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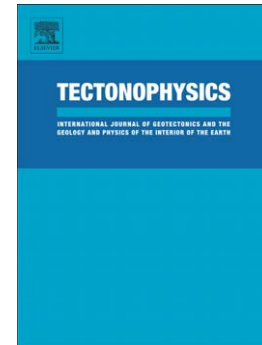
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Kinematic reconstructions and magmatic evolution illuminating crustal and mantle dynamics of the eastern Mediterranean region since the late Cretaceous

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Abstract

The relationship between subduction dynamics and crustal deformation in the Mediterranean region has been recently studied using three-dimensional (3D) numerical models. Such models require, however, a detailed information concerning the past geological evolution. We use stratigraphic, petrologic, metamorphic, structural, paleomagnetic and magmatic data to build new kinematic reconstructions of the eastern Mediterranean region since the late Cretaceous using the principle of non-rigid domains. The motions of the 56 deforming domains defined in this work are calculated based on published paleomagnetic rotations, the directions and amounts of displacement on crustal-scale shear zones and the burial and exhumation histories of the main metamorphic units. Extracted from these reconstructions, paleotectonic maps and lithospheric-scale cross-sections illustrate that the present-day subduction zone has been continuously retreating southward since the late Cretaceous and has accreted several small continental domains in the process. We find evidence for two back-arc-related extensional events: (1) slow extension along the Balkans and the Pontides in the late Cretaceous while the trench was long and linear and (2) faster extension in the Rhodope-Aegean-west Anatolian region since the Eocene-Oligocene. Rapid rotation of the Hellenides between 15 and 8 Ma probably indicates a slab tearing event below western Anatolia that could have further accelerated this extensional kinematics. Spatial distribution and the geochemical signature of magmatic centers integrated in these reconstructions allow us to trace mantle-related processes revealing the deep dynamics that controls both the magma genesis and the crustal deformation.

Keywords

Kinematic reconstructions; Eastern Mediterranean region; Oceanic and continental subduction; Back-arc opening; Subduction-related magmatism; Crustal and mantle processes

1. Introduction

Recent studies of Mediterranean subduction zones and back-arc basins have prompted the development of new models of interactions between slab dynamics, asthenospheric flow and lithospheric deformation [Faccenna *et al.*, 2003; Kreemer *et al.*, 2004; Jolivet *et al.*, 2009, 2013; Brun and Sokoutis, 2010; Faccenna and Becker, 2010; Sternai *et al.*, 2014]. These include (1) conceptual models based on the tectonic history recorded in subduction zones or mountain belts and their back-arc regions and on geophysical data, such as seismic anisotropy or seismic tomography, (2) instantaneous three-dimensional (3D) mantle velocity field calculations and (3) long-term 3D models of subduction using fully-coupled thermo-mechanical codes or analog experiments. The latter two types of model do not usually fit sequences of geological events on long timescales primarily due to computational limitations, while conceptual models do not include any quantitative physical constraints. The 3D complexity and evolution of the Mediterranean subduction zones, however, require to be constrained as precisely as possible to serve as tests for these models.

Several sets of kinematic and paleogeographic reconstructions have thus been proposed to constrain the geological evolution of the eastern Mediterranean region (figure 1) [Dercourt *et al.*, 1986, 1993; Ricou, 1994; Jolivet *et al.*, 2003; Barrier and Vrielynck, 2008; Stampfli and Hochard, 2009; van Hinsbergen and Schmid, 2012]. All these reconstructions show that the geodynamics of this region has been governed by the convergence between Africa and Eurasia and by the interactions between several oceanic domains belonging to the Neo-Tethys Ocean and continental blocks previously rifted from the African margin and belonging to the Apulian (or Adria) plate. However, apart from van Hinsbergen and Schmid [2012], these reconstructions generally consider the entire western Tethyan active margin and do not specifically focus on the smaller-scale eastern Mediterranean region, where subduction, collision, obduction, slab roll-back and tearing processes have occurred since the Mesozoic [Le Pichon and Angelier, 1979; Şengör and Yilmaz, 1981; Jolivet and Faccenna, 2000; Jolivet *et al.*, 2003; Faccenna *et al.*, 2006, 2013; Brun and Faccenna, 2008; Ring *et al.*, 2010; Brun and Sokoutis, 2010; Jolivet and Brun, 2010]. Detailed kinematic reconstructions

are therefore necessary to describe this apparently complex tectonic evolution especially as there is no consensus about the lithospheric behavior and the crustal-scale structures accommodating deformation. W-E and N-S trending detachments have thus been successively proposed to accommodate the Oligocene-Miocene back-arc extension in the Aegean domain [Jolivet and Brun, 2010, Grasemann *et al.*, 2012; van Hinsbergen and Schmid, 2012]. To tackle this problem, robust data provided by the large number of field studies performed in this region have to be considered. Integrating magmatism and its sources in these reconstructions is also particularly useful to discuss the interactions between crustal and mantle processes. *Pe-Piper and Piper*, [2006] thus added to their simplified kinematic reconstructions information about the distribution and geochemical characteristics of magmatic centers through time and then discussed the respective roles of slab roll-back and break-off, asthenospheric upwelling, as well as major crustal-scale structures, such as strike-slip faults and detachments.

In this paper, we reconstruct the tectonic and magmatic evolution of the eastern Mediterranean region in unprecedented detail. First, we review and discuss stratigraphic, petrologic, metamorphic, structural, paleomagnetic and magmatic data, providing information about the tectono-metamorphic history of various crustal units and their paleogeographic environments. Then we propose new paleotectonic maps and lithospheric-scale cross-sections illustrating the 3D evolution of the subduction zone geometry. The full set of tectonic reconstructions is also available as a movie (see Supplementary Materials, movie S1). Based on these reconstructions, we describe the continuous evolution of the Tethyan subduction zone since the late Cretaceous characterized by two back-arc-related extensional periods in the late Cretaceous and the Oligocene-Miocene. An increasing rate of trench retreat since 15 Ma in the Aegean-western Anatolian region suggests a slab tearing event, resulting in curvature of the trench and large-scale block rotations. Integrated in these reconstructions, the space and time evolution of magmatism is used to discuss the importance of subduction dynamics and related asthenospheric flow on controlling the migration of magmatic centers as well as crustal deformation.

2. Methodology

Plate-tectonic reconstructions consider the displacement of rigid polygons (representing different continental or oceanic crustal blocks) relative to one of the polygons considered stable or within an absolute framework. In this case, strain is considered to be entirely localized at the edges of these polygons. However, at the scale of the eastern Mediterranean region, deformation is widely distributed [Le Pichon and Kreemer, 2010; Pérouse *et al.*, 2012] and therefore the definition of only rigid polygons is partially misleading. One possible way to approximate this non-rigid behavior at the scale of the studied region is to consider (1) many small rigid polygons or (2) rigid boundaries of deforming polygons. The space between these different polygons is then considered as a fully deforming zone (i.e. principle of “continuously closed plates” [Gurnis *et al.*, 2012] or “deformable topologies” [van Hinsbergen and Schmid, 2012]). We use both strategies by defining 56 independently moving domains comprising both rigid lines and intervening deforming areas (figure 2). Due to their non-rigid behavior, the domain boundaries are not necessarily defined by rigid lines and can be implicitly defined between two deforming areas. These domains are integrated into a global plate kinematic model consisting of 30 rigid polygons [Barrier and Vrielynck, 2008] including Africa, Arabia, Eurasia and the Moesian platform (figure 2). Motions of the domains are then constrained by numerous geological data (figure 3) including (1) paleomagnetism to constrain the amount of vertical axis rotation, (2) directions and amounts of displacement on large-scale structures, such as detachments and strike-slip faults and (3) tectonic and metamorphic histories (i.e. pressure-temperature-time path) of main tectonic units. The latter also provides information on the burial and exhumation processes, constraining the vertical tectonic evolution of these units.

Using the interactive plate-tectonic reconstruction and visualization software *GPlates* (www.gplates.org) [Boyden *et al.*, 2011], we reconstruct the kinematics of the eastern Mediterranean region backward from the present-day plate configuration to the late Cretaceous. Relative motions of deforming domains are described by finite rotations given by the geographical

coordinates of the rotation pole and the rotation angle. We then obtain, for each defined domain, a succession of total reconstruction poles, corresponding to the finite rotation from the present-day position to the position at a given time in the past (positive and negative angles of rotation depicting, respectively, clockwise and counterclockwise rotations) (see Supplementary Materials, table S1). All these rotations are defined with respect to the Moesian platform, which is considered as having belonged to stable Eurasia since the late Cretaceous (figure 3) [van Hinsbergen *et al.*, 2008].

Finally, in order to discuss the space and time evolution of magmatism across the study area, we compile a GIS database of the magmatic products emplaced in this region since the late Cretaceous. We thus consider 512 magmatic centers, which are all characterized by their geochemical composition and their period of activity. Each magmatic center is then kinematically attached to the domain to which it belongs and moves with it in the process of reconstruction.

3. Geodynamic overview of the eastern Mediterranean region

The geodynamic history of the eastern Mediterranean region has been dominated by the convergence of Africa and Eurasia since the Mesozoic, inducing subduction, collision and obduction processes [Şengör and Yilmaz, 1981; Dercourt *et al.*, 1986; Ricou *et al.*, 1986; Ricou, 1994]. Accretionary wedges and associated orogenic belts, such as the Hellenides and the Anatolides-Taurides (figures 1 and 4), were formed along this active margin by the decoupling of crustal nappes from the subducting lithosphere and their accretion to the overriding plate [Le Pichon and Angelier, 1979; Jolivet *et al.*, 2003; van Hinsbergen *et al.*, 2005b]. These nappes were partially buried within the subduction zone, where they underwent high pressure-low temperature (HP-LT) metamorphism [Bonneau and Kienast, 1982; Okay, 1986; Avigad and Garfunkel, 1989; Schermer, 1993]. These metamorphic rocks were then partly exhumed in a syn-orogenic context within the subduction channel or the accretionary wedge, preserving HP-LT parageneses [Jolivet *et al.*, 2003; Ring and Layer, 2003; Ring *et al.*, 2007, 2010; Huet *et al.*, 2009]. Concomitantly with the migration

of the subduction zone, these accretionary wedges were transferred to the back-arc domain where buried rocks underwent medium to high temperature-medium pressure (MT- to HT-MP) metamorphism, overprinting the former HP-LT parageneses [Jansen and Schuiling, 1976; Altherr *et al.*, 1982; Buick and Holland, 1989; Gautier and Brun, 1994; Jolivet and Brun, 2010; Scheffer *et al.*, 2015]. As a result of various processes, such as gravitational collapse or slab roll-back, the subsequent collapse of the accretionary wedges and the opening of back-arc basins allowed for the post-orogenic exhumation of the lower parts of the stretched crust as metamorphic core complexes (MCCs), such as in the Rhodope and Menderes massifs and the Cyclades (figures 1 and 4) [Lister *et al.*, 1984; Sokoutis *et al.*, 1993; Jolivet *et al.*, 1994, 2004a; Lips *et al.*, 2001; Vanderhaeghe and Teyssier, 2001; Ring *et al.*, 2003; Brun and Sokoutis, 2007].

3.1. *Sedimentary and tectono-metamorphic constraints*

In the following section, we review the main stratigraphic, petrologic, metamorphic, structural and paleomagnetic data characterizing the various tectonic domains (figure 1) in order to constrain our kinematic reconstructions of the eastern Mediterranean region.

3.1.1. *The Balkans*

Displaying a present-day “L”-shape surrounding the Moesian platform, the Balkans have been part of the Eurasian margin since the Triassic Tethyan rifting [Georgiev *et al.*, 2001; Okay *et al.*, 2001a]. They consist of an orogenic belt in which four different areas can be distinguished: the Apuseni, Banat, Timok and Srednogorie regions (figure 1). A compressional period in the early Cretaceous was followed by a late Cretaceous phase of extension (figure 5) that included the opening of Gosau-type basins in the Banat and Apuseni regions [Berza *et al.*, 1998; Willingshofer *et al.*, 1999]. In the Srednogorie region, transtensional basins formed at the end of the early Cretaceous and evolved into rifting zones up to the beginning of the Cenozoic [Bergerat *et al.*, 2010].

An inversion of the Cretaceous extensional structures then occurred from the Paleocene and a N-vergent foreland fold-and-thrust belt developed throughout the Balkans, culminating in the middle-late Eocene (figure 5) [Doglioni *et al.*, 1996; Bergerat *et al.*, 2010].

Finally, paleomagnetic data acquired in the Apuseni mountains show an $\sim 60^\circ$ early Miocene clockwise rotation in this part of the belt (figure 3) [Pătraşcu *et al.*, 1994].

3.1.2. The Rhodope massif

Located south of the Balkans, the Rhodope massif consists mainly of a stack of nappes partially disrupted by several MCCs (i.e. the central and south Rhodope core complexes) (figure 6) [Burg *et al.*, 1990, 1996; Bonev *et al.*, 2006; Brun and Sokoutis, 2007]. Stacked units emplaced progressively southward [Burg *et al.*, 1996; Dinter, 1998] underwent HP metamorphism (i.e. 12-17 kbar, 750-811 $^\circ\text{C}$) with locally ultra-HP conditions no later than the late Cretaceous (U-Pb crystallization ages on zircon) [Liati *et al.*, 2002; Bonev *et al.*, 2006; Bauer *et al.*, 2007].

Afterwards, MT-MP metamorphism (i.e. 8-10 kbar, 560-650 $^\circ\text{C}$) was recorded in the Paleocene. Subsequent exhumation of these rocks as a MCC initiated first in the central Rhodope massif ~ 55 Ma ago below top-to-the NE ductile-brittle detachments (figures 3 and 5) [Burg *et al.*, 1996; Bonev *et al.*, 2006]. This exhumation is associated with an $\sim 11^\circ$ clockwise rotation (figure 3) [Brun and Sokoutis, 2007]. Contemporaneous sedimentation on the hanging wall of these detachments including olistholiths of the underlying metamorphic rocks indicates that the central Rhodope core complex was partially exposed at the surface in the late Eocene [Burchfiel *et al.*, 2003; Bonev *et al.*, 2006]. Exhumation below these detachments then ceased in the early Oligocene and high-angle normal faulting developed [Marchev *et al.*, 2004a; Wüthrich, 2009]. Further east, the Thrace basin opened in the middle Eocene-late Oligocene resulting in the deposition of sediments [Turgut and Eseller, 2000; Okay *et al.*, 2010a]. The amount of extension in the Thrace basin remains poorly constrained and van Hinsbergen and Schmid [2012] suggested an $\sim 2^\circ$ counterclockwise rotation of the southern part of this basin to describe its opening (figure 3).

Further south, exhumation of the southern Rhodope core complex started ~40 Ma ago, accommodated by the top-to-the SW Kerdyllion detachment whose hanging wall (i.e. the Vertiskos unit) has stayed in near-surface conditions since the early-middle Eocene (figures 3 and 6) [Sokoutis *et al.*, 1993; Brun and Sokoutis, 2007; Wüthrich, 2009; Kydonakis *et al.*, 2014]. Based on paleomagnetism [Kondopoulou and Westphal, 1986] and on the curvature of the stretching lineations in the MCC, Brun and Sokoutis [2007] hypothesize an ~30° clockwise rotation of the hanging wall of the Kerdyllion detachment (including the Chalkidiki Peninsula) with the major part of the rotation occurring in the final stage of exhumation of the south Rhodope core complex after 15 Ma [van Hinsbergen *et al.*, 2008]. Since the middle Miocene, sedimentary basins have also developed over the whole MCC, controlled by normal faults [Brun and Sokoutis, 2007].

3.1.3. The Hellenides

Located southwest of the Rhodope massif, the Hellenides is a W- to SW-vergent nappe stack constituted by several continental and oceanic units of low-metamorphic grade (figures 1, 4 and 6) [Aubouin, 1959; van Hinsbergen *et al.*, 2005b].

The Vardar suture zone is the most internal part of the Hellenides and represents the remnant of a wide oceanic basin that was partially obducted southward over the Pelagonian carbonate platform in latest Jurassic times [Ricou *et al.*, 1986; Schmid *et al.*, 2008]. This oceanic basin closed in the latest Cretaceous to Paleocene during the accretion of the ~200 km wide Pelagonian carbonate platform (figure 5) [Ricou *et al.*, 1998; van Hinsbergen *et al.*, 2005b].

Underthrust below the Pelagonian nappe in the late Paleocene-Eocene, the Pindos nappe was originally an oceanic basin with its continental margins [Aubouin, 1959] extending from the Dinarides in the northwest (i.e. the Budva-Cukali zone [Robertson and Karamata, 1994]) to the Aegean domain in the southeast and possibly as far as the Lycian basin in the Taurides [Bonneau, 1984]. This basin had an original width of at least 300 km [van Hinsbergen *et al.*, 2005b].

In the Oligocene, the Pindos nappe overthrust the Gavrovo-Tripolitza nappe consisting of an ~150 km wide carbonate platform (figure 5) [Aubouin, 1959; Sotiropoulos *et al.*, 2003; van Hinsbergen *et al.*, 2005b]. At the same time, the Gavrovo-Tripolitza nappe overthrust the Ionian platform whose original width is also estimated at ~150 km (figure 5) [van Hinsbergen *et al.*, 2005b].

Finally, forming the most external part of the Hellenides, the Paxos nappe (or Pre-Apulian nappe) consists of a carbonate platform that was underthrust below the Ionian nappe in the Miocene (figure 5). This platform has no lateral continuation in eastern Greece, where the Mesogean oceanic lithosphere has been subducting since the early-middle Miocene [Underhill, 1989; Le Pichon *et al.*, 2002]. Along this oceanic subduction zone, a sedimentary wedge called the Mediterranean ridge developed since ~15 Ma against the Hellenic thrust belt (figures 1 and 4) [Le Pichon *et al.*, 1982, 2002; Chaumillon *et al.*, 1996; Kopf *et al.*, 2003; Yem *et al.*, 2011]. At the junction between the Paxos platform and the Mesogean oceanic basin, the Kephallonia strike-slip fault has accommodated 100-120 km of dextral offset since ~5 Ma, (figure 3) [Finetti, 1982; Kreemer and Chamot-Rooke, 2004; Royden and Papanikolaou, 2011].

As shown by paleomagnetism, the Hellenides underwent significant rigid rotation during its accretion, with ~40° clockwise rotation between 15 and 8 Ma and an additional 10° clockwise rotation since the Pliocene (figure 3) [Kissel and Laj, 1988; van Hinsbergen *et al.*, 2005a]. A recent study shows, however, greater Pliocene-Quaternary clockwise rotation of ~23° in the southeastern Hellenides (figure 3), suggesting recent tectonic decoupling between these two parts of the orogenic belt [Bradley *et al.*, 2013].

3.1.4. The Cyclades

Located in the center of the Aegean domain, the Cyclades partly expose the buried parts of the Hellenides, exhumed since the Eocene as MCCs in low, medium and/or high temperature environments (figures 4 and 6) [Bonneau, 1984; Lister *et al.*, 1984; Faure and Bonneau, 1988;

Gautier and Brun, 1994; Jolivet et al., 1994, 2003, 2004a; Vanderhaeghe and Teyssier, 2001; van Hinsbergen et al., 2005c; Ring et al., 2010; Grasemann et al., 2012].

The buried Pindos oceanic basin and its continental margins underwent HP-LT metamorphism (i.e. ~15-20 kbar, ~500 °C) whose metamorphic peak is dated at ~50 Ma, forming the Cycladic Blueschists unit and the Cycladic basement [*Bonneau and Kienast, 1982; Bonneau, 1984; Wijbrans and McDougall, 1988; Tomaschek et al., 2003; Lagos et al., 2007*]. Exhumation of these metamorphic units first occurred between 45 and 35-30 Ma in a syn-orogenic context, following a cold retrograde path (figure 5) [*Altherr et al., 1979; Bonneau and Kienast, 1982; Avigad and Garfunkel, 1989; Trotet et al., 2001*]. These Eocene HP-LT parageneses are particularly well preserved in the Syros and Sifnos islands where syn-orogenic extensional shear zones, such as the Vari detachment in Syros (figure 6), show top-to-the ENE and top-to-the N senses of shear, respectively (figure 3) [*Trotet et al., 2001; Keiter et al., 2004; Ring et al., 2011; Roche et al., submitted*]. In the Ios island, a coeval top-to-the S structure called the South Cycladic Shear Zone (SCSZ) is observed at the base of the Cycladic Blueschists unit (figure 6). This structure was interpreted first as an extensional shear zone [*Lister et al., 1984*] then redefined as a thrust [*Huet et al., 2009*]. An extrusion model was proposed for the Eocene exhumation of the Cycladic Blueschists unit in the Cyclades with the Vari detachment in Syros and the SCSZ in Ios defining, respectively, the top and the base of the subduction channel or the extrusion wedge [*Jolivet et al., 2003; Ring et al., 2007, 2010; Brun and Faccenna, 2008; Huet et al., 2009; Jolivet and Brun, 2010*]. This unit also crops out within the Hellenides in the Olympos, Ossa and Pelion tectonic windows (figure 6), where it displays similar bivergent ductile deformation with both top-to-the NE and top-to-the WSW senses of shear associated with the exhumation of the HP-LT parageneses (figure 3) [*Godfriaux, 1968; Blake et al., 1981; Schermer, 1993; Lips et al., 1998; Lacassin et al., 2007*].

