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ORIGINAL PAPER

Cenozoic geodynamic evolution of the Aegean

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Abstract The Aegean region is a concentrate of the main geodynamic processes that shaped the Mediterranean region: oceanic and continental subduction, mountain building, high-pressure and low-temperature metamorphism, backarc extension, post-orogenic collapse, metamorphic core complexes, gneiss domes are the ingredients of a complex evolution that started at the end of the Cretaceous with the closure of the Tethyan ocean along the Vardar suture zone. Using available plate kinematic, geophysical, petrological and structural data, we present a synthetic tectonic map of the whole region encompassing the Balkans, Western Turkey, the Aegean Sea, the Hellenic Arc, the Mediterranean Ridge and continental Greece and we build a lithospheric-scale N-S cross-section from Crete to the Rhodope massif. We then describe the tectonic evolution of this cross-section with a series of reconstructions from ∼70 Ma to the Present. We follow on the hypothesis that a single subduction has been active throughout most of the Mesozoic and the entire Cenozoic, and we show that the geological record is compatible with this hypothesis. The reconstructions show that continental subduction (Apulian and Pelagonian continental blocks) did not induce slab break-off in this case. Using this evolution, we discuss the mechanisms leading to the exhumation of metamorphic rocks and the subsequent formation of extensional metamorphic domes in the backarc region during slab retreat. The tectonic histories of the two regions showing large-scale extension, the Rhodope and the Cyclades are then compared. The respective contributions to slab retreat, post-orogenic extension and lower crust partial melting of changes in kinematic boundary conditions and in nature of subducting material, from continental to oceanic, are discussed.

Keywords Mediterranean · Aegean · Exhumation · Subduction · Rollback · Extension

Introduction

The exhumation of HP–LT metamorphic rocks, despite recent efforts in many orogens, still remains partly enigmatic. Several mechanisms have been identified that can bring deep seated rocks to the surface: circulation in the subduction channel (Cloos and Shreve 1988; Gerya et al. 2002; Jolivet et al. 2003; de Sigoyer et al. 2004; Yamato et al. 2007), underplating and accretionary wedge tectonics (Platt 1986, 1993; Ring and Layer 2003), post-orogenic extension (Wernicke 1981, 1992; Lister et al. 1984; Gautier and Brun 1994a; Brun and Sokoutis 2007), erosion (Ring et al. 1999a). Each of these ingredients can account for a part of the history of a given blueschists or eclogite massif. However, their respective contributions are difficult to assess because the kinematic boundary conditions are often only partly constrained. Cenozoic Mediterranean (Fig. 1) orogens are located in a still active environment with reasonably well known boundary conditions (Dewey et al. 1989; Rosenbaum et al. 2002b). Among upper plate environments in convergent zones, the Aegean region has recently received much attention because it contains in a relatively restricted region many of the ingredients of

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subduction zones tectonics: exhumation of HP–LT metamorphic rocks in the subduction channel, formation of metamorphic core complexes in a backarc environment, and interaction between intracontinental large-scale strike–slip faults (North Anatolian Fault), active rifting in the backarc region (Corinth Rift) and the trench. Large-scale geophysical investigations give a rather precise idea of the 3D structure at the scale of the crust and mantle. Besides, the volcanic and plutonic record has been described in great detail from the Late Cretaceous to the present (Fytikas et al. 1984; Pe-Piper and Piper 2002, 2006, 2007). Based on a compilation of recently acquired data, this paper presents a series of lithospheric-scale 2D reconstructions along a N-S cross-section from the Rhodope massif to the Hellenic trench and Mediterranean Ridge as well as paleogeographic maps, spanning a 70-Ma-long period from the Late Cretaceous to the present. A synthetic tectonic map from the Balkans to the Mediterranean Ridge thrust front is also presented (Fig. 2). The reconstructions suggest a continuous northward subduction of the African plate below the southern margin of Eurasia from the Late Cretaceous suture of the Tethys ocean to the Present situation and a three-step evolution for HP–LT metamorphic rocks: (1) subduction with the sinking plate, (2) first exhumation within the subduction channel and (3) post-orogenic exhumation below low-angle normal faults in extensional metamorphic domes. A recent compilation of geological data and a series of reconstructions spanning a longer time range from the Paleozoic to the present were recently proposed by Papanikolaou et al. (2004). We instead focus our paper on the Cenozoic period and on the topic of the exhumation of HP–LT metamorphic rocks. Our reconstructions are partly based upon the hypothesis that a single subduction zone has been active through most of the Mesozoic and the entire Cenozoic as suggested by available tomographic models (Spakman et al. 1993; Bijwaard et al. 1998; Wortel and Spakman 2000). This situation implies an efficient delamination and a decoupling between the lithospheric mantle and lower crust below the main decollement and the upper crust above it (Jolivet et al. 2003). The same hypothesis was used in recent reconstructions by van Hinsbergen et al. (2005a). The present paper follows on this hypothesis and presents more detailed reconstructions where the geometry of the main geological structures, their kinematics and P–T–t evolution are taken into account. We show that the geological record is compatible with the hypothesis of a single subduction and that the subduction of continental blocks (Apulian and Pelagonian) has not induced slab break-off in this case.

Fig. 2 Compiled tectonic map of the Aegean region, Menderes massif, Rhodope massif and the Balkan, after (Creutzburg 1977; Bonneau 1982, 1984; Lyon-Caen et al. 1988; Burg et al. 1990; Armijo et al. 1992; Burg et al. 1995, 1996; Tzankov et al. 1996; Collins and Robertson 1997; Armijo et al. 1999; Collins and Robertson 1999; Okay and Tüysüz 1999; Okay and Satir 2000; Koukouvelas and Aydin 2002; Jolivet et al. 2004b; Papanikolaou et al. 2004; Chamot-Rooke et al. 2005) and the Geological map of Greece (IGME). Moho depth is after Tirel et al. (2004a) ▶

Geological and geodynamic setting

The Aegean Sea (Figs. 2, 3) is one of the Mediterranean Cenozoic backarc basins. The present-day Aegean Sea

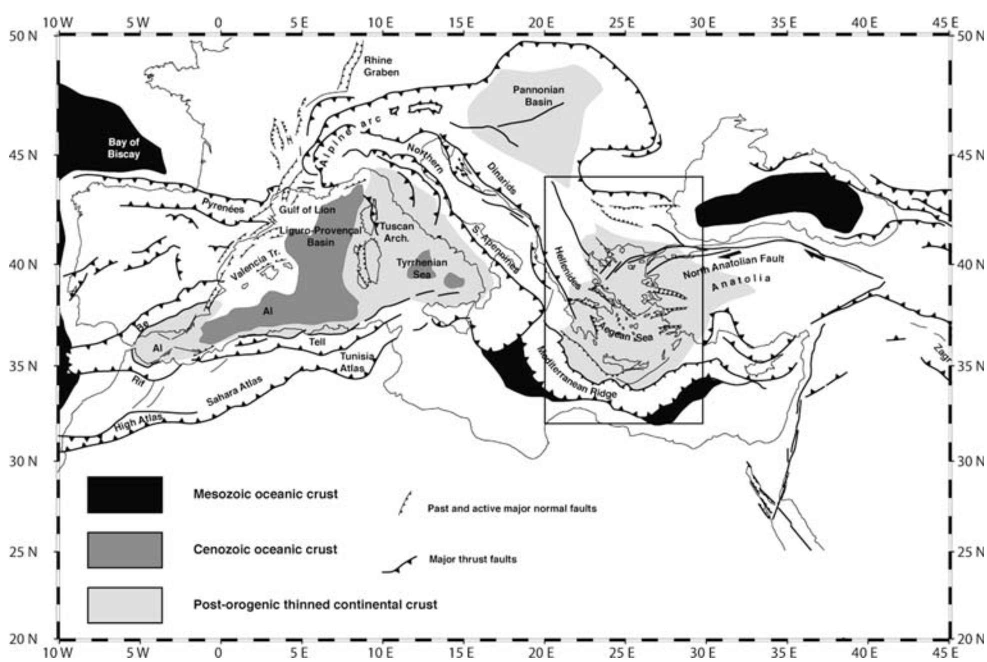
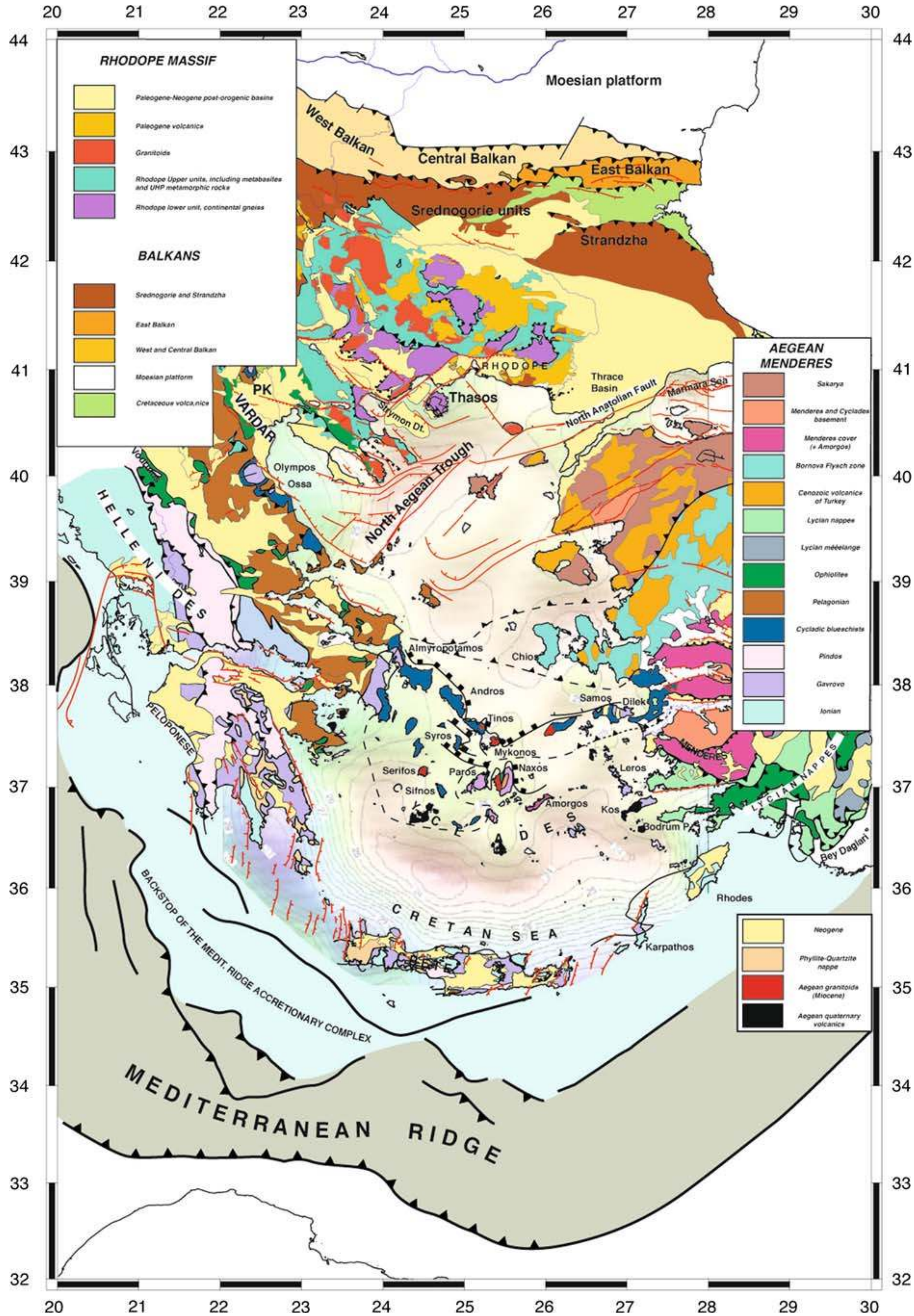


Fig. 1 Tectonic map of the Mediterranean region and the location of the studied area. The main compressional and strike–slip features are highlighted as well as the major extensional basins


RHODOPE MASSIF
Paleogene-Neogene post-orogenic basins
Paleogene volcanics
Granitoids
Rhodope Upper units, including metabasites and UHP metamorphic rocks
Rhodope lower unit, continental gneiss
BALKANS
Srednogorie and Strandzha
East Balkan
West and Central Balkan
Moesian platform
Cretaceous volca,nics
AEGEAN MENDERES
Sakarya
Menderes and Cyclades basement
Menderes cover (+ Amorgos)
Bornova Flysch zone
Cenozoic volcanics of Turkey
Lycian nappes
Lycian méelange
Ophiolites
Pelagonian
Cycladic blueschists
Pindos
Gavrovo
Ionian
Neogene
Phyllite-Quartzite nappe
Aegean granitoids (Miocene)
Aegean quaternary volcanics
Moesian platform
West Balkan
Central Balkan
East Balkan
Srednogorie units
Strandzha
RHODOPE
Thrace Basin
Marmara Sea
North Anatolian Fault
Thasos
Strymon Dt.
PK
VARDAR
Olympos
Ossa
North Aegean Trough
HELLENIDES
PELOPONESE
Almyropotamos
Chios
Andros
Samos
Dilek
Tinos
Syros
Mykonos
Naxos
Serifos
Paros
Sifnos
Leros
Amorgos
Kos
Bodrum P.
MENDERES
LYCIAN NAPPES
Bey Daglari
Rhodes
CRETAN SEA
Karpathos
BACKSTOP OF THE MEDIT. RIDGE ACCRETIONARY COMPLEX
MEDITERRANEAN RIDGE
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started to form at least in the Oligocene times above the retreating African slab now still sinking below Crete and the Mediterranean Ridge (Le Pichon and Angelier 1981a; Gautier et al. 1999; Jolivet and Faccenna 2000). But marine conditions are recognized in the Rhodope as soon as the Eocene in the Rhodope massif suggesting earlier extension (Burchfiel et al. 2003; Brun and Sokoutis 2007). The extended domain is superposed to a collapsed Early Cenozoic mountain belts, the Rhodope, the Hellenides and their eastern extension in the Taurides. Gravimetry and seismic profiles show a rather flat Moho around 25–26 km deep below most of the Aegean Sea (Vigner 2002; Tirel et al. 2004b), suggesting a weak lower crust during part of the extensional process as in the Basin and Range (Wernicke 1992). Part of the Aegean Sea and some of the metamorphic complexes like the island of Andros rests upon a thicker crust (30 km) transitional between the thicker crust of the Hellenides and the thin crust of the Aegean Sea (Karagianni et al. 2005). Other exceptions such as the North Aegean Trough in the north and the Cretan Sea in the south with thinner crust should be, however, emphasized. The crust in the Cretan Sea has been thinned down to ~15 km (Bohnhoff et al. 2001). The continental crust is thicker at the periphery, in Crete (30 km), in western Turkey and in continental Greece and in the Rhodope massif (up to 45 km) (Makris 1978; Makris et al. 2001). The thicker crust of the Rhodope is nevertheless associated with large-scale extensional structures and core complexes (Dinter and Royden 1993; Sokoutis et al. 1993).

