| Structure/Region | Sense | Amount | Age | Reference |
|---|---|---|---|---|
| Underthrusting of Pindos/Cycladic Blueschist unit | top-to-the-south | | ~60–35 Ma | *Degnan and Robertson* [1998] |
| Underthrusting Tripolitza unit | top-to-the-south | | ~35–23 Ma | *IGRS-IFP* [1966] and *Sotiropoulos et al.* [2003] |
| Underthrusting Internal/middle Ionian unit | top-to-the-south | | ~35–23 Ma | *IGRS-IFP* [1966] and *Sotiropoulos et al.* [2003] |
| Underthrusting external ionian zone | top-to-the-southwest | | ~23–15 Ma | *van Hinsbergen et al.* [2006] and *Broadley et al.* [2006] |
| Underthrusting Pre-Apulian zone | top-to-the-southwest | | ~15–4 Ma | *van Hinsbergen et al.* [2006] |
| Southeastward translation West Anatolian Taurides (distance Göçek window-Bey Dağları thrust front) | top-to-the-southeast | 75 km | 23–15 Ma | *Brunn et al.* [1970], *Hayward* [1984a], and *van Hinsbergen* [2010] |
| North Anatolian Fault Zone | dextral | 60 km, decreasing westward | 11–0 Ma | *Şengör et al.* [2005] |
| Thesprotiko Shear zone | dextral | 20 km | 7–3 Ma | *van Hinsbergen et al.* [2006] |
| Karditsa & Larissa basins | NE-SW extension | 10 km | 7–3 Ma | *Caputo and Pavlides* [1993] and *Caputo et al.* [1994]; see text |
| Gulf of Corinth | ~N-S extension | 15 km | 3.5–0 Ma | *Jolivet et al.* [2010a] and *Taylor et al.* [2011] |
| Amvrakikos-Sperchios graben | ~N-S extension | ~5 km[b] | 3.5–0 Ma | see text |
| Pyrgos-Gythion graben system | ~N-S to NE-SW extension | ~5 km[b] | 3.5–0 Ma | see text |
| Melimala-Kalamata graben system | ~N-S to NE-SW extension | ~5 km[b] | 3.5–0 Ma | see text |
| Kefallonia Fault Zone (i.e., Greece versus Adria) | dextral | 60 km[c] | 5–0 Ma | *van Hinsbergen et al.* [2006] |
| Zakynthos rotation | clockwise | ~20° | 1 Ma | *Duermeijer et al.* [1999] |
| Sea of Crete | N-S extension | 40 km | 10–0 Ma | *Angelier et al.* [1982] and *Zachariasse et al.* [2011] |
| South Aegean strike-slip zone (e.g., Pliny, Strabo Trenches) | sinistral | ~70 km[c] | | *Woodside et al.* [2000] and *Zachariasse et al.* [2008] |
| Rhodos rotation | counterclockwise | ~30° | 4–0 Ma | *van Hinsbergen et al.* [2007] |
| Büyük Menderes graben | N-S extens | 3 km | 5–0 Ma | *Çifçi et al.* [2011] |
| Gediz graben | N-S extens | 3 km | 5–0 Ma | *Çiftçi and Bozkurt* [2010] |
| Rotation of Chalkidiki versus Moesia[d] | clockwise | 17.9 ± 12.6° | mainly after 17 Ma | *van Hinsbergen et al.* [2008] |
| Vertical axis rotation western Greece and Albania[d] | clockwise | 38.0 ± 7.2° | mainly after 17 Ma | *van Hinsbergen et al.* [2008] |
| Rotation of SW Turkey versus NW Turkey[d] | counterclockwise | 27.3 ± 5.0° | 15–5 Ma | *van Hinsbergen et al.* [2010a, 2010b] |
| Displacement Simav detachment | NNE-SSW extension | 50 km | 25–15 Ma | *Ring et al.* [2003] |
| Extension Central Menderes Massif | N-S extension | increases westward | 15–5 Ma | *Gessner et al.* [2001a], *Ring et al.* [2003], and *van Hinsbergen et al.* [2010a] |
| Extension Rhodope metamorphic complex | NE-SW extension | increases southeastward | 35–4 Ma | *Brun and Sokoutis* [2007] and *Georgiev et al.* [2010] |
| Extension NW Cyclades | NNE-SSW to NE-SW extension | increases southeastward | 25–7 Ma | see text |
| Extension SE Cyclades | NNE-SSW to N-S extension | increases westward | 25–7 Ma | see text |
| Thinning upper nappes Crete | E-W extension[c] | factor ~10 | 25–11 Ma | *Zachariasse et al.* [2011] |
| Extension upper nappes Aegean fore arc | E-W extension | 650 km | 25–11 Ma | see text |

[a]See Figure 1 for locations of the items listed.
[b]Assumed, but no detailed documentation is available.
[c]Follow from the reconstruction and not used as primary input.
[d]Used to test the reconstruction and not taken as primary input.

*Ring and Layer*, 2003]. Given the similar lithostratigraphy and age, the Basal Unit is considered to be a lateral equivalent of the Tripolitza unit. On Crete and the Peloponnesos (Figure 1), anchimetamorphic [*Rahl et al.*, 2005] Tripolitza rocks are separated by an extensional detachment [*Fassoulas et al.*, 1994; *Jolivet et al.*, 1996] from the HP-LT metamorphic Phyllite-Quartzite unit (up to ∼18 kbar/400°C on western Crete, decreasing in metamorphic grade to the NW and east [*Jolivet et al.*, 1996, 2010c; *Klein et al.*, 2008]), which has similar peak metamorphic ages as the Basal Unit: ∼24–20 Ma [*Jolivet et al.*, 1996]. The Phyllite-Quartzite unit consists of a metamorphosed alternating marine series of mudstones, turbiditic sandstones and debris flow conglomerates of Carboniferous to Triassic age [*Krahl et al.*, 1983, 1986] overlying Paleozoic metamorphic basement [*Romano et al.*, 2004, 2006; *Xypolias et al.*, 2006; *Romer et al.*, 2008]. The Phyllite-Quartzite unit probably formed the stratigraphic underpinnings of the Tripolitza unit prior to their incorporation in the Aegean fold-thrust belt [*van Hinsbergen et al.*, 2005c].

[16] The Tripolitza nappe in the forearc is underlain by the Ionian nappe, which consists of Triassic evaporites and Jurassic to Eocene well-bedded, pelagic limestones and radiolarite intervals [*IGRS-IFP*, 1966; *British Petroleum Company*, 1971; *Underhill*, 1988]. Thrusts divide the Ionian zone in western Greece and Albania into external, central and internal parts. All parts are covered by flysch that becomes thicker and coarser grained from external to internal and is indistinguishable in facies and age from the Tripolitza flysch [*IGRS-IFP*, 1966; *Richter et al.*, 1978]. Underthrusting of the Tripolitza below the Pindos nappe and of the Ionian below the Tripolitza nappe occurred simultaneously throughout the Oligocene, as shown by growth structures in seismic lines across western Greece [*Sotiropoulos et al.*, 2003]. In the middle and internal parts, flysch deposition in a foreland basin setting ceased around 23 Ma (Table 1). In the central Ionian zone, an extensional half graben developed between ∼23 and 17 Ma [*Avramidis and Zelilidis*, 2001; *van Hinsbergen et al.*, 2005c], which became inverted after sedimentation and now forms the Botsara syncline (Figure 1). In the external Ionian zone, including the parts on the Ionian islands, deep marine clastic turbidite series continued to develop until the middle Miocene, ∼15 Ma [*Bizon*, 1967; *de Mulder*, 1975], after which they became folded. On the Peloponnesos and Crete, the Phyllite-Quartzite unit is underlain by a unit of well-bedded Mesozoic to Eocene "Plattenkalk" marbles and Oligocene metaflysch, which is interpreted as the metamorphic equivalent of the Ionian zone [*Bizon and Thiébault*, 1974; *Thiébault*, 1979; *Kowalczyk and Dittmar*, 1991; *Blümor et al.*, 1994; *Kowalczyk and Zügel*, 1997] (Figure 1).

[17] In western Greece and Albania, the Ionian zone is underlain by the pre-Apulian zone, which is generally considered to be the weakly deformed slope of the Apulian platform (Figure 1). It comprises a Cretaceous to Miocene sequence of platform-slope carbonates, with platform-derived mass-transported microbreccias [*Dermitzakis*, 1978; *Bornovas et al.*, 1980; *Underhill*, 1989]. It is overlain by a clastic foreland basin deposit that starts in the Burdigalian-Langhian (∼20–15 Ma) on Kephalonia and Levkas and in Serravallian to Tortonian times (∼12–10 Ma) on Corfu and Zakynthos [*Dermitzakis*, 1978; *Underhill*, 1989; *van Hinsbergen et al.*, 2006; *Drinia and Antonarakou*, 2012]. The Pre-Apulian zone was folded and uplifted in Pliocene time (∼4 Ma), and the western part of Kephalonia and Zakynthos experienced renewed uplift of at least several hundreds of meters in the Pleistocene [*van Hinsbergen et al.*, 2006]. The first uplift phase is considered as marking final collision of the Ionian and Pre-Apulian zones, and the latter phase relates to transpressional deformation along the right-lateral Kephalonia Fault Zone that bordered the Apulian platform in the south since the Pliocene [*van Hinsbergen et al.*, 2006] (see below). The Apulian platform did probably not extend far beyond the island of Zakynthos. South of Zakynthos, subduction of the Ionian basin underlain by oceanic lithosphere (not to be confused with the Ionian nappe which was probably underlain by continental crust) has been continuous since at least ∼13 Ma and formed the Mediterranean ridge [*Finetti*, 1982; *Underhill*, 1989; *Kastens*, 1991].

[18] The Apulian platform is part of the Adria microplate. This microplate consists of a largely submerged continental block that extends from southern Italy to the Alps. The Apennines, the Alps, and the Dinarides-Albanides-Hellenides fold-thrust belts bound it in the west, north, and east, respectively. It occupies a lower plate position in the Apennines and the Dinarides-Albanides-Hellenides. GPS data suggest that at present, the Adria microplate moves eastward with respect to the African plate. Continental lithosphere of Adria is bounded in the south from the Ionian oceanic lithosphere by the Apulia escarpment, and from the western Hellenides fold-thrust belt by the Kephalonia Fault Zone [*Oldow et al.*, 2002; *Battaglia et al.*, 2004; *D'Agostino et al.*, 2008; *Caporali et al.*, 2009] (Figure 1). The total amount and periods of motion of Adria relative to Africa are debated. Assuming continuous rigidity of Adria with respect to Africa would lead to overlaps of Adria and SW Europe in Pangea reconstructions [*Wortmann et al.*, 2001; *Handy et al.*, 2010]. In addition, Adria is presently separated from Africa by the oceanic crust of the Ionian basin, which probably started to open in the Triassic [*Frizon de Lamotte et al.*, 2011; *Gallais et al.*, 2011] (Figure 1). Several workers have suggested that Adria has been mobile relative to Africa at different episodes throughout the Cretaceous and Cenozoic [e.g., *Dewey et al.*, 1989; *Wortmann et al.*, 2001; *Ustaszewski et al.*, 2008; *Handy et al.*, 2010]. An important constraint for the relative motion of Adria with respect to the Albanides-Hellenides, relevant for our restoration, is provided by the fact that the eastern margin of the Adria plate facing the Albanides has been in contraction throughout the Neogene, without evidence of any extension [*Bertotti et al.*, 2001].

### 4.2. Western Turkey

[19] Similar to Greece, western Turkey contains a late Cretaceous suture (the İzmir-Ankara suture zone, equivalent to the Sava suture) that separates the southward facing fold-thrust belt of the "Anatolide-Tauride belt" from older continental fragments that accreted to Eurasia prior to the late Cretaceous. The deepest tectonostratigraphic units are exposed in the Menderes Massif, one of the extensional metamorphic complexes of the Aegean–West Anatolian region, in the central part of western Turkey (Figure 1).

[20] Pre-Late Cretaceous accreted continental fragments north of the İzmir-Ankara suture zone consist of the İstanbul

zone, which in sedimentary facies and age resembles the Pre-Balkan unit and the Moesian platform [*Chen et al.*, 2002; *Yiğitbaş et al.*, 2004; *Bozkurt et al.*, 2008]. It became separated from Moesia due to the mid-Cretaceous to Eocene opening of the western Black Sea [*Okay et al.*, 1994]. The İstanbul zone is bordered to the south by the Sakarya zone, a microcontinental fragment that can be traced all over northern Turkey [*Okay*, 2008]. In the Ordovician and Silurian, Sakarya belonged to Gondwana [*Dean et al.*, 2000], and paleomagnetic data show that Sakarya was part of Eurasia since at least Jurassic times [*Meijers et al.*, 2010a, 2010b]. The westernmost part of Sakarya is overlain by an ophiolite-bearing accretionary complex (Cetmi Mélange [*Beccaletto et al.*, 2005]), which formed in mid-Cretaceous to Cenomanian times and is associated with latest Cretaceous HP-LT metamorphic rocks. This complex was postulated to define the boundary between Sakarya and the Circum-Rhodope-Strandja complex [*Okay et al.*, 1991; *Okay and Satir*, 2000; *Beccaletto and Jenny*, 2004; *Okay*, 2004], but see *Okay et al.* [2010] for an alternative interpretation.

[21] The İzmir-Ankara-Erzincan suture zone demarcates the location of the former northern branch of the Neotethys Ocean in Turkey [*Şengör and Yilmaz*, 1981] (Figure 1). The suture zone comprises a mélange of ocean floor volcanics, oceanic sediments with Late Triassic and younger ages, and foreland basin deposits of Cretaceous age [*Gökten and Floyd*, 2007].

[22] Below and/or south of this mélange are non-metamorphic ophiolites and underlying ophiolitic mélange, which formed during Late Cretaceous ophiolite obduction over the Anatolide-Tauride microcontinental units [*Koepke et al.*, 2002; *Çelik et al.*, 2006]. The ophiolites and sub-ophiolitic mélanges are found both in northwestern as well as southwestern Turkey and the southeasternmost Aegean region (Figure 1). In westernmost Turkey, the ophiolites are adjacent to the Bornova Flysch Zone, which formed during the rapid foundering and destruction of the Anatolide-Tauride carbonate platform during the late Cretaceous. To the east, directly south of the İzmir-Ankara suture zone, the same ophiolites overlie the HP-LT metamorphic belt of the Tavşanlı zone. The Tavşanlı zone underwent peak metamorphic conditions up to 26 kbar/500°C [*Okay*, 1981; *Davis and Whitney*, 2006; *Whitney et al.*, 2011], dated by $^{40}Ar/^{39}Ar$ ages of 88–80 Ma [*Sherlock et al.*, 1999]. The Tavşanlı zone consists of Permo-Triassic metapelitic schists, Mesozoic marbles and metabasite, metachert and phyllite at the top, interpreted as detached from the subducted distal passive continental margin of the Anatolide-Tauride mega-unit and subsequently exhumed by return flow [*Okay*, 1984, 2008].

[23] The Tavşanlı zone overlies the Afyon zone, another continent-derived sedimentary sequence containing some Paleozoic basement slivers. It was metamorphosed under blueschist facies conditions (6–9 kbar, 350°C), but around 65–60 Ma [*Candan et al.*, 2005]. A similar series characterized by 65–60 Ma HP-LT metamorphism is found to the south of the Menderes Massif (the Ören unit) and is equivalent to the Afyon zone [*Güngör and Erdogan*, 2001; *Rimmelé et al.*, 2003a, 2005; *Pourteau et al.*, 2010].

[24] The Ören unit is underlain by a belt of still younger HP-LT metamorphic rocks that were classically interpreted as part of the "Menderes cover sequence" in the underlying Menderes massif [*Schuiling*, 1962]. However, because they experienced Eocene HP-LT metamorphism [*Whitney and Bozkurt*, 2002; *Rimmelé et al.*, 2003b], which is unknown from the core of the Menderes Massif proper [*Bozkurt and Oberhänsli*, 2001], it is more likely that these rocks belong to a separate nappe older than, and overlying the core of, the Menderes Massif. Because they correlate well in terms of tectonic position with the Cycladic Blueschist unit [*Gessner et al.*, 2001b], we regard these blueschists as a lateral equivalent to the Cycladic Blueschist, known in western Turkey as the "Dilek nappe" [*Ring et al.*, 1999]. We note, however, that the rocks of this "Menderes cover sequence" contain carpholite [*Rimmelé et al.*, 2003b], while this mineral is absent in the higher-grade "Dilek nappe" and the Cycladic Blueschist unit proper. For this reason these series were parallelized with a carpholite bearing unit on Amorgos, island that underlies the Cycladic Blueschist unit by *Jolivet et al.* [2004b] and *Chatzaras et al.* [2011]. In any case, the "Menderes cover sequence" appears to be a separate thrust unit, which we prefer to treat as equivalent to the Cycladic Blueschist in terms of its tectonic position.

[25] The Menderes Massif proper forms a large window below the ophiolites and the blueschist belts described above. It consists of a series of low- to high-grade, Pan-African basement-bearing nappes [*Gessner et al.*, 2011, and references therein] (Figure 1). There is evidence for Pan-African metamorphism and magmatism in the highest of these nappes (Çine nappe) [*Candan et al.*, 2001; *Gessner et al.*, 2004; *Oberhänsli et al.*, 2010], and the degree of Alpine metamorphism is debated [*Gessner et al.*, 2004; *Bozkurt et al.*, 2011]. There is clear evidence that the lowermost 'Bayındır' nappe only experienced Alpine greenschist facies metamorphism and contains uppermost Cretaceous carbonates [*Özer*, 1998; *Özer and Sözbilir*, 2003]. This nappe yielded a $^{40}Ar/^{39}Ar$ age of $36 \pm 2$ Ma, probably marking the end of nappe accretion in western Turkey [*Lips et al.*, 2001].