Following this, the Cycladic Blueschists unit underwent an early Oligocene MT-MP metamorphic overprint (i.e. ~9 kbar, 550-570 °C) followed by a greenschist-facies retrograde metamorphism observed in the Andros, Tinos and Kea islands, in Lavrion (figure 6) [*Altherr et al., 1982; Parra et*

al., 2002; Bröcker *et al.*, 2004; Jolivet and Brun, 2010; Grasemann *et al.*, 2012; Huet *et al.*, 2014; Scheffer *et al.*, 2015] and in the Olympos, Ossa and Pelion tectonic windows [Schermer, 1993; Lips *et al.*, 1999; Lacassin *et al.*, 2007]. In the lower Miocene, HT-MP metamorphism (i.e. 5-8.5 kbar, 500-700°C) and associated migmatites also developed in the central Cyclades, such as in the Naxos, Paros, Mykonos and Ikaria islands [Jansen and Schuiling, 1976; Lister *et al.*, 1984; Urai *et al.*, 1990; Keay *et al.*, 2001; Vanderhaeghe, 2004; Duchêne *et al.*, 2006; Ring, 2007; Kruckenberg *et al.*, 2011; Beaudoin *et al.*, 2015; Laurent *et al.*, 2015]. Exhumation of these rocks as MCCs was accommodated during the Oligocene-Miocene by several ductile-brittle detachment systems (figures 4 and 6) comprising the top-to-the NE North Cycladic Detachment System (NCDS) [Faure *et al.*, 1991; Lee and Lister, 1992; Gautier and Brun, 1994; Mehl *et al.*, 2005, 2007; Brichau *et al.*, 2007, 2008; Jolivet *et al.*, 2010a], the top-to-the N Naxos/Paros Extensional Fault System (NPEFS) [Urai *et al.*, 1990; Gautier *et al.*, 1993; Brichau *et al.*, 2006; Seward *et al.*, 2009; Bargnesi *et al.*, 2013] and the top-to-the SW West Cycladic Detachment System (WCDS) [Grasemann and Petrakakis, 2007; Brichau *et al.*, 2010; Iglseder *et al.*, 2011; Grasemann *et al.*, 2012]. The minimum offset of these extensional structures is estimated at 60-100 km for the NCDS [Jolivet *et al.*, 2004a; Brichau *et al.*, 2008] and ~50 km for the NPEFS (figure 3) [Brichau *et al.*, 2006]. The low-grade Upper Cycladic unit composing the hanging wall of these detachment systems is the equivalent of the Pelagonian nappe, essentially consisting of ophiolitic material related to the Jurassic Vardar ophiolites (e.g. in Crete) [Koepke *et al.*, 2002] or to the late Cretaceous Lycian and Tauride ophiolites (e.g. in Tinos) [Katzir *et al.*, 1996]. In Mykonos, Ikaria and Naxos, syn-tectonic sedimentary basins were also deposited in the hanging wall of these detachments during the Miocene [Photiades, 2002; Sánchez-Gómez *et al.*, 2002; Lecomte *et al.*, 2010; Laurent *et al.*, 2015]. The lack of clasts belonging to the Cycladic Blueschists unit indicates late exposure of this unit in the late Miocene [Sánchez-Gómez *et al.*, 2002].

Following this extensional tectonic regime, distributed strike-slip tectonics then E-W compression developed in the Cyclades in the late Miocene inducing folding and local reactivation of detachment planes as reverse faults (figure 5) [Ring *et al.*, 1999a; Avigad *et al.*, 2001; Menant *et al.*, 2013].

Paleomagnetic studies from middle Miocene intrusions in the Cyclades indicate an $\sim 20^\circ$ clockwise and an $\sim 30^\circ$ counterclockwise rotation in Mykonos and Naxos, respectively (figure 3) [Morris and Anderson, 1996]. To accommodate these two opposite block rotations, a large-scale dislocation fault was suggested by Walcott and White [1998] (i.e. the so-called Mid-Cycladic Lineament). A similar concept was used by Philippon *et al.* [2012, 2014], who proposed the existence of a dextral fault linking the Ikaria Basin and the Mirthes Basin (i.e. the Mirthes-Ikaria strike-slip fault). Recently, Jolivet *et al.* [2015] proposed instead a left-lateral transfer zone, accommodated as a gradient of finite extension across a large part of the central Aegean domain.

3.1.5. Crete and the Peloponnese

In the Peloponnese and Crete, located in the southern part of the Aegean domain, late Oligocene HP-LT metamorphism (i.e. 16-18 kbar, 400-550 °C) affected the basement of the Gavrovo-Tripolitza and the Ionian nappes, thus giving the Phyllite-Quartzite and the Plattenkalk units, respectively (figures 4 and 6) [Bonneau and Kienast, 1982; Seidel *et al.*, 1982; Theye *et al.*, 1992; Jolivet *et al.*, 1996; van Hinsbergen *et al.*, 2005c; Trotet *et al.*, 2006]. Exhumation of these units occurred between 24 and 9 Ma (Ar-Ar crystallization ages on white mica and fission track ages on zircon) in the footwall of the Cretan detachment, which accommodated more than 100 km of displacement (figures 3 and 5) [Jolivet *et al.*, 1996, 2010c; Ring *et al.*, 2001; van Hinsbergen and Meulenkamp, 2006; Papanikolaou and Royden, 2007; Marsellos *et al.*, 2010]. Associated deformation is characterized in Crete by a N-S stretching lineation with mainly top-to-the N sense of shear, while in the Peloponnese this stretching lineation varies from NE-SW in the north to E-W in the south with bivergent kinematic indicators (figure 3) [Jolivet *et al.*, 1996, 2010c]. The hanging wall of the Cretan detachment consists of the low-grade Gavrovo-Tripolitza unit and a Miocene to

late Pliocene sedimentary basin [Jolivet *et al.*, 1996; van Hinsbergen and Meulenkamp, 2006; Seidel *et al.*, 2007]. The lack of metamorphic clasts in the oldest sediments indicates exposure of the Phyllite-Quartzite nappe ~11-10 Ma ago at the earliest [van Hinsbergen and Meulenkamp, 2006; Seidel *et al.*, 2007].

Since the late Miocene, high-angle normal faults have crosscut the Cretan detachment and accommodated the uplift of Crete and the opening of the Cretan Sea [Angelier *et al.*, 1982; van Hinsbergen and Meulenkamp, 2006]. Finally, additional extensional basins opened in the Peloponnese and along the eastern coast of the Hellenides, within the so-called Central Hellenides Shear Zone [Papanikolaou and Royden, 2007; Royden and Papanikolaou, 2011]. These basins include the Gulf of Corinth, which has accommodated ~15 km of N-S extension since the Pliocene (figure 3) [Armijo *et al.*, 1996; Rohais *et al.*, 2007; Jolivet *et al.*, 2010b].

3.1.6. The Black Sea basin

Located north of the Pontides, the Black Sea basin is mainly composed of oceanic-type crust overlain by ~15 km of sediments (figures 1 and 4) [Zonenshain and Le Pichon, 1986; Nikishin *et al.*, 2015a]. Deep reflection seismic studies show two sub-basins whose rifting started in the early Cretaceous [Görür, 1988; Hippolyte *et al.*, 2010; Nikishin *et al.*, 2015b]. Intense crustal stretching ultimately led to oceanic spreading from Cenomanian to Santonian times (i.e. ~100 to 85 Ma) [Nikishin *et al.*, 2015b]. The Black Sea basin then underwent compressional deformation in the late Eocene (figure 5), identifiable in the observed reworking of pre-existing structures [Zonenshain and Le Pichon, 1986; Nikishin *et al.*, 2015b].

3.1.7. The Pontides

The Pontides is an E-W trending orogenic belt in the north of Turkey (figures 1 and 7) that has belonged to the Eurasian margin since the closure of the Paleo-Tethys Ocean in Jurassic times [Yilmaz *et al.*, 1997]. Since the late Cretaceous, the Pontides have recorded a succession of

extensional and compressional or transpressional events (figure 5). Late Cretaceous extension controlled the opening of sedimentary basins [Sunal and Tüysüz, 2002; Okay *et al.*, 2013]. Further south, a forearc basin developed in front of a succession of S-vergent thrust units where several HP-LT metamorphic events have been recorded during the Cretaceous [Okay, 1986; Okay and Satır, 2000a; Okay *et al.*, 2013] (figure 5). These units were subsequently exhumed as a syn-orogenic extrusion wedge, such as in the central Pontides and the Biga peninsula [Okay *et al.*, 2006; Beccaletto *et al.*, 2007].

Since the Paleocene, the tectonic regime in the Pontides has become compressional with the inversion of the late Cretaceous basins and the subsequent development of a N-vergent fold-and-thrust belt [Yılmaz *et al.*, 1997; Sunal and Tüysüz, 2002]. Oroclinal bending affected the central Pontides causing both clockwise and counterclockwise rotations in the latest Cretaceous-Paleocene [Meijers *et al.*, 2010]. In the eastern Pontides, there was rapid uplift in the Paleocene-early Eocene [Boztuğ *et al.*, 2004]. The late Cretaceous forearc basin also migrated southward up to the Eocene-Oligocene and evolved into a foreland basin, as attested by the deposition of deep-sea, shallow marine and, finally, continental sediments [Yılmaz *et al.*, 1997; Okay *et al.*, 2001b; Kaymakci *et al.*, 2009]. In the late Oligocene-early Miocene, HT-MP metamorphism (i.e. 5 ± 2 kbar, 640 ± 50 °C) was recorded in the western part of the Pontides (Rb-Sr crystallization ages on muscovite and biotite and fission track ages on apatite, figure 5) [Okay and Satır, 2000b; Cavazza *et al.*, 2009]. Exhumation of these rocks in a post-orogenic context then allowed the formation of the Kazdağ MCC below the top-to-the N Alakeçi and top-to-the S Şelale detachments (figure 7) [Okay and Satır, 2000b; Beccaletto and Steiner, 2005; Bonev and Beccaletto, 2007].

Finally, the dextral strike-slip North Anatolian Fault (NAF), initiated in the eastern Pontides 13-11 Ma ago, propagated westward as a result of the extrusion of Anatolia [Şengör *et al.*, 2005]. Although its exact timing is still debated [Armijo *et al.*, 1999; Yaltırak *et al.*, 2000; Şengör *et al.*, 2005; Le Pichon and Kreemer, 2010], the NAF likely localized in the Dardanelle Strait 6-5 Ma ago, opening the Marmara Sea as a pull-apart basin [Armijo *et al.*, 1999; Melinte-Dobrinescu *et al.*,

2009]. The total dextral offset of this structure in its western segment is estimated at 70-90 km (figure 3) [Armijo *et al.*, 1999; Şengör *et al.*, 2005; Le Pichon and Kreemer, 2010; Akbayram *et al.*, 2015].

3.1.8. The Central Anatolian Crystalline Complex (CACC)

The CACC is a continental domain separated from the Pontides by the Ankara-Erzincan oceanic suture zone (figures 1 and 7) and is mainly composed of HT metamorphic rocks and widespread plutons [Göncüoğlu *et al.*, 1991; Aydin *et al.*, 1998]. It also presents remnants of an ophiolitic unit (i.e. the central Anatolian ophiolites) emplaced from an intraoceanic subduction zone and obducted toward the south 90-85 Ma ago [Yaliniz and Göncüoğlu, 1998]. At the same time, regional HT-MP metamorphism (i.e. ~6 kbar, ~725 °C) affected the CACC, which subsequently underwent exhumation and cooling as MCCs along detachments (figure 5) [Whitney *et al.*, 2003; Lefebvre *et al.*, 2011].

The CACC then underwent compressional tectonics since the Paleocene, resulting in complete tectonic reorganization of the complex [Lefebvre *et al.*, 2013] and the oroclinal bending of the late Cretaceous-Oligocene forearc/foreland Çankırı basin, located in the northern part of the CACC [Kaymakci *et al.*, 2003, 2009; Meijers *et al.*, 2010].

A reheating of the southeastern part of the CACC dated at ~30 Ma (Ar-Ar crystallization ages on K-feldspar) has been attributed to a new burial stage associated with the activity of the neighboring transpressional central Anatolian fault (figure 7) [Koçyiğit and Beyhan, 1998; Jaffey and Robertson, 2001; Idleman *et al.*, 2014].

3.1.9. The Bornova flysch zone

Located in the westernmost part of Turkey between the Pontides and the Anatolide-Tauride block (figures 1 and 7), the Bornova flysch zone mainly consists of a non- or slightly metamorphosed mélange with sedimentary, ophiolitic and rare volcanic blocks embedded in a latest Cretaceous-

Paleocene terrigenous or ophiolite-derived sedimentary matrix [Okay *et al.*, 2001b, 2012; Robertson *et al.*, 2009]. This unit is folded and sheared toward the southwest and unconformably overlaid by undeformed Eocene sediments [Robertson *et al.*, 2009].

3.1.10. The Anatolide-Tauride block

Located south of the Izmir-Ankara and the Inner-Tauride oceanic suture zones (figures 1 and 7), the Anatolide-Tauride block consists of a series of metamorphic domains in the north (i.e. the Anatolides) and non-metamorphosed sediments in the south (i.e. the Taurides) [Okay, 1984]. During the Mesozoic, it consisted of a carbonate platform that became more pelagic since the Santonian (86-83 Ma) before it underthrust the Lycian and Tauride ophiolites [Şengör and Yilmaz, 1981; Dilek *et al.*, 1999; Celik *et al.*, 2006]. This obduction event resulted from intraoceanic subduction initiated in the Turonian (94-90 Ma) [Okay *et al.*, 2001b]. Similarly, the Antalya ophiolites overthrust the southern margin of the Tauride platform cropping out now south of Isparta region (figure 7) [Robertson, 2002]. Compressional front in the northern Tauride margin then migrated southward up to the Bey Dağları platform and its foreland basin, overthrust by the non-metamorphosed Lycian nappes in the Miocene [de Graciansky, 1967; Poisson, 1977; Gutnic *et al.*, 1979; Collins and Robertson, 1998; van Hinsbergen *et al.*, 2010a]. Further east, the late Cretaceous-early Cenozoic subduction of the Mesogean oceanic lithosphere below the southern Tauride margin was followed by the Oligocene-early Miocene continental subduction/collision of the Arabian plate allowing for the building of the Bitlis belt (figure 5) [Dercourt *et al.*, 1986; Hempton, 1987; Jolivet and Faccenna, 2000; Agard *et al.*, 2005; Allen and Armstrong, 2008; Barrier and Vrielynck, 2008; Okay *et al.*, 2010b; McQuarrie and van Hinsbergen, 2013].

Marking the initiation of the southward migration of the thrust front through the Tauride platform, the buried northern Tauride margin underwent HP-LT metamorphism, forming the Tavşanlı unit whose pressure peak conditions (i.e. 24 ± 3 kbar, 430 ± 30 °C) were reached between 88 and 78 Ma (figures 5 and 7) [Okay, 1986, 2002; Sherlock *et al.*, 1999; Seaton *et al.*, 2009; Plunder *et al.*,

2013]. Another HP-LT metamorphic event was then recorded in the underlying Afyon and Ören units between 70 and 65 Ma [Rimmelé *et al.*, 2003a, 2006; Candan *et al.*, 2005; Pourteau *et al.*, 2010, 2013]. These HP-LT metamorphic units were subsequently exhumed, reaching the surface in the Eocene [Özcan *et al.*, 1988]. Kinematic indicators related to this cold exhumation show mostly top-to-the NE sense of shear, notably in the Ören unit [Rimmelé *et al.*, 2003a]. Structurally below, the Mesozoic carbonate cover of the southern Menderes massif (also belonging to the Tauride platform) underwent similar HP-LT metamorphism in the Eocene forming the Kurudere-Nebiler unit (figures 5 and 7) [Rimmelé *et al.*, 2003b; Whitney *et al.*, 2008; Pourteau *et al.*, 2013]. At the same time, the major part of the Menderes massif underwent MT- to HT-MP Barrovian-type metamorphism (i.e. 5-8 kbar, 450-660 °C in the central Menderes massif) called the *Main Menderes Metamorphism* (figure 5) [Akkök, 1983; Şengör *et al.*, 1984; Bozkurt and Oberhänsli, 2001; Lips *et al.*, 2001; Okay, 2001]. Oligocene migmatites crop out in the northern Menderes massif, suggesting a HT event following the *Main Menderes Metamorphism* [Catlos and Çemen, 2005; Bozkurt *et al.*, 2010], while the rest of the massif experienced greenschist-facies metamorphism [Hetzel *et al.*, 1995a; Bozkurt and Oberhänsli, 2001]. Exhumation of the northern Menderes massif was accommodated by the top-to-the N Simav detachment, which displays a minimum offset of ~50 km up to the late Miocene (i.e. ~8 Ma) [Işık and Tekeli, 2001; Ring *et al.*, 2003; Bozkurt *et al.*, 2011]. In the southern Menderes massif, coeval top-to-the S extensional shear has also been recognized and volcanoclastic rocks were unconformably deposited in the early Miocene (figure 4) [Bozkurt and Satır, 2000; Gessner *et al.*, 2001a, 2001b; Ring *et al.*, 2003]. From the middle Miocene to the late Miocene-Pliocene, exhumation was localized in the central Menderes massif along the top-to-the NE Alaşehir and the top-to-the S Büyük ductile-brittle detachments (figure 7) [Hetzel *et al.*, 1995b; Gessner *et al.*, 2001b; Lips *et al.*, 2001; Ring *et al.*, 2003; Thomson & Ring, 2006; van Hinsbergen *et al.*, 2010b]. Additional syn-tectonic sedimentary basins were emplaced in the hanging walls of these structures [Sen and Seyitoğlu, 2009]. The final, Pliocene-Quaternary evolution of the Menderes massif was dominated by the N-S opening of several grabens controlled by high-angle

normal faults (figures 4 and 7) [Çiftçi and Bozkurt, 2009; Sen and Seyitoğlu, 2009]. Additional distributed strike-slip tectonics has been recorded in the western part of the Anatolide-Tauride block since the middle Miocene, mainly along the NE-SW Izmir-Balıkesir Transfer Zone [Ersoy *et al.*, 2011].

During its Cenozoic evolution, the Anatolide-Tauride block recorded several large-scale block rotations. While no significant rotation was recorded north of the Menderes massif, the central Menderes massif underwent 25-30° counterclockwise rotation during its exhumation (i.e. 15-5 Ma ago), concomitantly with the ~20° counterclockwise rotation recorded in the Lycian nappes and the Bey Dağları platform (figure 3) [Kissel *et al.*, 1993; van Hinsbergen *et al.*, 2010a, 2010b]. Further east, the central Taurides experienced an ~40° clockwise rotation between the middle Eocene-Oligocene (figure 3) [Kissel *et al.*, 1993] that also affects the Tavşanlı and Afyon units further north to a lesser degree [Pourceau *et al.*, 2010]. Finally, an ~40° counterclockwise rotation occurred in Eocene volcanic rocks in the eastern Taurides (figure 3) [Kissel *et al.*, 2003].

3.2. *Magmatic features*

The geodynamic evolution of the eastern Mediterranean region was accompanied by the emplacement of several magmatic provinces. The space and time distribution of magmatic centers and the evolution of their geochemical composition can be used to constrain the origin of magmas and associated geodynamic processes. In this section, we compile information on the nature and origin of the different magmatic provinces emplaced in this region since the late Cretaceous. Considering the huge amount of available data, we do not go into detail here about the geochemical and isotopic signatures of these magmas but refer instead to the appropriate studies.

3.2.1. *The Balkans-Pontides magmatic province*

During the late Cretaceous-early Paleocene, a magmatic province developed in the Balkans and the eastern Pontides where both plutonic and volcanic centers have been identified [Yılmaz *et al.*, 1997;

Berza et al., 1998; *Ciobanu et al.*, 2002]. In the Pontides, predominant volcanic rocks were mostly emplaced between the late Turonian and latest Campanian (~90-70 Ma) (figure 5) and seem to display southward migration through time [*Yılmaz et al.*, 1997; *Bektaş et al.*, 1999; *Okay et al.*, 2001b; *Eyüboğlu et al.*, 2010], similar to magmatic products in the Balkans [*Ciobanu et al.*, 2002; *von Quadt et al.*, 2005; *Kolb et al.*, 2012].

These rocks mainly display a medium- to high-K calc-alkaline magmatic trend with intermediate composition [*Yılmaz et al.*, 1997; *Berza et al.*, 1998; *von Quadt et al.*, 2005; *Boztuğ et al.*, 2006]. The geochemical and isotopic signatures of these magmas indicate a typical arc-related origin with an increasing mantle source component through time [*von Quadt et al.*, 2005]. An additional shoshonitic (i.e. high-K alkaline) magmatic trend has been recognized in the eastern Srednogorie region in the latest Cretaceous [*Boccaletti et al.*, 1978]. Finally, mafic alkaline volcanic rocks were emplaced in the eastern Srednogorie, Banat and Timok regions up to Eocene times, resulting from a small degree of melting of a depleted mantle source (figure 5) [*Cvetković et al.*, 2004].

3.2.2. *The CACC plutonic province*

In the CACC (figure 1), the late Cretaceous exhumation and cooling of HT metamorphic domes were accompanied by the emplacement of numerous intrusions in three successive stages [*Aydin et al.*, 1998; *Ilbeyli et al.*, 2004; *Boztuğ et al.*, 2009]. First, rare peraluminous two-mica leucogranites displaying a high-K calc-alkaline magmatic trend were emplaced as a result of the partial melting of crustal rocks [*Aydin et al.*, 1998]. Then, metaluminous high-K calc-alkaline intrusions developed, evolving from quartz monzonite to monzogranite, with geochemical and isotopic signatures indicating a subduction-related metasomatized mantle source [*Ilbeyli et al.*, 2004]. Finally, a small number of shoshonitic then silica-undersaturated alkaline intrusions were emplaced up to the beginning of the Maastrichtian, with no clearly constrained magmatic sources (figure 5) [*Aydin et al.*, 1998; *Ilbeyli et al.*, 2004].

3.2.3. *The Eocene calc-alkaline magmatic province*

During the Eocene, an E-W-trending magmatic province developed extending from the eastern Pontides to the Chalkidiki peninsula, south of the Rhodope massif (figure 5). This province is mainly represented by volcano-sedimentary rocks in the east [Arslan *et al.*, 2013] and plutonic bodies in the west [Harris *et al.*, 1994; Okay and Satır, 2006].

These rocks mainly follow a medium- to high-K calc-alkaline magmatic trend with both low-silica and high-silica end-members and geochemical and isotopic signatures typical of a subduction-related origin [Harris *et al.*, 1994; Ersoy and Palmer, 2013].