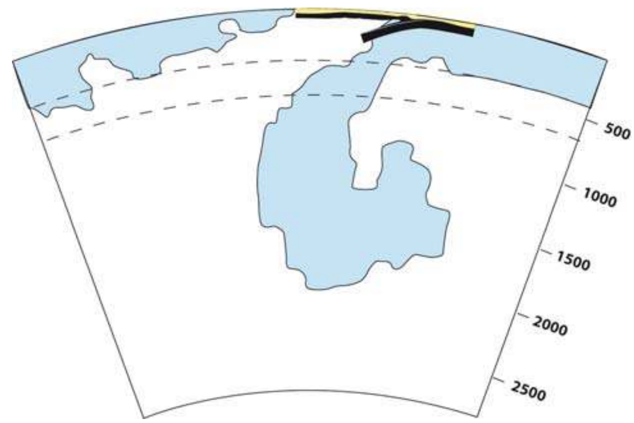


Fig. 4 A comparison of the lithospheric section shown in Fig. 3 with the geometry of the high seismic velocity anomaly below the Aegean region, after the tomographic model of Bijwaard et al. (1998)

Aegean slab

The Aegean slab is well imaged by seismic tomography and it can be followed down to the lower mantle (Spakman et al. 1988, 1993; Spakman 1990; Bijwaard et al. 1998; Wortel and Spakman 2000; Piromallo and Morelli 2003) (Fig. 4). This is a major difference with western Mediterranean subductions, where the slab has not yet sunk across the upper–lower mantle boundary (Faccenna et al. 2003). The long-lived Hellenic subduction is more matured and the slab is probably partly anchored in the lower mantle. Tomographic models also suggest that the Hellenic slab does not extend far

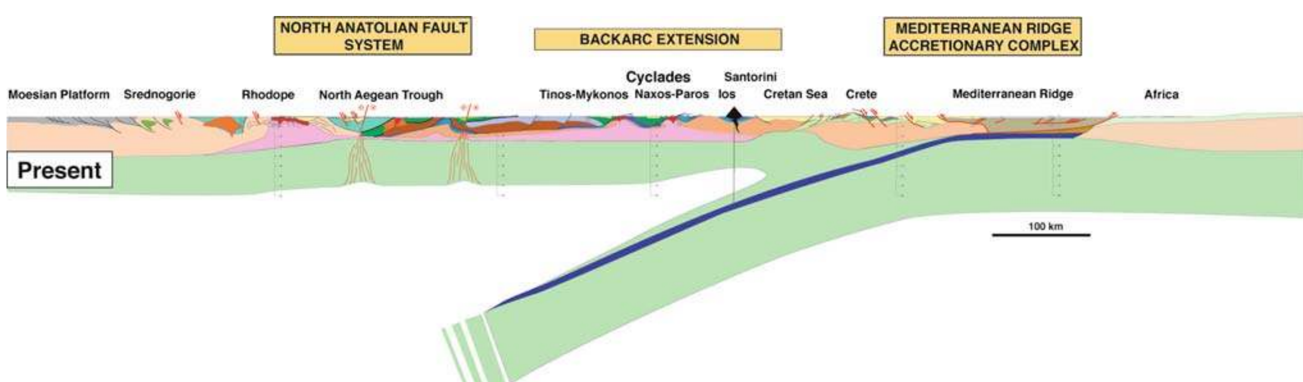


Fig. 3 Compiled lithospheric-scale cross-section of the Aegean region from the Balkan foreland to the African passive margin. This section is based upon surface geological observations as well as geophysical data: Moho depth (Papanikolaou et al. 2004 and references therein; Bohnhoff et al. 2001; Kuhlemann et al. 2004; Tirel et al. 2004b), geometry of the upper plate–lower plate contact in the subduction zone (Li et al. 2003). The base of the lithosphere has been set to approximately 100 km depth. This cross-section is strongly inspired by the cross-section published by the Transmed Team (Papanikolaou et al. 2004) for the northernmost part (Moesian platform and Srednogorie). The entire section has been otherwise totally reinterpreted based on the work of Ricou et al. (1998), Burg et al. (1996), Bonev et al. (2006) and Brun and Sokoutis (2007) for the Rhodope, on Gautier and Brun (1994a, b), Jolivet and Patriat (1999), Jolivet et al. (1996, 2004a, b) for the Cyclades and Crete, and Chaumillon et al. (1996), Kopf et al. (2003) and the DOTMED project (Chamot-Rooke et al. 2005) for the Mediterranean Ridge. The main crucial difference with the work of Papanikolaou et al. (2004) is that, we emphasize here the imprint of crustal extension during the formation of the Aegean Sea during the Oligocene and Miocene and the Eocene to present evolution of the Rhodope massif

eastward and that a tear is present below western Turkey. This tear though is to have allowed the fast southward retreat of the Hellenic slab during the Miocene (de Boorder et al. 1998; Govers and Wortel 2005). One of the main observations one can make from tomographic models is that a single slab, more then 1,500-km-long, is present. This observation suggests that a single subduction has been active throughout the Cenozoic in the area and that different paleogeographic domains (Vardar ocean, Pelagonian microcontinent, Pindos ocean, Apulian continent and the eastern Mediterranean oceanic domain) carried by the same plate were successively subducted and partly accreted to the southern margin of Eurasia (Jolivet et al. 2003; Van Hinsbergen et al. 2005a).

Active tectonics

The active kinematics is driven by two partly independent phenomenons: the retreat of the subducting African plate, and the westward motion of the Anatolian plate along the North Anatolian Fault (Le Pichon and Angelier 1981b; Jackson 1994; Jolivet 2001; Armijo et al. 2003; Flerit et al. 2004). Active extension is restricted to the periphery of the Aegean domain with two regions where it is faster: the Corinth rift and western Turkey. Normal faults are observed as far north as Sofia in Bulgaria or the northern part of the Balkans (Tzankov et al. 1996; Meyer et al. 2002). These northern normal faults accommodate an extension much slower than the 1.5 cm/year recorded across the Corinth Rift (Armijo et al. 1996; Briole et al. 2000). Much of the active deformation is furthermore controlled by the penetration of the dextral North Anatolian Fault in the Aegean Sea, an important event constrained around some 5 Ma (Armijo et al. 1999, 2003). Most of extension across the Corinth Rift has occurred since that time and most likely after 3–2 Ma (Armijo et al. 1996; Rohais et al. 2007).

Magmatism

Magmatic rocks were formed during subduction and slab retreat with a complex geographical pattern since the Late Cretaceous (Fytikas et al. 1984) (Figs. 5, 6). After the formation of a large magmatic arc in the Balkan (Srednogorie, Strandzha) in the Late Cretaceous (Jankovic 1997; Von Quadt et al. 2005), magmatism migrated to the Rhodope massif in the Paleocene and Eocene with the emplacement of large amounts of granitoids (Jones et al. 1992; Marchev et al. 2004, 2005). From 35–30 Ma to the present, the magmatic arc then migrated regularly toward the south to reach its present-day position (Fytikas et al. 1984; Pe-Piper and Piper 2002, 2006, 2007) (Fig. 6). A thorough synthesis of the geochemistry and petrology of Aegean magmatic rocks can be found in Pe-Piper and Piper (2006). The geochemistry and petrogenesis of the volcanic rocks suggest a heating of a subcontinental mantle and lower crust. The main mantle magma source was contaminated by a crustal contribution during extension (Pe-Piper and Piper 2006; Altherr and Siebel 2002). The distribution of volcanic rock types shows an unusually wide range of chemistry and petrology for a backarc environment, suggesting a range of geodynamic environments within the backarc region. This variety of situations led to the formation of variable magma-types from classical calcalkaline to alkaline or shoshonitic. Although the overall picture is compatible with subduction and arc and backarc environment, the presence of both oceanic and continental subduction, the existence of large-scale strike–slip faults such as the North Anatolian Fault, the roll-back of the Aegean slab and a possible slab tear below western Turkey (de Boorder et al. 1998) significantly complicate the panorama. In Western Turkey, an intense magmatic activity in the Miocene could be related to slab tearing. The region situated near the NW tip of Evia shows recent volcanic activity that may be linked with the propagation of the NAF. Other localized volcanic activity can also be controlled by strike–slip faults. Despite the variability of rock-type, the main cause of magmatism is subduction and the southward migration is quite constant through time.

Extension and block rotations

Extension in the Aegean Sea and subsequent localization of the North Anatolian Fault postdate the formation of the Hellenic and Dinaric chains. The pre-extension structure is partly preserved in continental Greece that shows the complete stack of nappes (Aubouin 1959; Godfriaux 1962; Jacobshagen et al. 1978; Doutsos et al. 1993; Papanikolaou et al. 2004; van Hinsbergen et al. 2005c). All other regions were strongly affected by post-orogenic extension, especially the Cyclades. The deep parts of the orogen with high-pressure metamorphic massives crop out in those regions severely extended. Part of the exhumation process was achieved during the Aegean extension (post-orogenic extension) and part of it before, during the formation of the Hellenides (syn-orogenic exhumation) (Jolivet et al. 1994b; Avigad et al. 1997; Jolivet et al. 2003, 2004a). In a first approach along the transect studied in this paper, syn-orogenic exhumation took place within the subduction zone itself (subduction channel tectonics), during the Eocene in the Cyclades, during the Oligocene and Early Miocene in Crete and Peloponnese, whereas post-orogenic extension and the final exhumation took place in the backarc region in a warmer regime. Previous studies showed that syn-orogenic exhumation in Crete and post-

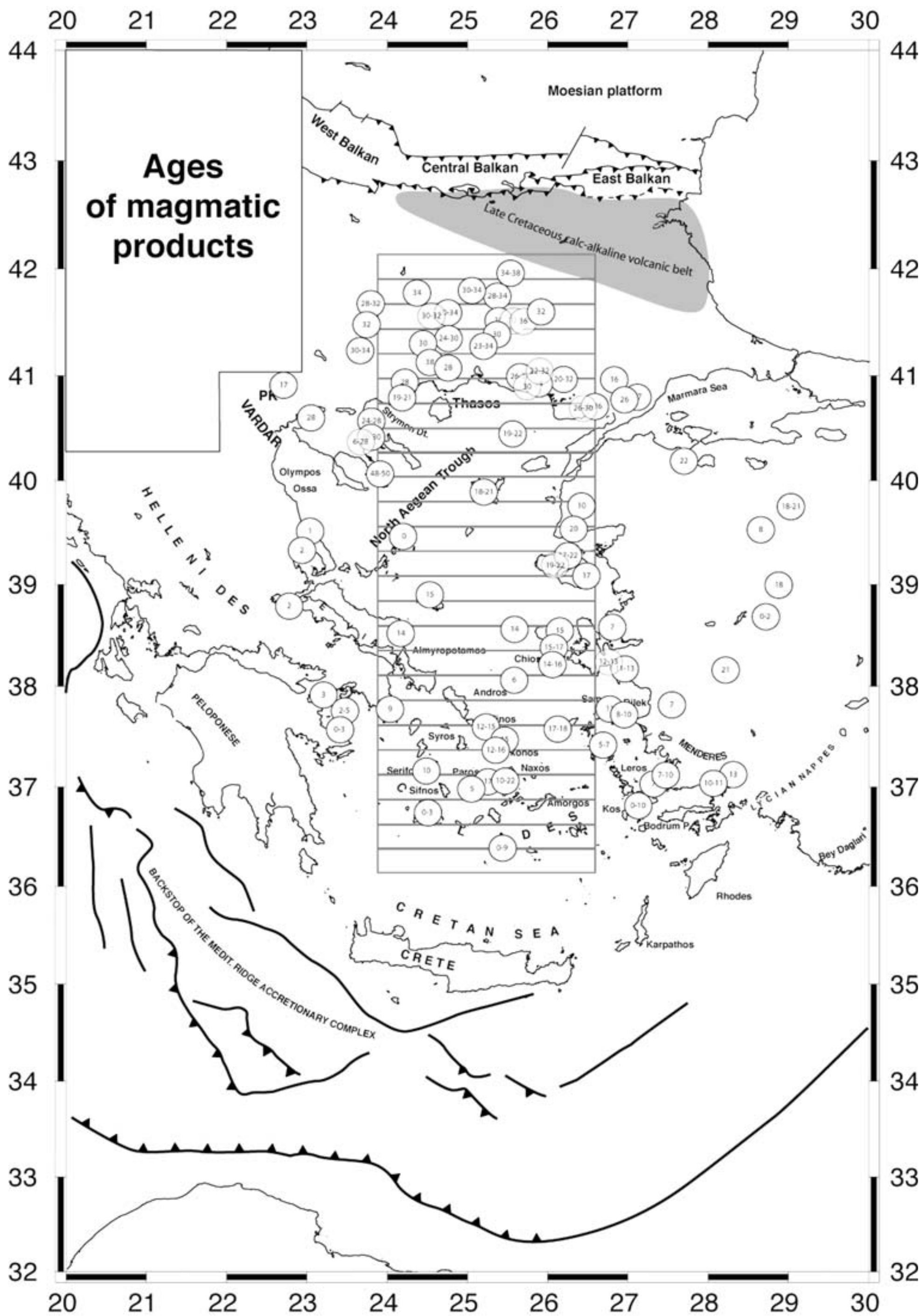


Fig. 5 Location and ages of the main magmatic events in the Aegean region. Ages are given in My and the complete time range of magmatic activity is given at each location. Data are from Pe-Piper and Piper (2002, 2006, 2007) and von Quadt et al. (2005). The grid shows the data considered to construct the diagram shown in Fig. 6

orogenic exhumation in the Cyclades were active during the same time span during the southward retreat of the African slab. Further north, Eocene extension in the Rhodope took place while the Cycladic Blueschists were being subducted and exhumed to the south (Krohe and Mposkos 2002; Brun and Faccenna 2007; Brun and Sokoutis 2007).