[26] To the south of the Menderes massif, Upper Cretaceous ophiolites were obducted in Late Cretaceous times onto nonmetamorphic thrust sheets that are classically attributed to the "Lycian nappes." We find the published terminology, in which the Lycian nappes sometimes include, and sometimes exclude the overlying ophiolites, rather confusing. We therefore abandon the term "Lycian Nappes" and use the term "West Anatolian Taurides" for the non-metamorphic thrust slices that underlie the ophiolites and ophiolitic mélange. The West Anatolian Taurides are in front of and underneath the Ören and Afyon metamorphics but above the Cycladic Blueschists. In many places, Upper Cretaceous to upper Eocene foreland basin clastics overstep nonmetamorphic Paleozoic to lower Cenozoic carbonate-dominated sediments [*de Graciansky*, 1972; *Bernoulli et al.*, 1974; *Poisson*, 1977; *Okay*, 1989; *Collins and Robertson*, 1999]. Because the ages of these foreland basin clastics in the West Anatolian Taurides appear to cover the entire time span of metamorphism of the Tavşanlı and Afyon-Ören units as well as that of the underlying Menderes massif, they may represent nonmetamorphic equivalents of the metamorphic units described above, which remained in a fore-arc position and escaped metamorphism.

[27] The West Anatolian Taurides, together with previously obducted ophiolites, overthrust the "autochtonous" Bey

Dağları platform to the SE (Figure 1). The Bey Dağları platform consists of a Jurassic to Eocene, locally Oligocene carbonate platform [*Poisson*, 1967; *Farinacci and Köylüoglu*, 1982; *Hayward and Robertson*, 1982; *Sarı and Özer*, 2002; *Sarı et al.*, 2004]. The Bey Dağları platform is generally considered to have formerly been physically connected to the Mesozoic-Cenozoic sedimentary rocks of the Menderes massif, which together formed a continuous Menderes-Bey Dağları platform [*Collins and Robertson*, 1997, 2003]. The Bey Dağları platform is overlain by a lower Miocene (∼23–16 Ma) foreland basin series in front of and below the frontal thrust of the West Anatolian Taurides. These foreland basin clastics constrain the timing of thrusting of the West Anatolian Taurides toward Bey Dağları [*Hayward*, 1984a; *van Hinsbergen et al.*, 2010b]. Langhian (∼16–13 Ma) sediments seal the thrust [*Poisson et al.*, 2003; *Flecker et al.*, 2005]. A window near Göçek exposes Burdigalian sediments of similar foreland basin facies [*Brunn et al.*, 1970; *Hayward*, 1984b] demonstrating at least 75 km of southeastward displacement of the West Anatolian Taurides relative to Bey Dağları between 23 and 16 Ma (Table 1). *van Hinsbergen* [2010], and earlier *Hayward* [1984b] and *Collins and Robertson* [2003], restored the West Anatolian Taurides back toward the NW on top of the Menderes Massif at 25 Ma and suggested that southeastward translation between ∼25 and ∼15 Ma was gravity driven and only affected a few kilometers thick tectonostratigraphic sequence of West Anatolian Taurides and overlying ophiolites. We adopt this reconstruction here and refer to *van Hinsbergen* [2010] for a detailed discussion.

[28] Finally, the present-day Aegean subduction zone is marked by an accretionary wedge, the "Mediterranean ridge" [*Finetti*, 1976; *Mascle et al.*, 1986; *Chaumillon et al.*, 1996; *Huguen et al.*, 2001; *Zitter et al.*, 2003; *ten Veen et al.*, 2004]. There is no evidence for any accretion of nappes since the late Eocene at the active continental margin south of Turkey, other than the Mediterranean ridge.

## 5. Reconstruction of the Aegean Region

[29] We now reconstruct the deformation that affected the above described nappe stack as a result of Cenozoic compressional and extensional deformation. To this end, we apply a step-by-step reconstruction approach. The bulk of extension in the area has been accommodated by the opening of the Rhodope, the eastern mainland Greece–Cyclades, the Menderes and the South Aegean extensional metamorphic complexes between Eocene and latest Miocene time. These are frequently referred to as "metamorphic core complexes," but because these areas typically contain multiple core complexes and extensional detachments [e.g., *Jolivet et al.*, 2004a; *Tirel et al.*, 2009], we prefer the term "extensional metamorphic complex." After the late Miocene, these complexes were in turn deformed by smaller extensional and strike-slip deformation.

[30] We first summarize and reconstruct the late Miocene to Recent deformation, after which we will restore the extensional metamorphic complexes one by one. Finally, we restore the accretion of nappes in the time window of our reconstruction (∼35 Ma), and show paleotectonic maps that display the final result of the reconstruction.

### 5.1. Late Cenozoic Deformation

#### 5.1.1. Kinematic Constraints

[31] The kinematic constraints described below are summarized in Table 1 and Figure 5. The most prominent young deformation feature in the Aegean–West Anatolian region is the North Anatolian Fault Zone (NAFZ), connecting westward to the North Aegean Trough. This fault system accommodates westward motion of Anatolia relative to stable Eurasia, which continues today [*Reilinger et al.*, 2010]. The NAFZ is a >1000 km long, northward convex, curved, dextral strike-slip fault system in northern Turkey [*McKenzie*, 1972; *Şengör and Canitez*, 1982]. Its total displacement was estimated by *Armijo et al.* [1999] and *Hubert-Ferrari et al.* [2002] at ∼85 km, based on offset riverbeds and geological markers. *Şengör et al.* [2005], however, suggested that the maximum displacement may be smaller, $\sim 60 \pm 20$ km. The same authors, moreover, argued that the total displacement budget along the North Anatolian Fault decreases westward, and may be minimal in the Sea of Marmara (4 km since 200 kyr). Basins that developed along the eastern part of the NAFZ started forming at ∼11 Ma, following upon collision of Arabia and Anatolia in SE Turkey [*Şengör et al.*, 2003, 2005; *Faccenna et al.*, 2006; *Hüsing et al.*, 2009]. *Hubert-Ferrari et al.* [2009] showed that a ∼4 Ma old volcano along the eastern part of the NAFZ was displaced by ∼50 km, but part of this displacement may have been accommodated within Turkey. Given the variety in NAFZ displacement estimates, we estimate the total amount of displacement based on the deformation pattern in Greece, and return to this issue in the discussion.

[32] In northern Greece, the NAFZ connects through the Sea of Marmara to the North Aegean Trough (Figure 1), a transtensional basin that branches into a transtensional horsetail splay marking the western termination of the NAFZ offshore eastern mainland Greece [*Papanikolaou et al.*, 2002]. An age of ∼5 Ma was inferred for the onset of transtensional opening of the Sea of Marmara [*Armijo et al.*, 1999], based on the absence of Messinian gypsum (6.0–5.3 Ma) and no evidence for a major transgression (Pliocene flooding) that is typical for the Mediterranean region at 5.3 Ma [*Krijgsman et al.*, 1999].

[33] Northern Greece is decoupled from western Greece, which currently moves southwestward relative to stable Eurasia [*Reilinger et al.*, 2010] along a curved system of (half) grabens. These include, from north to south, the Amvrakikos-Sperchios graben, the Gulf of Corinth graben, the Pyrgos-Gythion graben system and the Melimala-Kalamata graben system [*van Hinsbergen et al.*, 2006] (Figure 5). Parts of them were combined into the conceptual "Central Hellenic Shear Zone" by *Royden and Papanikolaou* [2011]. Detailed estimates of ages and displacements are only available for the Gulf of Corinth. The Gulf of Corinth half graben has a total extensional displacement of up to ∼15 km [*Jolivet et al.*, 2010a; *Taylor et al.*, 2011] since ∼3.5 Ma [*Rohais et al.*, 2007]. Southeastward, the Gulf of Corinth connects through the Isthmus of Corinth, which exposes lower Pliocene marine sediments [*Leeder et al.*, 2008], to the Saronic Gulf, in which the island of Aegina underwent strong subsidence since ∼5 Ma [*van Hinsbergen et al.*, 2004]. This suggests that extension may have started slightly earlier in the east than in the west. Displacements associated with the

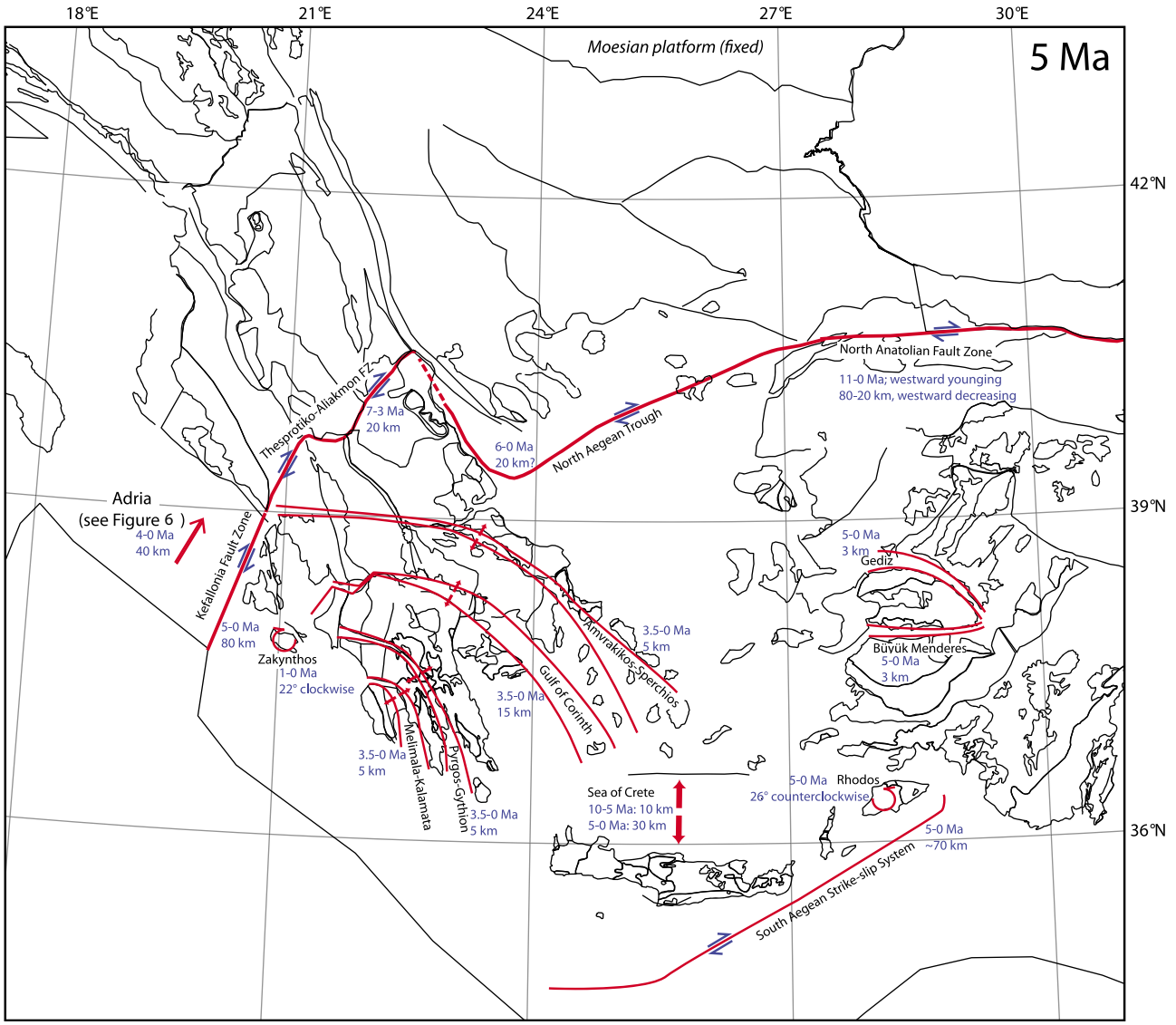



**Figure 5.** Reconstruction at 5 Ma, detailing the young brittle structures that deformed the Aegean–West Anatolian region that were used in this reconstruction, with their ages and displacements. See also Table 1.

formation of the Amvrakikos-Sperchios [*Kilias et al.*, 2008], Pyrgos-Gythion and Melimala-Kalamata graben systems [*van Hinsbergen et al.*, 2006] remain undocumented but are probably smaller. We assume up to 5 km of extension orthogonal to their trends along each of these graben systems, and apply a minor clockwise rotation of ~0.5–1° about a pole around their (north) western terminations to account for their eastward increasing widths [*van Hinsbergen et al.*, 2006] (Figure 5).

[34] Prior to the late Pliocene initiation of the Gulf of Corinth and Amvrakikos-Sperchios extensional systems, motion of western Greece relative to northern Greece was accommodated along the dextral Thesprotiko strike-slip system [*Jordan et al.*, 2005], which offsets lower Miocene sediments in the Botsara syncline of western Greece by ~20 km (Figures 1 and 5) [*van Hinsbergen et al.*, 2005c, 2006]. The age of displacement is probably younger than 7 Ma, based on the age of vertical axis rotations along the Thesprotiko- Kephalonia Fault Zone [*van Hinsbergen et al.*, 2005b]. Deformation along the Thesprotiko shear zone must have continued until after deposition of faulted and strongly rotated ~4.5 Ma old sediments within the shear zone [*van Hinsbergen et al.*, 2006]. To the northeast, the Thesprotiko shear zone probably connected to the Aliakmon fault zone [*Doutsos and Koukouvelas*, 1998; *Jordan et al.*, 2005], a diffuse system of strike-slip and transtensional faults in northern Greece that may have transferred displacement along the North Aegean Trough to the subduction zone in the west. The intramontane Karditsa and Larissa basins located south of the Aliakmon transtensive fault system formed in response to late Miocene–Pliocene NE-SW extension, followed by minor late Plio-Pleistocene ~N-S extension [*Caputo and Pavlides*, 1993; *Caputo et al.*, 1994]. These basins gradually disappear southward and are roughly triangular shaped. The amount of extension is unknown, but these basins serve to illustrate that the 20 km of displacement along the Thesprotiko shear zone is a maximum estimate for

the amount of displacement that was transferred from the North Aegean Trough to the subduction zone.

[35] The Kephalonia Fault Zone is a transform fault that has been active since the collision of Apulia with the western Aegean forearc around ~5 Ma [*van Hinsbergen et al.*, 2006; *Royden and Papanikolaou*, 2011]. In map view this transform fault offsets the presently active Aegean subduction zone by some 100–120 km [*Royden and Papanikolaou*, 2011] (Figure 1). Because the Kephalonia Fault Zone is a transform fault this does by no means imply an offset of a formerly contiguous subduction zone as suggested by *Royden and Papanikolaou* [2011]. Absence of offset markers across the Kephalonia Fault Zone precludes estimating the displacement it accommodated. We estimate this displacement by summing the offset along the Thesprotiko shear zone (~20 km) and summing the extensional displacements in the Amvrakikos-Sperchios and Gulf of Corinth graben systems (~20 km). We return to this issue below. We restored a local, 20° clockwise rotation of the island of Zakynthos in the last 1 Ma, following *Duermeijer et al.* [1999], which probably relates to dragging along the Kephalonia Fault Zone.

[36] In the south, the Sea of Crete hosts the largest graben in Greece. It formed due to ~N-S extension associated with a $\beta$ factor of ~1.7 [*Angelier et al.*, 1982], i.e., ~40 km N-S extension. Approximately 75% of the subsidence in the Sea of Crete is younger than 5 Ma [*Wright*, 1978; *Meulenkamp et al.*, 1994]. The high-angle normal faults of the Sea of Crete crosscut the Cretan Detachment (Figure 2), which was active until ~10.4 Ma [*van Hinsbergen and Meulenkamp*, 2006; *Zachariasse et al.*, 2011]. We apply ~30 km N-S extension in the Sea of Crete since 5 Ma, and 10 km between 10 and 5 Ma (Figure 5).

[37] The island of Crete became since ~9.6 Ma dissected by high-angle basin-bounding normal faults that accommodated both N-S and E-W stretching [*Fassoulas*, 1998, 2001; *ten Veen and Postma*, 1999; *ten Veen and Kleinspehn*, 2003; *van Hinsbergen and Meulenkamp*, 2006; *Zachariasse et al.*, 2011]. There is no estimate of the total amount of displacement along the high-angle E-W extensional normal faults, but it was probably limited. We model Crete as a rigid block for the last 10 Ma.

[38] In the southeast, the Aegean region is bounded from the African plate by a sinistral strike-slip zone, which includes the Pliny and Strabo trenches (Figure 1) [*Peters and Huson*, 1985; *Mascle et al.*, 1999]. Onshore, Crete and Rhodos contain transpressional strike-slip faults parallel to the Pliny and Strabo trenches [*ten Veen and Meijer*, 1998; *ten Veen and Kleinspehn*, 2002, 2003]. Both islands underwent local counterclockwise Plio-Pleistocene vertical axis rotations [*Duermeijer et al.*, 1998; *van Hinsbergen et al.*, 2007], suggesting they are included in the wider South Aegean strike-slip system. This strike slip zone is interpreted as a subduction-transform (STEP) fault that accommodates southwestward retreat of the Hellenic trench relative to Eurasia [*Govers and Wortel*, 2005; *van Hinsbergen et al.*, 2007, 2010c; *Zachariasse et al.*, 2008]. The South Aegean strike-slip system terminates in the transtensional, 4 km deep Rhodos basin (Figure 1), which does not contain Messinian evaporite deposits [*Woodside et al.*, 2000]; hence, the onset of the South Aegean strike slip system is younger than 5.3 Ma [*Krijgsman et al.*, 1999]. Crete underwent ~1 km uplift between ~5 and 3 Ma [*ten Veen and Kleinspehn*, 2003; *van Hinsbergen and Meulenkamp*, 2006; *Zachariasse et al.*, 2008], which was interpreted as a rebound effect of the subduction transform fault [*Zachariasse et al.*, 2008].

[39] The stratigraphy of the island of Rhodos suggests that it has been connected to mainland Turkey until the late Pliocene [*Meulenkamp et al.*, 1972]. It experienced late Miocene to Pliocene extension [*ten Veen and Kleinspehn*, 2002] associated with ~30° counterclockwise vertical axis rotation since ~4 Ma [*van Hinsbergen et al.*, 2007], which is incorporated in the reconstruction.

