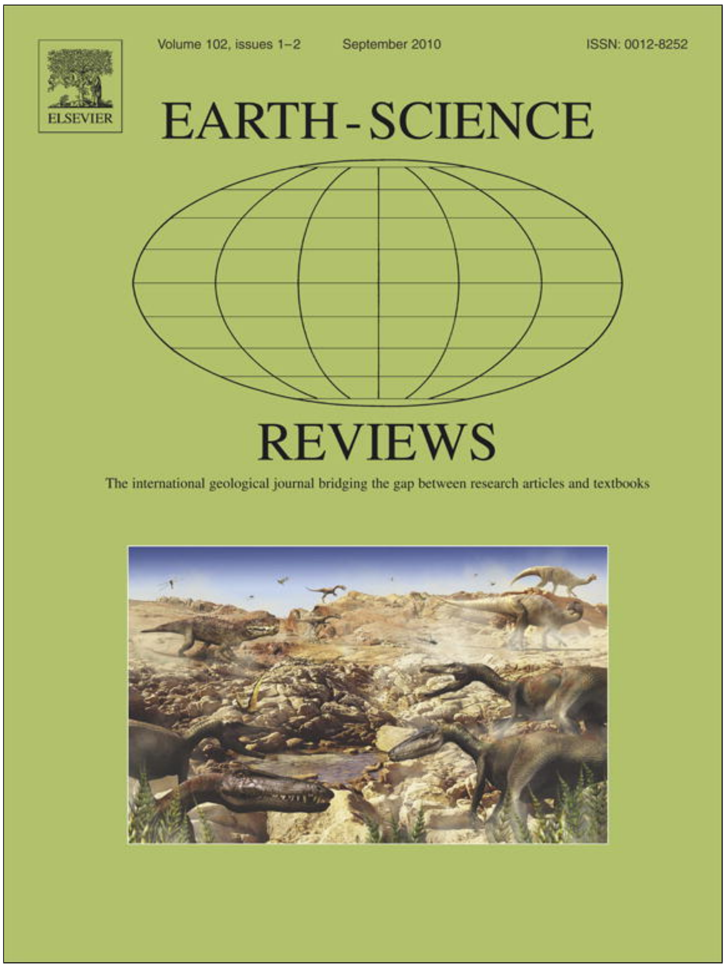
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A key extensional metamorphic complex reviewed and restored: The Menderes Massif of western Turkey

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ABSTRACT

This paper provides a review of the structure and metamorphism of the Menderes Massif in western Turkey, and subsequently a map-view restoration of its Neogene unroofing history. Exhumation of this massif — among the largest continental extensional provinces in the world — is generally considered to have occurred along extensional detachments with a NE–SW stretching direction. Restoration of the early Miocene history, however, shows that these extensional detachments can only explain part of the exhumation history of the Menderes Massif, and that NE–SW stretching can only be held accountable for half, or less, of the exhumation.

Restoration back to ~15 Ma is relatively straightforward, and is mainly characterised by a previously reported 25–30° vertical axis rotation difference between the northern Menderes Massif, and the Southern Menderes Massif and overlying HP nappes, Lycian Nappes and Bey Dağları about a pivot point close to Denizli. To the west of this pole, the rotation was accommodated by exhumation of the Central Menderes core complex since middle Miocene times, and to the east probably by shortening.

At the end of the early Miocene, the Menderes Massif formed a rectangular, NE–SW trending tectonic window of ~150×100 km. Geochronology suggests unroofing between ~25 and 15 Ma. The north-eastern Menderes Massif was exhumed along the early Miocene Simav detachment, over a distance of ≤50 km. The accommodation of the remainder of the exhumation is enigmatic, but penetrative NE–SW stretching lineations throughout the Menderes Massif suggest a prominent role of NE–SW extension. This, however, requires that the eastern margin of the Menderes Massif, bordering a region without significant extension, is a transform fault with an offset of ~150 km, cutting through the Lycian Nappes. For this, there is no evidence. The Lycian Nappes — a non-metamorphic stack of sedimentary thrust slices and an overlying ophiolite and ophiolitic mélange — have been previously shown to thrust to the SE between 23 and 15 Ma over at least 75 km. This is contemporaneous with, and orthogonal to stretching along the Simav detachment. I here argue that the amount of SE-wards displacement of the Lycian Nappes was twice the minimum amount of 75 km, which would restore them back on top of most of the Menderes Massif, apart from the ~50 km unroofed along the Simav detachment. A decollement was likely formed by a high-pressure, low-temperature metamorphosed nappe immediately underlying the Lycian Nappes in the north — the Ören unit. Latest Oligocene to early Miocene fission track ages of the Menderes Massif, as well as NE–SW trending lower Miocene grabens on the Massif are in line with this hypothesis.