Paleomagnetic rotations are recorded in the whole Aegean domain (Kissel and Laj 1988; Kissel et al. 1995, 2002; Haubold et al. 1997; Dimitriadis et al. 1998; Platzman et al. 1998; Duermeijer et al. 2000). They are mostly clockwise (CW) rotations in the Hellenides and Rhodope and a more complex pattern is seen in the Cyclades and western Turkey. A recent reinvestigation in continental Greece and the Peloponnese confirms this general pattern and brings a more precise timing (van Hinsbergen et al. 2005b). The whole of western Greece including the external Albanides has rotated CW by some 50°, 40°

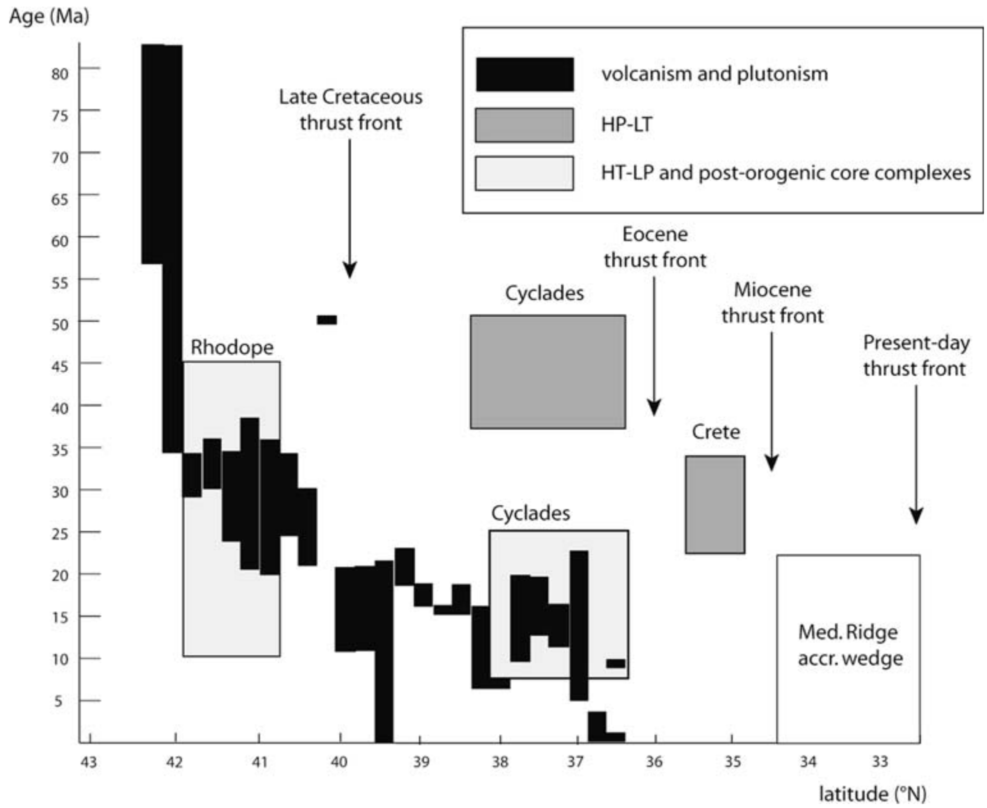


Fig. 6 Latitude versus age diagram of magmatic events in the Aegean region. Data are taken from Fig. 5 and the time range is taken in the grid shown on Fig. 5. The age and position of the main tectono-metamorphic events are also shown (see text for references)

between 15 and 8 Ma and the last 10° after 4 Ma. In the Rhodope massif clockwise rotations of some 10–20° are recorded after the Mid-Oligocene and they continued after the Early Miocene (Haubold et al. 1997; Dimitriadis et al. 1998). Walcott and White (1998) compiled the directions of stretching lineations throughout the Aegean domain and compared them with the pattern of paleomagnetic rotations. They suggested that, before rotation, the strike of extension was N023°E in the western and eastern Aegean domains during the Late Oligocene and Early Miocene. Then the area was divided into two more or less rigid blocks, among which the West Aegean domain limited to the southeast by the Mid-Cycladic Lineament that is a partly extensional and partly strike–slip fault zone. This fault zone would have ceased its activity in the Late Miocene. The existence of the Mid-Cycladic lineament is, however, not totally ascertained. It is based on the observation that block rotations were not coherent on either sides but its exact location is not known. Little information is available at sea where the fault zone should be present. Recent estimations of the crustal thickness in the Aegean region (Tirel et al. 2004b) do not seem to show a contrast across the postulated lineament that remains to be studied.

Sedimentary basins

Sedimentary basins were formed during this evolution. On land, the two main basins are the Meso-Hellenic basin that is a piggy-back-type basin, where up to 5 km of Eocene to Miocene sediments were deposited on top of the Pelagonian block during the subduction of the Pindos oceanic domain (Ferrière et al. 2004; Vamvaka et al. 2006), and the Thrace Basin, where up to 9 km of sediments were deposited during the same period (Huvaz et al. 2007; Siyako and Huvaz 2007). Offshore Aegean basins are rather poorly known. Thick accumulations are observed in the Thermaikos Gulf, the Orfanos Basin, the North Aegean Trough and the Cretan Sea (Martin 1987; Mascle and Martin 1990). In those regions, a large part of the sedimentary record is quite recent, Late Miocene to Quaternary. Earlier deposits are poorly known and most of the earliest sediments are of fluvial origin and were deposited during the Early and Middle Miocene (Sanchez-Gomez et al. 2002; Kuhlemann et al. 2004). However, the first known post-orogenic sediments are sometimes marine and of Aquitanian age in the Cyclades (Angelier et al. 1978) and Middle-Late Eocene in the Rhodope massif (Burchfiel et al. 2003).

A regional-scale cross-section

The present-day cross-section of Fig. 3 is based upon geophysical data for the Moho depth (Bohnhoff et al. 2001; Kuhlemann et al. 2004; Papanikolaou et al. 2004; Tirel et al. 2004b), and the geometry of the upper plate–lower

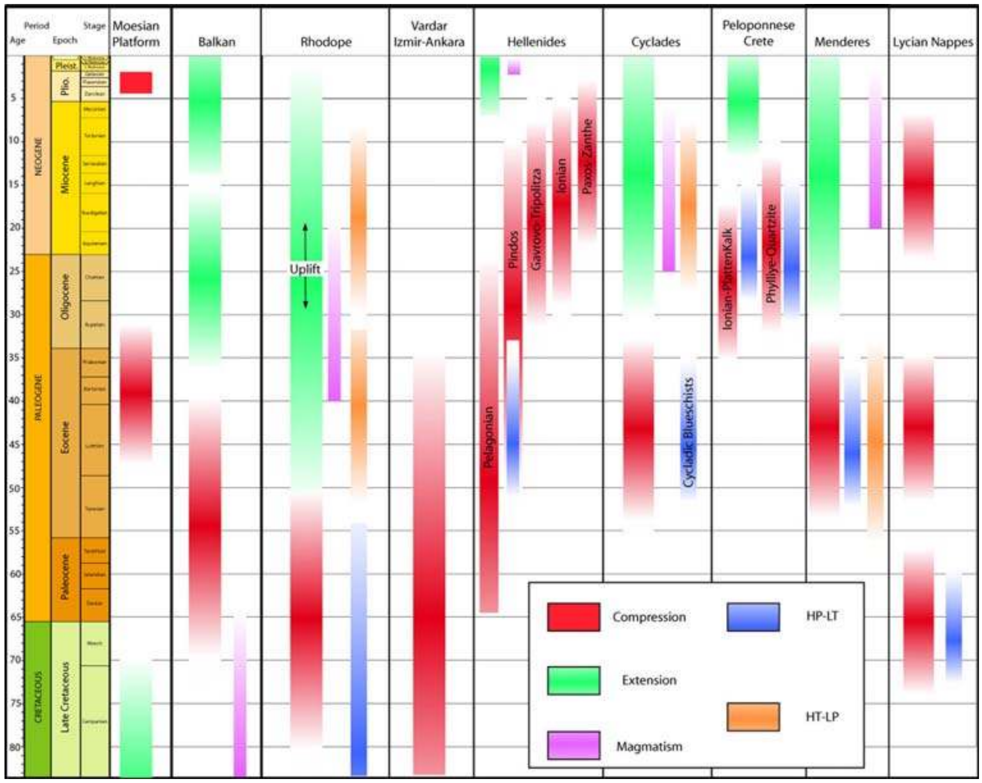


Fig. 7 The main tectonic, magmatic and metamorphic events in the Aegean region (see text for references)

plate contact in the subduction zone (Li et al. 2003). The base of the lithosphere has been set to approximately 100 km. For the northernmost part (Moesian platform and Srednogorie) The cross-section is strongly inspired by the cross-section published by the Transmed Team (Papanikolaou et al. 2004). The entire section has been otherwise totally reinterpreted based on the work of Ricou et al. (1998), Burg et al. (1996), Bonev et al. (2006) and Brun and Sokoutis (2007) for the Rhodope, on Gautier and Brun (1994a, b), Jolivet and Patriat (1999), Jolivet et al. (1996, 2004a, b) for the Cyclades and Crete, and Chaumillon et al. (1996), Kopf et al. (2003) and the DOTMED project (Chamot-Rooke et al. 2005) for the Mediterranean Ridge. The main crucial difference with the work of Papanikolaou et al. (2004) is that, we emphasize here the imprint of crustal extension during the formation of the Aegean Sea during the Oligocene and Miocene and the Eocene to present evolution of the Rhodope massif. In the following sections, the main tectonic, magmatic and metamorphic events are summarized from N to S (Fig. 7), following the tectonic map of Fig. 2.

The Balkanides

The Moesian Platform is the foreland of the Balkanides (Papanikolaou et al. 2004). It consists of a Paleozoic basement overlain by 4–5 km-thick shallow marine deposits of Mesozoic age, and finally by south-dipping Paleogene and Neogene clastics deposited in a foredeep. The platform partly records the succession of tectonic stages that affected the Balkanides. A Late Cretaceous phase of extension can be related to the opening of the Black Sea and its western equivalent north of the Strandzha massif (Bergerat et al. 1997). Then a Late-Middle Eocene compression is contemporaneous of the main phase of thrusting in the Balkanides. The last compressional events are mainly Oligocene in age (Doglioni et al. 1996). South of the Moesian Platform, the alpine belt is formed of a stack of north-verging thrust-sheets (Papanikolaou et al. 2004). The Balkan Zone overthrusts the Kamchya foredeep. It contains deeper water Mesozoic series overlain by an early Tertiary flysch and only minor amounts of Late Cretaceous volcanic products. The main contraction phase culminated in the Mid-Eocene and started in the Late Cretaceous. The Srednogorie and Strandzha zones form a portion of the Late Cretaceous volcanic and plutonic belt that formed above the subduction of the Vardar ocean. Large amounts of volcanic products were produced during this period. This zone is also characterized by intense Cu–Au mineralizations (Von Quadt et al. 2005). A Late Jurassic–Early Cretaceous shortening phase is recorded before the formation of a backarc setting (e.g., Black Sea rifting) and the emplacement of the magmatic arc (Okay et al. 2001a). The nappe stack is finally reworked by a series of grabens that formed from the Late Miocene to the Present (Sub-Balkan graben system) (Tzankov et al. 1996). Together with the Sofia graben and the steep normal faults cutting the Rhodope massif further south (Meyer et al. 2007), they represent the northernmost evidence of the Aegean extension.

The Rhodope massif

Originally considered a paleozoic rigid core, the Rhodope massif is now understood as a part of the Alpine orogen (Zagorchev 1998). Alpine nappes have now been recognized and a complex evolution involving nappe stacking and crustal thickening followed by post-orogenic extension has been described (Ivanov et al. 1985; Burg et al. 1990, 1996; Dinter and Royden 1993; Dinter 1998; Ricou et al. 1998; Krohe and Mposkos 2002; Bonev et al. 2006; Brun and Sokoutis 2007). Figures 2 and 3 show a simplified structure. Several stacked units have been recognized whose present-day geometrical arrangement strongly results from post-orogenic extension and, in particular, the formation of the large-scale metamorphic core complex of the Southern Rhodope (Dinter and Royden 1993; Sokoutis et al. 1993; Bonev et al. 2006; Brun and Sokoutis 2007). Despite core complex-type extension, the Moho is still at 40–45 km there and the transition with the Aegean Sea is quite abrupt.