[40] Finally, the West Anatolian Taurides and overlying ophiolites in southwestern Turkey were crosscut by several high-angle normal faults [e.g., *Alçiçek*, 2010], which we do not reconstruct in detail. We restore minor, i.e., 3 km, N-S extension since 5 Ma associated with the Büyük Menderes and Gediz grabens, following the studies by *Çiftçi and Bozkurt* [2010] and *Çifçi et al.* [2011].

#### 5.1.2. Reconstruction

[41] Figure 5 shows the reconstructed configuration of the Aegean region as a result of brittle deformation since the late Miocene. Most of the reconstruction is straightforward and of minor magnitude. Only two issues deserve some discussion: the amount of displacement that was transferred from the North Anatolian Fault to the Kephalonia Fault Zone, and the motion history of Adria relative to Africa.

[42] GPS measurements show that Anatolia moves westward relative to Eurasia, and the Aegean region moves to the SW at a higher rate. In other words, Anatolia and the Aegean region presently share a motion path that, going west, is curving to a southwestwardly direction. Superimposed on this displacement field, the Aegean region experiences net ~NE-SW extension [*McClusky et al.*, 2000; *Reilinger et al.*, 2006, 2010; *Caporali et al.*, 2009; *Reilinger et al.*, 2010] (Figure 2). When investigating a longer timescale, there may be some ~E-W shortening in the Cyclades expressed as late Miocene folds [*Avigad et al.*, 2001] but most Anatolian motion must have transferred to the Hellenic trench, for the last 5 Ma to the Kephalonia Fault Zone [*Armijo et al.*, 1999, 2004; *van Hinsbergen et al.*, 2006; *Royden and Papanikolaou*, 2011]. In the south, Anatolian extrusion is transferred to the South Aegean strike-slip system.

[43] At most 20 km of right-lateral strike-slip motion of the Aegean region relative to Eurasia was accommodated along the Thesprotiko-Aliakmon fault zone, active between ~7 and 3.5 Ma [*van Hinsbergen et al.*, 2006] (Table 1). The Gulf of Corinth and the Amvrakikos-Sperchios basins added an additional ~20 km to displacement of the Kephalonia Fault Zone, but this displacement seems to result from extension within the Aegean region, rather than transfer of motion from the Anatolian Fault to the Kephalonia Fault Zone. We therefore conclude that not more than 20 km of North Anatolian Fault Zone displacement was transferred to Greece. The conceptual "Central Hellenic Shear Zone" was postulated to transfer in a diffuse way much more (80 km) of North Anatolian Fault displacement to the trench [*Papanikolaou and Royden*, 2007; *Royden and Papanikolaou*, 2011]. Since as little as 15 km of extension is able to create the Gulf of Corinth graben [*Jolivet et al.*, 2010a; *Taylor et al.*, 2011], and 20 km of displacement of the Thesprotiko Shear Zone is easily recognized [*Jordan et al.*, 2005], we do not see how an additional 60 km of

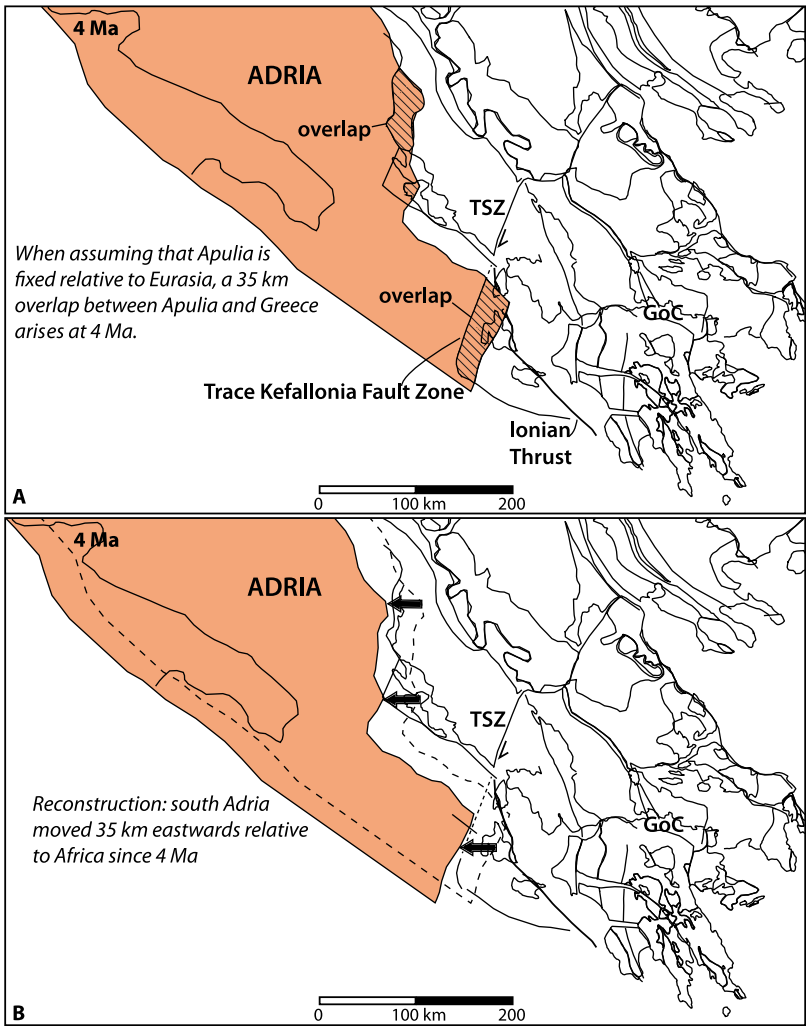



**Figure 6.** Reconstruction showing the overlap between Greece and Adria that arises when one assumes that Adria was fixed relative to Africa in the last 4 Ma. This overlap can be solved by reconstructing Apulia ~35 km to the west relative to Africa, consistent with the directions shown in GPS surveys [*D'Agostino et al.*, 2008]. We suggest that that current kinematic pattern has existed for the last 4 Ma. Prior to 4 Ma, there is no overlap and Adria can be fixed relative to Africa for as far Greece is concerned. See text for further discussion. TSZ = Thesprotiko Shear Zone; GoC = Gulf of Corinth.

right-lateral shear, even if distributed over multiple fault zones, would have remained unnoticed in western Greece. Rather, a considerable part of NAFZ displacement must have been accommodated within Anatolia (e.g., in the Çankırı region in north central Turkey [*Kaymakci et al.*, 2009]) and hence was not transferred into the Aegean region. This inference is consistent with the conclusion of *Şengör et al.* [2005] that NAFZ displacements substantially decrease, and almost disappear, westward.

[44] If one assumes a rigid Adria-Africa plate, a problem arises in the area NW of the Kephalonia Fault Zone for the last 4 Ma. This assumption, in combination with retrodeformation indicated in Figure 5, would lead to an overlap of the Adria plate in the area of the Albanian coast when reconstructing back to 4 Ma (Figure 6). If that overlap would be real, it would require post-4 Ma ~N-S extension between Adria and Albania/western Greece, which is inconsistent with the evidence for NE-SW compression in the area [*Meco and Aliaj*, 2000; *Bertotti et al.*, 2001; *van Hinsbergen et al.*, 2006]. GPS data demonstrate that Adria presently moves northeastward relative to Eurasia, along a small circle parallel to the Kephalonia Fault Zone, and almost eastward relative to Africa [*Battaglia et al.*, 2004; *D'Agostino et al.*, 2008]. Following the GPS directions, we therefore restore Adria ~35 km to the west relative to Africa at 4 Ma (Figure 6), so that its southeastern margin aligns with the Kephalonia Fault Zone. This leads to a 40 km northeastward translation of Adria relative to Eurasia in the last 4 Ma, leading to an average rate of ~9 mm/yr relative to Africa (in agreement with a retrodeformation by *Ustaszewski et al.* [2008] but approximately twice the GPS-determined velocity [*D'Agostino et al.*, 2008]). This displacement is accommodated along the Kephalonia Fault Zone, in addition to the 40 km of displacement of southwestern relative to northwestern Greece summarized above. This leads to a total displacement since the late Miocene along the Kephalonia Fault Zone of 80 km. Prior to 4 Ma, the assumption that Adria was fixed relative to Africa does not create inconsistencies in the

reconstruction of Greece. It is possible that Adria was also mobile with respect to Africa before, as postulated by *Ustaszewski et al.* [2008]; our restoration of the Aegean region cannot confirm or exclude this.

### 5.2. Rhodope Extensional Metamorphic Complex

#### 5.2.1. Kinematic Constraints

[45] Exhumation in the Rhodope by extensional unroofing led to the formation of a triangle shaped metamorphic complex (Figure 1), with extension-related stretching lineations gradually changing from NNE-SSW in the north, to NE-SW in the southwest of the complex [*Walcott and White*, 1998; *Brun and Sokoutis*, 2007; *Burg*, 2011] (Figure 2). An important extensional detachment in the Rhodope is the top-to-the-southwest Strymon detachment (Figure 1), which was dominantly active from early Miocene time (~25–20 Ma) onward as suggested by $^{40}Ar/^{39}Ar$ cooling ages [*Dinter and Royden*, 1993; *Sokoutis et al.*, 1993; *Dinter et al.*, 1995; *Dinter*, 1998; *Wawrzenitz and Krohe*, 1998; *Lips et al.*, 2000], lasting until the late Miocene time as suggested by fission track ages in the footwall of the Strymon detachment that cluster between 5 and 10 Ma [*Hejl et al.*, 2010], and even until the earliest Pliocene as indicated by ongoing basin subsidence in the Strymon valley [*Snel et al.*, 2006]. Based on the patterns of cooling ages across the Rhodope, *Brun and Sokoutis* [2007] suggested that the Rhodope underwent a two-stage extensional exhumation evolution, whereby the Miocene Strymon detachment developed within a footwall of an older top-to-the-SW extensional detachment that was active since ~40 Ma. Late Eocene extension is also evident from the upper Eocene Mesta half graben in the northwestern tip of the triangular Rhodope extensional complex in Bulgaria [*Burchfiel et al.*, 2003, 2008] and the development of the extensional Thrace basin to the east between early/middle Eocene and Oligocene time, with a stratigraphic thickness of up to 9 km (Figure 1) [*Turgut and Eseller*, 2000; *Siyako and Huvaz*, 2007; *Okay et al.*, 2010]. Because of the triangular shape and the gradually curving trend of the stretching lineations within the Rhodope, *Brun and Sokoutis* [2007] suggested that the post-middle Eocene evolution of the Rhodopian extensional metamorphic complex is best described by a clockwise rotation of the entire southwestern hanging wall of the metamorphic Rhodopes, including the Chalkidiki peninsula, with respect to the footwall. They determined an Euler pole for this rotation near the northwestern tip of the Rhodope triangular complex, and estimated a rotation angle of ~30°. Using the rotation pole of *Brun and Sokoutis* [2007] and the shape of the Rhodope extensional metamorphic complex, we estimate a slightly smaller finite rotation of ~25°. Such a rotation is consistent with paleomagnetic evidence, which shows 17.9 ± 12.6° clockwise rotation of Chalkidiki with respect to the Moesian platform since the Eocene [*Kondopoulou and Westphal*, 1986; *van Hinsbergen et al.*, 2008]. Western Greece and Albania rotated 38.0 ± 7.2° relative to the Moesian platform since the early Oligocene [*van Hinsbergen et al.*, 2008]. The bulk of this rotation occurred after ~15 Ma, and prior to ~3.5 Ma [*van Hinsbergen et al.*, 2005b] and pre-middle Miocene rotations in western Greece fall within typical paleomagnetic error bars of 5–10° (which could still lead to considerable extension far away from the rotation poles). Therefore, *van Hinsbergen et al.* [2008] suggested that significant vertical axis rotations in the Rhodope occurred mainly in Miocene times. This is consistent with the conclusions of *Georgiev et al.* [2010], who showed that especially the last phase of extension accommodated by the Miocene Strymon detachment was associated with major vertical axis rotation. Because the southeastern part of the Rhodope extensional complex is submerged, the structure of the eastern end of the Rhodope metamorphic complex remains unknown. The Biga peninsula of NW Turkey, however, was only locally (in the north (Kemer micaschist [*Beccaletto et al.*, 2007]) and south (Kazdağ massif [*Bonev et al.*, 2009; *Cavazza et al.*, 2009])) affected by Eocene and younger extension, and NW Turkey was not affected by major vertical axis rotations [*Beck et al.*, 2001; *van Hinsbergen et al.*, 2010a]. The extension history of the Rhodope therefore seems to have undergone two stages. In first stage, since early Eocene time, extension affected the area from the Mesta Graben in the northwest to the Thrace basin in the east. In a second stage, major extension (~120 km [*Brun and Sokoutis*, 2007]) occurred along the Strymon detachment and was associated with ~25° of vertical axis rotation of the Chalkidiki peninsula and the areas to the south and west. This phase of extension presumably started around 25 Ma ago, when extension in the Thrace basin ceased and the basin became inverted and compressed instead [*Turgut and Eseller*, 2000], whereas rapid cooling of the Rhodope metamorphics began [*Dinter and Royden*, 1993; *Sokoutis et al.*, 1993; *Dinter et al.*, 1995; *Dinter*, 1998; *Wawrzenitz and Krohe*, 1998; *Lips et al.*, 2000]. The rotating domain must have been bordered by a structure to the west of the Biga peninsula and Lesbos [see also *Brun and Sokoutis*, 2010].

#### 5.2.2. Restoration: Rhodope Extensional Metamorphic Complex

[46] Figure 7 shows the first-order reconstruction of the Rhodope extensional metamorphic complex, indicating the two stages of exhumation summarized above. In a first stage, extension affected a wide region from the Mesta Graben in the NW Rhodope to the Thrace basin in the east. The total amount of extension associated with this exhumation is poorly constrained, but was probably limited: it did not produce vertical axis rotations beyond paleomagnetic error bars, and probably did not exhume large metamorphic domains in Eocene-Oligocene time. There are no documented large strike-slip faults at the tips of the extensional system that bound the extensional system. Therefore, we restore the pre-~25 Ma extension history by opening a small lense-shaped extensional domain, with extension decreasing toward rotation poles in the NW and east, with a somewhat arbitrary but small maximum amount of 15 km of extension in the central Rhodope, accommodated by opposite, small rotations of ~2° around poles in the NW tip of the Mesta Graben and the eastern Thrace basin. At ~45 Ma, around which time Cenozoic extension probably started [*Brun and Sokoutis*, 2010], we still indicate a zone of exhumed rocks in the northern Aegean region: the Thrace basin unconformably overlies blueschists with ~86 Ma Rb/Sr cooling ages resulting from late Cretaceous to Paleogene exhumation that predates the time window of our reconstruction [*Topuz et al.*, 2008].

[47] Between 25 and 5 Ma, we restore the bulk of extensional exhumation in the northern Aegean region, accommodated along the Strymon detachment (Figure 7). Opening

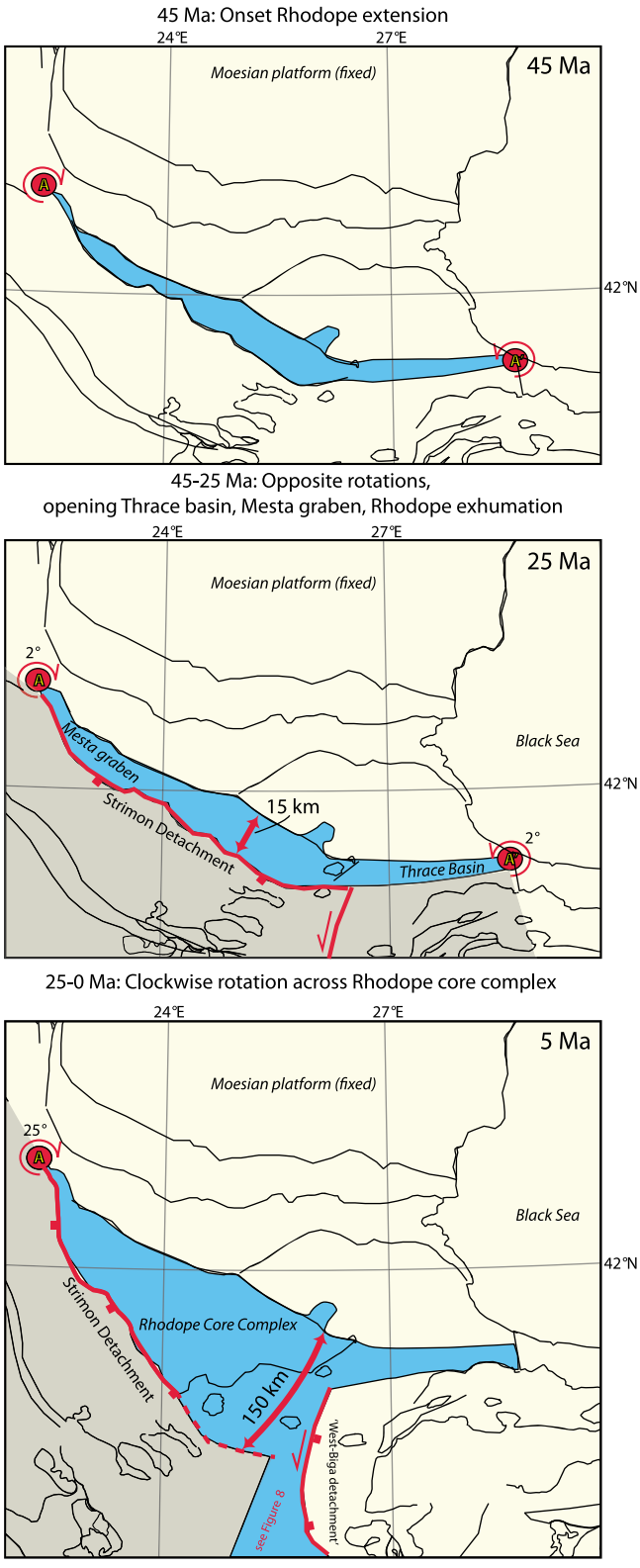



**Figure 7.** Reconstruction of the Cenozoic history of the Rhodope extensional metamorphic complex relative to the Moesian platform. See text for explanation.

the exhumed area presently covered by the Rhodope metamorphics around the pole in the NW of the Mesta Graben, as postulated by *Brun and Sokoutis* [2007], leads to a clockwise rotation of 25°. Toward the east, the Strymon detachment must have been bounded by a ~N-S trending structure (Figure 7), i.e., originally approximately orthogonal to the Strymon detachment, because no significant extension or rotation has been documented on the Biga peninsula. At the inception of major extension along the Strymon detachment, ~25 Ma ago, this structure must have been dominated by left-lateral translation, transferring the extension of the Rhodope to structures in the Cyclades [see *Brun and Sokoutis*, 2010], but upon ongoing rotation, such a structure would become increasingly transtensional and we conceptually draw this structure in Figure 7 as the "West Biga Detachment." We return to the effects of this structure below after introducing the kinematic constraints on extension in western mainland Greece and in the NW Cycladic region.