The main implications of this restoration are that 1) the eastern part of the Aegean back-arc accommodated not more than 50 km of NE–SW extension in the early Miocene, and 2) any pre-Miocene exhumation of the Menderes Massif cannot be attributed to the known extensional detachments. The restoration in this paper suggests that most of the Menderes Massif already resided at upper crustal levels at the inception of extensional detachment faulting, a situation reminiscent of the role of extensional detachments on the island of Crete.

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1. Introduction

Much remains to be understood about continental extension and exhumation of metamorphic rocks to the surface, particularly in subduction zone settings. The eastern Mediterranean orogenic belt between the Eurasian and African–Arabian plates (Fig. 1) has been instrumental in the development and the testing of exhumation scenarios of high-pressure, low temperature (HP–LT), as well as high-temperature, low pressure (HT–LP) metamorphic rocks (Lister et al., 1984; Buick and Holland, 1989; Thomson et al., 1999; Jolivet et al., 2003, 2009, 2010a; Ring et al., 2007a, b, 2010; Brun and Faccenna, 2008; Ring and Kumerics, 2008; Jolivet and Brun, 2010). There is a general agreement that exhumation of metamorphic rocks in subduction systems normally occurs in two stages, the first one driven by buoyancy (or upward extrusion) of metamorphosed rocks along the subduction zone (Chemenda et al., 1995; Jolivet et al., 2003; Ring and Layer, 2003; Ring et al., 2010), followed by 'classical' core complex style exhumation during crustal extension (Crittenden et al., 1980; Wernicke, 1981; Davis, 1983) to transport the rocks from mid-crustal depth to the surface.

Critical to the analysis of the geodynamic and kinematic evolution of crustal extension in the eastern Mediterranean is the Aegean–west Anatolian extensional region, comprising the Greek Cycladic Massif, and western Turkish Menderes Massif, which is among the best studied continental extensional provinces in the world. It is the purpose of this paper to provide the first map-view restoration of western Turkey, and in particular of the portion of exhumation of the Menderes Massif that can be ascribed to extensional detachments since the late Oligocene.

The Menderes Massif is exposed in a tectonic window of approximately 200 × 100 km exposing metamorphic rocks that were derived from a micro-continental block (the Anatolide–Tauride Block) that underwent Eocene underthrusting below, and collision with the Sakarya continent of northwestern Turkey (belonging to Eurasia since the Mesozoic) (Şengör and Yılmaz, 1981; Kaymakci et al., 2009; Torsvik and Cocks, 2009). It exposes, and is overlain by metamorphosed rocks that provide evidence for periods of HP–LT, as well as HT–LP metamorphism (Oberhänsli et al., 1997; Candan et al., 2001; Rimmelé et al., 2003b; Whitney et al., 2008). Timing of the metamorphic events is highly controversial, and will be reviewed below, but consensus exists that during the Cenozoic, the Menderes Massif experienced at least regional greenschist-facies metamorphism (Gessner et al., 2001c; Ring et al., 2003; Bozkurt, 2007). Neogene exhumation of these rocks to the surface has been attributed to the activity of extensional detachments, with preserved NE–SW to N–S stretching lineations (Şengör et al., 1984; Hetzel et al., 1995a, b; Bozkurt, 2000; Bozkurt and Oberhänsli, 