3.2.4. *The Rhodope magmatic province*

From the late Eocene to the Miocene, widespread magmatic activity developed in the Rhodope massif during the last stages of exhumation of the central and south Rhodope core complexes [Jones *et al.*, 1992; Marchev *et al.*, 2004b, 2005; Ersoy and Palmer, 2013]. Migmatites and associated leucogranitic veins have been observed in the cores of several metamorphic domes constituting the central Rhodope core complex [Burg *et al.*, 1996; Marchev *et al.*, 2004b]. Emplaced as soon as the late Eocene (figure 5), volcanic and plutonic rocks typically crosscut detachment faults [Marchev *et al.*, 2005], with some intrusions displaying a strong marginal foliation associated with an extension-related top-to-the N sense of shear [Jones *et al.*, 1992]. The composition of most of these magmas evolves from high-K calc-alkaline to shoshonitic, including both mafic and silicic rocks [Marchev *et al.*, 2004b; Ersoy and Palmer, 2013]. Their geochemical and isotopic signatures reflect both subduction-related metasomatized mantle and crustal source components, the latter decreasing from northwest to southeast [Marchev *et al.*, 2004b, 2005; Ersoy and Palmer, 2013]. The end of magmatic activity in the central Rhodope core complex is marked by the emplacement of late Oligocene alkaline dykes originating from an asthenospheric source contaminated by a depleted lithospheric mantle [Marchev *et al.*, 2004b].

In the south Rhodope core complex, migmatites are also observed [Brun and Sokoutis, 2007]. In addition, syn-tectonic intrusions were emplaced in the early Miocene (figure 5) with a similar composition and origin as the central Rhodope magmatic centers and displaying an extensional mylonitic fabric with a top-to-the SW sense of shear [Dinter and Royden, 1993; Dinter *et al.*, 1995; Ersoy and Palmer, 2013].

3.2.5. The west Anatolian-Aegean magmatic province

In the Biga peninsula (northwest Anatolia, figure 7) and the nearby north Aegean islands, volcanic activity developed since the Oligocene then migrated southward, reaching the Menderes massif in the Miocene [Yilmaz *et al.*, 2001; Dilek and Altunkaynak, 2009]. At the same time, several intrusions were emplaced during the exhumation of MCCs, such as in the Kazdağ massif in the late Oligocene-early Miocene [Black *et al.*, 2013] and in the Menderes massif and the Cyclades in the Miocene (figure 5) [Faure and Bonneau, 1988; Altherr and Siebel, 2002; Pe-Piper and Piper, 2006, 2007; Dilek *et al.*, 2009; Iglseder *et al.*, 2009; Denèle *et al.*, 2011; Laurent *et al.*, 2015; Rabillard *et al.*, 2015]. Migmatites associated with S-type granites have also been described in the cores of these metamorphic domes [Jansen and Schuiling, 1976; Lister *et al.*, 1984; Okay and Satır, 2000b; Vanderhaeghe, 2004; Ring, 2007; Bozkurt *et al.*, 2010; Kruckenberg *et al.*, 2011; Laurent *et al.*, 2015; Beaudoin *et al.*, 2015]. These intrusions are exposed in the footwall of major detachments where they show strong foliation with extensional kinematic indicators as well as cataclastic zones at the contact with the detachments [Faure and Bonneau, 1988; Gessner *et al.*, 2001a; Dilek *et al.*, 2009; Iglseder *et al.*, 2009; Denèle *et al.*, 2011; Rabillard *et al.*, 2015]. They were emplaced first in the northeastern Menderes massif with the Egrigöz and Koyunoba plutons both dated at ~21 Ma (U-Pb crystallization ages on zircon) in the footwall of the Simav detachment [Ring and Collins, 2005]. Then, a few intrusions developed below the Alaşehir detachment in the central Menderes massif, such as the ~15 Ma Salihli pluton (U-Pb crystallization ages on zircon), although these ages could be affected by dissolution-precipitation reactions [Glodny and Hetzel, 2007; Catlos *et al.*, 2010]. In

the Cyclades, finally, the intrusions were emplaced and exhumed in the footwall of the NCDS (in Ikaria, Mykonos and Tinos), the NPEFS (in Naxos) and the WCDS (in Serifos and Lavrion), with ages decreasing from east to west, from ~15 Ma in Ikaria (U-Pb crystallization ages on zircon) [Bolhar *et al.*, 2010] to ~8 Ma in Lavrion (K-Ar crystallization ages on biotite) [Altherr *et al.*, 1982].

Geochemical data for these volcanic and plutonic rocks indicate a composition evolving from high-K calc-alkaline to shoshonitic [Ersoy and Palmer, 2013; Seghedi *et al.*, 2013]. In the Menderes massif and the Cyclades, most of the Miocene intrusions are I-type granites or granodiorites [Altherr *et al.*, 1982; Altherr and Siebel, 2002; Dilek *et al.*, 2009; Stouraiti *et al.*, 2010] although some S-type granites and pegmatite dikes associated with migmatites are also observed in the central Cyclades [Vanderhaeghe, 2004; Ring, 2007; Kruckenberg *et al.*, 2011; Beaudoin *et al.*, 2015; Laurent *et al.*, 2015]. As for the Rhodope massif, Ersoy and Palmer [2013] suggest a subduction-related metasomatized mantle source component for these magmas, although the sole contribution of crustal sources has also been proposed [Altherr and Siebel, 2002; Stouraiti *et al.*, 2010]. Another interesting observation is the decrease through time, from the northeast (in Ikaria) to the southwest (in Serifos), of the crustal source component of the Cycladic intrusions with respect to their subduction-related metasomatized mantle source component [Jolivet *et al.*, 2015].

Additional scattered alkaline volcanism, represented by alkali basalts, occurred from the late Miocene to the Quaternary in an area extending from the Thrace basin in the north to the Patmos island and the Isparta region in the south (figure 5) [Agostini *et al.*, 2007; Ersoy and Palmer, 2013]. Their geochemical and isotopic composition involves a small degree of melting of a depleted mantle source [Agostini *et al.*, 2007].

Finally, active volcanism has occurred since the Pliocene along a curved narrow arc, such as in the islands of Aegina, Milos, Santorini and Nisyros (figure 5). This typical arc-related volcanism mostly has a medium-K calc-alkaline composition [Fytikas *et al.*, 1984; Pe-Piper and Piper, 2005; Ersoy and Palmer, 2013].

3.2.6. *The late Miocene-Quaternary eastern Anatolia volcanic province*

In eastern Anatolia, late Miocene-Quaternary volcanism developed progressively southward from the eastern Pontides to the Arabian foreland basin (figure 5) [Pearce *et al.*, 1990; Keskin, 2003, 2007]. The magmatic trend of these volcanics evolves from high-K calc-alkaline in the north to alkaline in the south with a composition ranging from basalt to rhyolite [Keskin, 2003]. The evolution of the geochemical and isotopic composition of these magmas suggests a decrease of their subduction-related metasomatized mantle source component from north to south, correlated with an increase of a depleted mantle source component [Keskin, 2003, 2007].

4. **Kinematic reconstructions of the eastern Mediterranean region**

Using all the sedimentary, tectono-metamorphic and magmatic data described above, our kinematic reconstructions of the eastern Mediterranean region were constrained from the present to the past. However, for clarity, this tectonic evolution is described in the following section from the late Cretaceous to the present day, following the geodynamic history of the region. The full set of paleotectonic maps is shown below (figures 8a-y), as well as the reconstructed lithospheric-scale cross-sections (figures 9 and 10). These kinematic reconstructions are also available as a movie (see Supplementary Materials, movie S1). In these reconstructions, the evolution of metamorphic units is shown from their time of formation to their exhumation at the surface with lighter or darker colors when these units are still at depth or close to the surface, respectively.

4.1. *Late Cretaceous (100-65 Ma, figures 8a-h)*

4.1.1. *A linear northward subduction zone along the Eurasian margin*

In the late Cretaceous, the Balkans, the Rhodope massif and the Pontides already belonged to the Eurasian margin. All these domains formed an E-W linear orogenic belt over 2000 km in length

when the early Miocene $\sim 60^\circ$ clockwise rotation of the Apuseni mountains is back rotated (figure 3) [Pătraşcu *et al.*, 1994]. Along this belt, the HP-LT metamorphic rocks exhumed as a syn-orogenic extrusion wedge [Okay, 1986; Liati *et al.*, 2002; Okay *et al.*, 2006; Beccaletto *et al.*, 2007] and the emplacement of wide calc-alkaline magmatic province [Yılmaz *et al.*, 1997; Berza *et al.*, 1998; Okay *et al.*, 2001b] indicate an active northward subduction zone below Eurasia accommodating the Africa-Eurasia convergence. The downgoing lithosphere consisted of a wide oceanic domain called the Vardar ocean along the Balkan and Rhodope margin and the Izmir-Ankara-Erzincan ocean along the Pontide margin (figures 9 and 10). Within the upper plate, a thrust belt developed progressively southward in the Rhodope massif and the Pontides [Burg *et al.*, 1990; Yılmaz *et al.*, 1997]. Behind this compressional front, transtensional to extensional tectonics allowed the opening of sedimentary basins [Willingshofer *et al.*, 1999; Sunal and Tüysüz, 2002]. At the same time, the magmatic activity seems to migrate southward [Yılmaz *et al.*, 1997; Ciobanu *et al.*, 2002; von Quadt *et al.*, 2005; Eyüboğlu *et al.*, 2010] suggesting a southward retreat of the oceanic slab [von Quadt *et al.*, 2005; Kolb *et al.*, 2012]. Further north, the opening of the Black Sea basin was associated with an intense stretching of the crust and oceanic spreading in Cenomanian to Santonian times [Zonenshain and Le Pichon, 1986; Görür, 1988; Hippolyte *et al.*, 2010; Nikishin *et al.*, 2015b]. An additional shoshonitic and alkaline volcanism was simultaneously emplaced in the eastern Srednogie region, just west of this extensional domain (figures 8g-h) [Boccaletti *et al.*, 1978].

4.1.2. Obduction and subduction involving Africa-derived continental blocks

Also governed by the Africa-Eurasia convergence, several continental blocks previously rifted from Africa (i.e. the CACC and Apulia including the Pelagonian, Gavrovo-Tripolitza, Ionian and Tauride carbonate platforms) moved northward during the late Cretaceous. Several obduction events started in the Turonian-Coniacian (95-85 Ma) probably initiated as intraoceanic subductions (figures 8b-d). A part of the Ankara-Erzincan oceanic domain then overthrust the CACC in a southward direction forming the central Anatolian ophiolites [Yaliniz and Göncüoğlu, 1998]. Further south, a

part of the Inner-Tauride oceanic domain was obducted over the Tauride and the eastern Pelagonian platforms forming the Lycian and Tauride ophiolites [Şengör and Yilmaz, 1981; Katzir *et al.*, 1996; Dilek *et al.*, 1999]. A similar process also occurred over the southern Tauride margin with the northward overthrusting of the Antalya ophiolites belonging to the Mesogean oceanic domain [Robertson, 2002].

As a result of this obduction event, the buried northern Tauride margin underwent HP-LT metamorphism forming the Tavşanlı then the Afyon-Ören units (figure 10) [Okay, 1986, 2002; Candan *et al.*, 2005; Plunder *et al.*, 2013; Pourteau *et al.*, 2013]. At the same time, HT-MP metamorphism was recorded in the overriding plate (i.e. the CACC), where several MCCs were subsequently exhumed in an extensional context [Whitney *et al.*, 2003; Lefebvre *et al.*, 2011]. Several high-K calc-alkaline, then shoshonitic and finally alkaline intrusions were emplaced in this context [Aydin *et al.*, 1998; Boztuğ, 2000; Boztuğ *et al.*, 2009].

4.2. *Paleocene-Eocene (65-35 Ma, figures 8h-n)*

4.2.1. *Continental accretion along the Eurasian margin*

As early as the latest Cretaceous, an inversion of extensional structures occurred throughout the Balkans and the Pontides [Yilmaz *et al.*, 1997; Sunal and Tüysüz, 2002; Bergerat *et al.*, 2010]. The eastern Pontides then underwent a rapid uplift in the Paleocene-early Eocene [Boztuğ *et al.*, 2004]. These features, together with the cessation of magmatic activity along this orogenic belt, indicate a major change in subduction dynamics, probably due to the collision of the Pelagonian platform and the CACC with Eurasia (figures 8h-i and 10). The indentation of the CACC along the Izmir-Ankara oceanic suture zone induced the oroclinal bending of the central Pontides [Kaymakci *et al.*, 2003, 2009; Meijers *et al.*, 2010] as well as the tectonic reorganization of the CACC [Lefebvre *et al.*, 2013]. In the late Paleocene-early Eocene, a second collisional event between the Eurasian-CACC margin and the Anatolide-Tauride block resulted in the closure of the Inner-Tauride and the Izmir-Ankara oceanic domains as well as the marine corridor separating the Pelagonian and Tauride

platforms (figures 8i-j). In this area, the latest Cretaceous-early Paleocene sedimentary and ophiolitic mélangé was then folded and sheared toward the south, forming the Bornova flysch zone [Robertson *et al.*, 2009; Okay *et al.*, 2012]. Another consequence of this second collisional event was the development of a north-vergent fold-and-thrust-belt along the eastern Balkans and the Pontides since the middle-late Eocene (figures 9 and 10) [Doglioni *et al.*, 1996; Yılmaz *et al.*, 1997]. Further south, a northward subduction zone initiated along the southern Tauride margin inducing the building of the Bitlis belt and the progressive closure of the Mesogean oceanic domain [Dercourt *et al.*, 1986; Barrier and Vrielynck, 2008].

4.2.2. *A continuous subduction migrating southward*

Immediately after the collision of the Pelagonian platform with Eurasia, the main thrust front shifted to the southern margin of the Pelagonian platform, inducing the subduction of the partly oceanic Pindos basin (figures 8i-j and 9) [Aubouin, 1959]. The latter was partly accreted to the Pelagonian nappe contributing to the building of the Hellenides, while its buried part underwent early Eocene HP-LT metamorphism recorded in the Cycladic Blueschists unit and the Cycladic basement [Bonneau and Kienast, 1982; Bonneau, 1984; Wijbrans and McDougall, 1988; Tomaschek *et al.*, 2003; Lagos *et al.*, 2007]. These units were then progressively exhumed up to the end of the Eocene in the Cyclades and in the Olympos, Ossa and Pelion tectonic windows as a syn-orogenic extrusion wedge bounded by the Vari detachment on top (in Syros) and the SCSZ at the base (in Ios) (figure 9) [Jolivet *et al.*, 2003; Ring *et al.*, 2007, 2010; Brun and Faccenna, 2008; Huet *et al.*, 2009; Jolivet and Brun, 2010].

Across the Tauride platform, the main thrust front migrated similarly southward carrying the remnants of the late Cretaceous Lycian and Tauride ophiolites [de Graciansky, 1967; Gutnic *et al.*, 1979; Collins and Robertson, 1998]. The late Cretaceous HP-LT Tavşanlı and Afyon-Ören units were then exhumed along a cold retrograde path reaching the surface in the early Eocene [Özcan *et al.*, 1988; Sherlock *et al.*, 1999; Rimmelé *et al.*, 2003a; Ring and Layer, 2003; Pourteau *et al.*, 2010,

2013]. Below these units, the major part of the buried Menderes massif underwent MT- to HT-MP metamorphism (i.e. the *Main Menderes Metamorphism*), except in its southern part where HP-LT metamorphism was recorded (i.e. the Kurudere-Nebiler unit, figures 8l-n and 10) [Şengör *et al.*, 1984; Bozkurt and Oberhänsli, 2001; Okay, 2001; Rimmelé *et al.*, 2003b, 2005; Whitney *et al.*, 2008; Pourteau *et al.*, 2013].

During the Eocene, while the subduction zone migrated southward across the accreted continental blocks following the retreat of the decoupled part of the downgoing lithosphere (figure 10), an E-W-trending magmatic province developed mainly displaying medium- to high-K calc-alkaline composition (figures 8j-n) [Harris *et al.*, 1994; Okay and Satır, 2006; Arslan *et al.*, 2013].

4.2.3. *Back-arc extension in the Rhodope massif*

Concomitantly with the southward retreat of the subduction zone, the lower part of the thickened Rhodope crust was strongly heated allowing the development of Paleocene MT-MP metamorphism associated with migmatites (figure 9) [Burg *et al.*, 1996; Mposkos, 1998; Jolivet and Brun, 2010]. In the central Rhodope massif, these metamorphic rocks were exhumed as several MCCs below top-to-the NE ductile-brittle detachments and reached the surface in the late Eocene (figures 8j-n) [Burg *et al.*, 1996; Burchfiel *et al.*, 2003; Bonev *et al.*, 2006]. This exhumation was accompanied by an $\sim 11^\circ$ clockwise rotation of the massif (figure 3) [Brun and Sokoutis, 2007]. At the same time, the Thrace basin started to open and was progressively filled by a thick sedimentary sequence [Turgut and Eseller, 2000; Okay *et al.*, 2010a]. At the end of the Eocene, this extension propagated southward within the southern Rhodope massif [Brun and Sokoutis, 2007, 2010; Wüthrich, 2009].

4.3. *Oligocene-early Miocene (35-15 Ma, figures 8n-r)*

4.3.1. *Continuing southward migration of the subduction zone*

In the eastern Mediterranean region, contemporaneously with the last indications of compression in the eastern Balkans in the Oligocene [Doglioni *et al.*, 1996] and the early Miocene $\sim 60^\circ$ clockwise

rotation of the western Balkans (figure 3) [Pătraşcu *et al.*, 1994], the main thrust front propagated southward, especially across the Tauride platform with the Lycian nappes reaching the Bey Dağları platform in the early Miocene (figure 10) [Poisson, 1977; van Hinsbergen *et al.*, 2010a]. Further west, the Gavrovo-Tripolitza and Ionian platforms started to subduct at the same time after the closure of the Pindos basin (figure 9) [Sotiropoulos *et al.*, 2003; van Hinsbergen *et al.*, 2005b]. The external parts of these carbonate platforms were accreted to the Hellenides, remaining at low pressure while their buried internal parts, comprising their basements, recorded HP-LT metamorphism forming the Phyllite-Quartzite and the Plattenkalk units in Crete and the Peloponnese [Bonneau and Kienast, 1982; Seidel *et al.*, 1982; Theye *et al.*, 1992; Trotet *et al.*, 2006; Jolivet *et al.*, 2010b]. These metamorphic units were then exhumed in the footwall of the top-to-the N Cretan detachment since ~25 Ma, following a cold retrograde path [Jolivet *et al.*, 1996, 2010c; Ring *et al.*, 2001]. Similarly to the Cycladic Blueschists unit, the exhumation of the Phyllite-Quartzite and the Plattenkalk units was associated with N-S to NE-SW extension as in Crete (figure 3) [Jolivet *et al.*, 1996, 2010c]. The E-W stretching lineation in the Peloponnese resulted from the subsequent middle-late Miocene clockwise rotation (see details in Section 4.4.2).

4.3.2. Back-arc extension in the Cyclades and western Anatolia

Behind this compressional front, back-arc extension propagated southward during the Oligocene-Miocene [Le Pichon and Angelier, 1979; Jolivet and Faccenna, 2000] leading to intense crustal thinning observed now in the Aegean-west Anatolian region [Tirel *et al.*, 2004; Karabulut *et al.*, 2013]. Thus, while the exhumation of the central Rhodope core complex ceased in the Oligocene [Marchev *et al.*, 2004a; Wüthrich, 2009], the southern Rhodope massif continued to be exhumed as MCC below the top-to-the SW Kerdylion detachment (figures 9 and 11) [Brun and Sokoutis, 2007]. Coeval with this exhumation, an ~30° clockwise rotation affected the hanging wall of the detachment (figure 3) [Kondopoulou and Westphal, 1986; Brun and Sokoutis, 2007].

Further south, Oligocene-early Miocene exhumation of MCCs mainly in greenschist-facies conditions occurred in the Cyclades and the Menderes massif [Altherr *et al.*, 1982; Lister *et al.*, 1984; Gautier and Brun, 1994; Jolivet *et al.*, 1994; Bozkurt and Oberhänsli, 2001; Vanderhaeghe and Teyssier, 2001; Ring *et al.*, 2003, 2010; Catlos and Çemen, 2005; Jolivet & Brun, 2010], associated with additional HT-MP metamorphism affecting the northern Menderes massif and the central Cyclades in the Oligocene and early Miocene, respectively (figures 8n-r and 11) [Jansen and Schuiling, 1976; Lister *et al.*, 1984; Gautier *et al.*, 1993; Keay *et al.*, 2001; Catlos and Çemen, 2005; Duchêne *et al.*, 2006; Bozkurt *et al.*, 2010; Beaudoin *et al.*, 2015]. In the Cyclades, the NCDS accommodated this exhumation since the Oligocene (figures 9 and 11) [Faure *et al.*, 1991; Lee and Lister, 1992; Brichau *et al.*, 2007, 2008; Jolivet *et al.*, 2010a] while the NPEFS and the WCDS became active in the early Miocene [Urai *et al.*, 1990; Gautier *et al.*, 1993; Vanderhaeghe, 2004; Brichau *et al.*, 2006; Seward *et al.*, 2009; Iglseder *et al.*, 2011; Grasemann *et al.*, 2012]. Syn-tectonic sediments started to be deposited in the hanging wall of these extensional structures [Photiades, 2002; Sánchez-Gómez *et al.*, 2002; Kuhlemann *et al.*, 2004; Lecomte *et al.*, 2010; Laurent *et al.*, 2015]. Further east, the exhumation of the Menderes massif was accommodated in the north by the top-to-the N Simav detachment recording at least ~50 km of displacement [Işık and Tekeli, 2001; Ring *et al.*, 2003; Bozkurt *et al.*, 2011] and in the south by a top-to-the S shearing just below the overlying Lycian nappes (figures 10 and 11) [Bozkurt and Satır, 2000; Gessner *et al.*, 2001a; Ring *et al.*, 2003]. Volcanoclastic rocks unconformably covering the southern Menderes massif indicate surface exposure of these metamorphic rocks in the early Miocene [Bozkurt and Satır, 2000]. Finally, the exhumation of the Kazdağ massif started in the late Oligocene in HT-MP conditions and was accommodated by the top-to-the N Alakeçi and top-to-the S Şelale detachments up to the early Miocene (figure 11) [Okay and Satır, 2000b; Beccaletto and Steiner, 2005; Bonev and Beccaletto, 2007].