### 5.3. Eastern Mainland Greece-Cycladic Extensional Metamorphic Complex

#### 5.3.1. Kinematic Constraints

[48] Almost all Cycladic islands contain evidence for the presence of extensional detachments and metamorphic core complex formation. The Cyclades contains multiple metamorphic domes interpreted as core complexes. The hanging wall of these detachments are typically formed by ophiolites, which, in the west, correlate to the Jurassic ophiolites that overlie the Pelagonian zone (West Vardar ophiolites in the sense of *Schmid et al.* [2008]) and in the east to the Cretaceous age obducted ophiolites that are widespread in western Turkey (e.g., on Ikaria [*Photiades*, 2002; *Pe-Piper and Photiades*, 2006]). For detailed reviews of the structure of the Cycladic core complexes, we refer to *Jolivet et al.* [2004b, 2010b], *Tirel et al.* [2009] and *Ring et al.* [2010]. For the purpose of our reconstruction, it is most important to note that the Cycladic extensional metamorphic complex is subdivided into a northwestern and a southeastern part, defined by markedly different stretching lineation trends [*Walcott and White*, 1998] (Figure 2).

[49] The northwestern Cyclades are characterized by consistently NE-SW trending stretching lineations, and an overall top-to-the-NE sense of shear [*Gautier and Brun*, 1994; *Walcott and White*, 1998]. The westernmost islands of Kea, Kythnos and Serifos, which have a top-to-the-SW sense of shear [*Grasemann and Petrakakis*, 2007; *Iglseder et al.*, 2009; *Grasemann et al.*, 2012], are the exception. The southeastern Cycladic islands expose extensional detachments with a consistent top-to-the-north sense of shear [*Gautier et al.*, 1993; *Kumerics et al.*, 2005]. The structure of the island of Ios, however, is dominated by top-to-the-south sense of shear, originally interpreted as an extensional detachment [*Lister et al.*, 1984; *Vandenberg and Lister*, 1996], but recently recognized as a nappe stacking-related thrust instead [*Huet et al.*, 2009]. Paleomagnetic data from synextensional middle Miocene granites on Tinos and Mykonos in the northwestern Cyclades (Figure 1) show evidence for ~20° clockwise rotation [*Morris and Anderson*, 1996; *Avigad et al.*, 1998]. The island of Naxos in the southeastern domain, however, underwent a contemporaneous counterclockwise rotation of ~30°. *Walcott and White* [1998] defined the boundary between the two domains as

the "Mid-Cycladic Lineament" (Figure 2), which accommodated opposite rotations on either side since at least middle Miocene time. The nature of this structure remains poorly resolved and enigmatic because it resides mainly offshore.

[50] The northwestern Cycladic metamorphic complexes are bounded in the northwest by a ~NE-SW trending, steep (subvertical) fault, which we refer to as Evvia-Attica fault, along which the Cycladic units were exhumed and became folded [*Xypolias et al.*, 2003; *Ring et al.*, 2007a]. This fault runs across the island of Evvia and the Attica peninsula (Figure 1), juxtaposing the Basal Unit (metamorphic equivalents of the Gavrovo-Tripolitza zone) in the southeast with the Pelagonian megaunit in the northwest. The Attica-Evvia fault trends more or less parallel to the dominant stretching lineation directions in the adjacent Cycladic units (Figure 2). Because of this, and because it cannot be traced further to the southwest into the Peloponnesos, we infer that this structure represents a transfer fault within the eastern mainland Greece–Cycladic metamorphic extensional complex. It transfers extension accommodated in the Cyclades as far southwest as west of the islands of Kea and Kythnos into extension in the Aegean Sea east of Evvia. As shown in Figure 1, we locate the southwestern limit of the northwestern Cycladic metamorphic complex somewhere between the westernmost Cycladic islands and the Peloponnesos (Figure 1).

[51] The Cycladic Blueschist unit, constituting the lateral equivalent of the Pindos Zone underwent a two-stage exhumation history. A first stage is associated with decompression from peak pressure conditions around ~20 kbar prevailing at around 55–45 Ma to ~8 kbar by about 30–25 Ma [e.g., *Parra et al.*, 2002; *Putlitz et al.*, 2005; *Bröcker and Franz*, 2006; *Lagos et al.*, 2007; *Ring et al.*, 2007b; *Jolivet and Brun*, 2010; *Philippon et al.*, 2011; *Schneider et al.*, 2011] (see overview by *van Hinsbergen et al.* [2005c]). The Basal Unit, laterally correlated to the Tripolitza megaunit, became buried since ~35 Ma, overlapping in time with the final stages of this first stage of exhumation the Cycladic Blueschist unit, and reached peak metamorphic conditions of ~8 kbar/400°C by 24–21 Ma [*Ring et al.*, 2001]. After underthrusting of the Basal Unit, ductile-to-brittle extensional detachments exhumed the Cycladic Blueschist unit in a second stage, during which also the Basal Unit came to the surface. This second stage of exhumation was associated with crustal stretching and the timing of its inception is inferred from the age of supradetachment basins preserved on Naxos and Paros, which started to form in the early Miocene, but only received metamorphic debris since late Miocene (~10 Ma) time [*Sánchez-Gómez et al.*, 2002]. Apatite fission track ages from the Cycladic rocks cluster between 5 and 15 Ma [*Hejl et al.*, 2002, 2008; *Kumerics et al.*, 2005; *Brichau et al.*, 2006, 2007, 2008, 2010]. The minimum amount of area gain associated with extension in the NW Cyclades is ~125 km parallel to the dominant NE-SW stretching direction. The total amount of extension accommodated in central Greece is probably larger since the nature of the rocks underlying the Aegean Sea remains unknown. We note that the North Aegean Sea is underlain by continental crust of only ~25 km thickness or less [*Tirel et al.*, 2004], as opposed to the 40–45 km thick crust in the western Aegean-Dinarides and SW Anatolian regions south of the Menderes Massif [*Mutlu and Karabulut*, 2011]. This suggests that a large part of the North Aegean Sea underwent significant extension.

[52] Northwest of the Attica-Evvia fault zone, the Attica and Evvia peninsulas expose no metamorphic rocks similar to the Cycladic complex. However, as mentioned in section 4.1, rocks corresponding to the Basal Unit and the Cycladic Blueschist reappear along the east coast of mainland Greece, in windows on the Pelion peninsula [*Lips et al.*, 1999; *Lacassin et al.*, 2007], and in the cores of Mount Ossa [*Lips et al.*, 1998] and Mount Olympos [*Godfriaux*, 1970; *Schermer*, 1990, 1993; *Nance*, 2010]. Considerable exhumation by extension of probably Miocene age is also observed within the Pelagonian units north of Mount Olympos [*Diamantopoulos et al.*, 2009] (Figure 1). These extensional windows are associated with consistently NE-SW oriented stretching lineations, i.e., parallel to the extension direction on the northwestern Cyclades, showing top-to-the-SW sense of shear along the western margin of the exhumed metamorphics and top-to-the-NE sense of shear along the east coast of mainland Greece (Figure 2). $^{40}Ar/^{39}Ar$ geochronology suggests that following an Eocene phase (~43–39 Ma) of rapid exhumation of rocks equivalent to the Cycladic Blueschist unit from blueschist to greenschist facies metamorphism, cooling was slow until early to middle Miocene (20–15 Ma) times. This was followed in a second stage by extensional exhumation, associated with dominantly top-to-the-NE shearing [*Lacassin et al.*, 2007]. Sparse apatite fission track ages from these rocks are ~18 Ma (Mount Olympos) and ~11 Ma (Pelion), suggesting middle Miocene final exhumation [*Hejl et al.*, 2008]. The extensional metamorphic system of eastern mainland Greece fades out toward the border region with the Former Yugoslav Republic of Macedonia, where metamorphism, partly high pressure, and subsequent exhumation are latest Jurassic to Cretaceous in age [*Kilias et al.*, 2010]. Low-temperature geochronological ages cluster in the Late Cretaceous to Eocene [*Most*, 2003; *Vamvaka et al.*, 2010] and no evidence for Cenozoic extension has been reported.

#### 5.3.2. Restoration: Eastern Mainland Greece and Northwestern Cycladic Metamorphic Complexes

[53] Because our restoration aims to provide a surface reconstruction of the Aegean region, we first need to discuss the surface expressions that different modes of exhumation may have. Since the recognition of metamorphic core complexes in the Aegean region [*Lister et al.*, 1984; *Buick and Holland*, 1989; *Avigad and Garfunkel*, 1991; *Gautier et al.*, 1993], many authors have equated *exhumation* with *crustal thinning* and suggested that all exhumation of metamorphic rocks was related to whole crustal extension in the Aegean region parallel to the roll-back direction, especially because exhumation rates are much higher than can be reasonably accounted for by erosion [e.g., *Forster and Lister*, 2009]. However, *Trotet et al.* [2001], *Parra et al.* [2002], *Jolivet et al.* [2003], *Ring and Layer* [2003] and later *Ring and Glodny* [2010] and *Ring et al.* [2010] pointed out that particularly the exhumation history of HP-LT metamorphic rocks may frequently occur in two stages. In a first stage, rocks typically exhume from their peak pressure conditions (of, e.g., ~20 kbar for the Cycladic Blueschist [*Jolivet et al.*, 2003]) to lower crustal levels (e.g., ~6–8 kbar [*Jolivet et al.*, 2003]) by return flow within the accretionary

prism above the subducting slab. In this stage, rocks moved toward the Earth's surface as a result of their buoyancy relative to the overlying mantle wedge and/or the ductile lower crust. Such return flow produces a normal fault at the top of the exhuming wedge that is unrelated to crustal thinning and produces what is referred to as "synorogenic extension" by *Jolivet et al.* [2003]. This stage does not necessarily have a surface expression and, in the absence of crustal thinning, we find the term "synorogenic extension" is misleading and prefer the term "exhumation by return flow." This stage is followed by "exhumation by extension" that leads to crustal thinning, during which metamorphic rocks exhume to the surface along extensional detachments ("postorogenic" extension in the sense of *Jolivet et al.* [2003]). This second stage of exhumation invariably has a surface expression and is relevant for our reconstructions since it involves quantifiable surface shape and area changes. To identify this second stage, it is to identify at what stage in history extensional sedimentary basins started forming is useful, and to investigate the cooling history reconstructed from low-temperature thermochronological methods, indicating when these rocks reached near-surface levels.

[54] Some authors suggested that the first stage of exhumation of the Cycladic Blueschist unit from peak pressure conditions around ~20 kbar prevailing at around 55–45 Ma to ~8 kbar by about 30–25 Ma were related to crustal thinning [e.g., *Forster and Lister*, 2009]. However, during the time interval related to this first exhumation phase, the Basal Unit, equivalent to the Tripolitza megaunit, only started to enter the subduction zone, and directly underthrusted below the Cycladic Blueschist, still being attached to, or in any case moving down with the subducting African slab. Pre-latest Oligocene exhumation of the Cycladic Blueschist therefore occurred immediately above and along the subduction zone that buried the Basal Unit, and below the previously accreted units in the upper plate, i.e., the Pelagonian megaunit and overlying ophiolites [*Ring et al.*, 2007a, 2007b; *Huet et al.*, 2009]. There are no extensional sedimentary basins known from this early period of time, and none of the low-temperature thermochronometers applied in metamorphic rocks of the Cyclades indicate near-surface cooling prior to the early middle Miocene. We therefore relate this first stage of exhumation of the Cycladic Blueschist, both recorded on the Cycladic islands as well as in Mount Olympos and Mount Ossa, to exhumation by return flow. We infer that the bulk of the crustal stretching in the Cyclades and the eastern mainland Greece extensional metamorphic complexes must be of middle to late Miocene in age and not older than the ~25 Ma end of underthrusting of the Basal Unit, which we therefore adopt as inception age for crustal extension in the Cycladic realm in our reconstruction.

[55] At first, in this section, we restrict the restoration to an area to the north of the enigmatic "Mid-Cycladic Lineament" (Figure 1) of *Walcott and White* [1998] (Figure 2) and restore the southeastern Cyclades together with the Menderes massif later. In the zone around the enigmatic Mid-Cycladic Lineament the stretching lineations change from N-S in the southeast to NE-SW in the northwest (Figure 2).

[56] As mentioned earlier, the exact location of the detachment separating a coherent nonmetamorphic hanging wall from the area of exhumed rocks in the central Aegean region, of which the Cyclades and eastern mainland Greece are a part, is unknown. However, it is certainly located somewhere within the North Aegean Sea. We estimate the position of the southwestern boundary of the extensional province by the western limit of the area covered by rocks that were exhumed from below extensional detachments in Miocene times. This western boundary is conceptually treated as a "break-away fault," marking the position where the detachment originally cut the surface at the inception of its activity. This line runs from northern Greece, ~60 km north of Mount Olympos following *Diamantopoulos et al.* [2009], along the western side of Mount Olympos and Mount Ossa and the Pelion, to the east coast of Evvia where no exhumed Miocene metamorphic rocks have been found (Figure 8). In southern Evvia, this breakaway fault is transferred southwestward along the Attica-Evvia transfer fault to a position between the eastern Peloponnesos and the westernmost islands of the Cyclades. Given the absence of major transform faults delimiting the extensional domain in the northwest, we reconstruct extension along eastern mainland Greece as a vertical axis rotation of mainland Greece relative to the Chalkidiki peninsula around a pole located ~60 km north of Mount Olympos, where the northernmost evidence for Miocene NE-SW extension was reported [*Diamantopoulos et al.*, 2009]. The maximum amount of Miocene rotation around this pole is ~15° (Figure 8): this aligns the breakaway fault west of Mts. Olympos and Ossa with the SW margin of the Chalkidiki peninsula.

[57] The vertical axis rotation of western Greece is the result of the combined rotations around the poles in the NW Rhodope (pole A in Figure 8) and the pole north of Mount Olympos (pole B in Figure 8). The ~40° of total clockwise rotation since ~25 Ma that we restored for western Greece as a result of these two stage-rotations is similar to the 38 ± 7.0° clockwise rotation of western Greece since Oligocene time relative to the Moesian platform [*van Hinsbergen et al.*, 2008] and a ~40° clockwise rotation of the eastern Peloponnesos [*Morris*, 1995] determined paleomagnetically. The timing of the western Aegean vertical axis rotations, however, suggests that the bulk of this rotation occurred in middle Miocene time. We will return to the distribution of rotation and extension through time later, when we combine the reconstruction of Greece with the reconstruction of western Turkey.

[58] As noted earlier [see also *Brun and Sokoutis*, 2010], the rotation of the Chalkidiki peninsula around the NW Rhodope pole must have been bounded in the east by an originally ~N-S trending transform fault. Restoring the superimposed rotation of western Greece around the pole north of Mount Olympos brings also the Evvia-Attica transform fault in a N-S orientation, parallel and adjacent to the island of Lesbos and the Biga Peninsula (Figure 8). We therefore suggest that the Evvia-Attica transform fault continued further north and connects to the transform structure bounding the Chalkidiki block in the east. Clockwise vertical axis rotation around the two poles mentioned above adds an extensional component to this transform fault that increases with ongoing rotation.

[59] The area covered by the northwestern Cyclades, as well as the Aegean Sea to the NE of the northwestern Cyclades, accommodated extension as a result of rotation around both these poles. The total amount of extension along the Mid-Cycladic Lineament, which limits the region that

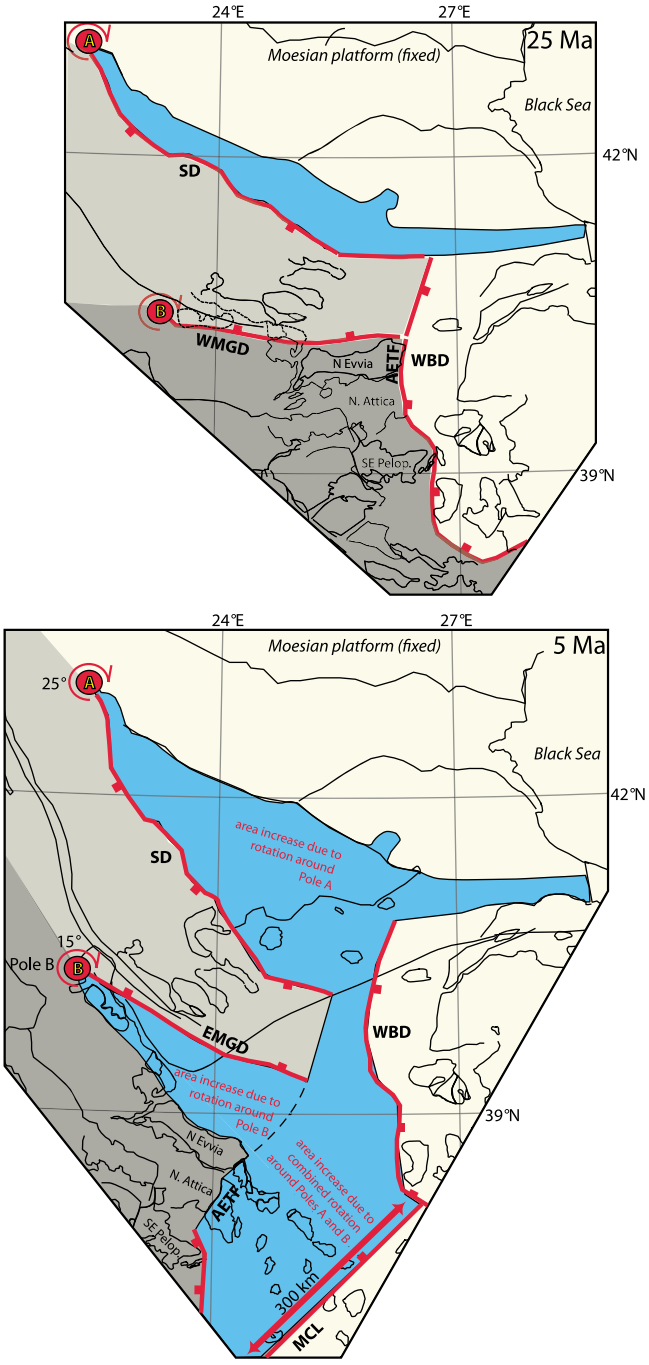



**Figure 8.** Reconstruction of the mainland western Greece and the western Cyclades extensional metamorphic complexes since 25 Ma relative to the Moesian platform, superimposed on the restoration of the Rhodope extensional metamorphic complex shown in Figure 7. AETF = Attica-Evvia Transfer Fault; EMGD = Eastern Mainland Greece Detachment; MCL = Mid-Cycladic Lineament; SD = Strymon Detachment; WBD = West Biga Detachment; WMGD = Western Mainland Greece Detachment.

underwent NE-SW stretching is ~300 km (Figure 8). Recently, *Jolivet et al.* [2010b] argued that the detachment bounding the islands of Tinos, Mykonos and southeast Evvia and the detachment along eastern mainland Greece are one and the same structure, connected through an inferred position of that structure offshore eastern Evvia (the "North Cycladic Detachment System"). Back rotating such a fault system according to the rotations mentioned above would, however, lead to a complete overlap of this fault system with the Sakarya zone of NW Turkey, which we find unacceptable. Rather, it appears that the detachment structure along which the northwestern Cyclades were exhumed is the N-S trending fault zone along the Turkish west coast, and the current coastline of northeastern Evvia and the Cyclades does not represent the boundary between a rigid hanging wall and an exhumed footwall. Restoring the extensional complexes of the Rhodope and eastern mainland Greece, and the rotation of western Greece, predicts instead that metamorphic rocks that were exhumed in Miocene time underlie the entire Aegean Sea. This is consistent with the limited crustal thickness of <25 km in this area [*Tirel et al.*, 2004].