2001; Işık and Tekeli, 2001; Ring et al., 2003; Seyitoğlu et al., 2004), in line with the general stretching trend of the greater Aegean extensional province (Jolivet et al., 2004; Jolivet and Brun, 2010b; Tirel et al., 2009; Ring et al., 2010). The Massif is surrounded by metamorphosed and non-metamorphosed older, structurally higher thrust slices. These are (1) the Bornova Flysch zone in the NW, a chaotic mélange of late Cretaceous age that formed during accretion and subduction prior to underthrusting of the Menderes Massif (Okay and Altıner, 2007); (2) the HP–LT metamorphic Afyon zone, comprising coherent thrust slices with metasediments and Pan-African basement, which reached blueschist facies metamorphic conditions in latest Cretaceous to Paleocene times (350 °C/6–9 kbar; Candan et al., 2005; Pourteau et al., 2010) and the overlying, older and higher pressure Tavşanlı zone, which consists of metasedimentary, and mélange-like HP–LT series that includes intervals metamorphosed under blueschist and eclogite-facies metamorphism, with climax P–T conditions up to ~430 °C/20 kbar, exhumed after 88 Ma (Okay et al., 1998; Sherlock et al., 1999). The Tavşanlı zone is overlain by ophiolites of the İzmir–Ankara suture zone that demarcate the suture of a strand of the Neo-Tethyan ocean that separated the Sakarya and Anatolide–Tauride blocks in the Mesozoic (Şengör and Yılmaz, 1981; Moix et al., 2008). Metamorphic soles of the İzmir–Ankara zone are ~95–90 Ma (Önen and Hall, 2000; Çelik et al., 2006) marking the minimum age for the onset of subduction between the Anatolide–Tauride and Sakarya blocks; (3) in the west, the Menderes Massif is overthrusted by the HP–LT Dilek Nappe (500 °C/15 kbar; Ring et al., 2007b) and overlying Selçuk ophiolitic mélange, which are correlated to the Cycladic Blueschist unit and overlying mélange of the Aegean region (Candan et al., 1997; Oberhänsli et al., 1997, 1998; Ring et al., 1999b). These give $^{40}Ar/^{39}Ar$ ages of 42–32 Ma (Ring et al., 2007b). The Dilek–Selçuk nappe is overlain by klippen of the Ören HP–LT unit (see below) (Rimmelé et al., 2006); (4) in the south, the top of the Menderes Massif is formed by a metasedimentary sequence including upper Paleozoic to lower Mesozoic rocks (Erdoğan and Güngör, 2004) with metamorphic conditions of up to 550 °C/6–8 kbar (Whitney and Bozkurt, 2002), termed the Selimiye unit (or nappe) by Gessner et al., 2001b; Régnier et al., 2003). This unit is overlain by magnesiocarpholite-bearing, Paleozoic to Eocene HP–LT (up to 470–500 °C/12–14 kbar, Rimmelé et al., 2003b; Whitney et al., 2008) metasediments that are either considered to belong to the Menderes Massif (e.g. Bozkurt, 2007), or alternatively as a separate HP-nappe, correlated with the Dilek Nappe/Cycladic blueschist unit (Régnier et al., 2007). This unit is separated by a metamorphosed ophiolitic mélange that may be correlated to the Selçuk mélange (Régnier et al., 2007), from an overthrusting series of Mesozoic to Eocene low-grade, magnesiocarpholite-bearing metasediments (up to 400 °C/10–12 kbar, Rimmelé et al., 2005), classically included as a metamorphosed part into the Lycian Nappes, but recently separately identified as the Ören unit, correlated with the Afyon zone to the north (Pourteau et al., 2010). Preliminary $^{40}Ar/^{39}Ar$ ages suggested a latest Cretaceous age of metamorphism (Ring and Layer, 2003). Finally, the Ören unit is overlain by the Lycian Nappes in the south and east