The upper unit is made of the Vertiskos complex of the Circum Rhodope belt to the southwest of the core complex and of the Kimi complex to the north (Krohe and Mposkos 2002). It is composed of a paleozoic basement and a mixture of meta-ophiolitic rocks and metasediments displaying HP–LT metamorphic parageneses and several occurrence of UHP conditions with metamorphic coesite and diamond (Mposkos and Kostopoulos 2001; Perraki et al. 2006). The HP (Liati and Mposkos 1990; Liati and Seidel 1996) and UHP metamorphism (Mposkos and Kostopoulos 2001), dated at 140 Ma (Liati and Gebauer 1999) and 184 Ma (Reischmann and Kostopoulos 2002), respectively, probably contemporaneous of the Lower Cretaceous compression recognized in the Balkan. This upper unit, characterized by the occurrence of UHP rocks, consists in the envelope of the Southern Rhodope Core Complex (Brun and Sokoutis 2007) that is made of repetitions of orthogneisses, marbles, pelitic schists and amphibolites resulting from slicing and imbrication by thrusting prior to extension. The protoliths of orthogneisses correspond to plutonic units of Hercynian-type age, that clusters between 270 and 300 Ma throughout the whole Greek Rhodope (Turpaud and Reischmann 2003). The stratigraphic age of marbles and associated meta sediments and amphibolites is unknown, but most likely they represent the Tethyan sedimentary and volcanic series deposited on top of the Hercynian basement. The exhumation of this unit is related to the core complex development with a SW-NE direction of stretching at regional scale. Extension driven exhumation occurred between 50 and 27 Ma in the southwestern and northern borders of the core complex and continued up to 11 Ma in the central part (Wawrzenitz and Krohe 1998). As summarized by Brun and Sokoutis (Brun and Sokoutis 2007), SW-NE extension started in Mid-Eocene times in the Southern Rhodope core complex as well as in surrounding regions of the Rhodope (Burchfiel et al. 2003; Kounov et al. 2004; Bonev et al. 2006). These timing and kinematics constraints are those taken into account in Figs. 2 and 3.

The Hellenides

The *Vardar suture zone*, which results from the convergence between Apulia (Hellenides) and Eurasia (Rhodope) continuously throughout the Cenozoic, is a complex assemblage of a magmatic arc belonging to the Eurasian margin and one or two distinct ophiolitic basins and a platform unit (Godfriaux and Ricou 1991a; Ricou et al. 1998) thrust onto the Pelagonian domain during the Paleogene. The gross structure of the Hellenides corresponds to a pile made of six main thin thrust units. From top to base or from north to south they are the Pelagonian, Pindos, Gavrovo-Tripolitza, Phyllite-Quartzite and Ionian and pre-Apulian thrust units, respectively.

The *Pelagonian thrust unit* is made of a paleozoic basement with a Paleozoic and Mesozoic carbonate cover overlain by a Jurassic ophiolite obducted toward the end of the Jurassic (Aubouin 1959; Celet and Ferrière 1978; Jacobshagen et al. 1978; Bonneau 1982, 1984; Jacobshagen 1986; Walcott and White 1998) whose original width is estimated around 200 km (Van Hinsbergen et al. 2005a). Outliers of the Jurassic ophiolite are found to the west, the largest of which being the Vourinos ophiolite resting upon the Pindos massif. Outliers of the Pelagonian nappe are also found on top of the edifice in the Cyclades (Upper Cycladic nappe) and in Crete (Asteroussia nappe) (Bonneau 1973, 1984; Reinecke et al. 1982; Papanikolaou 1987). On top of the Pelagonian block, the Meso-Hellenic basin that accumulated some 5 km of Cenozoic deposits (Ferrière et al. 2004; Vamvaka et al. 2006) started its evolution during the late Eocene during the subduction of the Pindos domain as a forearc basin filled with deep turbidites and evolved into a piggy-back filled with shallow marine sediments during the subduction of the external Gavrovo-Tripolitza platform. The basin was then progressively uplifted during the Miocene.

The *Pindos thrust unit* is partly an oceanic-type sequence made of pelagic limestones and siliceous deposits from the Late Triassic to the Paleocene covered with an Eocene-Oligocene flysch (Brunn 1956; Aubouin 1959; Stampfli et al. 2003). The continental margins of this ocean are also parts of the Pindos unit or its lateral equivalents (Jones and Robertson 1990; Robertson et al. 1991). The sedimentary parts of the Pindos domain are stacked in the external Pindos nappe in the Peloponnese (Skourlis and Doutsos 2003). A paleogeographic equivalent of the Pindos unit is found below the internal zones in the Olympos, Ossa and Pelion windows as well as in Evia and the Cyclades where they make the Cycladic Blueschists (Godfriaux 1962; Blake et al. 1981; Bonneau and Kienast 1982; Ferrière 1982; Schermer 1990, 1993; Schermer et al. 1990) that rest upon a paleozoic basement diversely metamorphosed intruded by Triassic granitoids (Engel and Reischmann 1997; Reischmann 1997). The original width of the Pindos ocean is difficult to precisely determine and in published reconstructions it ranges from 300 to 500 km (see "Discussion" in Van Hinsbergen et al. 2005a).

The *Gavrovo-Tripolitza thrust unit* was underthrusted below the Pindos units in the Oligocene. It consists in a series of pelitic sediments and volcanics of Triassic age (Tyros beds) overlain by a platform sequence of Late Triassic to Eocene age followed by a Late Eocene–Early Oligocene flysch (Aubouin 1959; Jacobshagen 1986). It crops out in continental Greece, the Peloponnese, Kithira, Crete and Karpathos mostly to the west of or below the external Pindos unit but also within the Olympos, Ossa and Almyropotamos windows, below the Pelagonian units and the Cycladic Blueschists (Godfriaux 1962, 1965; Godfriaux and Pichon 1980; Godfriaux and Ricou 1991b), where it displays a HP–LT metamorphism (Godfriaux and Pichon 1980; Schermer 1990; Shaked et al. 2000). The youngest sediments in the internal Gavrovo zone are at most early Eocene as attested by the presence of Nummulites in the flysch resting on top of the platform sequence (Dubois and Bignot 1979; Godfriaux and Pichon 1980; Godfriaux and Ricou 1991b), thus giving a maximum age for the HP–LT metamorphism. In the Cyclades, lateral equivalents of the Gavrovo (Bonneau 1984; Jolivet et al. 2004b) consist in metasediments overlying the anatectic basement of Naxos and Paros and in carbonates in Amorgos. Correlations can also be made with the metasedimentary cover of the Menderes massif (Candan et al. 1997), but, instead of a single unit, the Gavrovo-Tripolitza carbonate platform likely forms a stack of several thrust units (Jolivet et al. 2004b).

The *Ionian (or Plattenkalk)* and the *Pre-Apulian platform* (or *Paxos zone)* form the two most external thrust units. Moreover, the *Phyllite-Quartzite unit* is sandwiched between the Ionian and Gavrovo-Tripolitza units in Crete, Kithira and the Peloponnese (Creutzburg 1977; Bonneau 1984). The Ionian unit is composed of shallow-water carbonate sequence and gypsum until the Mid-Liassic, followed by a pelagic sequence until the Eocene and topped by a Late Eocene to Miocene flysch (Aubouin 1959; Sotiropoulos et al. 2003). Underthrusting of the Ionian unit and the Gavrovo-Tripolitza under the Pindos unit occurred during the Oligocene (Sotiropoulos et al. 2003). The Ionian unit is generally unmetamorphosed except in the southern Peloponnese where greenschist facies conditions are encountered and in Crete where HP–LT parageneses are found with Fe–Mg-carpholite and aragonite as index minerals (Seidel et al. 1982; Thiébault and Triboulet 1983; Theye et al. 1992; Theye and Seidel 1993; Jolivet et al. 1996). The Pre-Apulian platform is a continuous platform sequence until the Miocene with a transition to a flysch in the late early Miocene (van Hinsbergen et al. 2005a, c). The Phyllite-Quartzite unit is made of a detrital sequence including metavolcanics and limestones of Late Carboniferous to Mid Triassic age (Creutzburg 1977; Greilling 1982; Krahl et al. 1983; Hall et al. 1984) with some basement units in eastern Crete and Kithira that provided paleozoic and precambrian ages (Romano et al. 2004; Xypolias et al. 2006). HP–LT parageneses give a peak of pressure around 16 kbar and 400°C in Crete and 16 kbar and 500°C in the Peloponnese with ages around 25 Ma, for a final exhumation around 15 Ma (Fig. 8) (Seidel et al. 1982; Thiébault and Triboulet 1983; Theye et al. 1992; Theye and Seidel 1993; Bassias and Triboulet 1994; Jolivet et al. 1996; Thomson et al. 1998; Trotet 2000; Trotet et al. 2006). In Northern Peloponnese, peak pressure decreases to 10 kbar near Kastania and likely to not more than 5–6 kbar near Zaroukla (Trotet et al. 2006; Jolivet et al. 2007). The contact between the Phyllite-Quartzite nappe and the overlying Gavrovo-Tripolitza nappe corresponds to a strong pressure gap that can reach 10 kbar in western Crete and that is interpreted as an extensional detachment active from the late Oligocene to the Middle Miocene with a clear normal sense of shear only observed in Crete and principal directions of stretching trending N–S in Crete and NE–SW in Peloponnese (Fassoulas et al. 1994; Jolivet et al. 1994b, 1996). A sedimentary basin was deposited above the detachment with an internal geometry compatible with the northward displacement of the hangingwall (van Hinsbergen and Meulenkamp 2006; Seidel et al. 2007).

Metamorphism and deformation in the Cyclades and Menderes

The Cyclades archipelago is underlain by a thinned crust with a very constant thickness around 26 km (Vigner 2002; Tirel et al. 2004b) with however a tendency to a thicker crust toward the periphery below some of the "cold" metamorphic core complexes [30 km below Andros island,

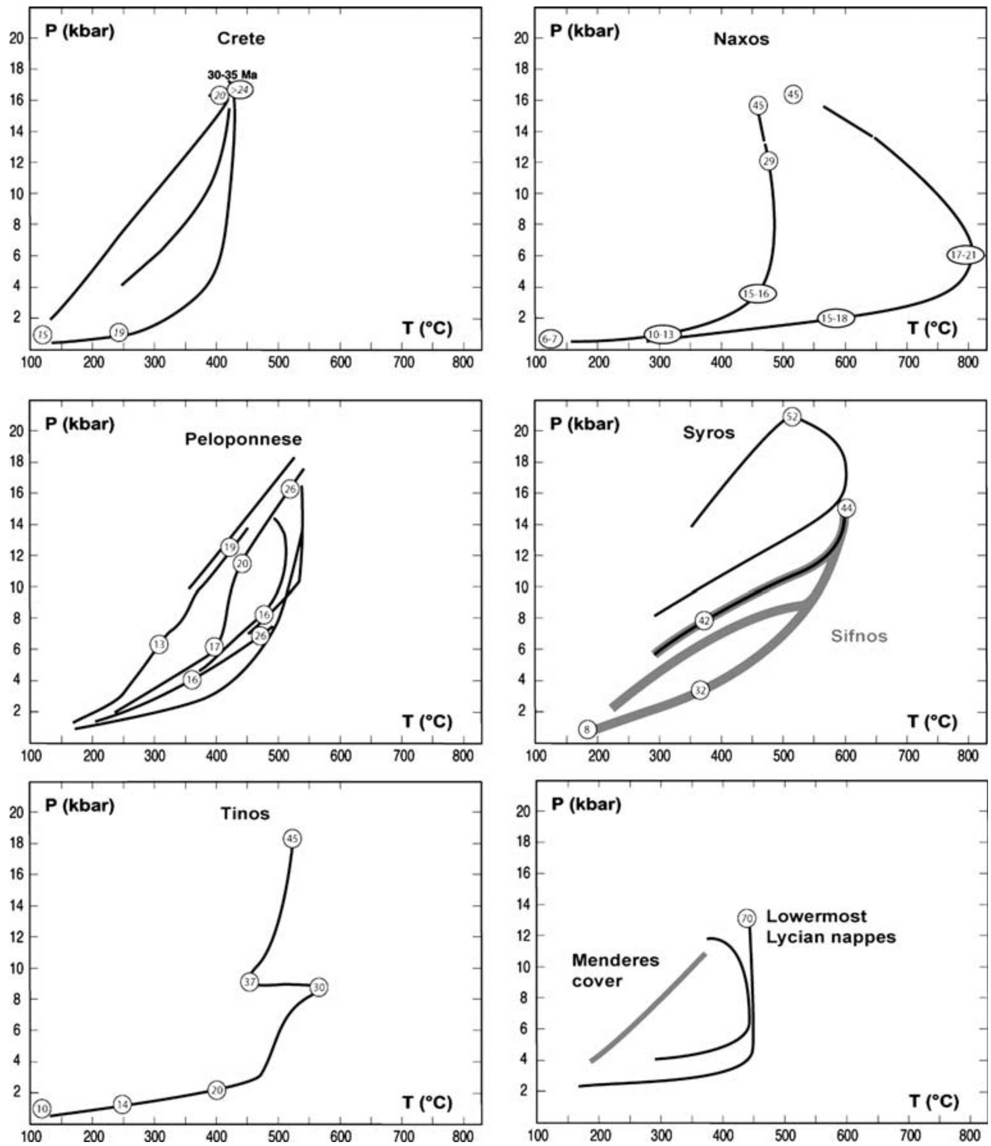


Fig. 8 A compilation of P–T–time paths in metamorphic units of the Cyclades, Peloponnese and Crete (Jolivet et al. 1996; Avigad 1998; Trotet 2000, 2001a, b, Trotet et al. 2006; Parra et al. 2002; Rimmelé et al. 2005; Duchêne et al. 2006; Ring et al. 2007)

(Karagianni et al. 2005)]. Seismic lines and gravimetric data converge toward a flat Moho below the center of the Cyclades suggesting a weak lower crust at least during the maximum of extensional activity. This region shows mainly metamorphic rocks that were exhumed in two successive stages (Bonneau and Kienast 1982; Avigad et al. 1997; Jolivet et al. 2003; Ring and Layer 2003) corresponding to two clusters of metamorphic ages (Altherr et al. 1979, 1982; Wijbrans and McDougall 1988; Bröcker et al. 1993; Wijbrans et al. 1993; Bröcker and Franz 1998; Keay et al. 2001; Bröcker et al. 2004; Lacassin et al. 2007) (Fig. 8). During the first stage, in the Eocene, blueschists and eclogites were partly exhumed while thrusting was still active further to the south, in the Hellenides, allowing the preservation of HP–LT parageneses (Trotet et al. 2001a, b). During the second stage, in Oligocene and Miocene times, exhumation occurred in Cordilleran-type metamorphic core complexes (Lister et al. 1984; Buick and Holland 1989; Buick 1991a; Gautier et al. 1993, Gautier et al. 1999; Gautier and Brun 1994a, b; Vanderhaeghe 2004) with NE dipping detachments showing a top-to-the-N or NE sense of shear (Faure and Bonneau 1988; Buick 1991a, b; Faure et al. 1991; Gautier et al. 1993; Jolivet et al. 1994a). The core of the high temperature metamorphic domes is affected by partial melting (Paros and Naxos, see Fig. 11) (Jansen and Schuilling 1976; Urai et al. 1990; Vanderhaeghe 2004; Duchêne et al. 2006). It must also be noted that a high temperature event of Late Cretaceous age is often dated in the Upper Cycladic unit that corresponds to the Pelagonian domain (Dürr et al. 1978; Reinecke et al. 1982; Maluski et al. 1987).