[60] In summary, our reconstruction of the Rhodope and northwestern Cycladic–eastern mainland Greece extensional metamorphic complexes suggests that the locations at which detachments root in the central Aegean region differ from those normally suggested [e.g., *Jolivet et al.*, 2010b] (Figure 8). We root the extensional detachments along which easternmost mainland Greece became exhumed (Eastern Mainland Greece Detachment, Figure 2) to the south and southwest of the Chalkidiki peninsula. The northwestern Cyclades, however, exhumed from below a detachment system that is rooted below a coherent hanging wall immediately west of the Sakarya zone of the Biga Peninsula, Lesbos and Chios, i.e., along a N-S oriented detachment we named West Biga Detachment (Figures 2 and 8). The Attica-Evvia "transfer fault" restores in line with this West Biga Detachment and is in fact the position of (or close to) the original breakaway fault; it connects to the transform fault bounding the Chalkidiki rotating block to the east. We emphasize that the existence of a West Biga detachment is hypothetical but required to kinematically balance the kinematic and extensional constraints of exposed parts of the Cyclades and western Greece. Moreover, we do not infer that *all* extension in the North Aegean Sea is accommodated along the West Biga detachment: the Cycladic islands expose a series of metamorphic core complexes with multiple extensional detachments [e.g., *Tirel et al.*, 2009]. Similarly, we consider it possible, or even likely, that multiple detachments have accommodated extension in the crust underlying the North Aegean Sea. The West Biga detachment is merely the easternmost of these detachments.

[61] In the following, we continue our reconstruction by restoring the Peloponnesos extensional metamorphic complex, after which we will complete the restoration adding the Menderes Massif, the southeastern Cyclades and the island of Crete.

### 5.4. South Aegean Extensional Metamorphic Complex

#### 5.4.1. Kinematic Constraints

[62] The South Aegean extensional metamorphic complex exposes HP-LT metamorphic rocks of the Phyllite-Quartzite and Plattenkalk units. It runs from the northern Peloponnesos, over the islands of Kythira and Crete to Kassos (Figure 1). Similarly to the Cyclades, two dominant lineation trends can be discerned in the South Aegean window: the Peloponnesos

and Kythira are dominated by ~NE-SW to ENE-WSW trending stretching lineations, whereas Crete consistently shows ~N-S trending lineations (Figure 2).

[63] Peak pressure conditions in the Phyllite-Quartzite unit are highest on western Crete and in the southern Peloponnesos, where P-T conditions reach 17 kbar/400°C and 16 kbar/500°C, respectively [*Jolivet et al.*, 2010c]. Peak metamorphic pressures gradually decrease to the northern Peloponnesos to ~8 kbar/450°C. On Crete, peak metamorphic conditions decrease eastward to ~8 kbar/300°C [*Theye and Seidel*, 1991; *Theye et al.*, 1992], or even lower peak temperatures of ~200°C [*Klein et al.*, 2008], consistent with the absence of Miocene annealing in zircon fission track ages [*Brix et al.*, 2002]. P-T estimates from the Kastania Phyllites at the base of the Plattenkalk series on the Peloponnesos yielded lower peak metamorphic estimates than in the overlying Phyllite-Quartzite: ~7–8.5 kbar and 310–360°C [*Blümor et al.*, 1994].

[64] The tectonostratigraphically higher Tripolitza nappe is separated from the Phyllite-Quartzite unit by a structure known as the "Cretan Detachment." Most authors suggest a dominant top-to-the-north sense of shear along the Cretan Detachment following *Jolivet et al.* [1996] and a top-to-the-northeast sense of shear for the equivalent detachment on the Peloponnesos [*Papanikolaou and Royden*, 2007], although there is also evidence for opposite senses of shear on the southwestern Peloponnesos and in southern Crete [*Fassoulas et al.*, 1994; *Jolivet et al.*, 2010c; *Papanikolaou and Vassikakis*, 2010] (Figure 2). The amount of displacement along the Cretan Detachment is subject of debate. *Rahl et al.* [2005] showed that the Tripolitza nappe could have been buried to ~20 km depth, based on geothermometry indicating peak temperature conditions of ~200°C, and by assuming a cool geotherm. This is consistent with the total cumulative thickness of the Tripolitza, Pindos and Hellenic ophiolite nappes in western Greece [*van Hinsbergen et al.*, 2005a]. *Jolivet et al.* [1996] argued that the Tyros beds at the base of the Tripolitza nappe did not experience pressures beyond 6 kbar based on the absence of carpholite, which is consistent with the inference of *Rahl et al.* [2005]. The peak pressures of the Phyllite-Quartzite on western Crete and the southern Peloponnesos then define a jump in pressures across the detachment on the order of ~10 kbar, decreasing to the east and west [*Jolivet et al.*, 2010c]. At present, the Tripolitza, Pindos and ophiolite nappes on Crete are extremely thinned and dismembered, and represent a cumulative thickness of no more than ~2 km, floating as klippen on top of the Cretan Detachment [*van Hinsbergen and Meulenkamp*, 2006; *Zachariasse et al.*, 2011]. This suggests that the upper nappes above the detachment of Crete have undergone extreme thinning by a factor of around 10. There are no structural geological data available from the upper nappes of Crete to constrain the direction of extension associated with this extreme thinning.

[65] The boundaries of the South Aegean extensional metamorphic complex can only be well defined on the Peloponnesos, where the window cuts through the Pindos and Tripolitza nappes (Figure 1), and on the southeastern Peloponnesos as well as on Crete, also through the Jurassic ophiolites of western Greece and/or the upper Cretaceous ophiolites known from western Turkey [*Baumgartner*, 1985; *Koepke et al.*, 2002] (Figure 1). On the Peloponnesos, the NE-SW width of the window gradually decreases northwestward (Figure 1). The northernmost exposures of the Phyllite-Quartzite unit occur in the Chelmos window on the northern Peloponnesos (Figure 1). The northern tip of the South Aegean extensional metamorphic complex is probably formed by the Itea-Amfissa detachment to the north of the Gulf of Corinth [*Papanikolaou et al.*, 2009] (Figure 1). This is an eastward dipping low-angle normal fault with a relatively small amount of E-W extension (~9 km) creating a continental to shallow marine basin of early to late Miocene age (18–8 Ma) [*Papanikolaou et al.*, 2009].

[66] $^{40}Ar/^{39}Ar$ white mica cooling ages of the Phyllite-Quartzite on Crete cluster between 24 and 20 Ma, and given the low peak temperature conditions of ~350–400°C, they probably at most only slightly postdate peak pressure conditions [*Jolivet et al.*, 1996]. $^{40}Ar/^{39}Ar$ white mica cooling ages on the Peloponnesos vary between ~26 and 13 Ma and the younger ages here probably reflect a longer cooling period following higher peak temperature conditions of ~500°C [*Jolivet et al.*, 2010c]. The latter authors also reported 50–30 Ma $^{40}Ar/^{39}Ar$ ages and interpreted these as reflecting inherited ages, either due to low-grade metamorphism due to burial, or, in the case of the Oligocene flysch of the Plattenkalk, detrital mica ages. Zircon fission track ages of the Phyllite-Quartzite unit vary from 15 to 12 Ma on Crete [*Thomson et al.*, 1998; *Marsellos et al.*, 2010] and ~13–9 Ma on Kythira and the Peloponnesos [*Marsellos and Kidd*, 2008; *Marsellos et al.*, 2010]. Hence, most of the exhumation of the Cretan nappe stack occurred between ~24 and ~12 Ma. On the Peloponnesos, the oldest rocks unconformably covering the exhumed metamorphic rocks are early Pliocene in age [*Frydas*, 1993], and the age of the first exposure of the Phyllite-Quartzite unit is poorly resolved. The oldest dated sediments on Crete form part of a short-lived supradetachment basin between 10.8 and 10.4 Ma, postdating most of the exhumation and extreme thinning of the Tripolitza and higher nappes [*van Hinsbergen and Meulenkamp*, 2006; *Zachariasse et al.*, 2011]. The oldest sediments that rework the Phyllite-Quartzite and Plattenkalk units are ~10.4 Ma old and found in the Ierapetra basin [*Fortuin*, 1978; *Fortuin and Peters*, 1984; *Zachariasse et al.*, 2011]. These ages more or less coincide with the youngest fission track ages.

#### 5.4.2. Restoration: Peloponnesos Extensional Metamorphic Complex

[67] We now restore the extensional history of the Peloponnesos (Figure 9). We will only return to the extension and exhumation history of Crete, where much less constraints are available on the boundaries of the extensional complex, after we restored the eastern side of the Aegean system in western Turkey and the southeastern Cycladic metamorphic complexes.

[68] There is considerable debate on the relative amounts of exhumation by return flow within a subduction channel and exhumation by extension in the case of the South Aegean extensional metamorphic complex. Exhumation has been postulated to have resulted from crustal thinning [*Ring et al.*, 2001; *van Hinsbergen et al.*, 2005b], leading to estimates of horizontal stretching of up to ~100 km in a NE-SW (Peloponnesos) or N-S (Crete) direction, respectively. It

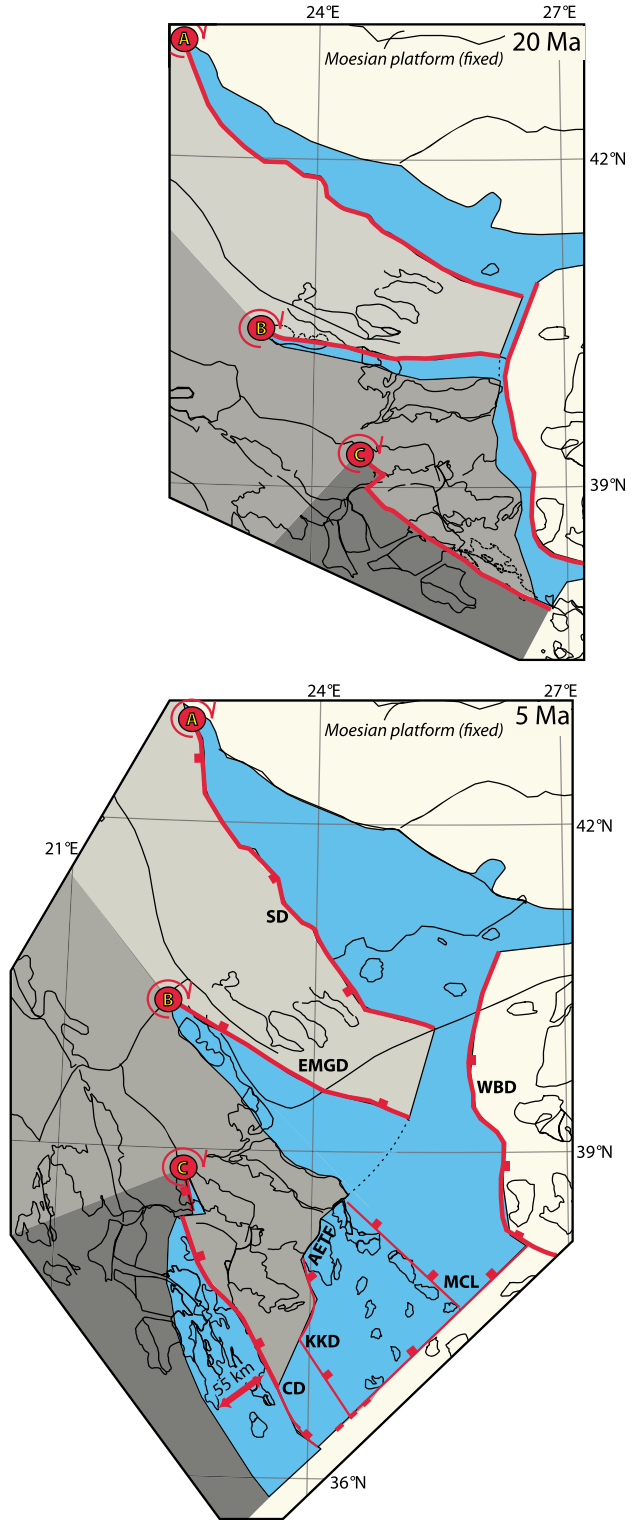



**Figure 9.** Reconstruction of the Peloponnesos extensional metamorphic complex since 20 Ma, relative to the Moesian platform, superimposed on the restorations of the Rhodope, eastern mainland Greece and northwestern Cycladic extensional metamorphic complexes shown in Figure 8. AETF = Attica-Evvia Transfer Fault; CD = Cretan Detachment; EMGD = Eastern Mainland Greece Detachment; KKD = Kea-Kythnos Detachment; MCL = Mid-Cycladic Lineament; SD = Strymon Detachment; WBD = West Biga Detachment.

seems more likely, however, that a significant part of the exhumation of these HP-LT metamorphic rocks should be viewed as occurring as a result of return flow within the subduction channel, be it driven by buoyancy or extrusion or a combination of both [*Jolivet et al.*, 1996, 2010c; *Thomson et al.*, 1999; *Doutsos et al.*, 2000; *Xypolias and Doutsos*, 2000; *Chatzaras et al.*, 2006; *Ring et al.*, 2010]. The fact that the Plattenkalk on the Peloponnesos yielded peak pressure conditions of 7–8.5 kbar, as opposed to ~17 kbar for the overlying Phyllite-Quartzite unit, shows that these units are separated by a Miocene thrust, along which the Phyllite-Quartzite unit moved upward, with a ~30 km vertical component (~10 kbar), quite similar to the Eocene-Oligocene exhumation of the Cycladic Blueschist above the underthrusting Basal Unit. This occurred at levels below the brittle upper crust, and may be have been accommodated by return flow above the subducting slab. *Jolivet et al.* [2003] and *Ring et al.* [2010] argued that exhumation of Aegean HP-LT rocks may have occurred as a result of extrusion between a thrust at the base and an normal fault type detachment at the top, similar to what is proposed for the Greater Himalaya [e.g., *Grujic et al.*, 1996; *Hodges*, 2000].

[69] Regardless of the questions after kinematics and forces that drive exhumation, the HP-LT wedge of the South Aegean had to exhume across a nappe stack that was originally at least ~20 km thick. The column of rock above the exhuming HP-LT wedge was either eroded or extended, or both, to allow the HP-LT wedge to reach the surface. In the case of Crete, there is strong evidence for crustal thinning: the originally ~20 km thick nappe stack above the Cretan Detachment has been dramatically thinned to ~2 km. In addition, the Itea-Amfissa detachment and the related overlying sedimentary basin at the northern tip of the SW Aegean extensional complex shows that there was ~ENE-WSW extension (in modern coordinates) between 18 and 8 Ma [*Papanikolaou*, 2009]. We infer that the modern width of the region that exposes the Phyllite-Quartzite and Plattenkalk units on the Peloponnesos probably resulted from NE-SW extension, facilitating the rise of the metamorphic units to the surface, and restore 55 km of NE-SW to N-S extension on the southern Peloponnesos (in addition to 5 km of post-3 Ma extension in the Gythion graben, see section 5.1 and Figure 5), decreasing to zero near a rotation pole at the northern tip of the Itea-Amfissa detachment (Figure 9). This amount of extension is sufficient to bury all the HP rocks of the Peloponnesos below the hanging wall of the "Cretan" detachment. It restores the Tripolitza and Pindos of the western Peloponnesos into single, NW-SE trending belts, which is consistent with vertical axis rotation patterns for the Tripolitza flysch on the northwestern and southwestern Peloponnesos [*van Hinsbergen et al.*, 2005b] (Figure 9). The timing of the end of extension is given by the youngest fission track ages on the southeastern Peloponnesos and Kythira (9 Ma [*Marsellos et al.*, 2010]). The age of onset, however, is somewhat arbitrary: extension was active at around 18 Ma [*Papanikolaou et al.*, 2009], and exhumation of the HP wedge started ~26–25 Ma ago based on $^{40}Ar/^{39}Ar$ cooling ages [*Jolivet et al.*, 2010c]. We tentatively suggest that the exhumation of the Phyllite-Quartzite unit between period between ~25 and 20 Ma occurred by return flow above the subducting slab, until juxtaposition with the Plattenkalk at pressure conditions of ~7.5 kbar. Thinning of the crustal

overburden facilitated by a clockwise rotation of the northwestern Peloponnesos relative to the northeastern Peloponnesos occurred thereafter; it led to final exhumation of the HP-LT metamorphic units around 9 Ma.