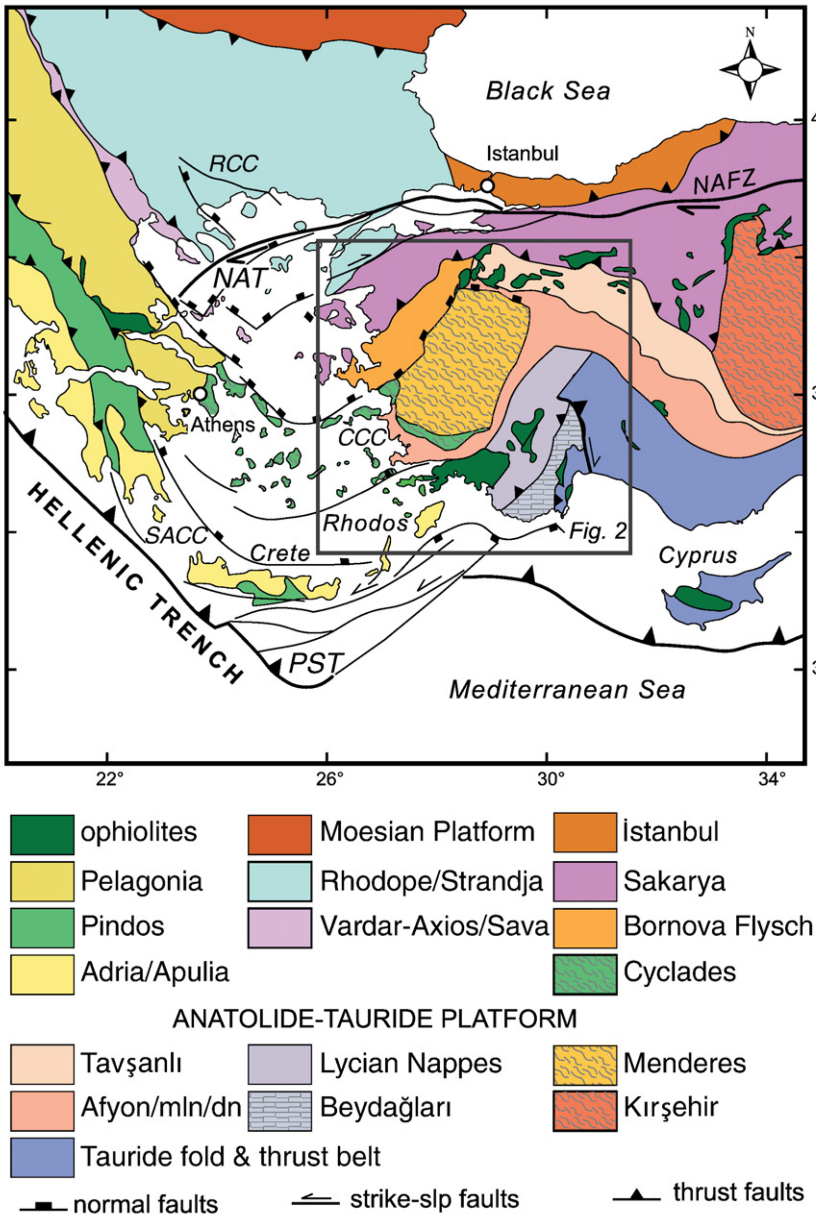

Fig. 1. Major terranes in the Aegean and Anatolian regions, modified after Moix et al. (2008). Afyon/mln/dn = Afyon zone, metamorphosed Lycian Nappes and Dilek Nappe; CCC = Cycladic Core Complex; NAFZ = North Anatolian Fault Zone; NAT = North Aegean Trough; PST = Pliny and Strabo Trenches (South Aegean left-lateral strike-slip system); RCC = Rhodope core complex; SACC = South Aegean Crystalline Complex.

(Bernoulli et al., 1974), characterised by NE–SW trending, SE verging thrust slices of Paleozoic to Paleocene sedimentary rocks, ophiolitic melange and ophiolites, correlatable with the Bornova Flysch zone and Izmir–Ankara ophiolites north and west of the Menderes Massif (Okay, 1989; Collins and Robertson, 1997, 1998, 1999; Rimmelé et al., 2003a, 2005, 2006; Candan et al., 2005). To the southeast the Lycian Nappes overthrust the Bey Dağları carbonate platform, which may be the unmetamorphosed equivalent of the Menderes Massif (Hayward, 1984b; Collins and Robertson, 1998; van Hinsbergen et al, 2010b; Fig. 2).

There is a wealth of published information on the structure, metamorphism, geochronology, stratigraphy and paleomagnetically determined vertical axis rotations of this region. This paper reviews the structure and metamorphism of the Menderes Massif and overlying HP–LT nappes, and attempts to provide a restoration of western Turkey back to ~25 Ma, which is generally considered as the time of the onset of extensional exhumation of the Menderes Massif (Bozkurt and Oberhänsli, 2001; Gessner et al., 2001a; Ring et al., 2003). The NE–SW trending stretching lineations (corrected for Miocene vertical axis rotations, van Hinsbergen et al., 2010a) that are found throughout the Menderes Massif, both on ductile-to-brittle extensional detachments as well as within only ductily deformed rock units support a generally accepted history of NE–SW extensional unroofing of the Menderes Massif (Bozkurt and Park, 1994; Régnier et al., 2003; Ring et al., 2003; Seyitoğlu et al., 2004). This paper, however, will show that such a simple NE–SW unroofing history of the Menderes Massif in early Miocene times is impossible to reconstruct because essential large-displacement transform faults that are required for such a scenario are absent. I will therefore present an alternative (adjusted) unroofing model, with major implications for the amount of exhumation that can be explained by the known extensional structures.

Structural, geochonological, stratigraphic and paleomagnetic constraints on the Neogene evolution of western Turkey are reviewed, followed by a reconstruction of the region back to ~15 Ma, which is relatively straightforward. This is followed by a new kinematic scenario for the early Miocene.