During the post-orogenic exhumation of these MCCs, high-K to shoshonitic magmatism developed and migrated southward. Thus, magmatic centers were first emplaced as early as the latest Eocene

in the central Rhodope massif (figures 9 and 11), then during the Oligocene in the Biga Peninsula and finally in the early Miocene in the south Rhodope and the northern Menderes massifs [Jones *et al.*, 1992; Altherr and Siebel, 2002; Marchev *et al.*, 2004b, 2005; Pe-Piper and Piper, 2006, 2007; Dilek and Altunkaynak, 2009; Dilek *et al.*, 2009; Ersoy and Palmer, 2013].

4.3.3. *Eurasia-Arabia collision*

Although its timing is debated, the continental subduction and subsequent collision of Arabia with the Eurasian margin seem to have occurred in the Oligocene-early Miocene, resulting in the building of the Bitlis-Zagros belt [Hempton, 1987; Jolivet and Faccenna, 2000; Agard *et al.*, 2005; Allen and Armstrong, 2008; Okay *et al.*, 2010b; McQuarrie and van Hinsbergen, 2013]. Such lithospheric-scale processes could also have induced major tectonic changes within the Anatolide-Tauride block. Thus, Pourteau *et al.* [2010] proposed that the post-Eocene $\sim 40^\circ$ counterclockwise rotation of the eastern Tauride platform [Kissel *et al.*, 2003] and the middle Eocene-Oligocene $\sim 40^\circ$ clockwise rotation of the central Tauride platform (figure 3) [Kissel *et al.*, 1993] resulted from the subduction/collision of the Arabian plate along the Eurasian margin. Rotation of these two rigid blocks could have been accommodated by the central Anatolian fault (figures 8n-r) [Koçyiğit and Beyhan, 1998; Jaffey and Robertson, 2001] as attested by the new, Oligocene burial of the Niğde massif in the southern CACC, attributed to the activation (or reactivation) of this transpressional structure [Idleman *et al.*, 2014]. Nevertheless, the timing of these block rotations, especially for the eastern Tauride platform, remains poorly constrained and Pourteau *et al.* [2010] do not exclude the possibility that the preceding Eocene collision of the Anatolide-Tauride block with the CACC further north could also have induced such block rotations.

4.4. *Middle-late Miocene (15-5 Ma, figures 8r-w)*

4.4.1. *Continuing syn-orogenic and post-orogenic exhumation*

In the middle-late Miocene, the main thrust front continued its southward migration resulting in the accretion of new continental terranes. The Paxos carbonate platform underthrust the previously accreted Ionian nappe in the western Hellenides [van Hinsbergen *et al.*, 2005b] while the Mesogean oceanic lithosphere started to subduct in the eastern Hellenides concomitantly with the growth of the Mediterranean ridge [Le Pichon *et al.*, 1982, 2002; Chaumillon *et al.*, 1996; Kopf *et al.*, 2003]. Further east, the non-metamorphosed Lycian nappes overthrust the Bey Dağları platform [Poisson, 1977; van Hinsbergen *et al.*, 2010a]. Within the accretionary wedge, the syn-orogenic exhumation of the HP-LT Phyllite-Quartzite and the Plattenkalk units continued up to 9 Ma in the footwall of the Cretan detachment which accommodated more than 100 km of northward displacement in Crete (figures 3 and 11). The metamorphic footwall was exposed at the surface 11-10 Ma ago at the earliest, while sediments were deposited in the hanging wall as soon as the lower Miocene [Jolivet *et al.*, 1996, 2010c; Ring *et al.*, 2001; van Hinsbergen and Meulenkamp, 2006; Seidel *et al.*, 2007; Marsellos *et al.*, 2010]. After 9 Ma, the activity of the Cretan detachment ceased and high-angle normal faults formed, inducing the uplift of Crete and the opening of the Cretan Sea (figure 11) [Angelier *et al.*, 1982; van Hinsbergen and Meulenkamp, 2006].

Behind this compressional belt, back-arc extension was still active. High-angle normal faults and associated sedimentary basins developed over the south Rhodope core complex which continued to be exhumed below the Kerdyllion detachment (figures 9 and 11) [Brun and Sokoutis, 2007]. In the Cyclades, the exhumation of MCCs ceased ~9-8 Ma ago below the NCDS and the NPEFS and ~6 Ma ago below the WCDS (figures 9 and 11) [Brichau *et al.*, 2006, 2008, 2010; Jolivet *et al.*, 2010a]. Surface exposure of the metamorphic rocks then occurred in the late Miocene (figure 11) [Sánchez-Gómez *et al.*, 2002]. Further east, displacement along the Simav detachment continued until ~8 Ma [Bozkurt *et al.*, 2011] and was contemporaneous with the exhumation of the central part of the Menderes massif below the top-to-the NE Alaşehir and the top-to-the S Büyük detachments, which occurred up to the end of the Miocene (figures 10 and 11) [Hetzel *et al.*, 1995a, 1995b;

Gessner et al., 2001b; *Lips et al.*, 2001; *Ring et al.*, 2003; *Thomson & Ring*, 2006; *van Hinsbergen et al.*, 2010b].

At the same time, numerous syn-tectonic intrusive bodies were emplaced in the footwalls of detachment systems in the central Menderes massif and the Cyclades [*Faure and Bonneau*, 1988; *Gessner et al.*, 2001a; *Dilek et al.*, 2009; *Iglseder et al.*, 2009; *Lecomte et al.*, 2010; *Denèle et al.*, 2011; *Laurent et al.*, 2015; *Rabillard et al.*, 2015].

4.4.2. Large-scale block rotations

Although the southward trench migration and exhumation processes remained similar from the Oligocene to the late Miocene, additional large-scale clockwise and counterclockwise block rotations occurred since 15 Ma both in the arc and back-arc domains (figures 3 and 11). Most of the $\sim 30^\circ$ clockwise rotation of the south Rhodope core complex took place after 15 Ma [*Brun and Sokoutis*, 2007; *van Hinsbergen et al.*, 2008]. Similarly, the major part of the Hellenides and the western Cyclades underwent, respectively, an $\sim 40^\circ$ and $\sim 20^\circ$ clockwise rotation between 15 and 8 Ma [*Kissel and Laj*, 1988; *Morris and Anderson*, 1996; *van Hinsbergen et al.*, 2005a]. In addition, in order to fit the NE-SW to E-W stretching lineation of the Phyllite-Quartzite unit in the Peloponnese (figure 3) with the regional N-S to NNE-SSW extension, a similar clockwise rotation has to be applied to this HP-LT unit during its final exhumation in the Peloponnese. In opposition, $20\text{-}30^\circ$ counterclockwise rotations have been recorded further east in the eastern Cyclades, the central Menderes, the Lycian nappes and the Bey Dağları platform (figure 3) [*Kissel et al.*, 1993; *van Hinsbergen et al.*, 2010a, 2010b]. These opposite rotations are related to the increasing curvature of the trench during the middle and late Miocene (figure 11) with no rotation recorded at the hinge, such as in Crete where the strike of the stretching lineation of the exhumed Phyllite-Quartzite and the Plattenkalk units fits with the regional N-S to NNE-SSW extension (figure 3).

4.4.3. Continuing collision and extrusion of Anatolia

In the middle-late Miocene, the ongoing collision of Arabia with Eurasia along the Bitlis belt (figures 8r-w) is associated with the uplift of eastern Anatolia [Dewey *et al.*, 1986; Şengör *et al.*, 2003; Okay *et al.*, 2010b] and the development of a wide magmatic province [Pearce *et al.*, 1990; Keskin, 2003, 2007]. About 12 Ma ago, the whole Anatolide-Tauride block started to move westward, mostly accommodated by the dextral strike-slip NAF that propagated throughout the Pontides and reached the Dardanelle Strait 6-5 Ma ago (figures 8t-w) [Armijo *et al.*, 1999; Şengör *et al.*, 2005; Melinte-Dobrinescu *et al.*, 2009]. However, as early as the late Miocene, distributed strike-slip then E-W compressive tectonics developed, probably as a result of this westward extrusion notably in western Anatolia along the Izmir-Balıkesir Transfer Zone [Ersoy *et al.*, 2011] and in the Cyclades with the folding and local reactivation of detachments with a reverse movement [Avigad *et al.*, 2001; Menant *et al.*, 2013].

4.5. *Pliocene-Quaternary (5-0 Ma, figures 8w-y)*

4.5.1. *Active collision and slab roll-back*

From the Pliocene to the present day, while Arabia is still colliding with Eurasia along the Bitlis belt in the east, the southward retreat of the Mesogean slab continues in front of the Aegean domain at the same time as the growth of the Mediterranean ridge that finally reaches the northern African margin (figures 8y and 9). Further west, the Paxos carbonate platform continues to accrete to the Hellenides. At the junction between the slowly subducting Paxos continental platform and the fast retreating Mesogean oceanic lithosphere, the dextral strike-slip Kephallonia fault has developed since ~5 Ma, inducing a segmentation of the main thrust front (figures 8w-y) [Finetti, 1982; Kreemer and Chamot-Rooke, 2004; Royden and Papanikolaou, 2011]. Behind this subduction front, a medium-K calc-alkaline volcanic arc has been active in the south Aegean domain (figures 8x-y) [Pe-Piper and Piper, 2005].

4.5.2. *Extrusion and localization of extension*

Since 6-5 Ma, the NAF has entered the Aegean domain, accommodating ~80 km of dextral strike-slip displacement in the Dardanelle Strait (figure 3) [Armijo *et al.*, 1999; Şengör *et al.*, 2005; Lacassin *et al.*, 2007; Le Pichon and Kreemer, 2010; Akbayram *et al.*, 2015]. At the same time, extension ceased in the Cyclades and has localized at the edges of the Aegean domain, where high-angle normal faulting has formed several grabens and half-grabens (figures 8w-y), such as in the Evvia island and the Corinth region (forming the Central Hellenic Shear Zone [Papanikolaou and Royden, 2007]), the Peloponnese, the Cretan Sea and the Menderes massif [Armijo *et al.*, 1996; Royden and Papanikolaou, 2011; Pérouse *et al.*, 2012]. This extensional tectonic regime has been accompanied in the Hellenides by a clockwise rotation of ~10° and ~23° north and south of the Central Hellenic Shear Zone [van Hinsbergen *et al.*, 2005a; Bradley *et al.*, 2013], increasing the curvature of the trench. This feature suggests a tectonic decoupling between these two rotating blocks that can be accommodated by the dextral shear distributed across the Central Hellenic Shear Zone [Royden and Papanikolaou, 2011].

5. Discussion

These new kinematic reconstructions provide many details about the late Cretaceous-Cenozoic geodynamic evolution of the eastern Mediterranean region that have not been considered in previous larger-scale reconstruction models [Dercourt *et al.*, 1986, 1993; Ricou, 1994; Jolivet *et al.*, 2003; Barrier and Vrielynck, 2008; Stampfli and Hochard, 2009]. The geodynamic scenario described above is subject to debate, especially for the late Cretaceous period, where the lack of geological constraints does not allow us to propose a tectonic evolution as accurately as for the Cenozoic period. We also acknowledge that different kinematic solutions could have been considered because other sources of error could arise from uncertainties in the exact amount of displacement on large-scale structures, in the timing of metamorphic events and in the inherent assumptions made in the definition of deforming domains in our reconstructions.

Extracting from our reconstructions, the motion paths of different points distributed over the whole region provide insights about past kinematics and, therefore, instantaneous surface deformation of this region (figures 12 and 13). This kinematic model allows for discussion of the different crustal and mantle processes driving the progressive back-arc opening and the present-day kinematics of the eastern Mediterranean region. The space and time distribution of the magmatic activity illustrated in our kinematic reconstructions (figures 8a-y) is also discussed in terms of crustal and mantle interactions.

5.1. Late Cretaceous and Oligocene to Present slab roll-back and back-arc opening

During the late Cretaceous-Cenozoic, two main episodes of southward trench retreat and related back-arc opening have been highlighted: the first during the late Cretaceous along the Balkans and the Pontides (figures 8a-h) and the second since the Oligocene in the Rhodope, Aegean and west Anatolian domains (figures 8n-w). The late Cretaceous southward migration of the trench is estimated in our kinematic reconstructions at 50-100 km (figure 12a), giving a very low rate of N-S extension (i.e. 0.1-0.3 cm/yr), if we make the assumption of a constant velocity in the late Cretaceous. This low trench retreat (figure 12b) contrasts with the large crustal stretching and subsequent oceanic spreading occurring at the same time in the Black Sea basin (figures 8a-h and 10) [Zonenshain and Le Pichon, 1986; Görür, 1988; Hippolyte *et al.*, 2010; Nikishin *et al.*, 2015b]. Although this retreat can explain the extensional tectonics, its amplitude is obviously not sufficient when considering the ~300 km north-south width of the Black Sea oceanic domain. However, the lack of further geological constraints on the structuration of the Black Sea basement, overlain by the ~15 km of sediments, limits the accuracy of our reconstructions in this region for this period. In contrast, the total amount of trench retreat since the Oligocene in the Aegean domain is estimated at ~500 km in a NNE-SSW direction (figure 12a) giving a rate of extension of 0.5-1 cm/yr in the Oligocene-early Miocene (figure 12b). During the curvature of the trench, this rate increases drastically up to 3.4 cm/yr between 15 and 8 Ma in the Aegean domain while it only reaches ~1.7

cm/yr along the Hellenides that form the flank of this bending trench. Intense crustal stretching behind the hinge of the trench then results in the exhumation of the deepest parts of the crust as migmatitic domes in the central Cyclades.

In our reconstructions the rate of trench retreat, however, increases as early as 50 Ma (figure 12b), concomitantly with inception of extension in the Rhodope massif and the subduction, accretion and exhumation of the Cycladic Blueschists unit, which belongs to the Pindos basin (figures 8k-n). However, the magmatic arc started to move southward only ~30 Ma, at the same time as the inception of the Aegean back-arc extension (figure 12o) (see also *Jolivet et al.* [2004a] and *Jolivet and Brun* [2010]). This discrepancy may be partially explained by the model of *Brun and Faccenna* [2008] suggesting that the velocity of slab roll-back (tracked by magmatic arc retreat) is modified by the nature of the lithosphere entering the trench. When continental lithosphere enters the trench, slab roll-back stops and the slab steepens while thrusts propagate within the subducting continental domain. Once the subducting continental crust has been decoupled from the subducting lithosphere by these thrusts, roll-back starts again with progressive shallowing of the slab, and back-arc extension ensues. The period between 50 and 35 Ma corresponds to the subduction of the partly oceanic Pindos basin that is progressively incorporated within the nappe stack. One may therefore assume that during this period the slab remained quite steady, progressively steepening, with little migration of the magmatic arc. Subsequently, decoupling was sufficient to allow an increase in the rate of roll-back, causing shallowing of the slab, back-arc extension and migration of the magmatic arc. Trench retreat and slab roll-back should therefore be considered separately. Slab roll-back seems to have been active in the eastern Mediterranean region not only since the Oligocene but also during the late Cretaceous, as suggested by the coeval trench retreat and magmatic arc migration in the Balkans and the Pontides (figures 8a-h) [*Yılmaz et al.*, 1997; *Ciobanu et al.*, 2002; *Eyüboğlu et al.*, 2010]. This result does not exclude the possibility that other processes, such as gravitational collapse of thickened crust [*Dewey*, 1988; *Vanderhaeghe*, 2012], could have partially controlled these extensional events.

Apart from the nature of the subducting lithosphere, other causes should be considered to explain the different extension rates associated with slab roll-back. Continuous subduction since the Mesozoic implies more subducted lithospheric material and therefore higher slab pull and faster slab roll-back since the Oligocene than during the late Cretaceous. Another parameter to consider is the slab width. Indeed, the resisting forces of the mantle to slab roll-back are higher with a wide slab [Faccenna *et al.*, 2007; Schellart *et al.*, 2007; Loiselet *et al.*, 2009], such as in the late Cretaceous along the Balkans and the Pontides. The width of this continuously subducting lithosphere was then probably reduced by several episodes of slab tearing, leading to an increase of its retreating rate. Since the middle Miocene, large-scale block rotations (figure 3) and evolution of magmatism in the Cyclades and western Anatolia (see details in *Section 5.5*) support the existence of a slab tear below western Anatolia, as suggested by several tomographic models [de Boorder *et al.*, 1998; Wortel and Spakman, 2000; Piromallo and Morelli, 2003; Dilek and Altunkaynak, 2009; Jolivet *et al.*, 2009, 2013; Brun and Sokoutis, 2010; Salaün *et al.*, 2012]. The reduced width of the slab thus could have induced its fast retreat and the fast rotation of the Hellenic trench from 15 to 8 Ma (figures 11 and 12) [Jolivet *et al.*, 2015], an effect that would probably have been accentuated by the toroidal mantle flow occurring on its edge. Similarly, the differential Pliocene-Quaternary large-scale counterclockwise rotations in the Hellenides and the Peloponnese (figure 3) and the second acceleration of trench retreat since ~5 Ma (figure 12b) could have been a consequence of a younger slab tear below the Corinth rift, decoupling the surface kinematics and crustal deformation between the Hellenides and the Peloponnese [Royden and Papanikolaou, 2011; Jolivet *et al.*, 2013]. It is also possible that an older segmentation of the subduction zone may have occurred, such as during the continental accretion in the late Cretaceous-Paleocene [e.g. Okay *et al.*, 2012]. However, the continuous E-W trending late Cretaceous Balkans-Pontides magmatic arc and Eocene calc-alkaline magmatic province emplaced from the Chalkidiki peninsula to the eastern Pontides suggest a continuous E-W trending subduction zone that progressively migrated southward through this time (figures 8a-g and k-n).

5.2. *Strain partitioning in the Aegean domain*

The use of deforming domains in these new kinematic reconstructions allows the consideration of the non-rigid behavior of the lithosphere, especially the stretching of the crust resulting from the dominant trench-perpendicular extension, such as during the Oligocene-Miocene back-arc extension in the Aegean domain (figure 11). The small component of trench-parallel extension [Reilinger *et al.*, 2006; van Hinsbergen and Schmid, 2012] resulting from the progressive curvature of the arc since the middle Miocene is also considered within this framework.

Stretching lineations recorded in metamorphic rocks in back-arc domains give information about the strain accommodating the exhumation of MCCs. The exhumation kinematics of the Rhodope, Kazdağ, Cyclades and Menderes MCCs in our reconstructions is consistent with these kinematic indicators (figure 13a). Although stretching lineations relate to strain and not to displacements, the overall parallelism between lineations and displacement trajectories suggests that N-S stretching deformation has been much more prevalent than perpendicular shortening in this region. Local abnormal trends of lineations are also observed, such as the WSW-ENE Eocene lineation in Syros. Considering the N-S lineation of the same age described in the neighboring Sifnos island (figure 3) [Roche *et al.*, submitted], this peculiar strike of the stretching lineation in Syros suggests that this island underwent local rotations.

Since the Oligocene, MCCs were exhumed in the footwalls of several ductile-brittle detachment systems (figures 9 and 11). In the Cyclades, for example, three large-scale structures with well-constrained kinematics and timing have been considered: the NCDS cropping out in the Tinos, Mykonos and Ikaria islands [Lee and Lister, 1992; Gautier and Brun, 1994; Mehl *et al.*, 2005; Brichau *et al.*, 2007; Jolivet *et al.*, 2010a; Lecomte *et al.*, 2010], the NPEFS cropping out in the Naxos and Paros islands [Lister *et al.*, 1984; Urai *et al.*, 1990; Gautier *et al.*, 1993; Seward *et al.*, 2009; Bargnesi *et al.*, 2013] and the WCDS cropping out in the Serifos island (figures 3 and 6) [Grasemann and Petrakakis, 2007; Brichau *et al.*, 2010; Iglseder *et al.*, 2011; Grasemann *et al.*,

2012]. In our reconstructions, the relative southward motion of the Tinos island in the footwall of the NCDS, with respect to the Pelagonian nappe in the Hellenides considered as the hanging wall of the NCDS, is estimated at ~70 km during the activity of this structure, fitting its estimated offset (i.e. ~60-100 km, figure 13b) [Jolivet *et al.*, 2004a; Brichau *et al.*, 2008].

The activity of these detachment systems continued after 15 Ma while the dynamics of back-arc extension changed with the increasing curvature of the trench related to slab tearing below western Anatolia (figure 11). The accommodation of the two opposite block rotations in the Cyclades was initially attributed to the hypothesized Mid-Cycladic Lineament [Walcott and White, 1998]. However, the existence of such a fault is mainly based on the deflection of stretching lineations (notably in the Paros island) and no clear strike-slip shear zone has ever been observed. In addition, the kinematics of such a fault is enigmatic, presented either as a dislocation zone [Walcott and White, 1998], as a detachment fault [van Hinsbergen and Schmid, 2012] or as a dextral strike-slip fault [Philippon *et al.*, 2012, 2014]. In our reconstructions, considering fully deforming domains, it appears that these opposite rotations could be accommodated by a gradient of finite extension, without the need of a localized fault [Jolivet *et al.*, 2015].

After this fast rotation and the probable slab tear, extension has developed further south with the ~120 km opening of the Cretan Sea since 10-9 Ma (figure 13b). At the same time, the stress regime in the Cyclades changed rapidly with the development of a distributed strike-slip then compressional deformation [Ring *et al.*, 1999a; Avigad *et al.*, 2001; Menant *et al.*, 2013]. The exact reason for this late Miocene E-W compression recorded in the Cyclades is unclear. Models of back-arc extension simply driven by slab roll-back do not predict compression in this region [Gautier *et al.*, 1999; Faccenna *et al.*, 2006; Philippon *et al.*, 2014; Sternai *et al.*, 2014]. Competition between slab roll-back, extrusion and mantle flow may render the distribution of stresses in the back-arc lithosphere quite complex, warranting new investigations on the interactions between extrusion tectonics and back-arc extension.