The blueschists-eclogites exhumation stage: evidence from Syros

In Syros, as well as in Sifnos, the Cycladic Blueschists crop out below a remnant of the Upper Cycladic Unit (Vari Unit, see Bonneau et al. 1980; Ring and Layer 2003) showing the best preservation of HP–LT parageneses (Fig. 10, 11). They are divided into three main units that display contrasting P–T–t evolutions but with the same P–T peak condition in the eclogite facies around 15 kbar and 550°C (Fig. 9) (Trotet et al. 2001a, b). Despite some Late Cretaceous U–Pb ages on zircons obtained in Syros island (Bröcker and Enders 1999), it is generally acknowledged that the peak of pressure occurred during the Eocene some 50 My ago. Radiometric ages, moreover, suggest that most of the exhumation was achieved during the Eocene before 42 Ma for units that have best preserved the HP–LT parageneses and before 32 Ma for greenschist units (Altherr et al. 1979; Maluski et al. 1987; Wijbrans et al. 1993; Bröcker and Enders 1999). Stretching lineations trend mostly E–W or NE–SW and the sense of shear is dominantly top-to-the-east or top-to-the-NE, synchronous with the retrogression of the eclogites in the blueschist facies. Greenschist facies is localized along lithological contacts (Trotet et al. 2001a, b; Keiter et al. 2004). An older deformation (prograde) suggests a top-to-the-southwest thrusting event (Keiter et al. 2004). The large pressure gap observed below the Vari unit and the colder evolution of HP units located immediately below indicate that the contact is a major fault with a normal sense of shear (Jolivet and Patriat 1999; Trotet et al. 2001a). High-pressure units were thus exhumed with little internal deformation (Keiter et al. 2004), except in the lowermost unit, in the southwest of the island, where greenschist facies deformation is predominant. This is at variance with Rosenbaum et al. (2002a) and Bond et al. (2007), who argued for HP rocks exhumation resulting from distributed coaxial stretching at crustal-scale. In summary, the Vari detachment can be considered as the roof of the exhuming Cycladic Blueschists extruding wedge (Ring and Layer 2003; Ring et al. 2007), or subduction channel (Jolivet et al. 2003). Similar ages have been documented in Samos island in the eastern Cyclades and the Dilek peninsula of western Turkey where the Cycladic Blueschists have also been exhumed during the Eocene below extensional detachments, while the Hellenic chain was under construction (Oberhänsli et al. 1998; Ring et al. 1999c; Ring and Layer 2003).

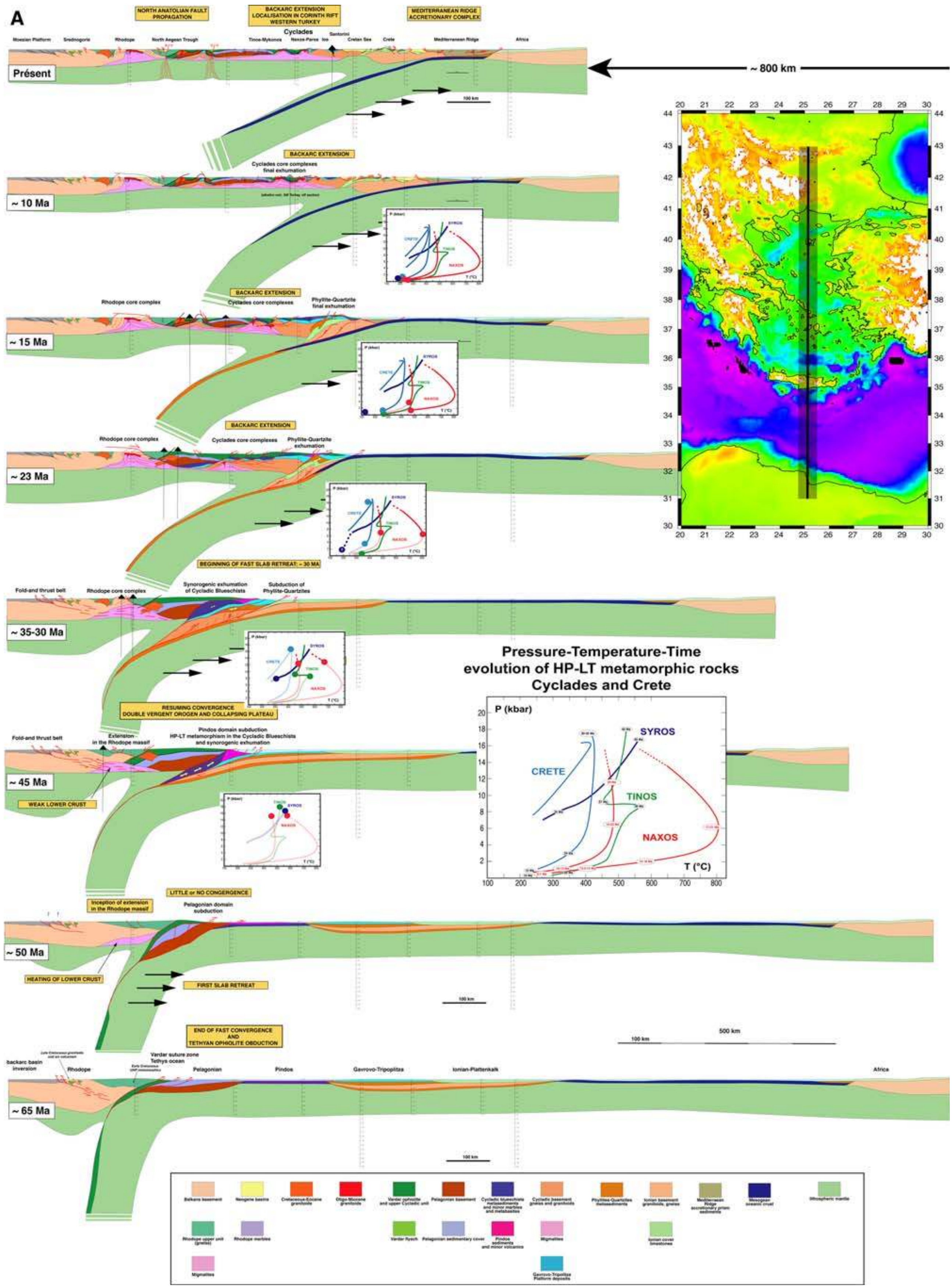
Fig. 9 **a** Tectonic reconstructions of the N-S section shown on Fig. 3 and the evolution of a few points along the P–T–t paths. **b** Details. See text for comments

High temperature core complexes: the example of Naxos

Naxos (Fig. 11) displays a high temperature metamorphic dome elongated in a NS direction. Uppermost micaschists, marbles and metabasites have locally preserved the HP–LT parageneses of the Eocene stage while the core is partially molten (Jansen and Schuilling 1976; Feenstra 1985; Buick and Holland 1989; Urai et al. 1990; Avigad 1998; Vanderhaeghe 2004; Duchêne et al. 2006). The dome is later intruded by a granodiorite during the Middle Miocene (Altherr et al. 1982; Altherr and Siebel 2002; Koukouvelas and Kokkalas 2003). Both the dome and the granodiorite show a pervasive N–S stretching lineation associated to a consistent top-to-the-north sense of shear (Buick 1991a, b; Gautier et al. 1993). A detachment separates the metamorphic dome from the overlying Upper Cycladic unit represented by serpentinites and non metamorphosed sediments of early Miocene age (Jansen 1973; Angelier et al. 1978; Kuhlemann et al. 2004) that are separated from the dome metamorphic rocks by a detachment. Approaching the detachment ductile deformation is locally overprinted by cataclasites and the upper unit displays only brittle deformation (Buick 1991a, b; Gautier et al. 1993). Eocene (45 ±5 Ma) radiometric ages have been preserved at the periphery of the dome and Oligo-Miocene ages are recorded everywhere else (Wijbrans and McDougall 1986; Avigad 1998; Duchêne et al. 2006). A fast exhumation is recorded between 20 and 8 Ma (Gautier et al. 1993; Duchêne et al. 2006) (Fig. 8). A recent investigation with low-temperature geochronological methods (apatite and zircon fission tracks and (U–Th)/He) points to a slip rate of the order of 6–8 mm/year of the detachment between 16 and 8 Ma (Brichau 2004). An analysis of the detrital content of the Early-Middle Miocene sediments resting on top of the upper Cycladic Nappe shows that the basin was mainly fed by the Pelagonian domain that must have widely outcropped before and during extension (Sanchez-Gomez et al. 2002; Kuhlemann et al. 2004). The HT rocks of the dome are not reworked in the basin showing that they arrived late at the surface.

Superposion of HT core complex evolution on HP rocks: inferences from Tinos

Tinos, as well as Andros, reveals an evolution stage that is somehow intermediate between Syros where Eocene HP events are well preserved and Naxos where Oligo-Miocene HT events have strongly overprinted the previous HP


A
NORTH ANATOLIAN FAULT PROPAGATION
BACKARC EXTENSION LOCALISATION IN CORINTH RIFT WESTERN TURKEY
MEDITERRANEAN RIDGE ACCRETIONARY COMPLEX
Cyclades
Présent
~ 800 km
BACKARC EXTENSION
Cyclades core complexes final exhumation
~ 10 Ma
Rhodope core complex
BACKARC EXTENSION
Cyclades core complexes
Phyllite-Quartzite final exhumation
~ 15 Ma
BACKARC EXTENSION
Rhodope core complex
Cyclades core complexes
Phyllite-Quartzite exhumation
~ 23 Ma
BEGINNING OF FAST SLAB RETREAT: ~ 30 MA
Fold-and thrust belt
Rhodope core complex
Synorogenic exhumation of Cycladic Blueschists
Subduction of Phyllite-Quartzites
~ 35-30 Ma
RESUMING CONVERGENCE DOUBLE VERGENT OROGEN AND COLLAPSING PLATEAU
Fold-and thrust belt
Extension in the Rhodope massif
Pindos domain subduction HP-LT metamorphism in the Cycladic Blueschists and synorogenic exhumation
~ 45 Ma
WEAK LOWER CRUST
Pressure-Temperature-Time evolution of HP-LT metamorphic rocks Cyclades and Crete
P (kbar)
T (°C)
CRETE
SYROS
TINOS
NAXOS
LITTLE or NO CONGERGENCE
Inception of extension in the Rhodope massif
Pelagonian domain subduction
~ 50 Ma
HEATING OF LOWER CRUST
FIRST SLAB RETREAT
END OF FAST CONVERGENCE AND TETHYAN OPHIOLITE OBDUCTION
500 km
Vardar suture zone Tethys ocean
Rhodope
Pelagonian
Pindos
Gavrovo-Tripolitza
Ionian-Plattenkalk
Africa
~ 65 Ma