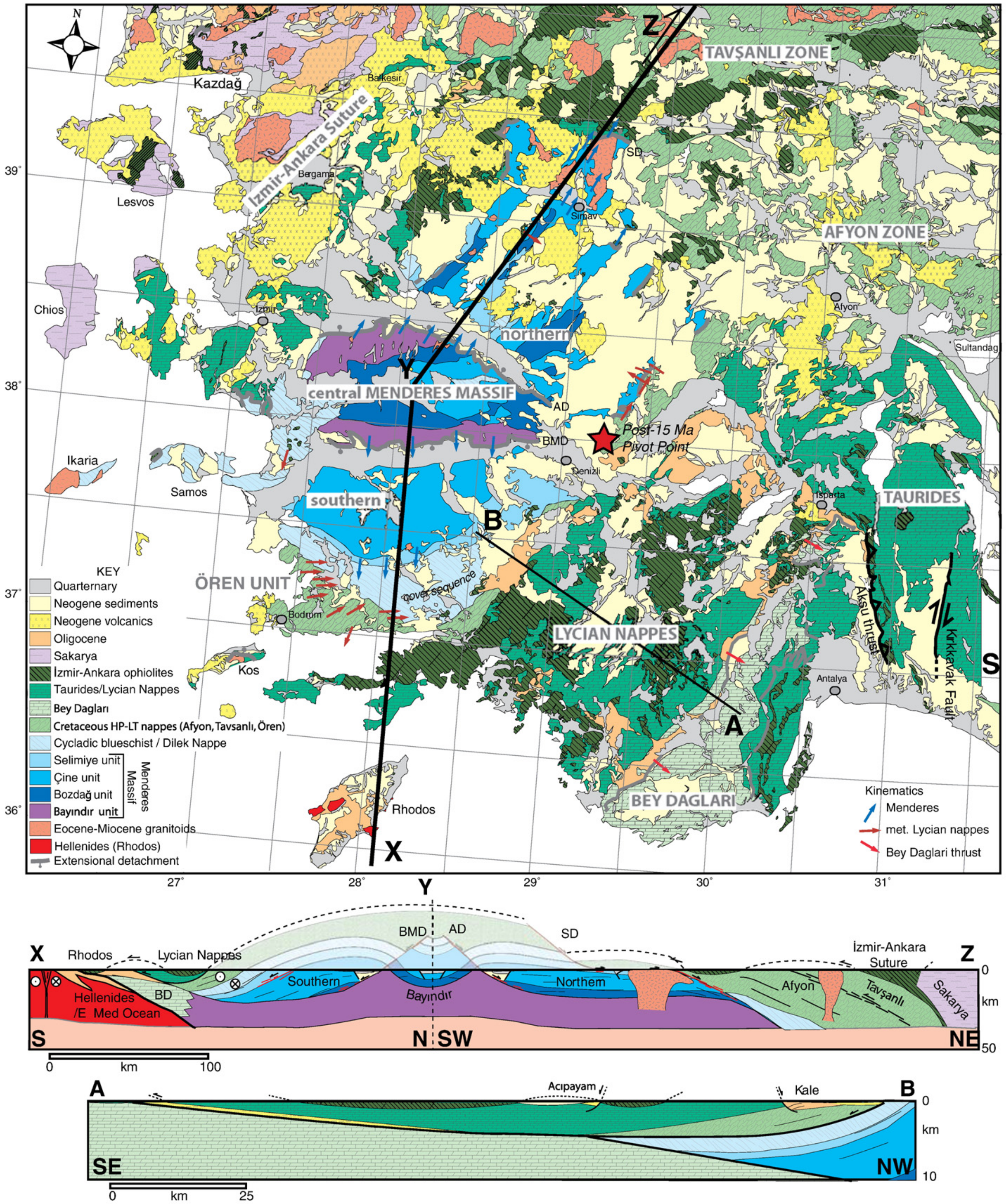

Fig. 2. Geological map of western Turkey, with schematic cross sections, based on the geological map of Turkey (MTA, 2002), with nappe subdivision in the Menderes Massif following Gessner et al. (2001b) and Régnier et al. (2007). The Dilek nappe in the southern Menderes Massif is based on the presence or absence of magnesiocarpholite and metabauxite, taken from the compilation of Rimmelé et al. (2003b). Arrows indicate shear sense with arrowhead pointing the direction of the tectonic top. AD = Alaşehir Detachment; BMD = Büyük Menderes Detachment; SD = Simav Detachment.

2. Regional geodynamic setting

The eastern Mediterranean region has been an active margin since at least Mesozoic times related to the closure of the Neo-Tethyan ocean, and the convergence between Africa and Eurasia since the Cretaceous (Dercourt et al., 2000; Stampfli and Borel, 2004; Barrier and Vrielynck, 2008; Torsvik et al., 2008). The structural and metamorphic history of the eastern Mediterranean region since at middle Cretaceous times resulted from convergence between Africa and Eurasia, and the accretion of rocks subducting continental or oceanic lithosphere (Şengör and Yılmaz, 1981; Faccenna et al., 2003; van Hinsbergen et al., 2005a; Hafkenscheid et al., 2006; Jolivet and Brun, 2010). The geology of western Turkey reflects the convergence since ~95–90 Ma between the Anatolide–Tauride block of Gondwanan origin (Kröner and Şengör, 1990; Hetzel and Reischmann, 1996; Gessner et al., 2004) in the south, and the Sakarya block that belonged to Mesozoic Eurasia (Topuz et al., 2007; Bozkurt et al., 2008; Okay et al., 2008a, b; Torsvik and Cocks, 2009) in the north (Figs. 1 and 2), and the accretion of thrust slices during subduction of the intervening branch of the Neo-Tethyan Ocean and the Anatolide–Tauride passive margin. The Anatolide–Tauride block collided with Sakarya in the Eocene (~40 Ma, Şengör and Yılmaz, 1981; Kaymakci et al., 2009) after which the Africa–Europe convergence was accommodated to the south, and western Turkey underwent exhumation and eventually Miocene extension (Şengör et al., 1984).