5.3. Successive metamorphic events in the Menderes massif

The Menderes massif is often interpreted as an Alpine nappe stack affecting a Pan-African basement and its sedimentary cover [e.g. Ring *et al.*, 1999b; Bozkurt & Oberhänsli, 2001; Gessner *et al.*, 2001a, 2013; Jolivet *et al.*, 2004b]. Following this accretionary event, this massif has been exhumed in a back-arc extensional context, forming a dome-like structure [e.g. Gessner *et al.*, 2001b; Lips *et al.*, 2001; van Hinsbergen *et al.*, 2010b]. This succession of tectonic events resulted in a complex metamorphic succession that is still largely debated today, notably due to the existence of a pre-Alpine tectono-metamorphic history [Şengör *et al.*, 1984; Candan *et al.*, 2001].

In the Eocene, two metamorphic events were recorded in the Menderes massif: (1) MT- to HT-MP Barrovian-type metamorphism observed in the major part of the Menderes massif and called the *Main Menderes Metamorphism* and (2) HP-LT metamorphism preserved in the southern part of the massif. Because of the staggered ages of these metamorphic events, some authors argue that the HP-LT metamorphism occurred before the *Main Menderes Metamorphism* [e.g. Rimmelé *et al.*, 2003b; Pourteau *et al.*, 2013]. However, the lack of HP-LT relics in the whole Menderes massif seems to contradict this hypothesis (although rock chemistry restrictions can also be considered [Rimmelé *et al.*, 2003b]) and the overlapping of the metamorphic ages of these two events suggests that they were coeval (figure 5). The question is, how is it possible? According to many authors, the *Main Menderes Metamorphism* resulted from the nappe stacking of the Menderes basement and cover during the Eocene [Akkök, 1983; Şengör *et al.*, 1984; Bozkurt and Oberhänsli, 2001; Okay, 2001]. In our kinematic reconstructions, this unit started to subduct in the late Paleocene-early Eocene and was quickly accreted to the Eurasian margin as the subduction front propagated southward (figure 10). The upper nappes, forming the present-day northern Menderes massif, could thus have undergone MT- to HT-MP metamorphism as a result of crustal thickening. Conversely, the lower nappes could have been buried within the subduction zone and experienced HP-LT metamorphism, observed in the southern part of the Menderes massif [Rimmelé *et al.*, 2003b; Whitney *et al.*, 2008; Pourteau *et al.*, 2013]. However, these synchronous events require a major

tectonic discontinuity that would have juxtaposed these units and further investigations are therefore needed to clarify this point [Whitney *et al.*, 2008].

Following this Eocene metamorphic stage, a HT metamorphic event has also been recorded in the northern Menderes massif, characterized by migmatites of Oligocene age (figure 5). The few geochronological data show that this event partially overlapped the *Main Menderes Metamorphism*, suggesting a relatively continuous MT to HT metamorphic event from the Eocene to the Oligocene [Catlos and Çemen, 2005; Bozkurt *et al.*, 2010]. Thus, similarly to the Aegean domain, the Menderes massif seems to have been progressively transferred in late Eocene-early Oligocene times from the compressional arc domain, where it underwent both subduction-related HP-LT and nappe stacking-related Barrovian-type metamorphisms, to the extensional back-arc domain where it was exhumed as a MCC in a HT environment (figure 10). Conversely, the exhumed parts of the southern Menderes massif only experienced greenschist-facies metamorphism [Hetzel *et al.*, 1995a; Bozkurt and Oberhänsli, 2001], suggesting an exhumation process either at a higher rate or in a colder environment.

5.4. Driving forces acting on the westward extrusion of Anatolia

Highlighted by GPS measurements, the present-day kinematics of the eastern Mediterranean region is dominated by an overall counterclockwise rotation of velocity vectors from the northward collision of Arabia, westward motion of Anatolia and southward retreat of the Hellenic trench with respect to Eurasia [Reilinger *et al.*, 2006; Le Pichon and Kreemer, 2010]. Several studies focused on the respective contributions of crustal and mantle processes in driving crustal deformation. For Armijo *et al.* [1999], the main force moving Anatolia is the push from the Arabia-Eurasia collision in a typical extrusion model, while in more recent contributions, the main force is slab roll-back [Faccenna *et al.*, 2006; Becker and Faccenna, 2011; Jolivet *et al.*, 2013; Sternai *et al.*, 2014].

Integrated into our kinematic reconstructions with a rotation pole situated close to the Nile delta, the extrusion of Anatolia started ~12 Ma according to the timing of the inception of the NAF [Şengör *et*

et al., 2005]. The resulting post-12 Ma surface kinematics fits the present-day kinematics deduced from GPS measurements, with a southward migration of the southern Aegean domain significantly faster than the westward motion of Anatolia (figure 12) suggesting a relatively steady kinematics in the eastern Mediterranean region since the late Miocene. In addition, this surface kinematics is in agreement with the pattern of seismic anisotropy Aegean domain (figure 12), which may be considered as a proxy for tracking mantle flow [Jolivet *et al.*, 2009, 2013; Paul *et al.*, 2014]. These results, supported by (1) the similarity between mantle-related seismic anisotropy directions and the pattern of deformation in thinned back-arc domains [Jolivet *et al.*, 2009], (2) mantle-related instantaneous 3D velocity field calculations [Becker and Faccenna, 2011] and (3) high-resolution 3D numerical modeling [Sternai *et al.*, 2014], suggest that slab roll-back and related mantle flow codetermine surface deformation in this region and notably the extrusion of Anatolia as soon as its inception ~12 Ma.

5.5. *Southward retreat of the subduction zone and magma genesis*

Our kinematic reconstructions indicate that magmatic activity in the eastern Mediterranean region has migrated southward together with the subduction zone since the late Cretaceous, showing the strong control of subduction dynamics on magma genesis (figures 8a-y). The distribution and petrological and geochemical variations of these magmas suggest that their sources have evolved through time with changes in the subduction geometry and dynamics.

Medium-K calc-alkaline magmatism with mostly intermediate composition (commonly called arc-related magmatism) results from (1) partial melting of a hydrous mantle source metasomatized during the subduction of an oceanic lithosphere and (2) a subsequent episode of magma fractionation and crustal assimilation in the melting assimilation storage and homogenization (MASH) zone at the base of the crust [Gill, 1981; Hildreth and Moorbath, 1988; Harangi *et al.*, 2006]. In the eastern Mediterranean region, this medium-K calc-alkaline magmatism is represented with the late Cretaceous Balkans-Pontides magmatic province (figures 8a-h) [Yılmaz *et al.*, 1997;

Berza et al., 1998], the Eocene calc-alkaline magmatic province (figures 8j-n) [*Harris et al.*, 1994; *Okay and Satir*, 2006] and the still active south Aegean volcanism (figures 8x-y) [*Fytikas et al.*, 1984; *Pe-Piper and Piper*, 2005]. In reconstructed cross-sections (figures 9 and 10), all these magmatic provinces are emplaced above a dehydrating oceanic slab (i.e. the Vardar and Izmir-Ankara-Erzican oceanic lithosphere in the late Cretaceous and Eocene and the Mesogean oceanic lithosphere since the Pliocene) allowing the formation of primary magmas in the mantle wedge. It thus appears that magmatism with arc-related medium-K calc-alkaline geochemical characteristics may occur in a back-arc-related transtensional or extensional context as soon as dehydration of the underlying oceanic slab triggers partial melting in the mantle wedge.

When the slab retreats, the processes of magma genesis change. In the back-arc region, the increased geothermal gradient due to lithospheric thinning and underlying asthenospheric flow allow for the partial melting of the lower crust and possibly of the lithospheric mantle, supplying the MASH zone. This lithospheric mantle is often highly metasomatized as a result of the subduction of both continental and oceanic materials [*Hawkesworth et al.*, 1995; *Harangi et al.*, 2006; *Ersoy and Palmer*, 2013]. These magmas then show a higher magmatic alkali content giving high-K calc-alkaline to shoshonitic compositions. In the eastern Mediterranean region, the late Cretaceous CACC plutonic province (figures 8b-g) [*Aydin et al.*, 1998; *Ilbeyli et al.*, 2004; *Boztuğ et al.*, 2009], the latest Cretaceous volcanic activity in the eastern Srednogie region (figures 8g-h) [*Boccaletti et al.*, 1978] and the Oligocene-Miocene Rhodope and west Anatolian-Aegean magmatic provinces (figure 11) [*Ersoy and Palmer*, 2013 and references therein] display this typical back-arc-related composition. This high-K magmatism often developed during the exhumation of MCCs (figures 9 and 10). In this context, detachment systems control the exhumation and cooling of syn-tectonic intrusions that show a pervasive co-magmatic to ductile-brittle deformation controlled by the regional extensional stress regime [*Faure and Bonneau*, 1988; *Dinter and Royden*, 1993; *Gessner et al.*, 2001a; *Denèle et al.*, 2011; *Laurent et al.*, 2015; *Rabillard et al.*, 2015]. With ongoing extension, the respective proportions of crustal and mantle source

components in these magmas change with an increase of the subduction-related metasomatized mantle source component through time, as observed in the CACC, the Rhodope massif and the Cyclades. In Naxos (central Cyclades), partial melting of metasediments is evidenced by field relationships [Vanderhaeghe, 2004; Kruckenberg *et al.*, 2011] resulting in dykes of S-type granites dated ~15 Ma and emplaced before the ~13 Ma I-type granite displaying a mantle source component (U-Pb crystallization ages on zircon) [Bolhar *et al.*, 2010]. The same relative timing is observed in the Ikaria island, where two S-type muscovite-bearing granites resulting from the partial melting of continental crust were emplaced, followed by a I-type granite [Altherr *et al.*, 1982; Ring, 2007; Bolhar *et al.*, 2010; Laurent *et al.*, 2015]. This change was likely due to the thinning and heating of the lithosphere, allowing for the partial melting of the subduction-related metasomatized mantle [Altherr and Siebel, 2002; Ilbeyli *et al.*, 2004; Marchev *et al.*, 2005; Stouraiti *et al.*, 2010; Ersoy and Palmer, 2013].

If slab roll-back continues, the extreme thinning of the upper plate can trigger the partial melting of the underlying asthenosphere, producing alkaline magmas with a major depleted mantle source component, typical for an intraplate-related magmatism. This magmatism developed in the Balkans (figures 8g-k), the central Rhodope massif and from the Thrace basin to the Isparta angle (figures 8t-y), succeeding to the high-K calc-alkaline to shoshonitic magmatism [Cvetković *et al.*, 2004; Ilbeyli *et al.*, 2004; Marchev *et al.*, 2004b; Agostini *et al.*, 2007].

This first-order spatial and compositional evolution of the arc-, back-arc- and finally intraplate-related magmatism has already been documented by *Pe-Piper and Piper* [2006]. They invoke several slab break-off or lithospheric delamination events during the earlier Cenozoic triggering the rise of hot asthenosphere and the change in magma genesis processes. In our opinion, although magma genesis processes remain the same, the subduction dynamics and related asthenospheric flow controlling these processes are different. Indeed, tomographic models show a continuous 1,500 km-long slab in the Rhodope-Aegean-west Anatolian region, suggesting continuous subduction since the late Cretaceous without any slab break-off [Wortel and Spakman, 2000; Piromallo and

Morelli, 2003]. The progressive southward retreat of the subduction zone and related lithospheric thinning and underlying asthenospheric flow, notably after 35-30 Ma are the most likely processes controlling the first-order evolution of the magmatism. However, other subduction-related or -unrelated processes should also be considered when variations of this *classical* magmatic evolution are observed.

5.6. *Modulation of magmatic evolution by slab tearing and mantle plumes*

The surge of magmatism in the eastern Aegean domain in the Miocene, with composition evolving from high-K calc-alkaline to alkaline (including shoshonitic composition), is not evenly distributed over the entire back-arc domain but instead is localized above the slab tear postulated from seismic tomographic models [de Boorder *et al.*, 1998; Wortel and Spakman, 2000; Dilek and Altunkaynak, 2009; Jolivet *et al.*, 2013; 2015]. In addition, the westward emplacement of syn-tectonic intrusions from the Menderes massif to the Cyclades in the Miocene (figure 11), associated with an increase of the mantle source component of these magmas [Stouraiti *et al.*, 2010; Jolivet *et al.*, 2015], can be correlated with this slab tear. Indeed, as shown by analog and numerical modeling [Funiello *et al.*, 2003, 2006; Piromallo *et al.*, 2006; Faccenda and Capitanio, 2012; Sternai *et al.*, 2014], slab tearing-related asthenospheric flow displays a significant toroidal component that could progressively heat the base of the stretched lithosphere inducing partial melting.

In eastern Anatolia, the late Miocene-Quaternary volcanism shows a similar evolution to the west Anatolian-Aegean magmatic province, with a southward migration of volcanism whose composition evolves from high-K calc-alkaline to alkaline (figures 8t-y) [Pearce *et al.*, 1990; Keskin, 2003]. However, since the Miocene, no significant crustal thinning is observed in this region that is instead supported by an ~45 km thick crust supporting a 2 km high plateau [Dewey *et al.*, 1986; Şengör *et al.*, 2003]. Suggested on tomographic models, slab break-off below the Bitlis belt and arrival of asthenospheric flow from the Afar hot spot could allow the preservation of a high-elevation plateau and the thermal delamination of the lithospheric mantle below eastern

Anatolia [Şengör *et al.*, 2003; Keskin, 2003, 2007; Faccenna *et al.*, 2013]. Successive partial melting of the lower crust, lithospheric mantle and finally asthenosphere could then control the emplacement of this high-K calc-alkaline to alkaline magmatism.

6. Conclusion

Integrating the principle of non-rigid domains, the new kinematic reconstructions presented in this paper describe the late Cretaceous-Cenozoic evolution of the eastern Mediterranean region including the history of rotations, the kinematics of the shear zones, the timing of exhumation of metamorphic units and the distribution and geochemical characteristics of magmatic centers.

Extension and related back-arc opening occurred during (1) the late Cretaceous along the Balkans and the Pontides and (2) since the Eocene-Oligocene in the Rhodope-Aegean-west Anatolian region, separated during the Paleocene by a compressional episode characterized by the accretion of Africa-derived continental blocks. Associated with an acceleration of the southward slab roll-back, extension in the Rhodope-Aegean-west Anatolian region is characterized by a faster kinematics culminating after 15 Ma probably as a result of a slab tearing below western Anatolia.

The kinematic reconstructions show the progressive deformation of the back-arc domain and the formation of the main detachments, such as the NCDS, the WCDS and the Kerdyllion detachment. They also show the progressive exhumation of the main metamorphic units from their maximum burial to their arrival at the surface.

The migration of magmatic activity, as well as the evolution of its composition from arc- to back-arc- and finally intraplate-like geochemical signature, is consistent with this evolution with (1) the first-order southward migration of the magmatic centers controlled by the southward retreat of the continuous Tethyan subduction zone since the late Cretaceous and (2) the second-order localization of high-K calc-alkaline and alkaline magmatism and the westward migration of plutonic bodies from the Menderes massif to the Cyclades during the Miocene, controlled by the tearing of the slab and related asthenospheric flow.

Finally, the driving forces of crustal dynamics may be also discussed in light of these kinematic reconstructions. The faster migration of the Hellenic trench than the westward extrusion of Anatolia since the late Miocene implies a steady-state kinematics in this region since the late Miocene mainly controlled by slab roll-back. At the same time, the emplacement of magmatic centers in eastern Anatolia with a high depleted mantle source component suggests the arrival of asthenospheric flow beneath this region, which could also drive crustal deformation by basal drag in regions where the crust is relatively thin and hot.

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Figure captions

Figure 1: Tectonic map of the eastern Mediterranean region highlighting the main tectono-metamorphic domains. Modified from *Jolivet et al.* [2013]. Base map made with GeoMapApp (<https://www.geomapapp.org>).

Figure 2: Map of the eastern Mediterranean region showing the main domains used in the kinematic reconstructions. Each domain has a specific ID (see details in Table S1) as well as an age of appearance and disappearance (in *italic*). Base map made with GeoMapApp (<https://www.geomapapp.org>).

Figure 3: Tectonic map showing the main data used to constrain the kinematic evolution of the eastern Mediterranean region, including the stretching lineations in the MCCs [*Rimmelé et al.*, 2006; *Bonev and Beccaletto*, 2007; *Jolivet et al.*, 2013], the offset of main structures [*Armijo et al.*, 1999; *Jolivet et al.*, 2004a, 2010b; *Brichau et al.*, 2006, 2008] and the vertical-axis rotations deduced from paleomagnetic studies [*Kissel and Laj*, 1988; *Kissel et al.*, 1993, 2003; *Pătraşcu et al.*, 1994; *Morris and Anderson*, 1996; *van Hinsbergen et al.*, 2005a, 2008, 2010a; *Brun and Sokoutis*, 2007; *Bradley et al.*, 2013]. CW: clockwise rotation. CCW: counterclockwise rotation. NAF: North Anatolian Fault. NCDS: North Cycladic Detachment System. NPEFS: Naxos/Paros Extensional Fault System. Base map made with GeoMapApp (<https://www.geomapapp.org>).

Figure 4: Present-day cross-sections in the eastern Mediterranean region. See figure 1 for location. Modified from *Jolivet and Brun* [2010]. Crustal thickness in the Aegean and western Anatolian region has been extracted from *Tirel et al.* [2004] and *Karabulut et al.* [2013].

Figure 5: Diagram showing the main tectonic, metamorphic and magmatic events since the late Cretaceous used to constrain the kinematic reconstructions of the eastern Mediterranean region (see text for references).

Figure 6: Detailed tectonic map of the Aegean region. NCDS: North Cycladic Detachment System. NPEFS: Naxos/Paros Extensional Fault System. SCSZ: South Cycladic Shear Zone. WCDs: West Cycladic Detachment System. Base map made with GeoMapApp (<https://www.geomapapp.org>).

Figure 7: Detailed tectonic map of the western and central Anatolian region. Base map made with GeoMapApp (<https://www.geomapapp.org>).

Figure 8: (a-y) Paleotectonic maps extracted from the kinematic reconstruction model. CACC: Central Anatolian Crystalline Complex. OOP: Olympos-Ossa-Pelion tectonic windows.

Figure 9: Reconstructed cross-sections of the Aegean region highlighting the different magma genesis processes. MASH zone: Melting Assimilation Storage and Homogenization zone. See figure 4 for detailed legend.

Figure 10: Reconstructed cross-sections of the western Anatolian region highlighting the different magma genesis processes. MASH zone: Melting Assimilation Storage and Homogenization zone. See figure 4 for detailed legend.

Figure 11: Detailed paleotectonic maps of the Rhodope-Aegean-west Anatolian region during the Oligocene-Miocene extension. White arrows: direction of stretching. Black arrows: main block rotations. The thick grey line shows the position of the slab at 150 km depth. NAF: North Anatolian Fault. NCDS: North Cycladic Detachment System. NPEFS: Naxos/Paros Extensional Fault System. OOP: Olympos-Ossa-Pelion tectonic windows. WCDS: West Cycladic Detachment System. See figure 8 for detailed legend.

Figure 12: (a) Motion paths deduced from the kinematic reconstructions of the eastern Mediterranean region. SKS fast splitting directions over the eastern Mediterranean region are also represented [Paul *et al.*, 2014]. (b) Time-velocity diagram, deduced from the kinematic reconstructions, showing the southward to southwestward trench retreat in the Hellenides and the Aegean domain. Main tectono-metamorphic events in the Hellenide-Rhodope-Aegean region are also shown (1: subduction of the Vardar oceanic lithosphere; 2: subduction and accretion of the Pelagonian platform; 3: subduction and accretion of the Pindos basin, formation of the HP-LT Cycladic Blueschists unit, extension in the Rhodope massif; 4: exhumation of the HP-LT Cycladic Blueschists unit; 5: subduction and accretion of the Gavrovo-Tripolitza platform, formation of the HP-LT Phyllite-Quartzite unit, extension in the Aegean domain; 6: exhumation of the HP-LT Phyllite-Quartzite unit; 7: first slab tear and curvature of the trench; 8: second slab tear?).

Figure 13: (a) Tectonic map showing the correlation between deformation (i.e. stretching lineations) and kinematics of exhumation of the MCCs in the Aegean domain since the Eocene (deduced from the kinematic reconstructions). (b) Time-displacement diagram, deduced from the kinematic reconstructions, showing the cumulative southward to southwestward displacement of the

Hellenides, the Cyclades (the Tinos island) and Crete since the Oligocene. NCDS: North Cycladic Detachment System.