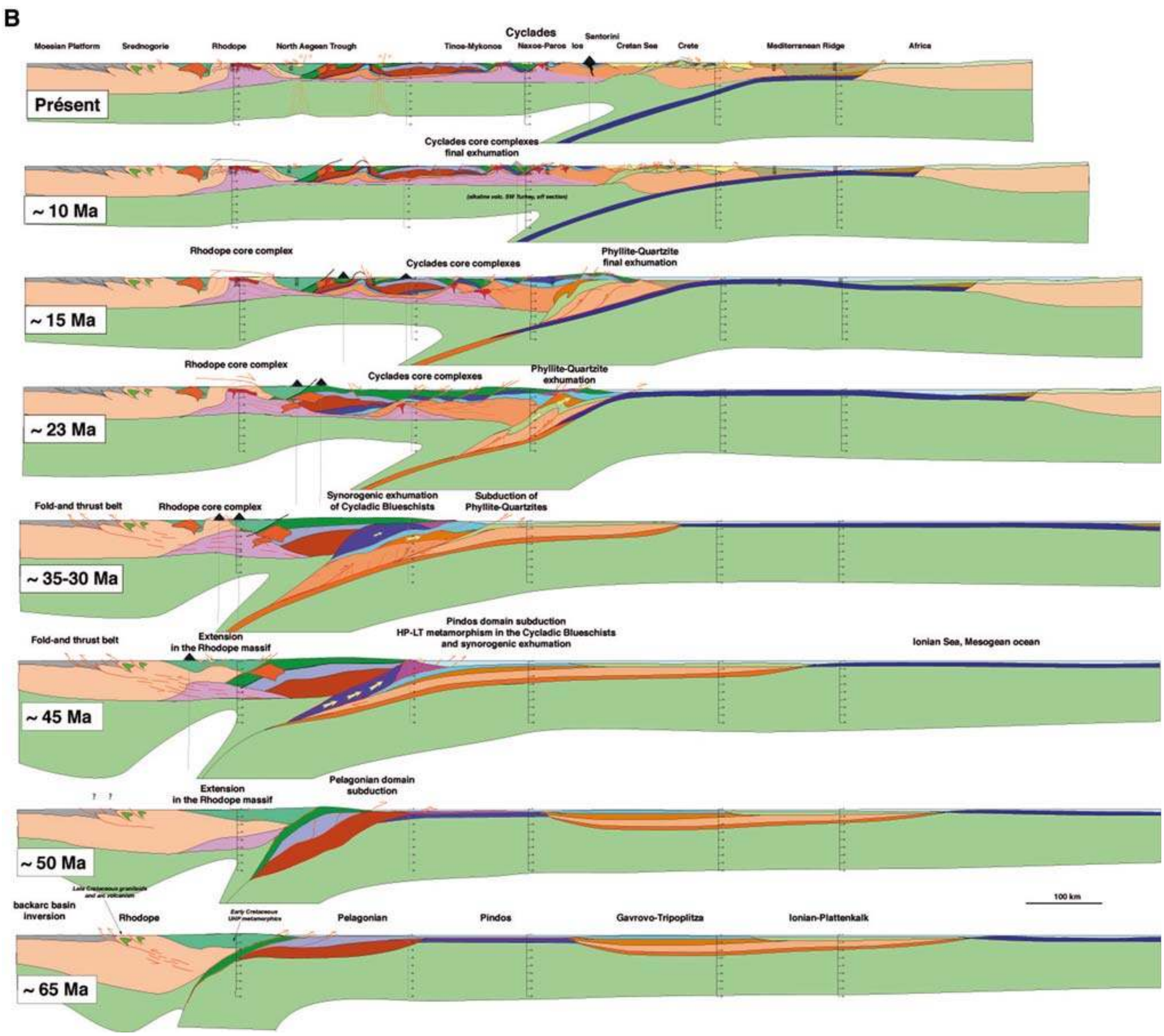


Fig. 9 continued

history (Parra et al. 2002; Jolivet et al. 2004a; Mehl et al. 2005) (Fig. 8).The Upper Cycladic unit is represented by several large units (Melidonis 1980; Avigad and Garfunkel 1989; Katzir et al. 1996; Stolz et al. 1997) with a basal contact defining a broad open antiform as well as the metamorphic foliation of metapelites, metabasites and marbles in the underlying Cycladic Blueschists. HP–LT parageneses are best preserved to the southwest but pristine blueschists still occur in small lenses in the rest of the island (Jolivet and Patriat 1999; Jolivet et al. 2004a). Greenschist retrogression increases from southwest to northeast associated to a gradient of strain intensity. Deformation that occurs on the southwest as narrow localized shear zones with mixed top-to-the-SW and top-to-the-NE senses of shear evolves toward the northeast into a pervasive shearing, with a consistent NE–SW-trending stretching lineation and a dominant top-to-the-NE shear sense (Gautier and Brun 1994a, b Jolivet and Patriat 1999). The contact between the Cycladic Blueschists and the Upper Cycladic unit in the northeast of the island is marked by a thick zone of cataclasites that results from a transition in time from ductile to brittle deformation. Associated to a pressure gap it is best interpreted as an extensional detachment (Gautier and Brun 1994a, b; Jolivet and Patriat 1999; Mehl et al. 2007). Radiometric data suggest an Eocene age for the HP–LT parageneses and an Early Miocene age for the greenschists retrogression (Bröcker et al. 1993; Bröcker and Franz 1998). A large granodiorite intrusion cuts trough the detachment (Avigad and Garfunkel 1989), but is still deformed by the top to NE

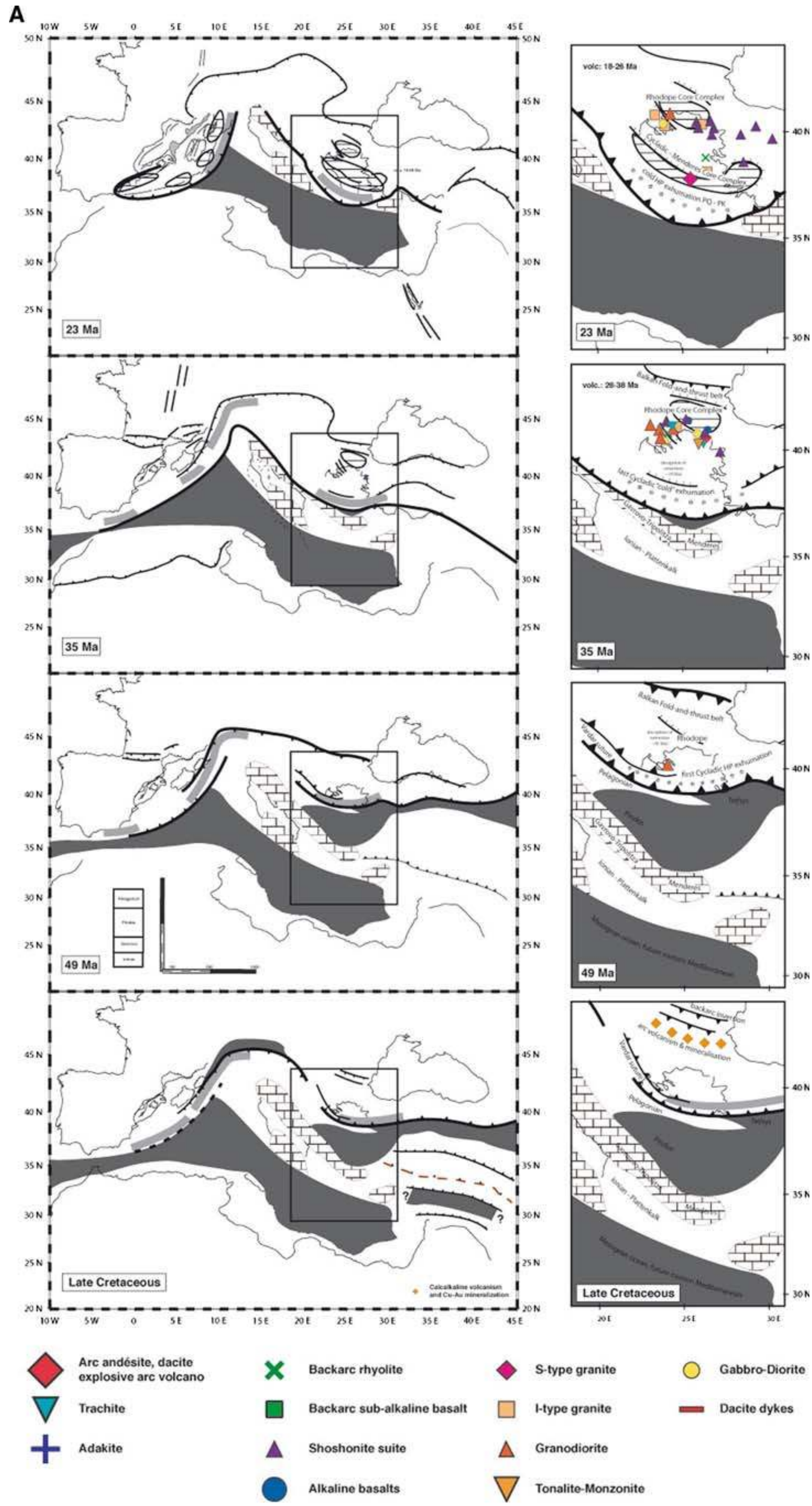


Fig. 10 Paleogeographic maps of the Mediterranean region and details for the Aegean region from the Late Cretaceous to the Early Miocene. The reconstructions are based upon the kinematics parameters given by Dewey et al. (1989) for the Africa-Eurasia motion. Reconstructions of the internal deformation is similar to Jolivet et al. (2003) and Lacombe and Jolivet (2005). The details for the Aegean region shows the main paleogeographic domains and the position and nature of the magmatic arc (Pe-Piper and Piper 2002, 2006, 2007)

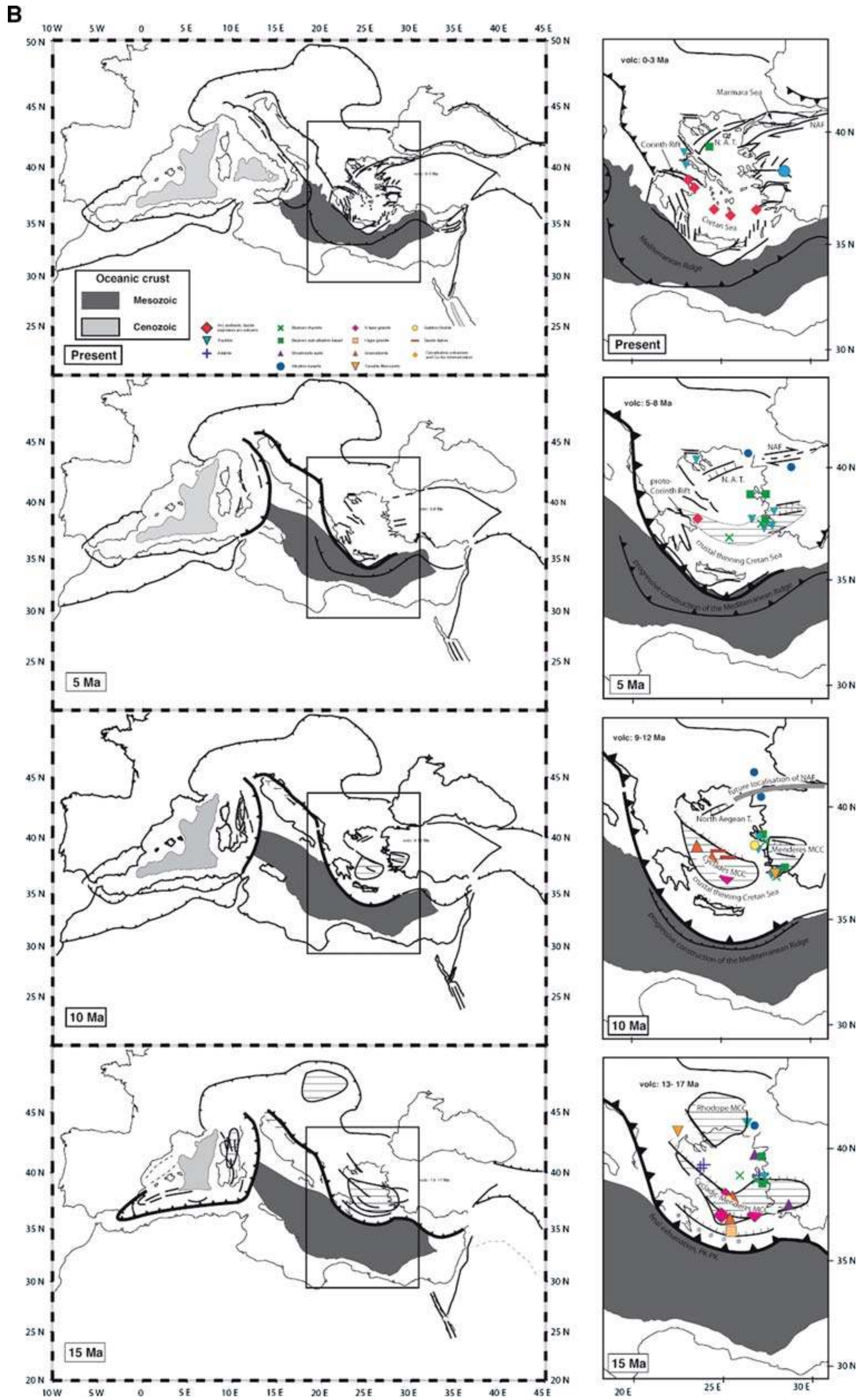


Fig. 10 continued

sense of shear (Faure et al. 1991; Gautier and Brun 1994a, b; Jolivet and Patriat 1999) indicating the synkinematic nature of the intrusion during the Early Miocene (from 19 to 14 Ma) (Bröcker and Franz 1994, 1998, 2000; Brichau et al. 2007). Comparable timing is observed in the nearby island of Mykonos where a migmatite dome and a

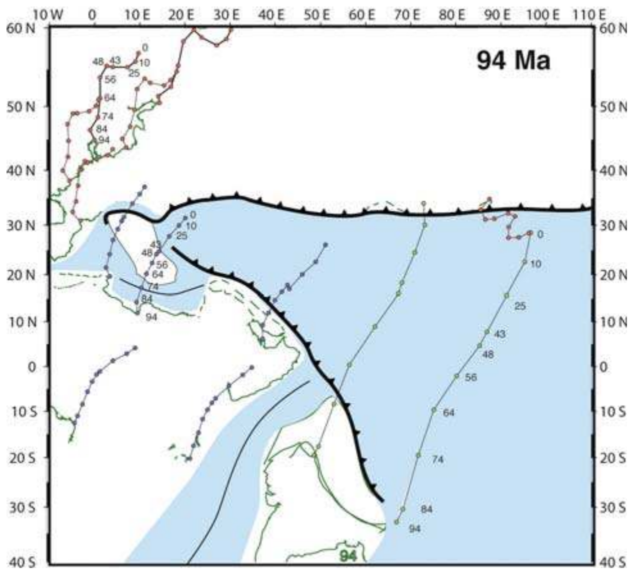


Fig. 11 Displacements of selected points of Africa, India and Eurasia from the late Cretaceous to the present in the hotspot reference frame (based on the kinematic parameters of Gordon and Jurdy (1986) and Lithgow-Bertelloni and Richards (1998)

10-Ma-old granodiorite were exhumed below a shallow-northeast-dipping detachment (Faure and Bonneau 1988; Faure et al. 1991; Gautier and Brun 1994a, b; Avigad et al. 1998). One can thus assume that the same detachment zone has been at work from ca. 25 to after 9 Ma in the northern Cyclades during the formation of the Aegean Sea. P–T estimates on Tinos show a two-staged evolution: a first exhumation along a cold gradient, probably between 45 and ~37 Ma, then an evolution at constant pressure until ~30 Ma and a second exhumation from 30 to 10 Ma (Parra et al. 2002). The first exhumation can be compared to the main deformation seen on Syros and the contact observed on the SW coast of Tinos is probably the same as the Vari detachment on Syros.