The general consensus is that lithospheric extension in the Aegean and west Anatolian region started approximately 25 Ma ago (Gautier et al., 1999; Jolivet, 2001; Tirel et al., 2009; Ring and Glodny, 2010; Ring et al., 2010), if not before, e.g. in the Rhodope (Burchfiel et al., 2008; Jolivet and Brun, 2010; Georgiev et al., 2010), and that extension relates to roll-back of the Aegean slab (Le Pichon et al., 1982; Meulenkamp et al., 1988; Spakman et al., 1988). There has been a long-standing debate on the role of westward extrusion of Anatolia as a result of Arabian indentation further east as cause for N–S extension in the region (Dewey and Şengör, 1979; Şengör et al., 1985), but the recent recognition of a late Miocene age for this collision (Keskin, 2003; Hüsing et al., 2009), an 11 Ma onset of formation of the North Anatolian Fault Zone (Hubert-Ferrari et al., 2002; Şengör et al., 2005), and the fact that the bulk of its displacement is Pliocene in age (Hubert-Ferrari et al., 2009) shows that this process has made little contribution to the early Miocene history of the Aegean and west Anatolian regions.

In modern times, the Aegean subducting slab is delimited in the southeast along a NE–SW trending subduction transform edge propagator (STEP) fault that accommodates southwestward roll-back of the Aegean slab with respect to the African lithosphere (Govers and Wortel, 2005). The surface expression of this STEP fault is formed by the left-lateral Pliny and Strabo strike-slip faults in offshore southeastern Greece (Peters and Huson, 1985; Mascle et al., 1999; Woodside et al., 2000; Fig. 2), which started some 5 Ma ago (Woodside et al., 2000; van Hinsbergen et al., 2007; Zachariasse et al., 2008). Based on the sharp transition of extension in the Menderes Massif to no Miocene extension to its east, van Hinsbergen et al. (2010a) postulated that roll-back of the Aegean slab was already bounded by this slab edge since the early Miocene, and that the eastern limit of the Menderes Massif may have been bounded by a STEP fault in this time.

3. Geological history of the Menderes Massif

3.1. Pre-Neogene structure and metamorphism of the Menderes Massif

The Menderes Massif is a series of crystalline rock units with a complex Alpine, and partly Pan-African metamorphic, igneous and structural history (Bozkurt et al., 1993; Bozkurt and Park, 1994, 1997; Satır and Friedrichsen, 1986; Hetzel and Reischmann, 1996; Loos and Reischmann, 1999; Bozkurt and Oberhänsli, 2001; Ring et al., 2003; Gessner et al., 2001b, 2004). Schuiling (1962) classically subdivided the Menderes Massif into a 'core' of augen gneisses, and a metasedimentary 'cover'. This simple subdivision has been shown to be incorrect (e.g. Bozkurt et al., 1993; Gessner et al, 2001b). Among other lines of evidence, the most compelling argument against the classical core-cover interpretation was found by Gessner et al. (2001b), who showed that the deepest structural level of the Menderes Massif, consists of (rudist-bearing, Özer and Sözbilir, 2001) metasedimentary rocks — termed the Bayındır Nappe — which have the lowest metamorphic grade in the Menderes Massif (greenschist-facies, biotite grade, Gessner et al., 2001b; Okay, 2001). These are overlain by higher-grade, and older units, the foliation of which was intruded by Triassic and Pan-African granitoids (Gessner et al., 2001c, 2004). The contact between the Bayındır unit and the overlying units is hence by definition a thrust postdating the late Cretaceous (based on the rudists). Three lithostratigraphic, regionally mappable units comprise the higher parts of the Menderes Massif, which were also considered to be Alpine nappes by Gessner et al (2001b): the metasedimentary Bozdağ unit, with metamorphic conditions of unknown age reaching ~530 °C/8 kbar (Okay, 2001). The foliation of the Bozdağ unit is intruded by Triassic granitoids (Gessner et al., 2001c). It is overlain by the Çine unit that contains the augen gneisses recognised by Schuiling (1962), and finally the Selimiye unit mentioned in the introduction.