### 5.5. Menderes Extensional Metamorphic Complex

#### 5.5.1. Kinematic Constraints

[70] We will only briefly summarize the key features of the Menderes Massif as far as relevant for this paper, and refer for a detailed reconstruction and review of western Turkey to *van Hinsbergen* [2010] and *van Hinsbergen et al.* [2010c]. Within the Menderes extensional metamorphic complex, the Alaşehir and Büyük Menderes ductile-to-brittle extensional detachment systems divide the Menderes metamorphic complex into a southern, central and northern part (Figure 1). These opposite verging detachments exhumed the central Menderes massif between ~15 and 5 Ma [*Hetzel et al.*, 1995; *Bozkurt*, 2001; *Bozkurt and Oberhänsli*, 2001; *Gessner et al.*, 2001a; *Lips et al.*, 2001; *Ring et al.*, 2003; *Catlos and Cemen*, 2005; *Glodny and Hetzel*, 2007]. The northern Menderes Massif, in addition, is bounded to the north by the Simav detachment, which separates it from rocks of the Afyon zone in the NE. This latter detachment completely omits the roots of the West Anatolian Taurides (Figure 1). The Simav detachment was active earlier, namely, between ~25 and 19 Ma, at most until 15 Ma, as suggested by $^{40}Ar/^{39}Ar$ cooling ages, ages of synextensional granites, fission track ages, and ages of unconformably covering sediments [*Seyitoğlu and Scott*, 1992; *Işik and Tekeli*, 2001; *Ring et al.*, 2003; *Işik et al.*, 2004; *Purvis and Robertson*, 2004; *Purvis et al.*, 2005; *Thomson and Ring*, 2006; *Karaoğlu et al.*, 2010].

[71] The second stage of exhumation between 15 and 5 Ma led to a triangular-shaped metamorphic core complex in the center of the Menderes massif, which exhumed along the eastward converging oppositely dipping Alaşehir and Büyük Menderes detachments (Figure 1). Exhumation of the Central Menderes Massif was associated with a counterclockwise vertical axis rotation of ~27.3 ± 5.0° of SW Anatolia relative to NW Anatolia, between 15 and 5 Ma, around a pole near Denizli (Figure 1) [*van Hinsbergen et al.*, 2010a] (Table 1). Extension along the Simav detachment between ~25 and 15 Ma may have amounted up to ~50 km [*Thomson and Ring*, 2006]. The surfacing of the remainder of the Menderes Massif in early Miocene time possibly resulted from southeastward translation of the West Anatolian Taurides away from the Menderes Massif, simultaneously with thrusting of the West Anatolian Taurides over the Bey Dağları foreland between ~23 and 16 Ma [*Collins and Robertson*, 2003; *van Hinsbergen*, 2010]. There are no Miocene extensional detachments known from the southern Menderes Massif, and the tectonic accommodation of its exhumation to only several kilometers depth remains an outstanding problem in the exhumation history of western Turkey [*van Hinsbergen*, 2010].

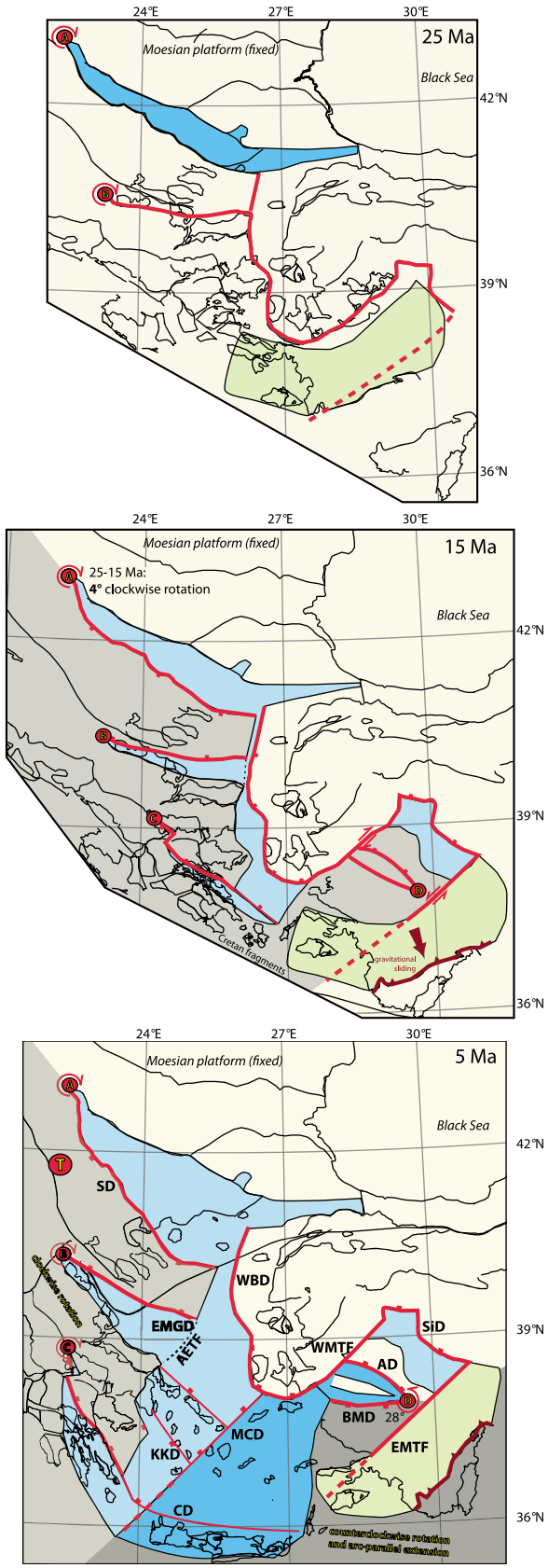
[72] The first stage of exhumation of the Menderes massif along the Simav detachment occurred without significant vertical axis rotations. Hence, given the top-to-the-NE kinematics of the Simav detachment [*Işik and Tekeli*, 2001], the Menderes metamorphic complex appears to be laterally bounded by displacement-parallel fault zones exhibiting considerable strike-slip components (transfer faults [*Gibbs*, 1984]). The NW tectonic boundary of the Menderes metamorphic complex juxtaposes the Menderes metamorphics, including remnants of the Dilek Nappe that corresponds to the Cycladic Blueschist, with the nonmetamorphic Cretaceous Bornova Flysch zone [*Erkül*, 2010; *Ersoy et al.*, 2011; *Sözbilir et al.*, 2011] (Figure 1). No major extensional detachment is found on the NW side of this structure, which consequently must be a dextral transform fault with a displacement equal to the displacement of the Simav detachment (~50 km, Table 1). Likewise, sinistral strike slip movement has to be invoked for the fault zone bounding the Menderes Massif in the SE. The latter marks the eastern boundary of the Aegean extensional system during the early Miocene [*van Hinsbergen et al.*, 2010a]. No such fault can be observed at the surface, however, nor can a southwestward continuation of such a transfer fault be recognized within the West Anatolian Taurides. This is one of the arguments to interpret that an up to ~2 km thick klippe comprising the West Anatolian Taurides gravitationally slid southeastward relative to the Menderes Massif and the Bey Dağları platform and covers this transfer fault that presumably exists at depth, as proposed by *van Hinsbergen* [2010]. This interpretation predicts that the Lycian Nappes (West Anatolian Taurides) have distributed ~50 km of left-lateral strike-slip displacement in early Miocene time as they were transferred southeastward over the location of the inferred SE Menderes transform. The detailed kinematic evolution of the West Anatolian Taurides in early Miocene time remains to be tested in the field.

#### 5.5.2. Restoration: The Menderes, Southeast Cycladic, and Cretan Metamorphic Complexes

[73] We now integrate the reconstructions of the Rhodope, eastern mainland Greece–northwest Cycladic and Peloponnesos extensional metamorphic complexes, characterized by superimposed clockwise rotations, with the available constraints from the southeastern Cyclades, Crete and the Menderes massif (Figure 10).

[74] The northern Rhodope and the northern Menderes Massif share NNE-SSW trending stretching lineations associated with the onset of extension in both complexes. Although extension in the Rhodope commenced earlier than in the Menderes, the majority of Rhodope extension occurred since ~25 Ma ago, simultaneous with extension in the northern Menderes Massif. Hence, the Rhodope and northern Menderes massifs represent the northernmost parts of the Aegean–West Anatolian extensional domains, and paleomagnetic data show that the areas to their north (i.e., the Moesian platform and NW Turkey) did not experience a significant (i.e., not more than ~5°) vertical axis rotation relative to stable Eurasia [*van Hinsbergen et al.*, 2008, 2010a]. Therefore the NNE-SSW oriented stretching lineations preserved in the northeastern Rhodope and in the northern Menderes Massif (Figure 2) likely represent the original stretching direction at the inception of major Aegean extension, some ~25 Ma ago. All deviations from this direction observed elsewhere in the Aegean region (Figure 2) are likely to have resulted from vertical axis rotations.

[75] The distribution of the rate of rotation and extension in the Rhodope during the Miocene is not known in detail. However, paleomagnetic data suggest that bulk of west




Aegean vertical axis rotation, i.e., beyond typical paleomagnetic error bars of ~5–10° occurred since middle Miocene time [*van Hinsbergen et al.*, 2005b]. The 25–15 Ma history of the Menderes massif on the eastern side of the extensional system was most likely bounded by a transfer fault with a sinistral strike slip component along its southeastern side (Figure 10). The transfer fault bounding the northern Menderes massif to the NW, as well as the inferred transfer fault to its southeast, is subparallel to a small circle around the NW Rhodope pole (pole A in Figure 10). Therefore, we model the 25–15 Ma extension history of the northern Menderes Massif as a "rigid" body rotation of the entire Aegean and West Anatolian region around the NW Rhodope pole, bounded in the east by the eastern Menderes transfer fault, in a "window sweeper" fashion. Rotations around the poles north of Mount Olympos (pole B) and north of the Peloponnesos (pole C in Figure 10) are superimposed on this wholesale Aegean–West Anatolian clockwise rotation (Figure 10). The amount of NNE-SSW extension in the northern Menderes massif of ~50 km [*Ring et al.*, 2003; *van Hinsbergen*, 2010] corresponds to a 4° clockwise rotation around the Rhodope pole. After 15 Ma, the West Anatolian region underwent extension associated with counterclockwise rotations, and the northern Aegean and southern Aegean–West Anatolian regions decoupled (see below). We equally distribute the remaining ~20° of total Miocene rotation between Chalkidiki and Moesia [*Kondopoulou and Westphal*, 1986; *Brun and Sokoutis*, 2007; *van Hinsbergen et al.*, 2008; *Georgiev et al.*, 2010] around pole A the 15 to 4 Ma period (Figure 10). The end of rotation is based on the fact that western Greece still experienced ~10° clockwise rotation in the Pliocene [*van Hinsbergen et al.*, 2005b], and extension in the Strymon valley continued until at least the early Pliocene [*Snel et al.*, 2006].

[76] The island of Crete, the southeastern Cyclades and the southern Menderes massif consistently share N-S trending stretching lineations (Figure 2). In western Turkey, it was demonstrated that the change from NNE-SSW lineations in the northern Menderes Massif to N-S in the southern Menderes Massif occurs gradually over the 15–5 Ma old Central Menderes metamorphic core complex, and was associated with a 27.6 ± 5.0° counterclockwise rotation of SW Turkey [*van Hinsbergen et al.*, 2010a]. A comparable middle Miocene and younger counterclockwise rotation of ~30° was paleomagnetically determined from the island of Naxos [*Morris and Anderson*, 1996; *Avigad et al.*, 1998]. From the island of Crete, there are no paleomagnetic data from rocks

**Figure 10.** Reconstruction of the Menderes, southeastern Cyclades and Cretan extensional metamorphic complexes since 25 Ma relative to the Moesian platform and northwestern Turkey, superimposed on the restoration of the western and northern Aegean region shown in Figure 9. AD = Alaşehir Detachment; AETF = Attica-Evvia Transfer Fault; BMD = Büyük Menderes detachment; CD = Cretan Detachment; EMGD = Eastern Mainland Greece Detachment; EMTF = East Menderes Transfer Fault; KKD = Kea-Kythnos detachment; MCD = Mid-Cycladic Detachment; SD = Strymon Detachment; SiD = Simav Detachment; WBD = West Biga Detachment; WMTF = West Menderes Transfer Fault.

older than ~8 Ma, and after 8 Ma rocks on Crete only experienced minor local counterclockwise rotations as a result of motion along the South Aegean strike-slip system [*Duermeijer et al.*, 1998]. *Walcott and White* [1998] suggested that the entire area that exposes N-S stretching lineations forms one single, albeit internally deforming, rotating domain. Given the consistency of the stretching direction and the similarity of paleomagnetic results of Naxos with those of SW Turkey, we restore the counterclockwise rotation that formed Central Menderes metamorphic core complex for the entire region of the southeastern Cycladic and Cretan extensional metamorphic complexes, as well as southwestern Turkey, with respect to the northern Menderes massif. In western Turkey, the triangular Central Menderes Massif, bounded by the Büyük Menderes and Alaşehir detachments, accommodates this rotation. In the central Aegean region, the boundary between the counterclockwise and clockwise rotating domains is discrete and formed by the Mid-Cycladic Lineament. Restoring opposite block rotations across this "lineament" leads us to model it as an extensional detachment (Mid-Cycladic Detachment, Figure 10). Eastward, this hypothesized detachment abuts against the junction between the N-S trending transtensional detachment to the west of Chios, Lesbos and the Biga Peninsula, and the E-W trending southeastern Cycladic Detachment located between the Ikaria and Samos islands in the south and the island of Chios in the north (Figure 10). Further to the northeast this boundary is traced along the Büyük Menderes and Alaşehir detachments that flank the northwestern margin of the Cycladic (Dilek) blueschists and the Central Menderes Massif (Figure 2).

[77] Originally the Mid-Cycladic Lineament was inferred to be a transform fault accommodating opposite rotations of the NW and SE Cyclades [*Walcott and White*, 1998]. However, field evidence, as acknowledged by *Walcott and White* [1998], has not been able to demonstrate such a transform fault. Our reconstruction requires instead that the NW and SE Cyclades became separated by a top-to-the-N extensional detachment, which is located somewhere within (perhaps between) the Basal Unit and Cycladic Blueschist sequences. For simplicity, we modeled the structure as a discrete detachment, being aware that presently available field data cannot confirm this, in part since its location may be offshore. It is possible that the inferred extensional deformation is distributed in a wider zone, leading to changing orientations of the stretching directions in a wider zone around the traced Mid-Cycladic Detachment: future data need to further constrain the kinematics of this extensional corridor.

[78] During the 25–15 Ma period, Aegean–West Anatolian extension was distributed asymmetrically by clockwise rotations around poles in the NW of the system, associated with a small ~4° rotation angle. After 15 Ma, however, the Aegean–West Anatolian extensional system was more or less symmetric, with the eastern and western side undergoing opposite rotations around poles in the NW and east (Figure 10). Restoration aligns all stretching lineations in the extensional metamorphic complexes of the Aegean to an original NNE-SSW direction, and lead to a ~40° clockwise rotation of western Greece relative to the Moesian platform since 25 Ma, ~30° of which occurred since ~17 Ma. This is consistent with paleomagnetic evidence, which gives a 38.0 ± 7.2° post-Eocene clockwise rotation of western Greece relative to the Moesian platform, the bulk of which occurs after 17 Ma [*van Hinsbergen et al.*, 2008]. For southwestern Turkey, the southeastern Cyclades and Crete, we restored a 28° counterclockwise rotation relative to NW Anatolia, consistent with the 27.3° ± 5.0° determined paleomagnetically in western Turkey [*van Hinsbergen et al.*, 2010a, 2010b] and with ~30° measured on Naxos [*Morris and Anderson*, 1996; *Avigad et al.*, 1998].

[79] These opposite rotations are inevitably associated with trench-parallel extension, the amount of which increases southwestward. Trench-parallel extension in the Cycladic region occurs parallel to small circles around the rotation pole of the SE Aegean domain, and leads to N-S trending stretching lineations associated with extension inferred to occur along the Mid-Cycladic Detachment. We restore the southeastern Cyclades into a position *below* the northwestern Cyclades, and the Mid-Cycladic Lineament is restored as a top-to-the-north oblique extensional detachment accommodating trench-parallel extension induced by these opposite rotations. Extreme extension in the central Aegean region around the Mid-Cycladic Detachment is consistent with the observation that the northwestern Cyclades is underlain by the thinnest crust of the Aegean (18 km) [*Tirel et al.*, 2004].

[80] We note that many authors [*Kissel and Laj*, 1988; *Gautier et al.*, 1999; *Kissel et al.*, 2003; *van Hinsbergen et al.*, 2005b; *Brun and Sokoutis*, 2010] have drawn the main western Aegean-Eurasia (i.e., Moesia) rotation pole at the Scutari-Pec line in northern Albania (Figure 1) that separates the nonrotated Dinarides [*Kissel et al.*, 1995] from the ~40° rotated Albanides [*Mauritsch et al.*, 1995; *Speranza et al.*, 1995] and Hellenides [*Kissel and Laj*, 1988; *van Hinsbergen et al.*, 2005b; *Broadley et al.*, 2006]. This is incorrect in the sense that the Scutari-Pec pole merely describes the rotation of western Greece/Albania with respect to the Dinarides; the latter moved northward with respect to Eurasia throughout the Cenozoic, which is accommodated by shortening in the eastern Alps [*Schmid et al.*, 2004, 2008; *Ustaszewski et al.*, 2008; *Handy et al.*, 2010]. Hence we postulate that rotation of the western Hellenides and Albanides relative to Eurasia occurred around the stage poles in the northern Rhodope and north of Mt Olympos instead, as described above. This results in a total reconstruction pole of the rotation of the western Aegean region relative to Moesia at latitude 41.47°, longitude 22.34° (pole T in Figure 10). The rotating western Aegean–Albanian domain, however, extended further to the NW, toward the Scutari-Pec fault. To the NW of pole T, a rigid body rotation of the western Aegean-Albanian region would lead to net NE-SW shortening. As postulated by *van Hinsbergen et al.* [2008], this was probably accommodated by northward extrusion of units from between Albania and the Moesian platform, accommodated along the Timok fault, perhaps in combination with reactivation of thrusts within the Albanian and Macedonian nappe stack.