In summary, the exhumation of the Cycladic Blueschists proceeded in two stages. It was first accommodated during the Eocene along a cold P–T gradient bringing some units close to the surface, e.g., Syros and Sifnos, whereas some other units remained at a deeper positions in the crust undergoing moderate heating, e.g., Tinos and Andros, or heating strong enough for partial melting to develop, e.g., Naxos. During the second stage in Oligocene-Miocene, those units that previously stayed longer at depth were brought to the surface by core complex-type extension forming a series of metamorphic domes cored by migmatites and/or granites, as exemplified in Naxos and Tinos. Both stages are accommodated by a series of shallow-NE-dipping detachments or a single detachment that migrated upward with time until the Late Miocene or even later.

The Menderes massif and Lycian nappes

The Menderes massif is located between the zones of Istanbul, Sakaria and Tavsanli, to the north, and by the Lycian thrust units and the Bey Daglari platform, to the south. The Istanbul zone is made of a Cadomian basement and its Paleozoic cover, in turn covered with Mesozoic and Cenozoic sediments (Abdülselamoglu 1963; Kaya 1973; Okay 1989; Görür et al. 1997; Dean et al. 2000). To the south, it is separated by the Intra-Pontide suture from the Sakarya zone that consists in a series of tectonic units made of volcanic and clastic rocks with a Late Paleozoic and Late Triassic overprint, covered with unconformable Jurassic and Cretaceous sediments. The Izmir-Ankara suture separates the Sakarya zone from the Anatolide-Tauride platform made of several units separated by major thrusts. The Tavsanli zone (Okay 2001) is made of blueschists overthrusted by the non-metamorphic ophiolitic units of the suture. The Bornova flysch zone is an olistostrome of Maastrichtian-Paleocene age. The Afyon zone is a Cadomian basement overlain by Triassic sediments metamorphosed in the blueschist facies (Candan et al. 2002, 2005). Then the Menderes massif and the overlying Lycian units overthrust the external Bey Daglari platform.

Following the synthesis of Bozkurt and Oberhänsli (Bozkurt and Oberhänsli 2001), the most striking characters of the Menderes massif can be summarized as follows. A Late Paleozoic to Late Cretaceous sedimentary cover, deposited on top of late Proterozoic (Cadomian) basement, shows a progressive deepening of depositional environments ending with a Late Paleocene–Early Eocene olistostrome (Önay 1949; Dürr 1975; Caglayan et al. 1980; Özkürt and Koçyigit 1983; Sengör et al. 1984; Konak et al. 1987; Özer 1998; Özer et al. 2001). In the Dilek peninsula, rock assemblages closely resembling the Cycladic Blueschists that have given Eocene ages (40 Ma) are thrust over the Menderes massif (Candan et al. 1997; Candan and Dora 1998; Oberhänsli et al. 1997, 1998; Okay et al. 2001b; Cetinkaplan 2002; Ring et al. 2007). The so-called Main Menderes Metamorphism (MMM) that is recorded in the whole massif corresponds to Barrovian-type conditions (9 kbar, 550°C) (Ashworth and Evirgen 1984; Okay 2001) in the lower parts of the massif and greenschist conditions above with early Tertiary Rb–Sr and ^{40}Ar–^{39}Ar ages (Sengör et al. 1984; Satir and Friedrichsen 1986; Lips et al. 2001). The cover sequence and the olistostrome have recorded HP–LT parageneses (10–12 kbar, 440°C) (Rimmelé et al. 2003). The last metamorphic event in Oligocene-Miocene times corresponds to the greenschist retrogression during the exhumation below shallow-dipping detachments with a N-S direction of stretching associated to senses of shear either top-to-the north or to the south (Hetzel et al. 1995a, b; Seyitoglu and Scott 1996; Ring et al. 1999b; Bozkurt 2001; Bozkurt and Oberhänsli 2001; Lips et al. 2001; Ring et al. 2007). However, the total amount of extension is likely less than in the Cyclades as indicated by a thicker crust.

The Lycian thrust units rest on top of the Menderes massif (Brunn et al. 1970, 1976; Graciansky 1972; Collins and Robertson 1997, 1999; Okay et al. 2001b; Rimmelé et al. 2002). To the south the lowest thrust unit is characterized by P–T estimates in the range 8–10 kbar for 350–400°C (Oberhänsli et al. 2001; Rimmelé et al. 2002; 2005). They are covered by a melange, then by a HT metamorphic sole and finally by the Lycian ophiolite, obducted at the end of the Cretaceous (Gutnic et al. 1979; Collins and Robertson 1997). Available Ar–Ar do not show clear plateaux but the spectra step up from 35 to 90 Ma and the authors suggest a Late Cretaceous age for the HP–LT metamorphism (Ring and Layer 2003).

The Lycian thrust units, including the ophiolite, correlate with the Izmir-Ankara suture units to the north such as the Menderes massif that appears as a tectonic window entirely buried below the Cycladic Blueschists in Cretaceous-Eocene times (Rimmelé et al. 2006). The Menderes massif is therefore a lateral equivalent of the Apulian units underthrust below the Cycladic Blueschists in Naxos or Ios (Jolivet et al. 2004b).

Cenozoic evolution of the Aegean domain

Using the structural and chronological data summarized above, a reconstruction of the tectonic evolution leading to the present-day geometry is presented here as a series of step-by-step cross-sections (Fig. 9) and regional-scale maps (Fig. 10). The width of initial palaeographic domains is based upon van Hinsbergen et al. (2005a) (and references therein). The position of the volcanic centers and the main plutons is after Pe-Piper and Piper (2002, 2006, 2007). Plate kinematics is after Dewey et al. (1989). Using the kinematic parameters proposed by Rosenbaum et al. (2002b) would change only marginally the reconstructions. Both sets of parameters agree about a very slow convergence, if any, between Africa and Eurasia between 67–65 and 52 Ma (this slow convergence period starts earlier in Dewey et al. 1989), an increase of the convergence velocity from 52 to approx. 20 Ma and a slower convergence afterward. In terms of kinematics, the reconstructions for the whole Mediterranean region are similar to those proposed by Jolivet et al. (2003). In the reconstructions, Apulia is rigidly linked to Africa as suggested by paleomagnetic data (Van der Voo 1990). The migration of the Hellenic trench is estimated using the southward migration of the volcanic arc that amounts to approx. 2 cm/year. The beginning of fast slab retreat has been set to 30–35 Ma as suggested by the migration of the volcanic arc (Fig. 6) and a limited amount of slab retreat has been assumed during the no-convergence period.

The observation that a single subduction controlled the regional evolution from the Mesozoic to the present (Jolivet et al. 2003; Van Hinsbergen et al. 2005a) strongly suggests that the slab was anchored in the lower mantle over the whole period covered by the reconstruction. Active backarc extension recorded during part of the Cretaceous in the Black Sea and Balkan regions would suggest a steep slab, as in modern subduction zones, upper plate deformation directly depending on slab dip: extension above steep slabs versus compression above shallow-dipping slabs (Heuret and Lallemand 2005; Heuret et al. 2007). The present-day slab geometry being rather shallow-dipping, i.e., ~40–45° at depth and ~30° in the upper mantle, the slab dip has likely decreased through time since the Cretaceous–Tertiary boundary.

From the base of Paleocene (65 Ma) to the Early Eocene (50 Ma)

In the end of the Cretaceous, the Tethys is still an open oceanic space that subducts along its northern margin below the southern margin of Eurasia east of our section. A magmatic arc has developed in the present-day Balkans. The backarc basin that formed earlier in the Cretaceous is being partly inverted. The Pelagonian domain and the Tethyan ophiolite that was obducted at the end of the Jurassic are now engaged in the subduction zone. The ultrahigh pressure domain of the Rhodope has been exhumed in earlier stages and is now in the upper crust. The last remnants of the Tethys are thrust over the northern margin of the Pelagonian domain. Further east, the final closure of the Tethys will take place a little later during the Paleocene. The position of the Lycian thrust units is not totally ascertained. They surely belong to an oceanic domain located north of the Menderes massif and they likely root in the Izmir-Ankara suture (Robertson et al. 1991, 1996; Stampfli and Borel 2002). The Pindos ocean started to subduct around 50 Ma. The Gavrovo-Tripolitza platform that is carried by a continental block related to Apulia grades southward into the Ionian basin. The Phyllite-Quartzite thrust unit likely belongs to the basement of the Gavrovo-Tripolitza platform.

In the Late Cretaceous, the final closure of the Vardar ocean led to crustal thickening north and south of the suture. Compressional stresses were transmitted as far as the Balkans where the extensional structures formed earlier during the Cretaceous were inverted forming the first compressional structures of the Balkan fold-and-thrust belt. After a period of fast convergence that finally led to obduction in the eastern Mediterranean region and further east, the relative motion between Africa and Eurasia comes almost to a halt at $\sim$65–67 Ma. The anchored slab likely starts to retreat with respect to the overriding plate, as slab sinking is then only driven by its own weight. Therefore, during this period, subduction continues without convergence. The Pelagonian domain is dragged within the subduction zone by the negative buoyancy of the subducting lithosphere. During slab rollback, the cold lithosphere being replaced by a hot asthenosphere, the Middle-lower crust is strongly heated and submitted to partial melting, leading to considerable strength drop at crustal scale and thus preparing the collapse of the Rhodope massif.

From the Early Eocene (50 Ma) to the Late Eocene-Early Oligocene (30–35 Ma)

Convergence accelerates reinstalling an overall compressional regime. The Pindos oceanic domain and its margins are quickly subducted forming at depth the Cycladic Blueschists and eclogites. Thrusting propagates southward understacking progressively the thrust slices of the sedimentary Pindos and of the Gavrovo-Tripolitza, whose deeper metamorphosed parts appear today within the Olympos, Ossa and Almyropotamos windows. As shown in Syros and Sifnos, the Cycladic Blueschists start exhuming according to cold retrograde P–T paths that preserve HP–LT parageneses below low angle detachments. Despite active compression in the frontal zones of the Balkans during this period, the Rhodope is submitted to extension and core complexes start to develop. Topography soon returns to sea level in some regions of the Rhodope and, from the Lutetian, marine sediments are deposited in extensional basins. The mechanics of these simultaneous extension in the Rhodope and compression in the Balkans until late Eocene remains to be explained (see Sect. "Discussion" below). Synextensional volcanism and granite plutonism occur in the northern Rhodope and in Chalkidiki peninsula (e.g., Sithonia granodiorite, 48–50 Ma).

From the late Eocene–Early Oligocene (30–35 Ma) to the Late Oligocene–Early Miocene (23 Ma)

A drastic change in the boundary conditions occurs at 30–35 Ma. The absolute northward motion of Africa decreases and the subducting slab started to retreat southward at high pace (Jolivet and Faccenna 2000). Southward propagation of trusting continues in the Apulian platform up to the Early Miocene with the understacking of the Phyllite-Quartzite and Ionian units. These units then start to quickly exhume below the Cretan detachment with cold P–T paths, allowing the preservation of HP–LT parageneses. The magmatic arc continues to migrate southward together with the trench. In the backarc region of the retreating slab, rheological weakening due to partial melting enables vigorous lower crustal flow that simultaneously contributes to develop HT core complexes in the extending Cyclades, i.e., Naxos and Mykonos, and to maintain a flat Moho. The marine sediments are deposited on top of the Cyclades core complexes in the Aquitanian.

From the Late Oligocene–Early Miocene (23 Ma) to the Early Late Miocene (10 Ma)

As slab retreat continues backarc extension and detachment activity in the Cyclades reaches a peak. Granodioritic plutons intrude core complexes and detachments migrate upward in the crust eventually up to the detachment proper, e.g., Tinos. Extension is then taken up by a more superficial detachment. In the Rhodope, brittle extension affects the core complex allowing the development of sedimentary basins. The same occurs in the Menderes massif. The Cretan detachment still continues to accommodate the exhumation of HP–LT metamorphic units up to the surface and the synchronous deposition of sediments up to $\sim$11–10 Ma. To the south, sediments are accreted to form the first stages the Mediterranean ridge accretionary complex.

From the Early Late Miocene (10 Ma) to the present

The thrust wedge front reaches the north African margin as the Mediterranean Ridge grows. With ongoing slab retreat, extension localizes in the Cretan Sea giving a local crustal necking, whereas a constant thickness of $\sim$26 km is kept below the hottest part of the Cyclades and around 30 km below islands located closer to the continent. This suggests a change in either crustal rheology or in boundary conditions. In the Rhodope, continental sedimentation continues during the Pliocene-Pleistocene in extensional basins on top of the core complex. At $\sim$5 Ma the North Anatolian Fault enters the Aegean domain and extension localizes in the North Aegean Trough and connects to the extensional domains of Evia and Corinth. Active extension becomes then located mostly in the outer Aegean arc, the Peloponnese and Crete and in western Turkey within E–W trending grabens.