The augen gneisses of the Çine unit have Pan-African (~550 Ma) protoliths (Hetzel and Reischmann, 1996; Loos and Reischmann, 1999) and have intrusive relationships with metasedimentary rocks of the Çine unit (Bozkurt et al., 1993). Both lithologies have a generally well-developed foliation that is regionally consistent across most of the Menderes massif, and is flat-lying, with a very consistent N(E)–S(W) stretching lineation associated with top-to-the-north sense of shear (Fig. 2) (Bozkurt and Oberhänsli, 2001). This foliation is in places intruded by weakly deformed Pan-African granites (Gessner et al., 2004). The Çine unit experienced regional HT–LP metamorphism (see e.g. Akkök, 1983; Bozkurt and Park, 1999; Bozkurt and Oberhänsli, 2001). Locally, the augen gneisses contain mafic inclusions metamorphosed to eclogite (640 °C/15 kbar), overprinted under Barrovian conditions (620 °C/7 kbar), associated with the regional HT–LP metamorphism of the Çine unit (Oberhänsli et al., 1997, in press; Candan et al., 2001). Both the eclogites, as well as the metasediments of the Çine unit provide geochronological evidence for Pan-African metamorphism, e.g. through dating of monazite inclusions in garnet (Catlos and Çemen, 2005; Oberhänsli et al., in press). $^{40}Ar/^{39}Ar$ and Rb/Sr cooling ages, which are exclusively Eocene or younger, date cooling below ~350 °C (Satır and Friedrichsen, 1986; Bozkurt and Satır, 2000; Lips et al., 2001), and are unable to discern between Alpine greenschist or amphibolite-facies metamorphism of the Çine and deeper units. No conclusive evidence exists in support of Alpine amphibolite-facies metamorphism in the Menderes Massif to date.

The Selimiye unit has a Permo-Triassic stratigraphy (Erdoğan and Güngör, 2004) and its metamorphism — with conditions of up to 550 °C/6–8 kbar (Whitney and Bozkurt, 2002), or slightly lower pressures (4 kbar, Régnier et al., 2003) — is associated with the Alpine orogeny (Gessner et al., 2001b). The Çine–Selimiye contact may have been an unconformity between Pan-African basement and Tethyan sediments, but is at present a top-to-the-south shear zone containing Eocene granitoid sills (Bozkurt, 2004), that may be extensional (Bozkurt and Park, 1994), compressional (Ring et al., 2003) or both (Bozkurt, 2007). Aside from the Pan-African eclogites, the four units of the Menderes Massif are devoid of evidence for high-pressure metamorphism. Overlying the Selimiye unit, however, is a metasedimentary sequence with a Paleozoic to Eocene stratigraphy (Konak et al., 1987; Özer, 1998; Özer et al., 2001), which is classically incorporated in the 'Menderes cover sequence' (Oberhänsli et al., 1998; Bozkurt and Oberhänsli, 2001; Rimmelé et al., 2003a, b; Whitney et al., 2008). This sequence has magnesiocarpholite-bearing assemblages and contains metabauxite with a metamorphic field gradient from corund in the north to diasporite in the south (Schuiling, 1962; Konak et al., 1987; Yalçin et al., 1993; Hatipoğlu et al., 2010). Given its stratigraphic age, its metamorphism must be Alpine, and reached HP–LT conditions (up to 470–500 °C/12–14 kbar, Rimmelé et al., 2003b; Whitney et al., 2008). This unit is separated by a metamorphosed ophiolitic mélange that may be correlated to the Selçuk mélange of the overlying the Dilek nappe in the west (Ring et al., 2007a, b; Régnier et al., 2007), from another magnesiocarpholite-bearing metasedimentary unit (up to 400 °C/10–12 kbar, Rimmelé et al., 2005). The latter is classically included as a metamorphosed part into the Lycian Nappes, but recently separately identified as the Ören unit, correlated with the Afyon zone to the north (Pourteau et al., 2010). Foliations in the Selimiye, 'Dilek', and Ören units are more or less parallel, and dip to the south, below the Lycian Nappes.

The contrasting metamorphic grades in the metasedimentary units overlying the Çine unit led to two hypotheses: (1) The Menderes Massif underlying the carpholite-bearing 'cover sequence' has experienced Alpine HP metamorphism, which was entirely overprinted by a regional HT–LP overprint (e.g. Bozkurt, 2007; Whitney et al., 2008). This hypothesis hence predicts that the Çine unit not only underwent Pan-African, but also Alpine amphibolite-facies metamorphism; (2) Alternatively, the contrast in peak pressure between the Selimiye unit and overlying carpholite-bearing successions led Gessner et al. (2001b) and Régnier et al. (2007) to postulate that the carpholite-bearing rocks should be correlated with the Dilek nappe of western Turkey.

In favour of the first hypothesis is the parallelism between the syn-HT–LP stretching lineations and the Miocene stretching lineations associated with Neogene detachment faults (see below). In addition, an Eocene high-temperature pulse was also concluded from the intrusion of a large number of granitoids in and volcanism on the nappes overlying the Menderes Massif in the north between ~50 and 35 Ma (Aldanmaz et al., 2000; Delaloyle and Erguzer, 2000; Dilek and Altunkaynak, 2009). This hypothesis, however, is troubled by absence of conclusive evidence for Alpine HT metamorphism in the Çine unit, and the presence of the weakly deformed pan-African and Triassic granites intruding the Çine and Bozdağ units' foliations (Gessner et al., 2001c, 2004).