[81] Finally, our reconstruction (e.g., Animation S1 in the auxiliary material) treats the nonmetamorphic domains with structural and paleomagnetic coherence, such as western Greece, Chalkidiki, Moesia, Biga or Bey Daglari, as semi-rigid blocks that undergo Euler rotations.[1] The areas now

occupied by extensional metamorphic provinces (e.g., Cyclades, northern Aegean region) was almost entirely generated during extension. The metamorphic rocks in these areas were hence exhumed from underneath adjacent more rigid domains such as NW Turkey. Although extensional detachments played an important role in the exhumation of these rocks to the surface, we envisage that lower crustal flow must have played an important role on the scale of the entire study area, to produce the modern 20–25 km thick central Aegean crust with a flat Moho. Such lower crustal flow was already inferred on the scale of a series of core complexes [e.g., *Tirel et al.*, 2009] but may here be important on a much larger scale.

### 5.6. Area Losses Due to Post ~35 Ma Africa-Europe Convergence

[82] We now complete our kinematic restoration of the Aegean region by also restoring the area losses that resulted from subduction of the African-Adriatic plate below the Aegean that was contemporaneous with core complex formation further to the north: throughout the extension history reconstructed above, subduction and accretion continued along and immediately north of the Hellenic subduction zone, respectively. There are no detailed balanced cross sections available in Greece for estimating the amount of shortening within and across individual nappe complexes, with the exception of the Pindos zone [*Skourlis and Doutsos*, 2003]. Even if there were, such reconstructions would only provide a minimum amount of convergence.

[83] Therefore, we chose a different approach. We estimated the onset and end of accretion of individual nappes from estimated from the age of the oldest and youngest foreland basin sediments in the nonmetamorphosed parts of the nappes in the fore arc. This we combine with the youngest sedimentary ages of metamorphosed portions of the same nappe exposed in extensional metamorphic complexes, marking their maximum age of onset of underthrusting, and their age of peak pressure metamorphism, representing the moment when those rocks ceased to be buried together with the down going plate (i.e., the end of their underthrusting) [*van Hinsbergen et al.*, 2005c]. For most of the nappes, these two approaches provide similar age ranges for the same nappe [see *van Hinsbergen et al.*, 2005c]. To restore the tectonic units into their presubduction configuration and dimension, we assume that until the end their accretion, all nappes were rigidly attached to Africa, and we estimate the pre-35 Ma original dimension of the paleogeographic domains of the Pre-Apulian, the Ionian and Tripolitza zones by the amount of Africa-Europe convergence that occurred during their underthrusting, diminished by the amount of extension in the overriding plate we reconstructed as described above (see Figures 3d and 3e). Note that in the time window covered by our reconstruction, only the Pre-Apulian, Ionian and Tripolitza zones were involved in underthrusting in the Aegean region. In western Turkey, there are no known accretion events after 35 Ma (see section 4.2).

[84] Stratigraphic and structural evidence suggests that the middle internal Ionian zone and the Tripolitza zone became underthrusted more or less simultaneously from ~35 to ~23 Ma. We therefore partition the contemporaneous Africa-Greece convergence equally over these two nappes in our reconstruction. Furthermore, following *van Hinsbergen et al.* [2010c], we assume that there were no accretion events during the last 35 Ma in western Turkey. Hence, all lithosphere that was subducted in western Turkey since this time is interpreted to have been oceanic.

[85] *Broadley et al.* [2006] showed that the external Ionian zone underwent a smaller (~25°) clockwise rotation in the Neogene than the rest of the western Aegean nappe stack (38.0° ± 7.2° [*Kissel and Laj*, 1988; *van Hinsbergen et al.*, 2005b]) and interpreted this as evidence that the external Ionian zone accreted later, around 15 Ma, than the rest of middle and internal Ionian zones, in line with the continuation of foreland basin deposition on the external Ionian zone until this time. We follow this suggestion, and let the external Ionian zone be part of the African plate until 15 Ma, after which we reconstruct it as part of the western Aegean nappe stack. Finally, we fix the Pre-Apulian zone to the Apulian platform of the Adriatic microplate until it accreted 4 Ma ago [*van Hinsbergen et al.*, 2006].

[86] Combining the nappe accretion reconstruction with the restoration of the Aegean region leads to the paleogeological and paleogeographic maps of the Aegean–West Anatolian region of Figure 11, and in Animation S1 in the auxiliary material.

## 6. Discussion

[87] We now explore the most important implications of our restoration. At first, we show that our restoration highlights the importance of trench-parallel extension for the extension and exhumation evolution of the southern Aegean region, on Crete. Then, we study the prime implications regarding the paleogeographic correlations between Aegean and Anatolian geological units.

### 6.1. Trench-Parallel Extension and Implications Regarding the Extension and Exhumation History of Crete

[88] Our reconstruction suggests that the total amount of length change due to trench-perpendicular NNE-SSW extension in the Aegean region since 25 Ma amounts to some 400 km (Figure 12). The amount of trench-parallel extension, such as accommodated along the Mid-Cycladic Detachment, increases radially outward, and amounts to as much as 650 km change of length parallel to the trench, measured from the SW Peloponnesos, along southern Crete, to Rhodos on a present-day map. As shown in Figure 9 for the Peloponnesos, and by *van Hinsbergen* [2010] for SW Turkey and Rhodos, the geology of the southwestern and southeastern parts of the Aegean region are still sufficiently coherent to allow kinematic restorations into the undeformed state. Our reconstructions show that the original distance between the SW Peloponnesos and Rhodos was a mere 100 km 25 Ma ago, as opposed to the 750 km measured today parallel to the trench (Figure 12). The island of Crete in the sense of a more or less coherent tectonic block came into existence only at ~10 Ma, by which time the Phyllite-Quartzite unit cooled to below the zircon fission track annealing temperature [e.g., *Marsellos et al.*, 2010]. The oldest sediments on the island were deposited at 10.8 Ma

[1]Auxiliary materials are available in the HTML. doi:10.1029/2012TC003132.

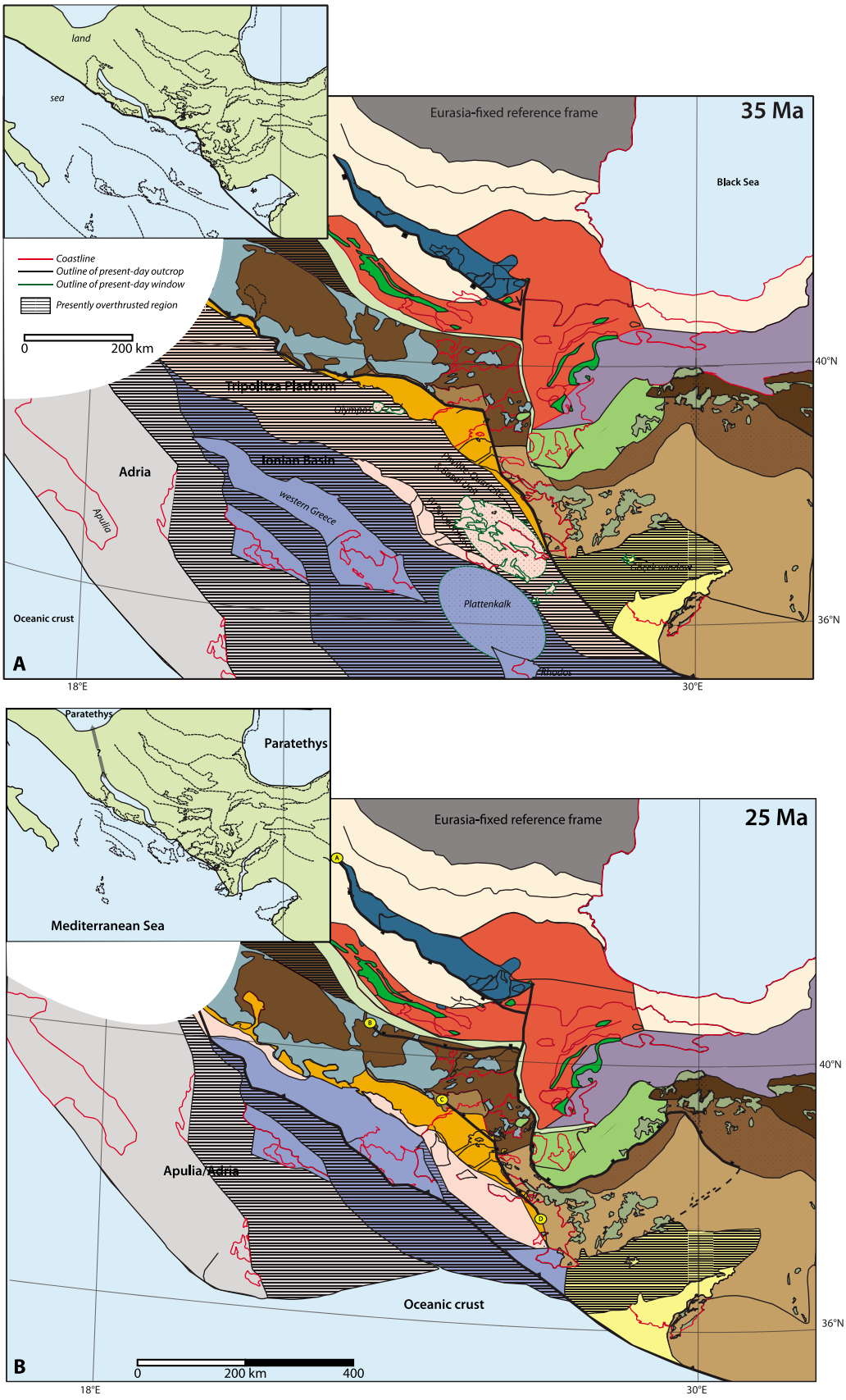



**Figure 11.** Tectonic and paleogeographic reconstructions of Greece. Paratethys paleogeography for the last 10 Ma modified after *Popov et al.* [2006]. For key, see Figures 1 and 11a. See also Animation S1 of the reconstruction in the auxiliary material.

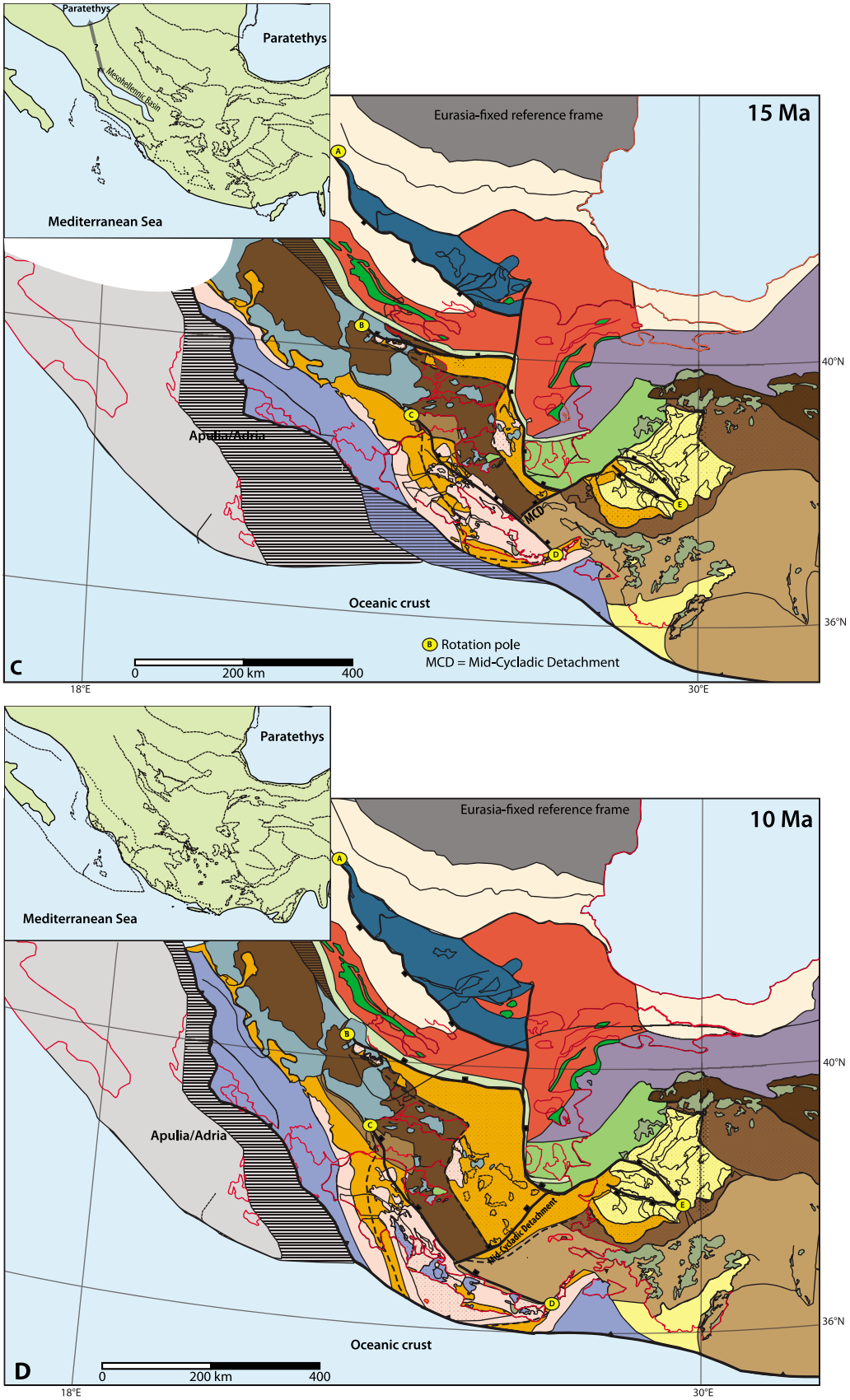



**Figure 11.** (continued)

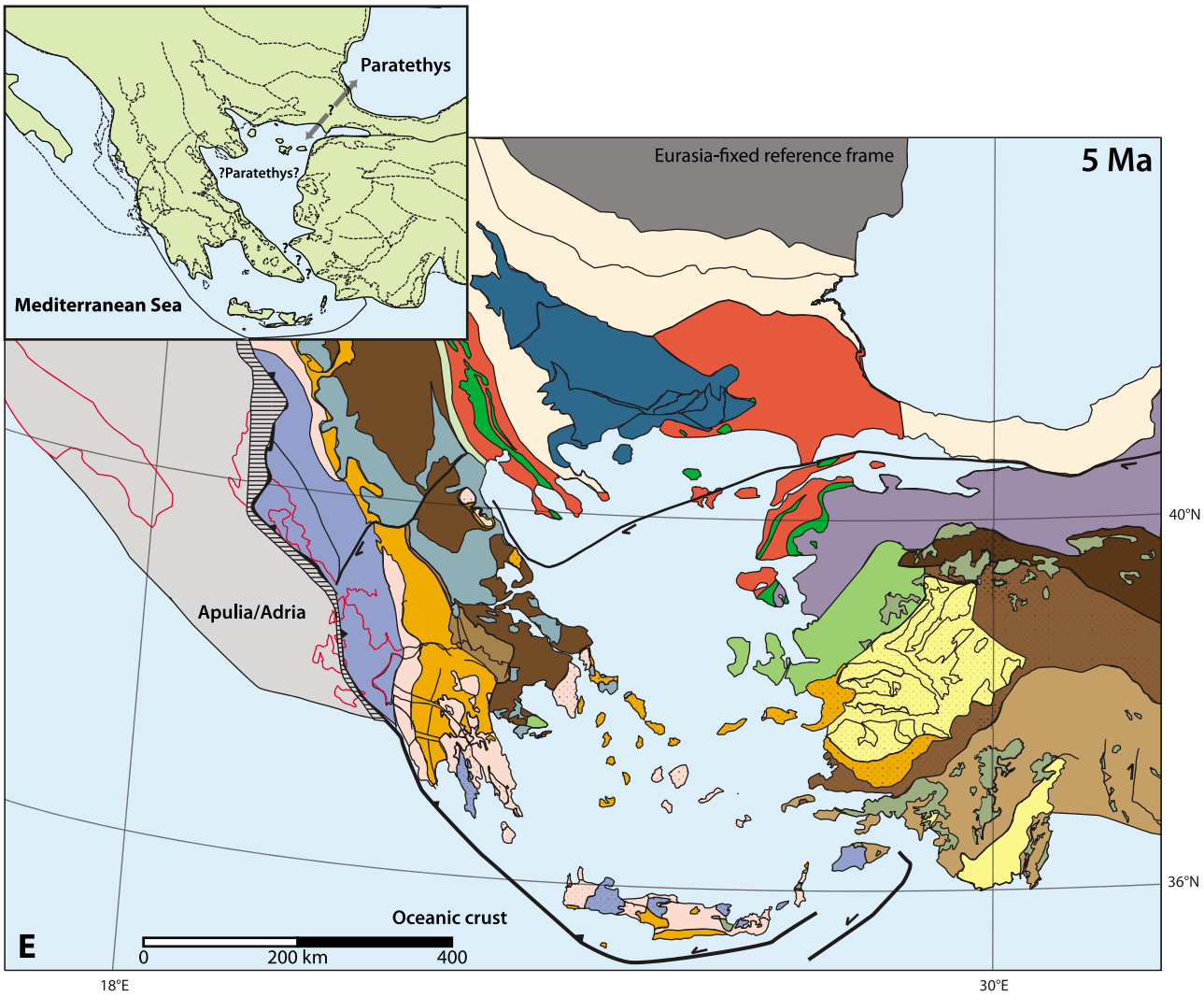
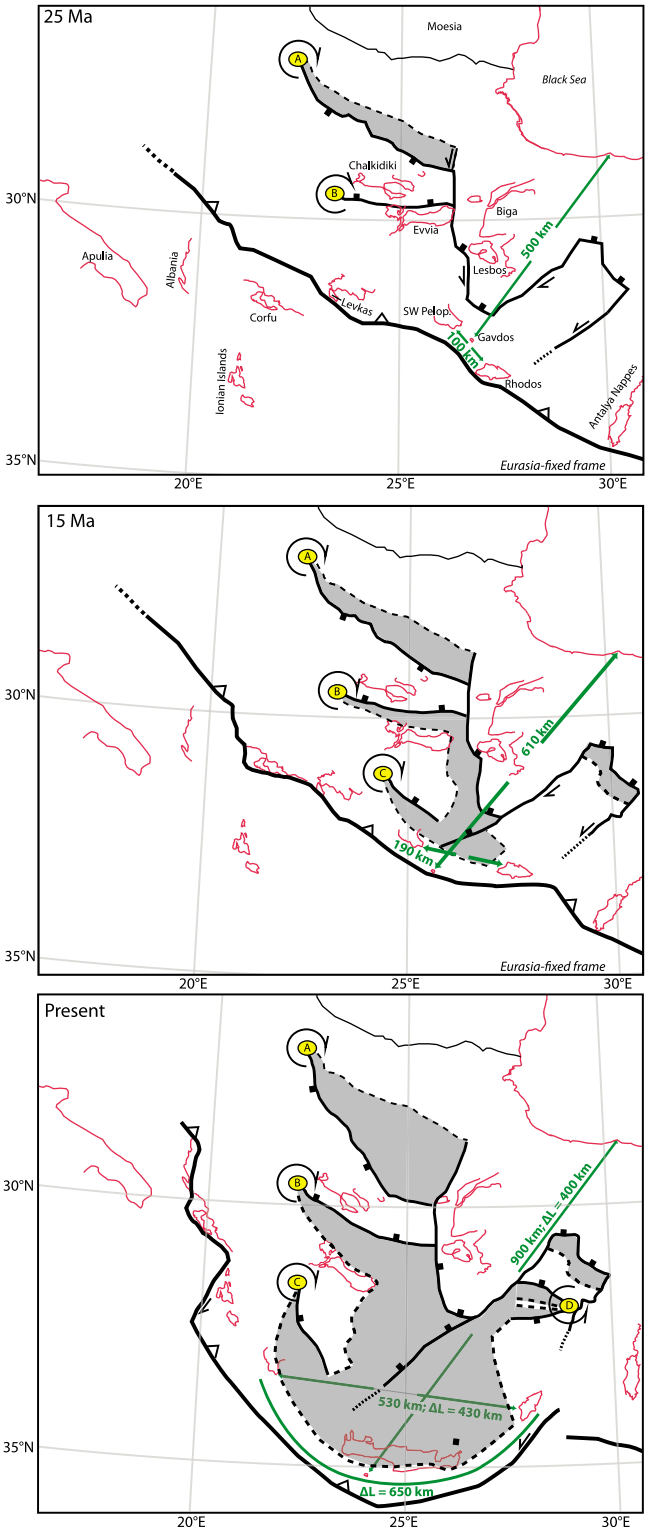