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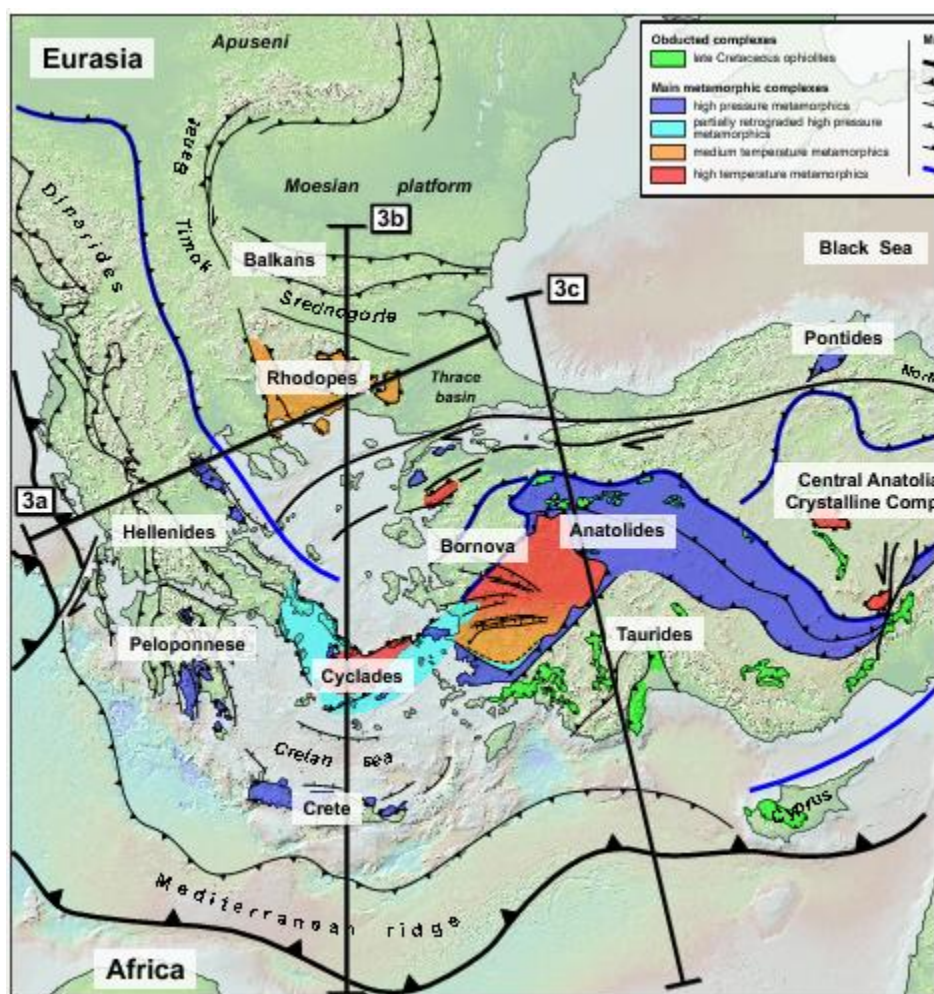


Figure 1

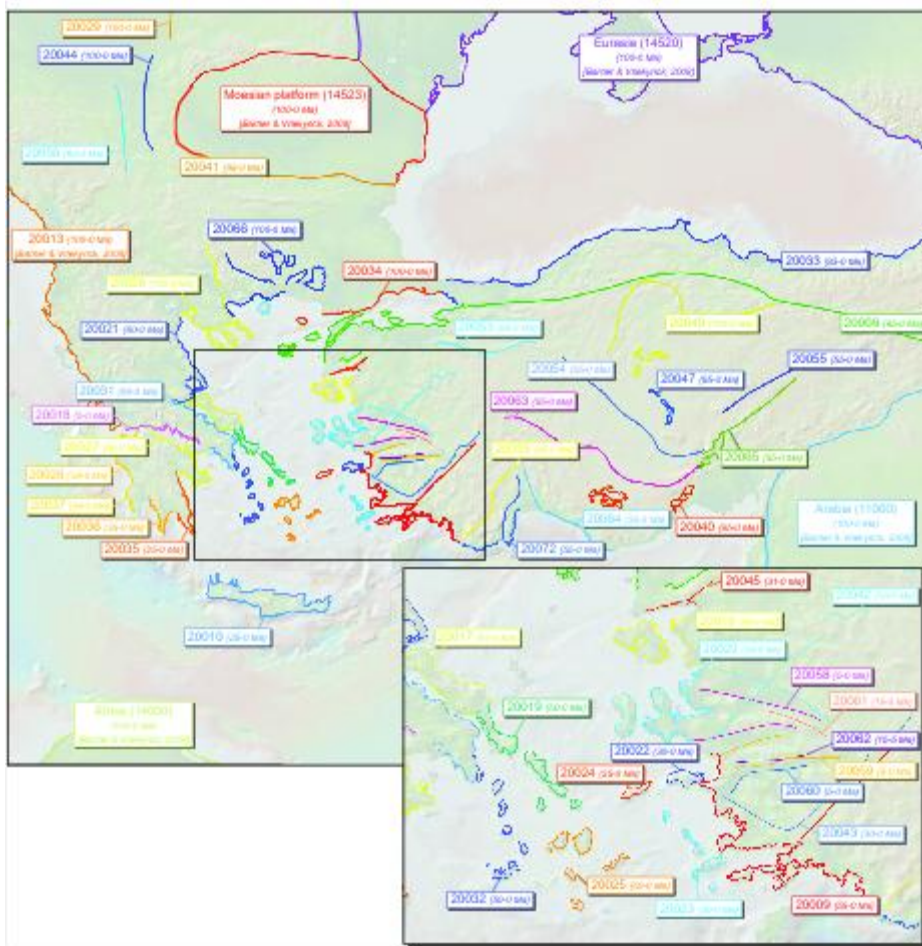


Figure 2

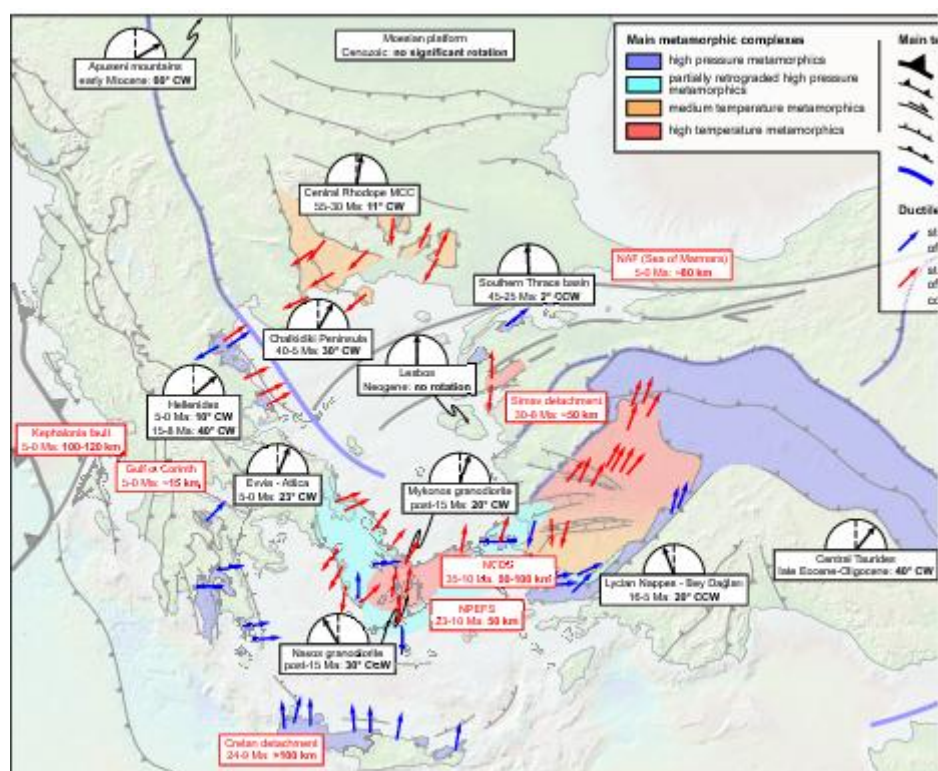


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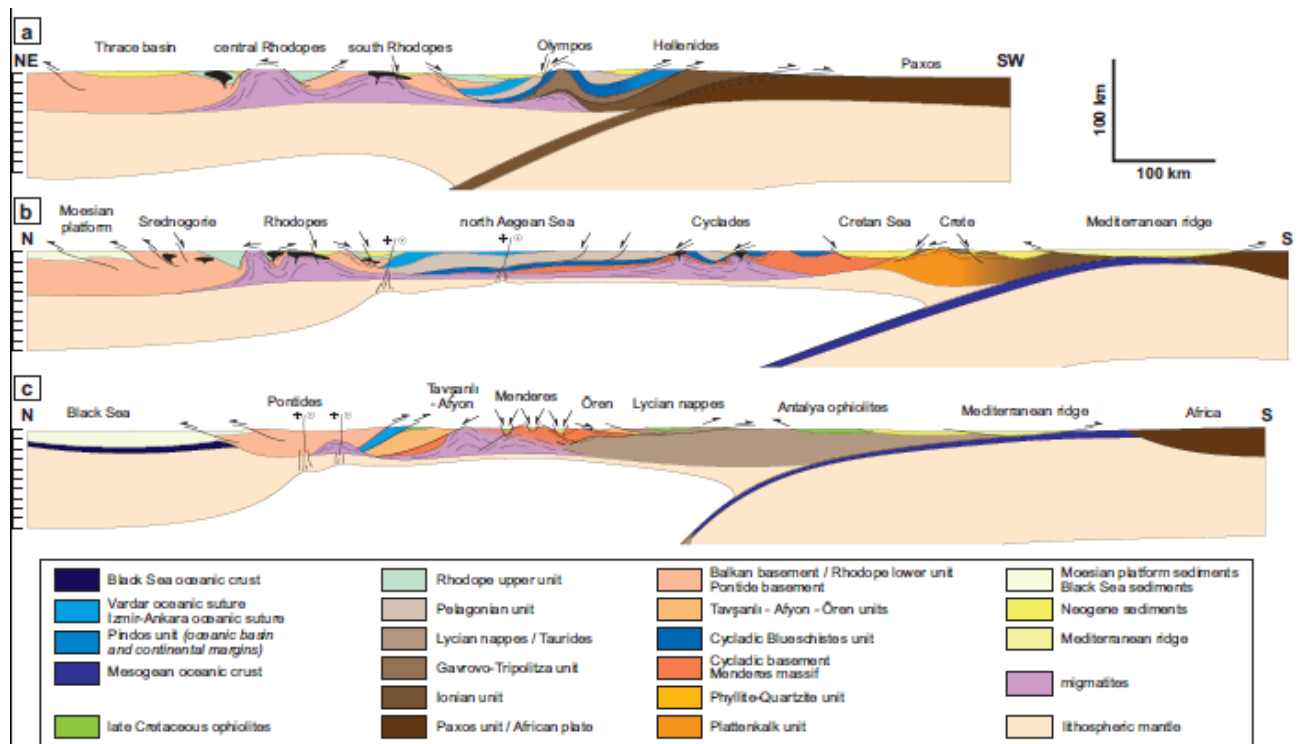


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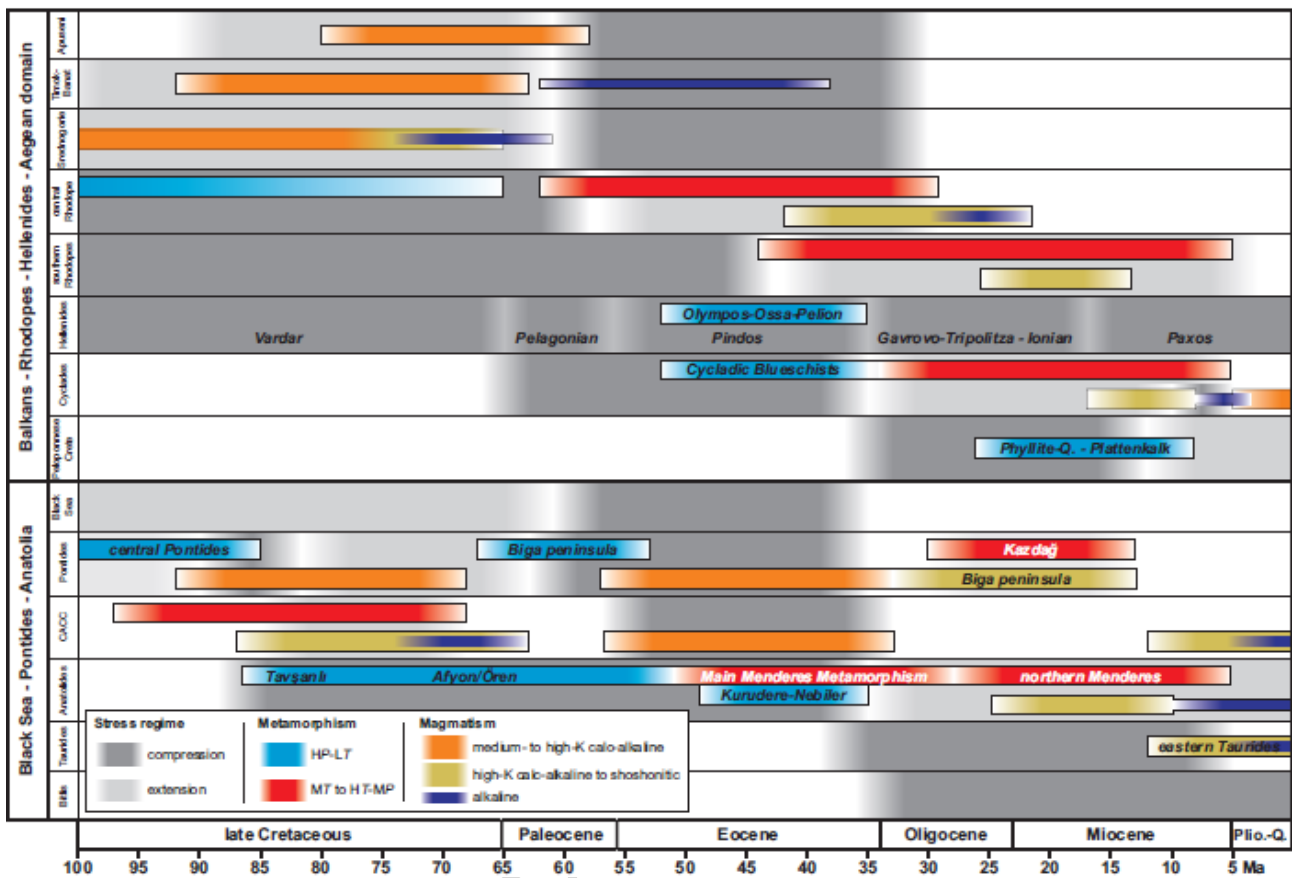


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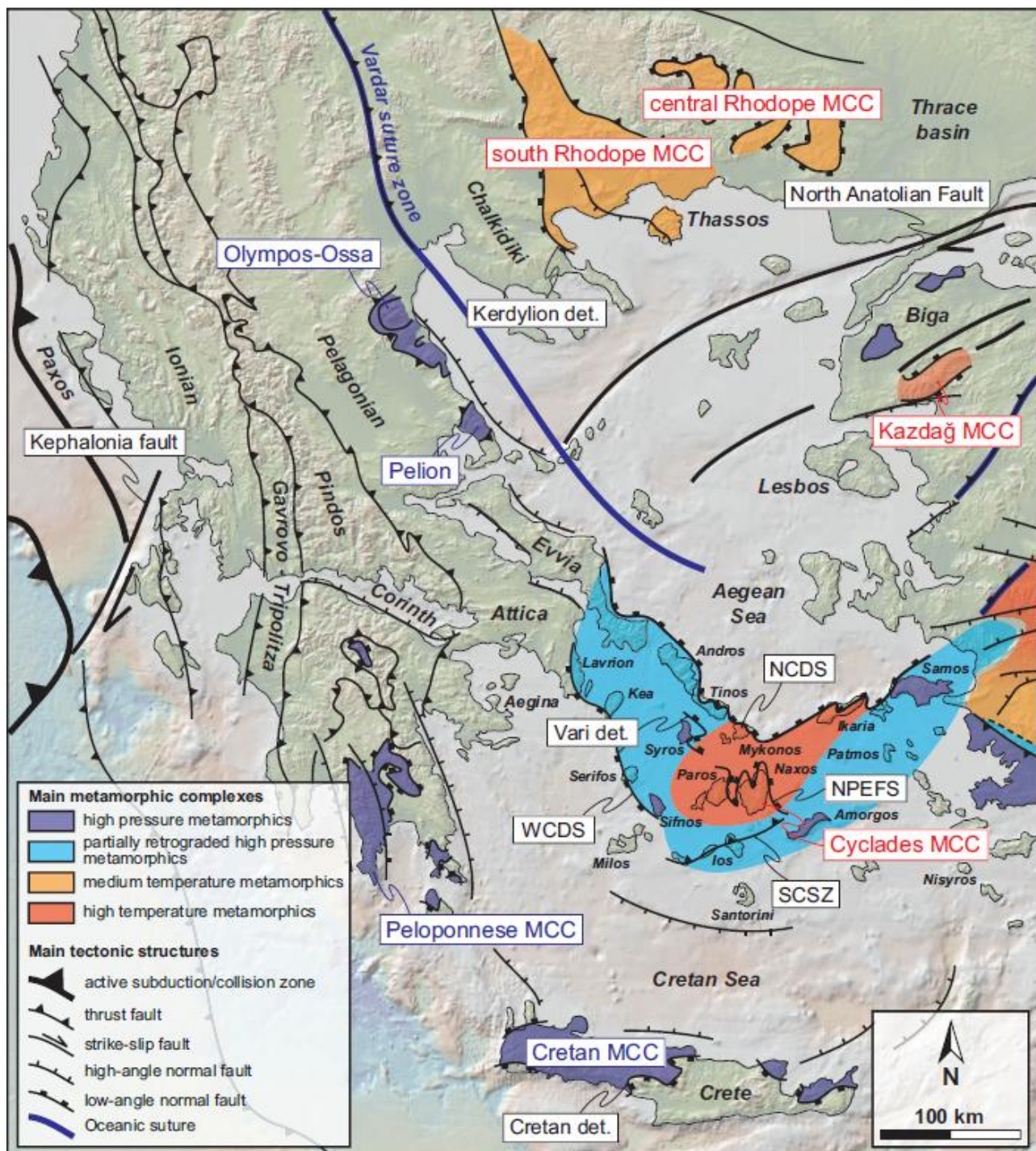