Discussion

A single subducting slab and no slab break-off

From the Late Cretaceous to the present, all the tectonic and magmatic events recorded in the Aegean region can be related to the subduction of the African slab plate below the southern margin of Eurasia. Nappe stacking and subsequent extension occurred while a single lithosphere slab was subducted. The amount of subducted material in our reconstructions is based upon the relative positions of plates deduced from plate kinematics data. Since the Late Cretaceous, approximately 1,500 km of lithosphere were subducted in our reconstructions which is slightly more than the amount proposed by Van Hinsbergen et al. (2005a), but of the same order of magnitude. The average subduction rate is thus around 2.3 cm/year. The total amount of convergence between Africa and Eurasia during the same period is of the order of 800 km parallel to our transect (1.3 cm/year in average). This leaves about 700 km for slab retreat and backarc extension, the total amount of subduction being the sum of convergence and backarc extension. If we assume that ~50–100 km of retreat were accommodated during the Eocene, i.e., first stage of Rhodope extension, this leaves ~600–650 km of retreat since 30–35 Ma, thus an average rate of ~2.0 cm/year. This is likely not much different from the present-day rate of extension at Aegean scale as 1.5 cm/year occurs across the Corinth Rift and that additional components certainly occur in smaller active rifts. During the same period the volcanic arc has migrated southward of some 550–600 km thus at approximately the same rate of ~2 cm/year.

It is usually considered that continental subduction leads almost inevitably to slab break-off (Davies and von Blanckenburg 1995; see also the experiments of Regard et al. 2003, 2005). Gravitational forces tend to stretch the oceanic slab and separate it from the continental lithosphere that resists subduction. However, in the Aegean example the Pelagonian and Apulian blocks of continental lithosphere have not been able to lead to slab break-off as evidenced by tomographic models. For such continental blocks that are of limited size, i.e., few hundred kilometers, the upper continental crust is more likely able to entirely delaminate from the underlying mantle.

Two episodes of simultaneous blueschist exhumation and core complex extension

The reconstructions presented here show two stages of slab rollback and upper plate extension, probably at two different rates. First, core complex-type extension starts in the Rhodope partly synchronous with the first stage of formation and exhumation of the Cycladic Blueschists and with the subduction of the Pindos oceanic domain, in Eocene times. Second, core complex-type extension starts in the Cyclades synchronous with the burial and exhumation of blueschists in Crete and Peloponnese, and with the subduction of the Mediterranean oceanic lithosphere, in Oligocene–Miocene times. Several effects can lead to slab rollback in this case:

(1) The timing suggests that trench retreat, which is a suitable boundary condition for extension to occur, is partly controlled by the transition between subduction of continental and oceanic lithospheres. The subduction of an oceanic lithosphere after a continental lithosphere gives a buoyancy change that results in slab rollback and, consequently, to trench retreat. This effect has been recently illustrated by modelling, both analog (Martinod et al. 2005) and numerical (Royden and Husson 2006). In other terms, the subduction of the Pindos oceanic domain could have stimulated slab rollback and thus favored Aegean extension. Following the Pindos subduction, the entrance of the Apulian block into the subduction zone temporarily interrupted the slab rollback until the arrival of the Mediterranean oceanic lithosphere. Then slab rollback and associated trench retreat started again responsible for the second extension event marked by HT core complex-type extension, continuing in the Rhodope and starting in the Cyclades, and coeval exhumation of blueschists in Crete and Peloponnese.

This first stage of extension is contemporaneous with compression in the foreland of the Balkans and this problem still remains to be solved. One could argue that slab rollback induced by subduction of the Pindos oceanic domain has partly released the compressional stresses thus leading to crustal collapse but not enough to induce whole plate extension. Part of the collapse zone was thus bounded on either sides by two zones of compression. An additional possibility to consider is that the intense partial melting of the lower crust has decoupled the behaviors of the crust and mantle, the motion of the crust being mostly controlled by gravitational spreading (Bird 1991; Royden 1996; Vanderhaeghe and Teyssier 2001). This hypothesis would hold if the major part of the first stage of extension was achieved before the deposition of Lutetian marine sediments on top of the Rhodope.

Could this first episode of extension and slab rollback also be controlled by larger scale processes? It follows the very strong decrease of convergence velocity between 67–65 and 55–52 Ma between Africa and Eurasia. At variance to what happened some ~30 Ma velocity this very slow convergence is probably not due to a reduction of the absolute velocity of Africa. Figure 11 shows the successive positions of the main plates surrounding the Tethyan region from the Late Cretaceous to the present in the hot spot frame of reference, calculated from Gordon and Jurdy (1986) and Lithgow-Bertelloni and Richards (1998). The motion of Africa does not show any reduction of velocity in the concerned period. It seems, however, that Eurasia moves faster northward during the same period. Recent studies of modern subduction zones suggest that the tectonic regime is primarily controlled by two main forces: slab pull and slab suction (Chase 1978; Conrad et al. 2004; Faccenna et al. 2007). The second force, long underestimated, is strong when the two plates move in opposite direction in the hot spots reference frame, weak when they move in the same direction. This slab suction force is responsible for a strong downward drag of the mantle surrounding the slab in the case of two converging plates. When this force is weak, this drag is weaker. The observations suggest that backarc extension corresponds to this situation. The presence of an old, steep and dense slab with a strong slab pull will play in the same sense leading to a retreat of the slab with respect to the upper plate. This episode of slow convergence could explain the initiation of extension and the heating of the lower crust that would reduce the overall resistance of the Rhodope edifice and easing the subsequent post-orogenic extension.

(2) The second stage of extension seems faster than the first one and it leads to a more pronounced crustal thinning. It is furthermore contemporaneous throughout the whole Mediterranean (Jolivet and Faccenna 2000). It is more clearly associated with a fast southward migration of the slab. It has been explained earlier as a consequence of the Africa–Eurasia collision in the some 30–35 Ma ago and the coeval reduction of the absolute velocity of Africa, that then subducted under its own weight and started to retreat efficiently leading to backarc extension in the whole Mediterranean. The distribution of magmatic products suggests that it started some 35–30 Ma ago when the Rhodope was rapidly uplifted. The flat Moho observed

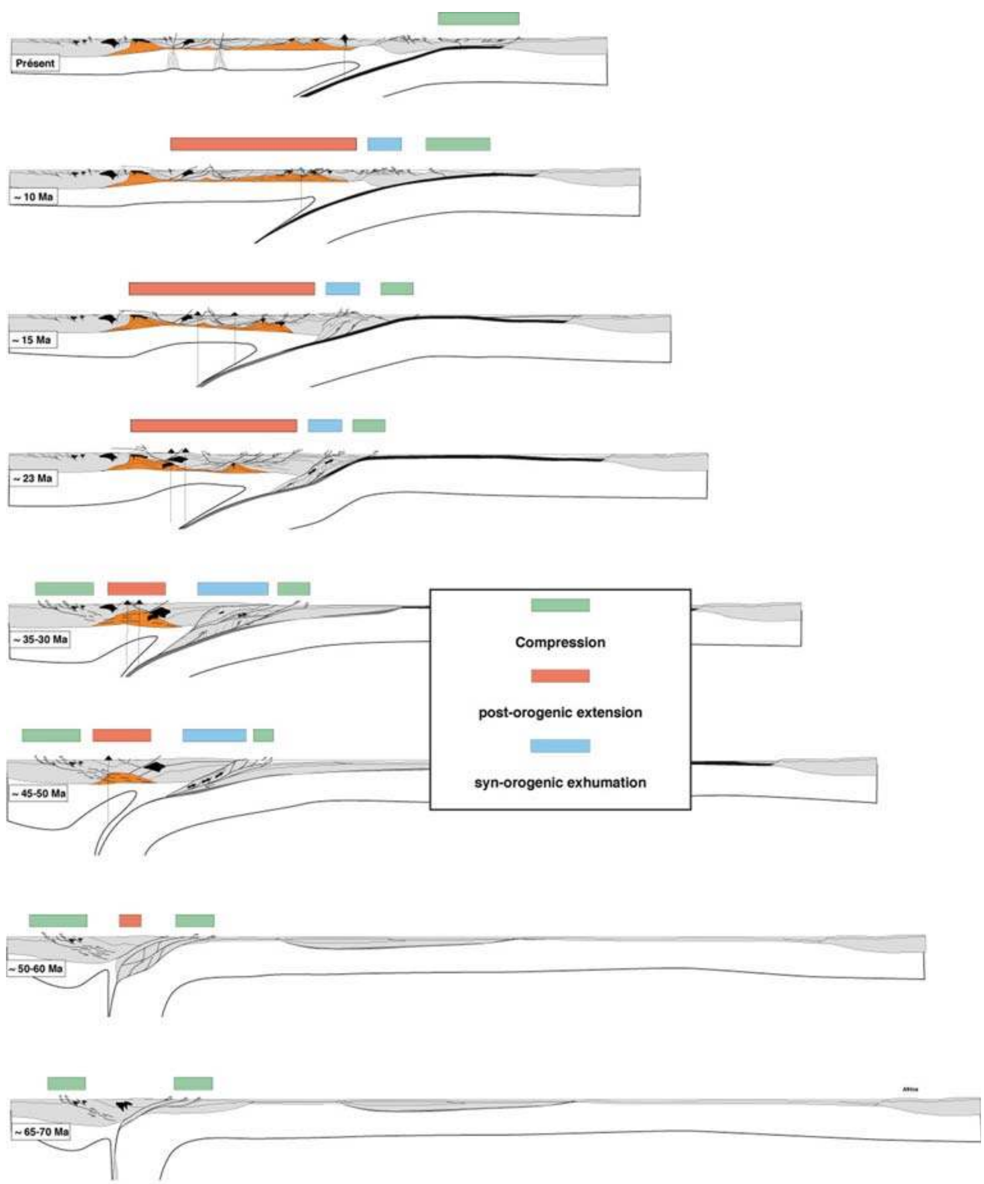


Fig. 12 The same reconstructions as on Fig. 8 showing the position of lower crust partial melting and the migration of compression, post-orogenic extension and syn-orogenic exhumation

below the present-day Aegean domain suggests that the crust has been weak during extension, further suggesting that lower crustal partial melting was intense (Fig. 12). In the reconstructions, we have assumed that crustal melting was triggered by the uplift of the lithosphere–asthenosphere boundary during slab retreat, due either to the passage of the volcanic arc or to the removal of lithospheric mantle material by ablative subduction. The lithosphere has cooled down afterward and has probably thickened back as suggested by the rigid behavior of the Aegean domain during the last 5–7 Ma (Sonder and England 1989).

The two-stage exhumation of high-pressure metamorphic rocks

Regional-scale inferences, as depicted by our reconstructions, strongly sustain that convergence has been accommodated by a single subducting slab since the Mesozoic. As the trench progressively migrated southward, the continental units were delaminated from their underlying mantle and accreted to the overriding plate. During this transfer from the subducting plate to the overriding plate, the crust undergoes a three-stage history in terms of pressure and temperature variations. First, pressure increases during subduction giving prograde HP metamorphism. Second pressure decreases during exhumation and accretion to the upper plate giving a retrogression of HP parageneses. During these first two stages, the crust undergoes moderate temperature variations. Third, because the asthenosphere flows in the wedge that opens between the delaminated crust and the subducting mantle, the crust is strongly and rapidly heated from below leading to partial melting and thermal weakening of the crust, creating suitable rheological conditions for the development of core complexes in the extending upper plate. This sequence of pressure and temperature variations explains the fact that HP rocks like those of Tinos (Fig. 10) can display well defined threefold PT histories: first fast decompression up to around 9–10 Kb, second heating at constant pressure and depth and third, slower decompression and cooling. In the frame of the Aegean, it is noteworthy that the two stages of exhumation are separated by an isobaric heating event that ends at the same time as thrusting of the most external Ionian units. The first part of the exhumation is at least partly simultaneous with the subduction of the Pindos oceanic domain, whereas the second part is obviously related to the second Aegean extension.

Conclusion

Using available plate kinematic, geophysical, petrological, structural data, we construct a synthetic tectonic map of the whole region encompassing the Balkans, Western Turkey, the Aegean Sea, the Hellenic Arc, the Mediterranean Ridge and continental Greece and we build a lithosphere-scale N-S cross-section from Rhodope to Crete. The relationships in space and time between subduction rollback and thrusting and extension in crustal units are described using a step-by-step restoration of this cross-section, since the late Cretaceous. This provides a suitable reference frame, in both time and space, to discuss the mechanisms that likely control the circulation of HP–LT metamorphic rocks within the subduction zone and during subsequent core complex-type extension in the backarc region during trench retreat.

The reconstructions illustrates the continuous character of deformation during the northward subduction of the African plate below the southern margin of Eurasia and a three-step evolution for HP–LT metamorphic rocks: (1) subduction with the sinking plate, (2) first exhumation within the subduction channel and (3) post-orogenic exhumation in extensional metamorphic domes, synchronous with a HT metamorphism overprinting. The presence of at least two continental blocks in the subducting plate has not prevented a continuous subduction that proceeded without slab break-off.

The example of the Rhodope-Aegean transect shows that a succession of tectonic stages, and a variety of structural settings can be observed during a single episode of subduction, depending upon the relative and absolute motions of the subducting and overriding plates and the nature of the subducting material. The succession in time from syn-orogenic to post-orogenic exhumation episodes in the Cyclades is well explained by the migration of the subduction zone during slab rollback. The presence of several parallel belts of high-pressure metamorphic rocks does not require either multiple subduction zones or trench jump. The progressive building of the Hellenic chain simply results from the delamination between the upper crust and the lower crust and lithospheric mantle. With slab migration, the domains that were accreted to the orogenic wedge are then affected by backarc extension when the compressional “wave” has passed.

The model we propose is testable. Its timing is, however, based upon a limited number of radiometric ages that suggest pulses of high-pressure (Early Cretaceous in the Rhodope, Eocene in the Cyclades, Oligo-Miocene in Crete and the Peloponnese) and high temperature metamorphism (Eocene in the Rhodope, Oligo-Miocene in the Cyclades). These episodic events of HP metamorphism and exhumation, during a subduction history that is continuous, could likely be related to the subduction of alternating continental and oceanic domains, in the Eocene for the Pindos and Oligocene-Miocene for the Mediterranean oceanic lithosphere that also partly controls the two main peak of extension. Further more detailed geochronological studies will say whether the real evolution is really discontinuous or more progressive.

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