**Figure 11.** (continued)

[*Zachariasse et al.*, 2011], and only relatively low-displacement, high-angle normal faulting affected the island since then. Prior to this time, the Phyllite-Quartzite and Plattenkalk units resided at depth, moving upward along the subducting slab. In the 25 Ma configuration (Figure 11b), the relicts of the Tripolitza, Pindos and ophiolite nappes now dispersed all across Crete, can only have occupied the very narrow region of ~100 km wide mentioned above (Figure 12) between the restored positions of Rhodos and the SW Peloponnesos. Trench-parallel stretching amounts to a factor >6 for Crete and the seaways between Crete and the Peloponnesos in the west, and between Crete and Rhodos in the east. This is consistent with the fact that the nappe stack above the Cretan Detachment is no more than ~2 km thick, whereas their original thickness was probably on the order of 20 km [*Zachariasse et al.*, 2011]. Despite the major trench-parallel extension that is required to accommodate the trench-normal Aegean extension and the opposite rotations of the west Aegean and southeast Aegean–southwest Anatolian domains, the stretching lineations measured in the HP-LT rocks of Crete are consistently N-S oriented [e.g., *Jolivet et al.*, 1996]. We propose that this apparent paradox results from the fact that the HP-LT rocks of Crete were not part of the overriding plate at the inception of, and during trench retreat until ~11 Ma. Instead, they were part of the subducting slab and moved with that slab, while at the same time exhuming by return flow expressed by their cooling between ~24–21Ma ($^{40}$Ar/ $^{39}$Ar ages [*Jolivet et al.*, 1996]) all the way to 10–12 Ma ago (zircon fission track ages [e.g., *Marsellos et al.*, 2010]) (see Figure 13). As a result, the Cretan HP-LT rocks carry only trench-perpendicular stretching lineations. Rocks of the upper units (Tripolitza, Pindos and Uppermost Unit), however, accreted to the overriding plate prior to the Oligocene, i.e., prior to the inception of oroclinal bending due to roll-back. These rocks, as a result, were stretched, with trench-parallel extension dominating over trench-normal extension.

[89] The Cretan HP-LT rocks nowhere actually reach the surface until 10.4 Ma, the oldest sediments on Crete (~10.8 Ma old) being devoid of metamorphic debris. Metamorphic debris first appears in conglomerates at around ~10.4 Ma when activity along the Cretan Detachment ceased [*Zachariasse et al.*, 2011]. By 10.8 Ma, rocks belonging to the nappes above the Cretan Detachment were

reworked in, and unconformably overlain by sedimentary basins. Hence, E-W oriented tectonic thinning of the series above the Cretan Detachment must also have occurred largely before 10.8 Ma [*Zachariasse et al.*, 2011]. Only during the very last stage of exhumation was mechanical coupling between the HP-LT wedge and the rocks above the Cretan Detachment established (Figure 13).




[90] So far there are is only a limited amount of kinematic data available from the Tripolitza, Pindos and Uppermost Units of Crete. Late Neogene activity documented by a relatively younger fault set and by earthquake fault plane solutions [*Mountrakis et al.*, 2011] is indeed characterized by E-W extension. But before a larger data set is available our inference of primarily trench-parallel (i.e., E-W) extension of the supradetachment units remains a prediction. We note, however, that a very similar scenario to the one we infer for Crete applies to the Oligo-Miocene evolution of the Calabrian fore arc [*Séranne*, 1999]. Similar to Crete, a stack of units existed in the original fore arc prior to the inception of back-arc extension (consisting of the so-called Africo-Polsi and Aspromonte-Stilo units) that contemporaneous with the onset of back-arc extension started to deform along extensional detachments that accommodated trench-parallel (NE-SW to NNE-SSW) extension from ~29 to 15 Ma [*Heymes et al.*, 2008, 2010]. Simultaneously, however, a HP-LT wedge was formed below this forearc during ongoing northwestward oceanic subduction: the so-called "Ligurian ophiolites" [*Rossetti et al.*, 2001]. These exhumed by return flow below the trench-parallel extending fore arc, and their stretching lineations, including those at the extensional detachment between the Ligurian ophiolites and the overlying pre-Oligocene forearc units, are consistently trench-normal (i.e., WNW-ESE [*Rossetti et al.*, 2001]). Whereas Oligocene to early Miocene sedimentation occurred in the "Stilo-Capo d'Orlando" basin [*Weltje*, 1992] that developed as supradetachment basin above a top-to-the-NE detachment [*Heymes et al.*, 2008], final (Serravallian) exhumation of the Ligurian ophiolites was associated with the (Amantea) supradetachment basin resulting from trench-normal (WNW-ESE) extension [*Mattei et al.*, 2002]. Although the geology of Crete has not preserved sedimentary basins related to the trench-parallel extension phase, we argue the South Aegean forearc evolution in early to middle Miocene times is equivalent to that of the Oligocene to early Miocene history of Calabria.

### 6.2. Implications for Correlations Between Greek and Turkish Tectonic and Paleogeographic Units

[91] The restored configuration of the Aegean region at 25 Ma allows making few first-order inferences on the correlation of Anatolian and Aegean tectonic units. In the north, the Sakarya microcontinental block of northern Turkey cannot be correlated with any units in Greece. Since the rocks of

**Figure 12.** Stepwise evolution of the Aegean–West Anatolian extensional complexes. From 25 to 15 Ma, extension results from clockwise rotation of western Greece around stage poles north of Mount Olympos and in the northwest of the Rhodope. Since 15 Ma, extension resulted from clockwise rotation in western Greece and counterclockwise rotation of SW Anatolia. The Mid-Cycladic Detachment separated the two rotating domains. Trench-perpendicular extension amounted to a maximum of 400 km in total since 25 Ma (110 km between 25 and 15 Ma; 290 km since 15 Ma) in a NNE-SSW direction and was associated with up to 650 km of trench-parallel extension between Rhodos and the SW Peloponnesos. See the auxiliary material for Animation S1 of the reconstruction, highlighting area loss due to subduction and area gain due to extension.

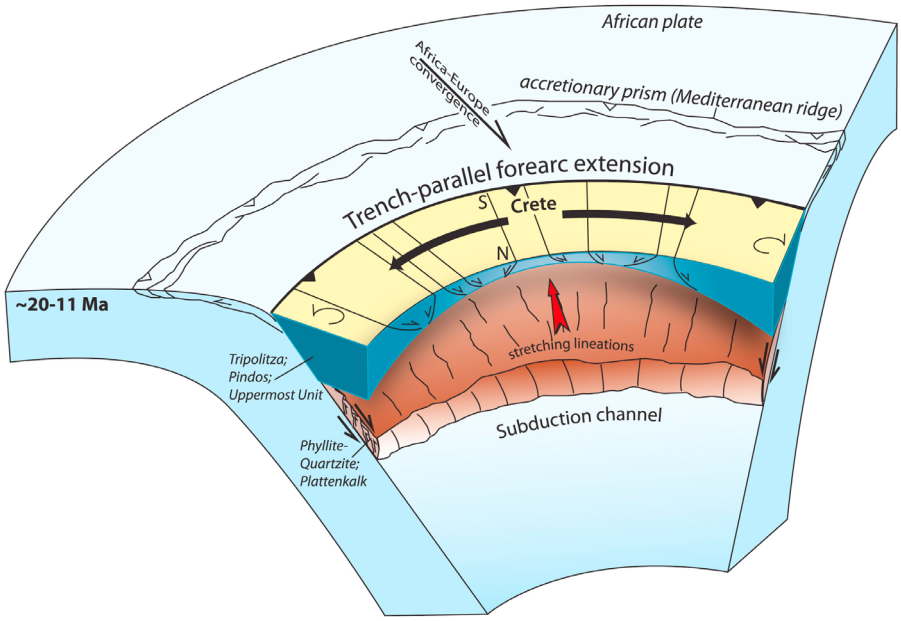



**Figure 13.** Three-dimensional block diagram illustrating the Miocene to recent tectonic evolution of Crete. E-W extension of the pre-Miocene fore-arc units above the Cretan Detachment (e.g., Tripolitza and Pindos units) dramatically thins the fore-arc crust. Simultaneously, the Cretan HP-LT metamorphic units (Phyllite-Quartzite and Plattenkalk) exhume by return flow above the retreating slab. Because these HP units are decoupled from the overriding plate, they do not primarily experience trench-parallel extension but mainly record deformation induced by their motion relative to the trench and the mantle wedge, leading to N-S stretching lineations. See text for further explanation.

the Circum-Rhodope belt appear to have been thrust over both Sakarya and Moesia, we infer that the Biga peninsula forms the western boundary of Sakarya. A transpressional transfer fault bounded the Sakarya block in the west probably during the Cretaceous, dextrally offsetting the Circum-Rhodope belt (Figures 8d and 8e). It seems likely that this compressional transfer fault was reactivated later as a Miocene age transtensional structure that corresponds to the restored position of the Attica-Evvia transform fault (corresponding to the West Biga Detachment; Figure 8). This same lineament may have had an earlier phase of activity associated with the mid-Cretaceous to Eocene opening of the Black Sea, which accommodated some 150–170 km of N-S extension [*Munteanu et al.*, 2011].

[92] The Pelagonian zone restores as a lateral equivalent of the continental lithosphere that carried the Bornova Flysch and Tavşanlı zones, and their nonmetamorphic equivalents in the West Anatolian Taurides. Both may have been physically connected to each other. They were bordered in the south by the Pindos/Cycladic Blueschist/Dilek units and the Afyon zone. In mainland Greece and the Aegean area the Pindos zone was deep marine and wide, probably underlain by thinned continental [e.g., *Schmid et al.*, 2008], possibly partly oceanic [e.g., *Channell and Kozur*, 1997; *Degnan and Robertson*, 1998, 2006; *Stampfli and Borel*, 2002] crust. However, in Turkey this Pindos pelagic realm wedges out eastward and it may paleogeographically change into the continental Afyon zone. Likewise, the pelagic sediments of the Pindos zone wedge out in the NW near Dubrovnik [*Schmid et al.*, 2008], NW of the Scutari-Pec line and just off the map presented in Figure 1.

[93] The younger age of the widespread obducted ophiolites with ~95–90 Ma metamorphic soles [*Çelik et al.*, 2006] in Turkey can be traced eastward all the way to Oman [*Dilek and Furnes*, 2011]. This represents a major difference in respect to the geology of the Hellenides and Dinarides, where obducted ophiolites are markedly older, namely of Jurassic age. At the northern margin of the Taurides-Anatolides of Turkey, subduction below an oceanic upper plate must therefore have occurred much later. The configuration of the subduction zone(s) active between Pelagonian-Tauride elements in a lower plate position and the Eurasian margin in an upper plate position must therefore have changed considerably and rather abruptly going from west to east during the Late Cretaceous and earlier. It is interesting to note that also the Cretaceous ophiolites of western Turkey appear to be bounded by the same N-S trending structure bounding Sakarya in the west (Figure 8e).

[94] As a result of the eastward disappearance of the Pindos pelagic basin, the neritic to hemipelagic continental realms to the south, carrying the Tripolitza and Ionian zones in Greece, and the Menderes Massif, Bey Dağları platform and Taurides in western Turkey, had a position oblique to the subduction zone. These external realms arrived in the subduction zone already at around ~50 Ma in western Turkey, but only around ~35 Ma in western Greece [*van Hinsbergen et al.*, 2005a, 2010c]. There is no evidence for a major discontinuity between the Menderes/Bey Dağları and Tripolitza/Ionian zones. Oceanic subduction has been continuous since ~35 Ma in western Turkey, whereas in southern Greece, oceanic subduction only started after the consumption of the Ionian zone, perhaps 23 Ma ago, or later.

In western Greece, collision occurred with the Apulian platform since the Pliocene.

## 7. Conclusions

[95] In this paper, we provide the first detailed kinematic reconstruction of the Aegean region of Greece and western Turkey, relative to Eurasia and since 35 Ma, embedded in the Atlantic plate circuit. The Aegean–West Anatolian region consists of stacked upper crustal slices (nappes) that were decoupled from the subducting African-Adriatic lithospheric slab. These are bounded by major thrusts and characterized by internally coherent stratigraphy and sedimentary facies. Each nappe represents a paleogeographic entity such as a deep basin or a carbonate platform, some of them overlain by previously obducted ophiolites. During the last ~35 Ma, and especially since 25 Ma, the nappe stack was cut by extensional detachments along which metamorphosed portions of the nappes were exhumed in metamorphic core complexes. Our reconstruction carefully restores these extensional complexes in order to correct for post-35 Ma extension. The following conclusions summarize our findings:

[96] 1. The Aegean region underwent a maximum of 400 km of trench-perpendicular (NE-SW) length change due to extension since 25 Ma, decreasing to the NW and east. Before that, between 45 and 25 Ma, some extension was accommodated in the Rhodope.

[97] 2. Aegean extension occurred in two stages. During the first stage, from 25 to 15 Ma, the western, southern and southeastern Aegean region rotated around stage poles in the NW Rhodope, and north of Mount Olympos This rotation was accommodated along a transfer fault bounding the Menderes massif in the SE. The maximum amount of extension in this time window was 110 km.

[98] 3. Most of the extension occurred after 15 Ma (max 290 km), and was accommodated by simultaneously operating opposite rotations of the western and eastern parts of the Aegean region.

[99] 4. Our reconstruction clearly identifies a normally underestimated element in oroclinal bending: Opposite rotations of the western and southeastern Aegean regions was associated with up to 650 km of trench-parallel extension between the SW Peloponnesos and the island of Rhodos. This is consistent with extreme thinning of the Cretan nappe pile.

[100] 5. Trench-parallel extension during the opposite rotation phase since 15 Ma in the central Aegean was accommodated along the Mid-Cycladic Detachment, which exhumed the SE Cyclades from below the NW Cyclades.

[101] 6. In using previously proposed detachment configurations for the submerged Central Aegean region we were unable to kinematically restore to the undeformed state. Instead we propose the alternative detachment configuration shown in Figure 12.

[102] 7. The Sakarya zone of Turkey, located to the north of the İzmir-Ankara suture zone, cannot be correlated to megaunits in Greece and the Balkans. It appears to be bounded in the west by a pre-35 Ma compressional transfer fault zone that became reactivated since 25 Ma as an extensional detachment.

[103] 8. The ophiolites of Turkey (and further east) that were obducted in the Cretaceous can also not be traced beyond western Turkey and easternmost Greece. They appear to be bounded by the same structure that limits the Sakarya zone in the west.

[104] 9. The units south of the İzmir-Ankara-Sava suture can be correlated. Bringing the Aegean region back into its 35 Ma configuration suggests that the Tripolitza zone of the Aegean paleogeographically corresponds to the Bey Dağları-Menderes platforms. The Pindos zone and Dilek nappe may have paleogeographically interfingered with the Afyon zone. The Pelagonian zone may correspond to the large blocks embedded in the Bornova Flysch and to the Tavşanlı zone.

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