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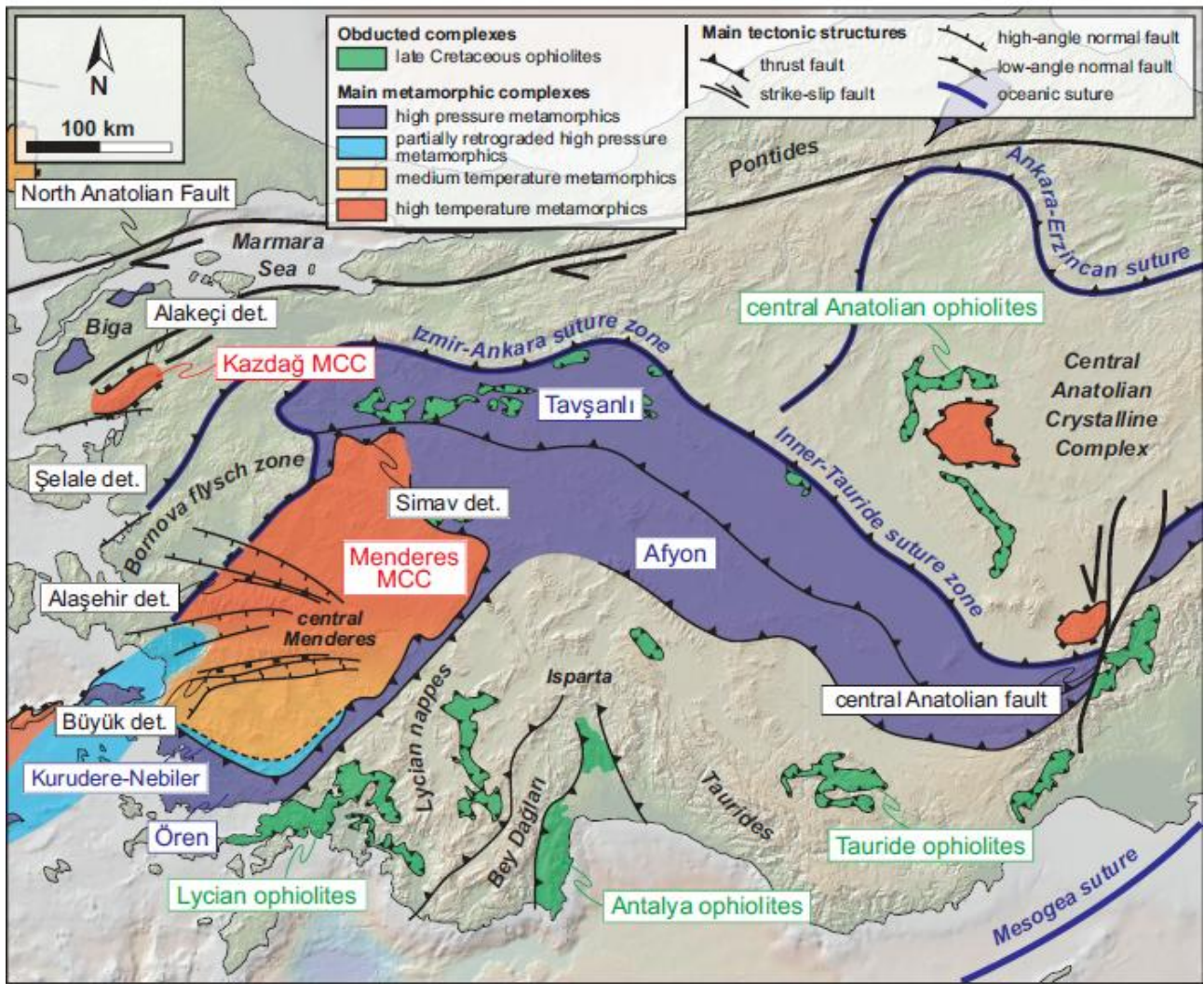
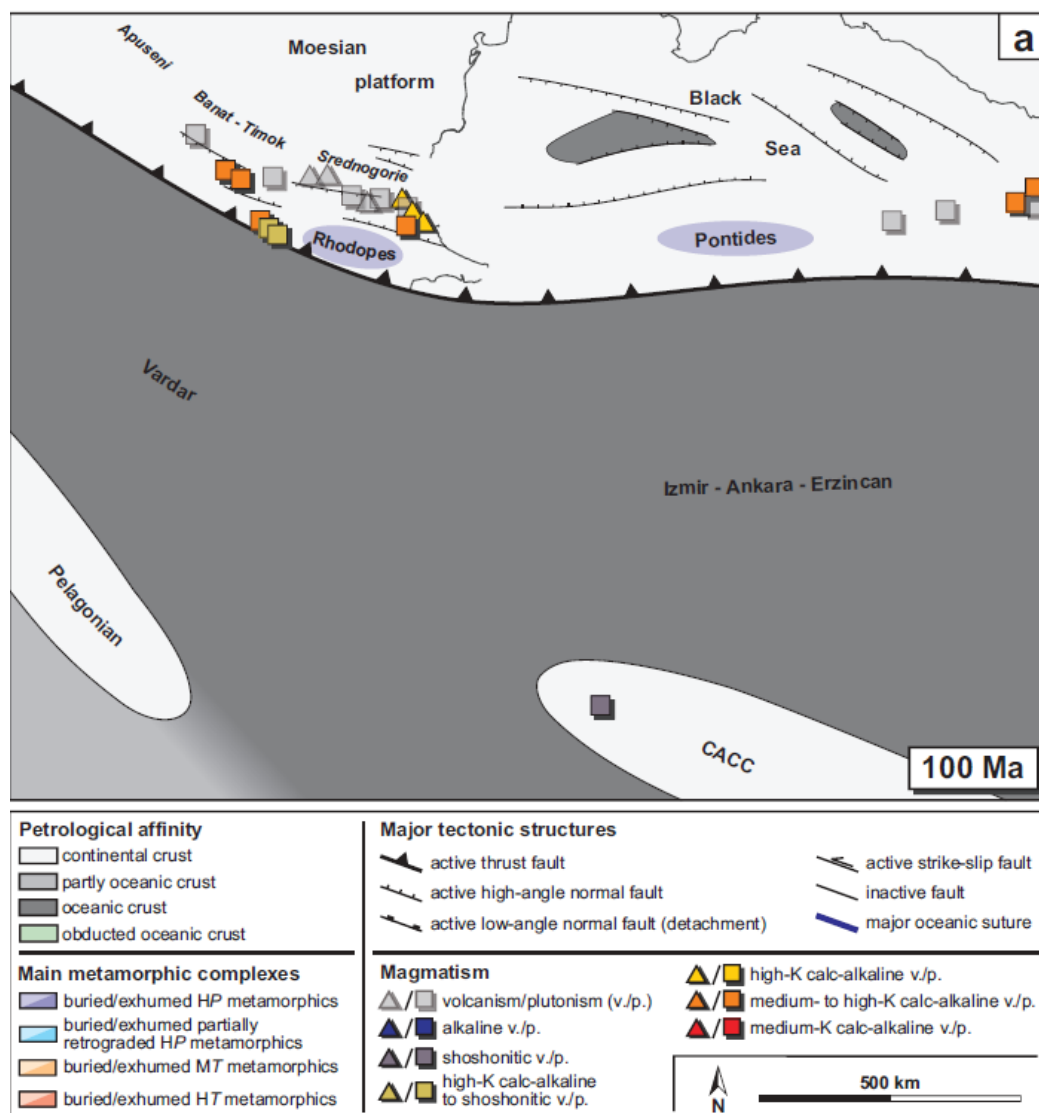
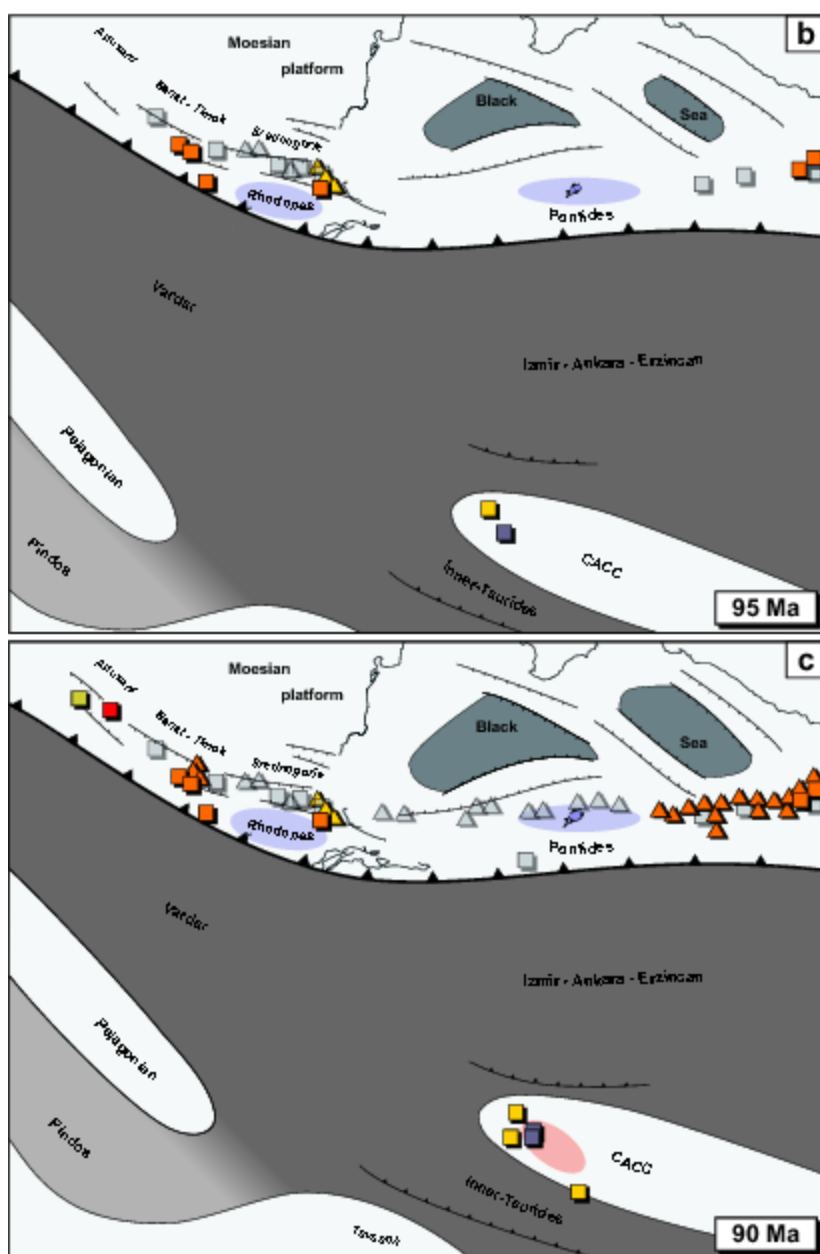
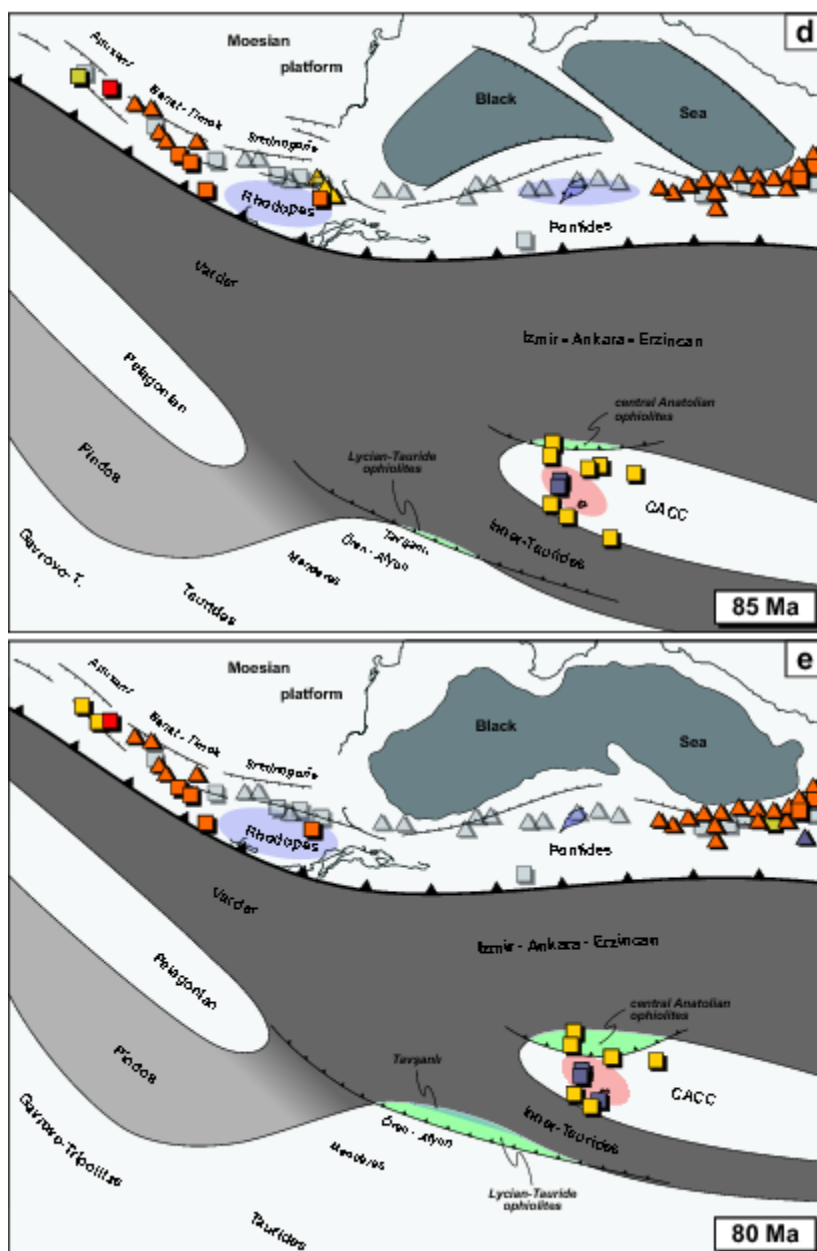
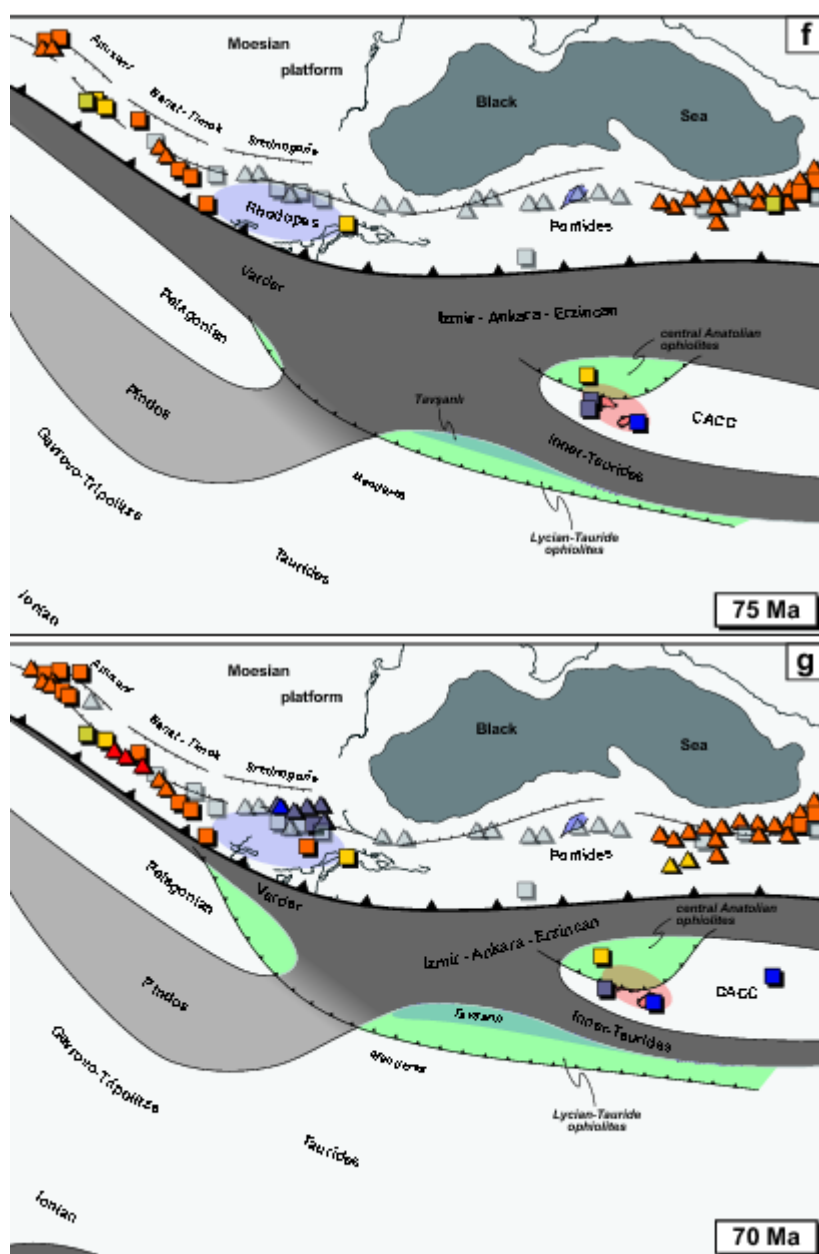


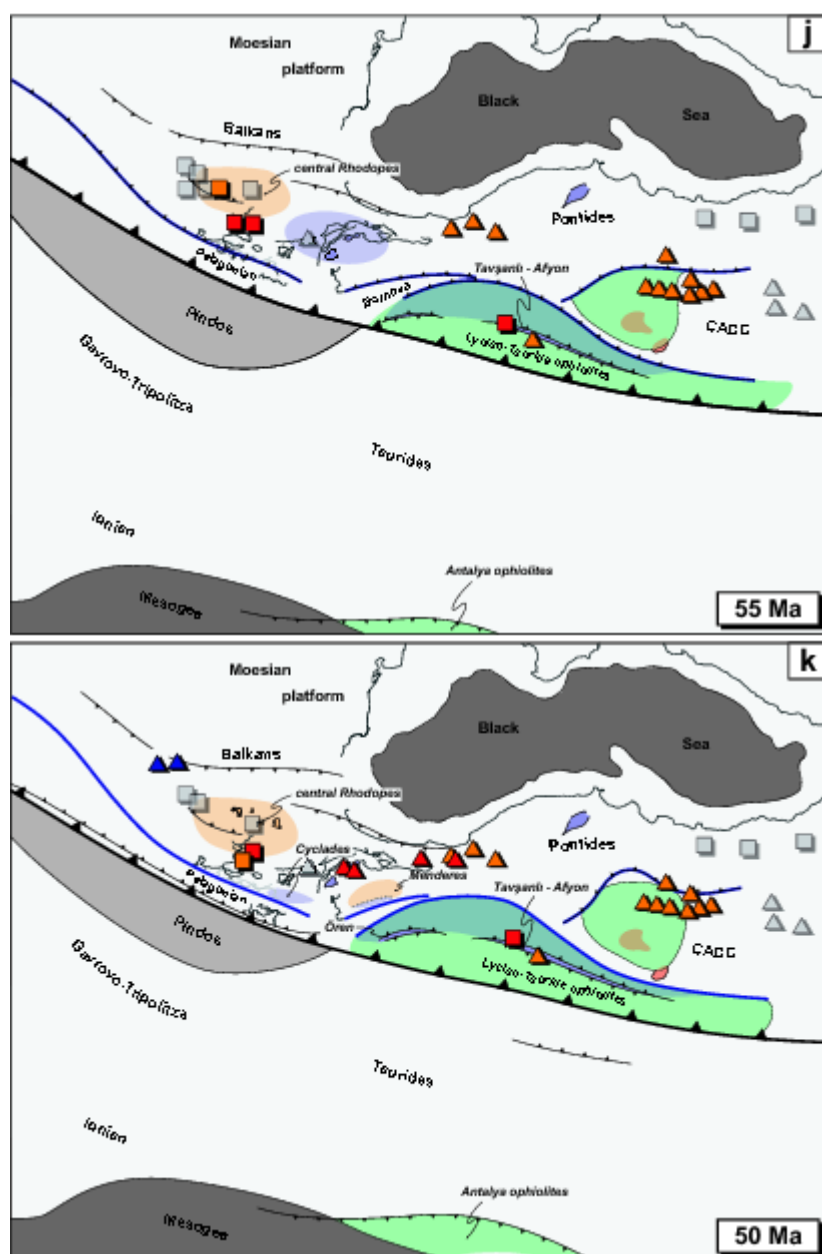
Figure 7

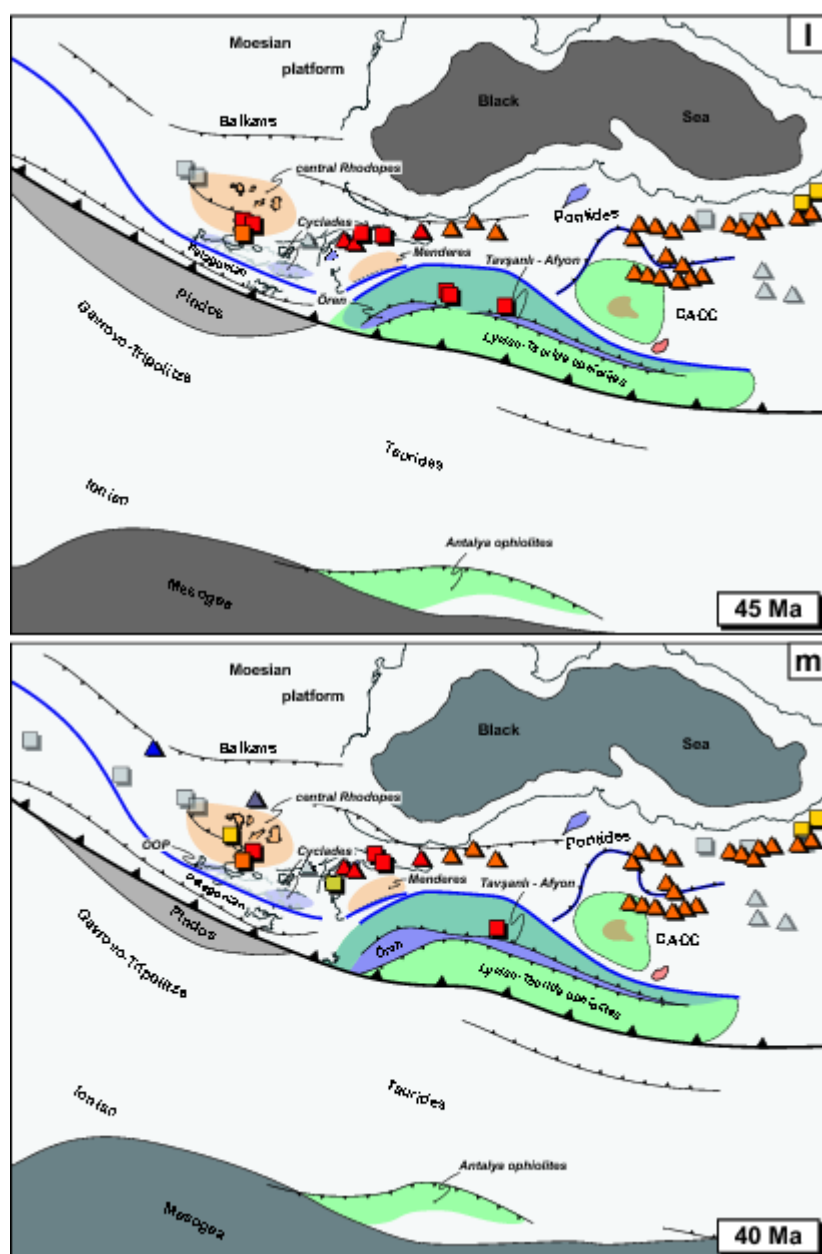


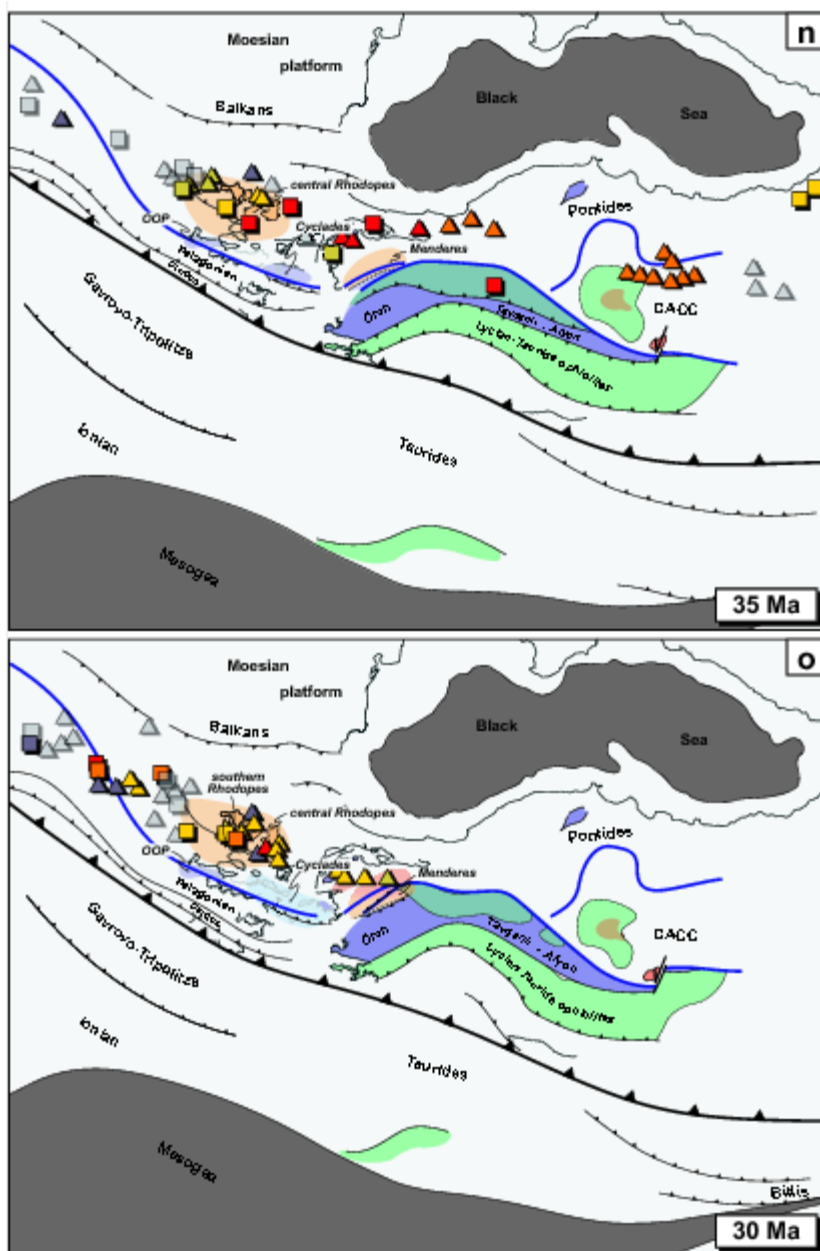


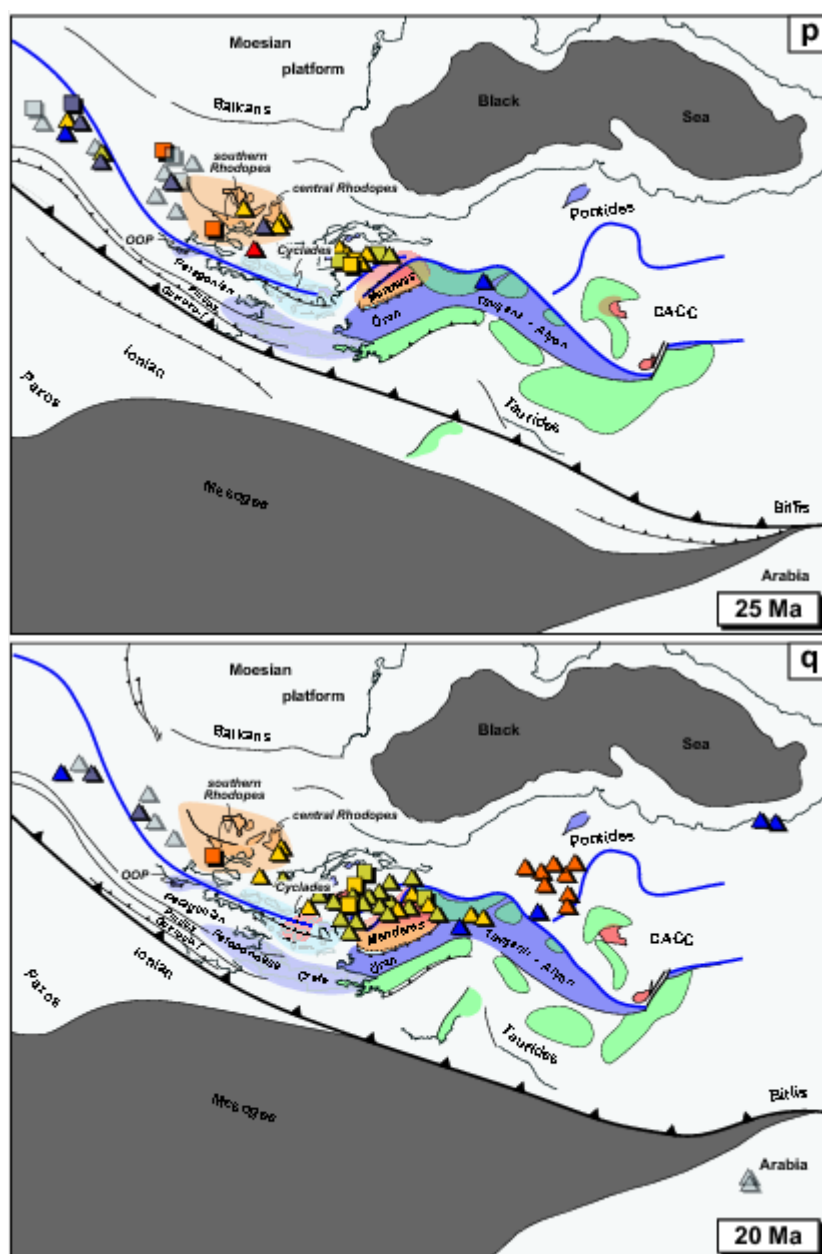












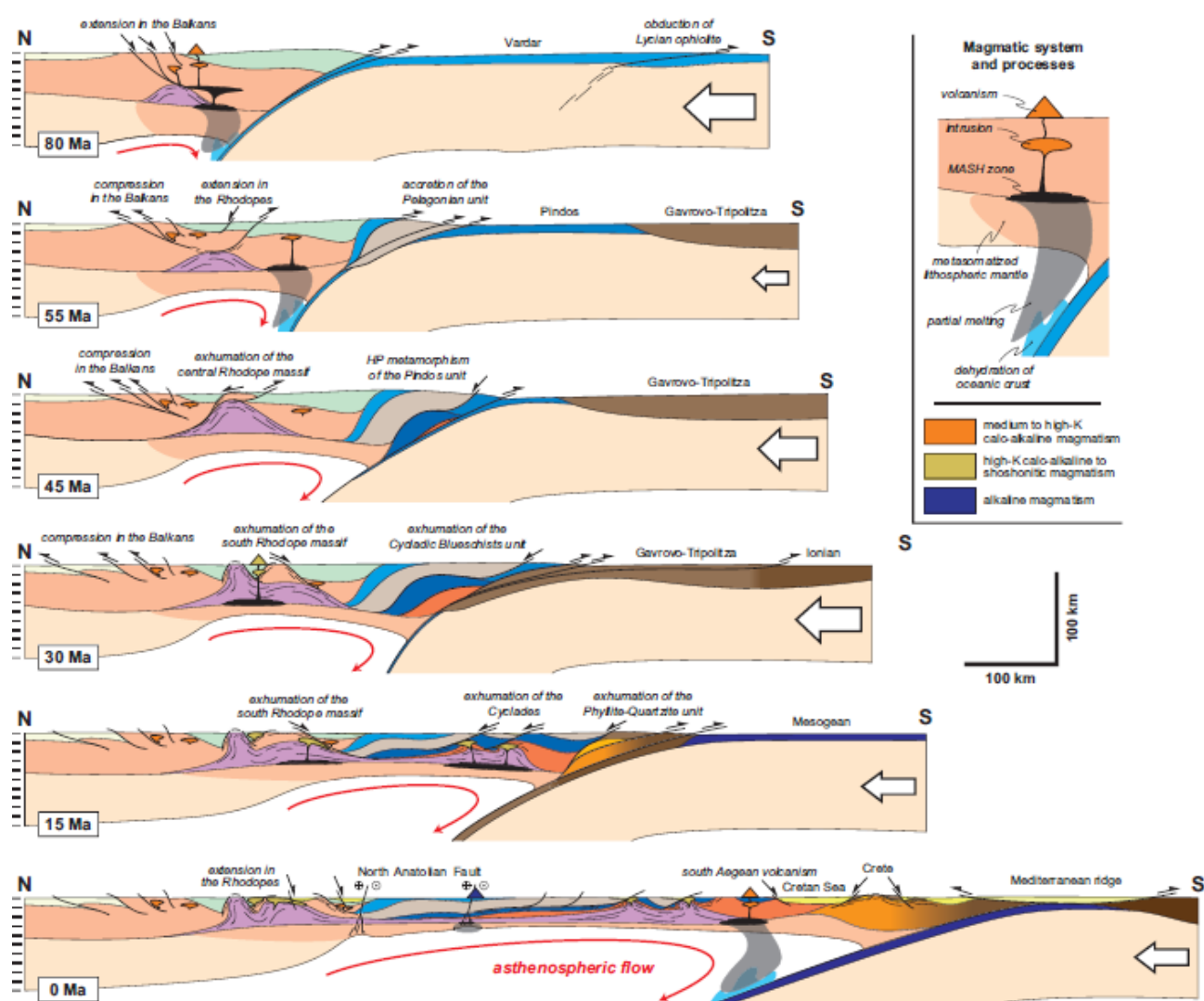


Figure 9

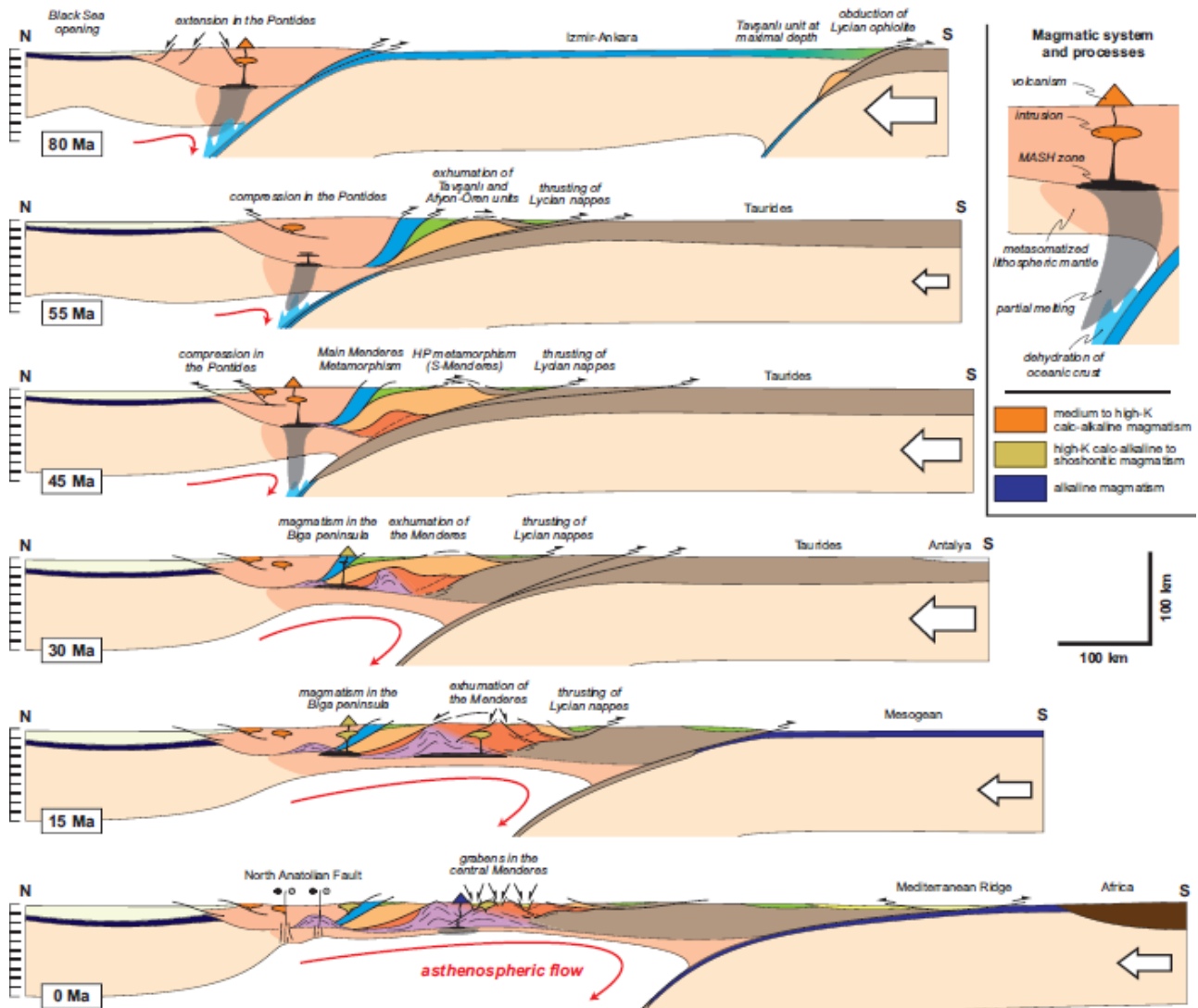


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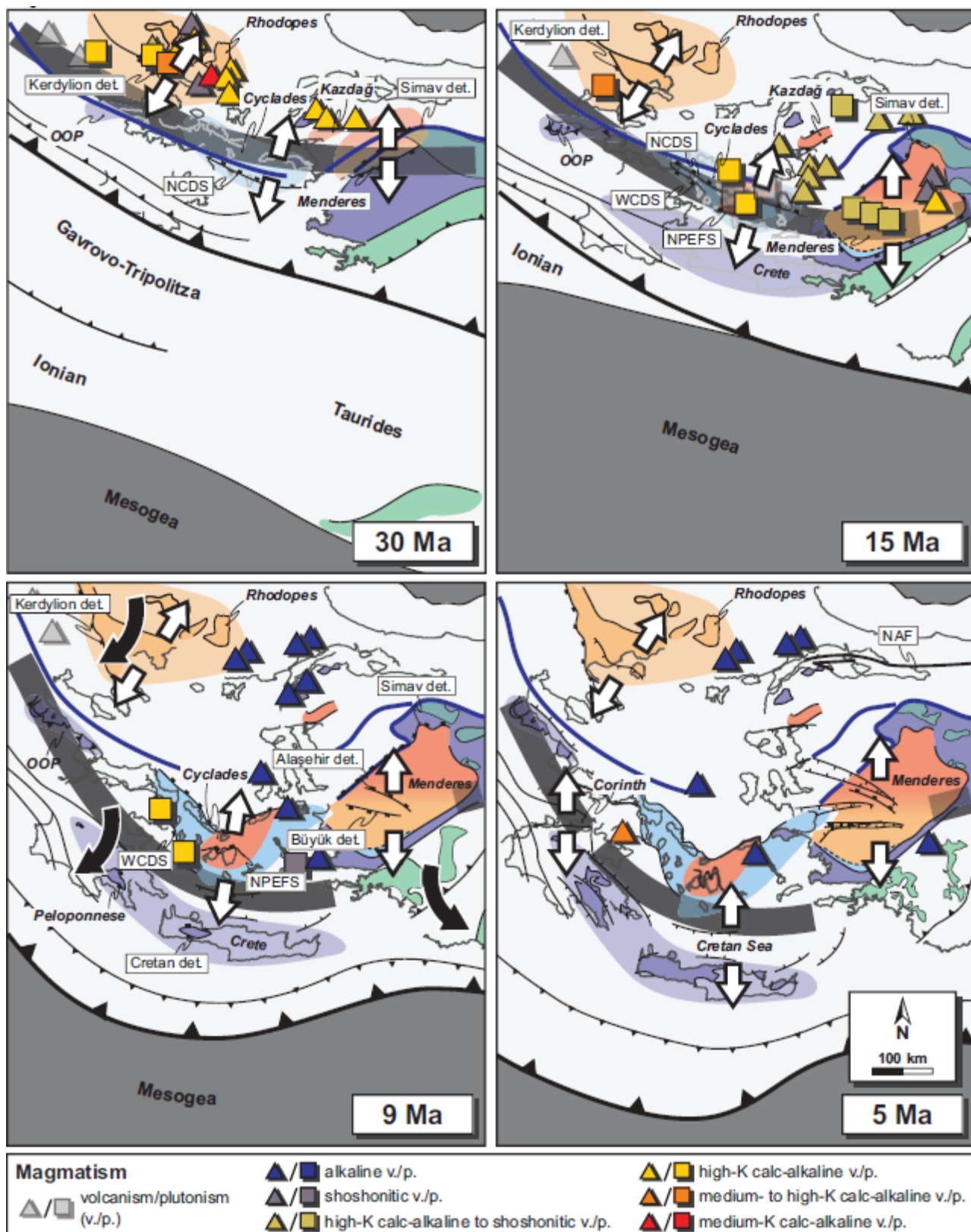


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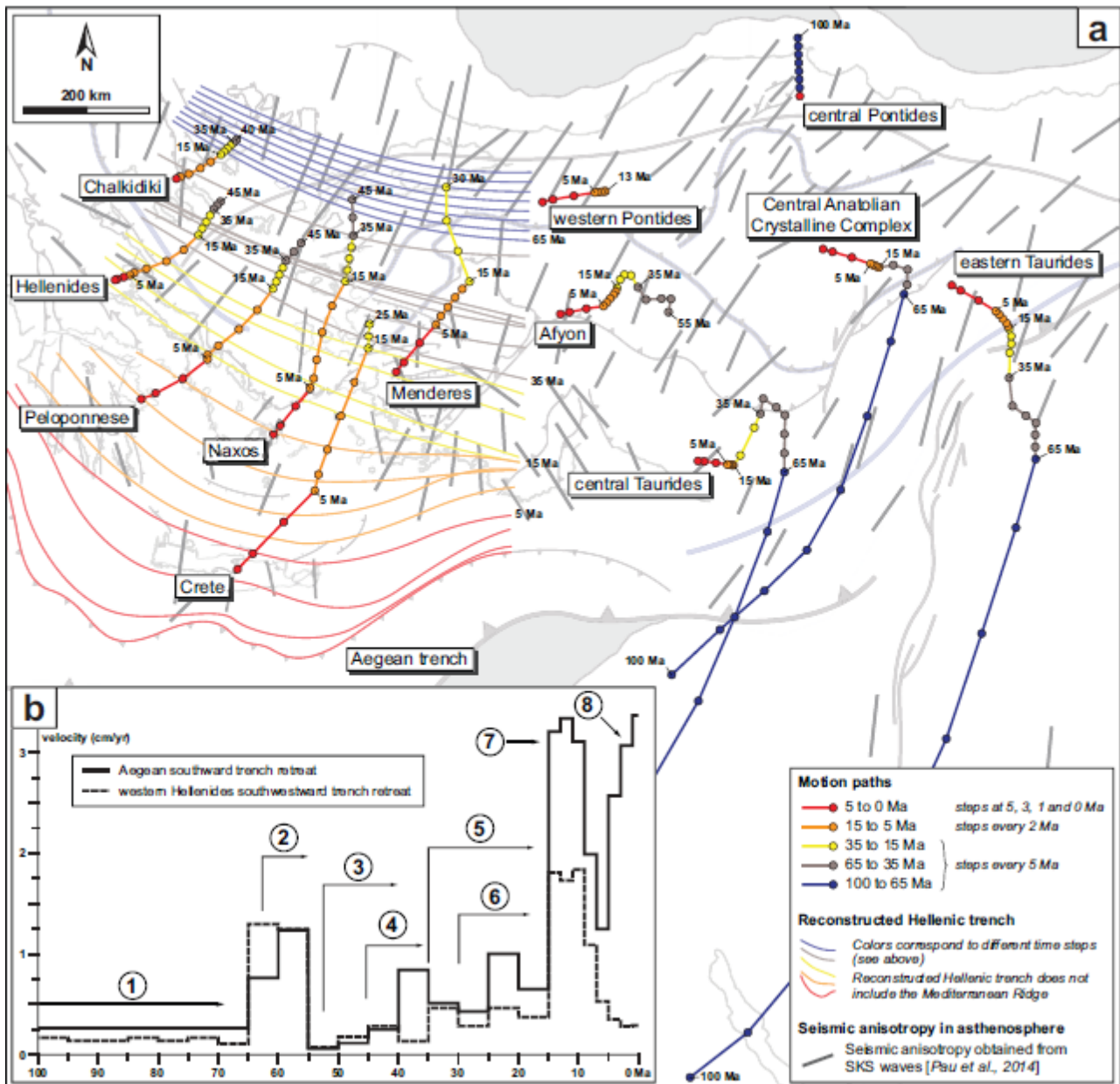


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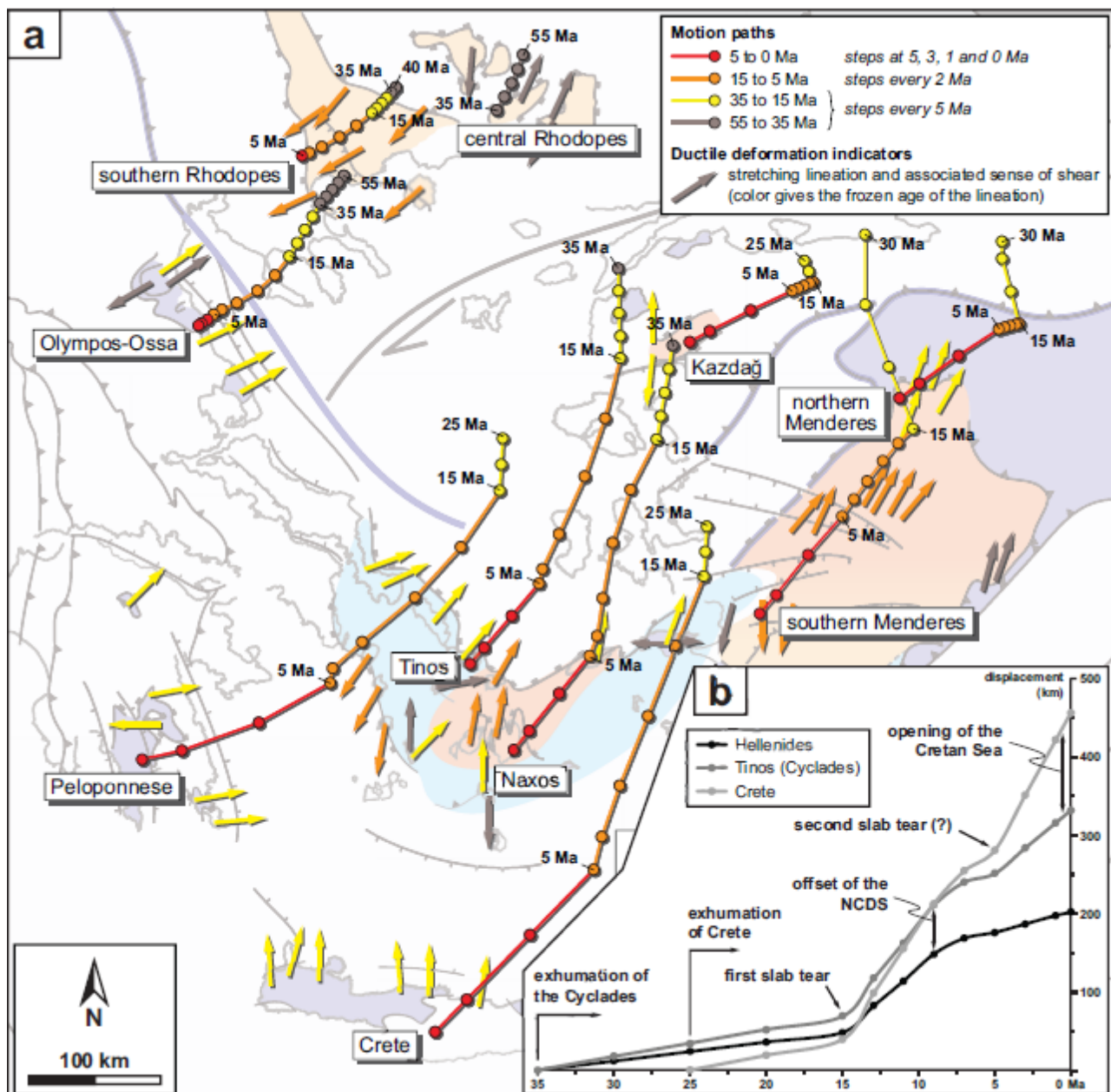


Figure 13

Highlights

- New kinematic reconstructions of the eastern Mediterranean since the late Cretaceous
- Formation and exhumation of metamorphic units are considered in reconstructions
- 512 volcanic and plutonic centers are integrated in reconstructions
- Reconstructed kinematics fits MCCs exhumation kinematics
- Reconstructions allow discussing relations between crustal and mantle processes