In favour of the second hypothesis is the similar metamorphic grade of the Dilek nappe (500 °C/15 kbar; Ring et al., 2007b) and the carpholite-bearing successions south of the Menderes Massif (470–500 °C/12–14 kbar; Rimmelé et al., 2003a, b; Whitney et al., 2008), the presence of metabauxites in both (Schuiling, 1962; Konak et al., 1987; Yalçin et al., 1993; Ring et al., 2007b; Hatipoğlu et al., 2010), the presence of metamorphosed ophiolitic mélange overlying both, and the absence of evidence for HP metamorphism in the remainder of the Menderes Massif. In addition, it is in line with the intrusive relationship between weakly deformed Pan-African and Triassic granites and the Çine and Bozdağ unit's foliations (Gessner et al., 2001c, 2004). However, it includes the peculiar coincidence of parallel Pan-African and Alpine stretching lineations and foliations, and is difficult to reconcile with absence of evidence for intense deformation of the Çine and Bozdağ unit's foliations, despite Alpine thrusting and greenschist-facies metamorphism.

Both hypothesis, however, have to face the fact that rudist-bearing, low-grade rocks are at the lowermost structural level of the Menderes Massif (Gessner et al., 2001b), which inevitably indicates that the highest three units of the Menderes Massif were thrusted upon the Bayındır nappe in Alpine times, regardless of the grade of their Alpine metamorphism and deformation. The contacts between the Bozağ, Çine and Selimiye units may or may not be thrusts, and may or may not be Alpine.

In the map of Fig. 2, I have indicated the presumed Dilek nappe south of the Menderes Massif, based on the presence or absence of carpholite findings (mainly based on Rimmelé et al., 2003a, b), as well as the four units of the Menderes Massif identified by Gessner et al. (2001b). It is up to the reader to decide which hypothesis is most likely. For the remainder of this paper, which focuses on the Neogene

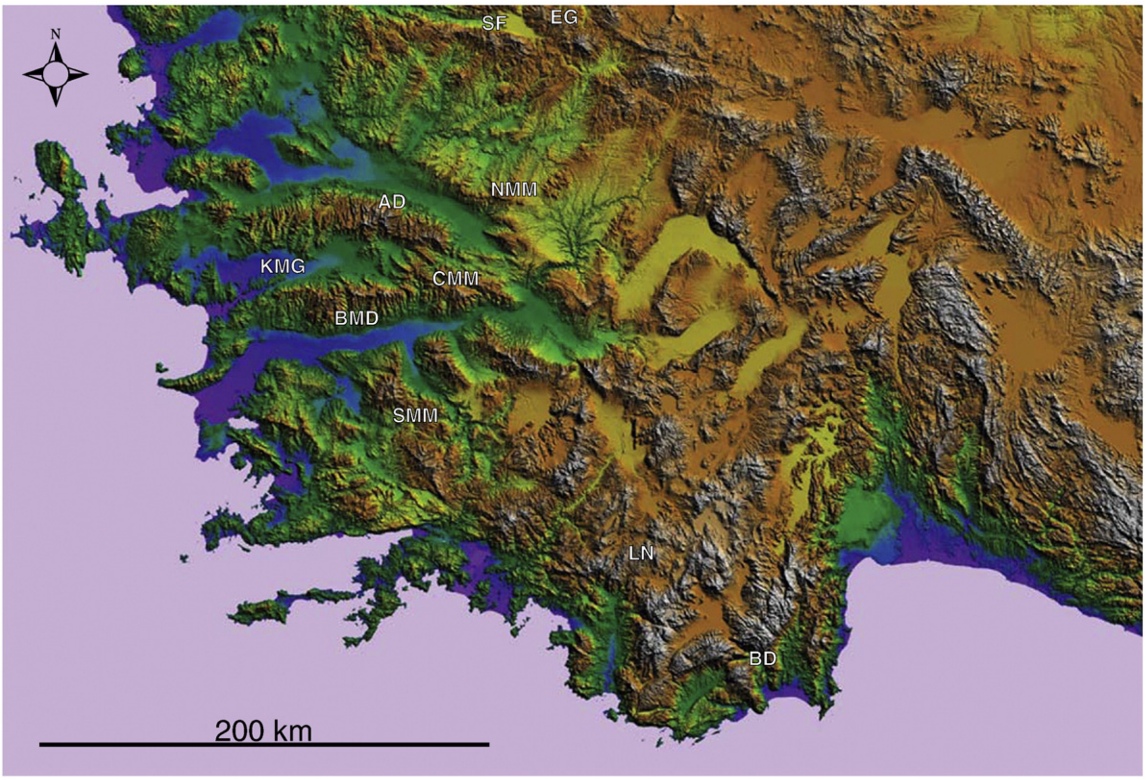

Fig. 3. Topography of western Turkey, with some reference features for comparison with the geological map of Fig. 2. AD = Alaşehir detachment; BD = Bey Dağları; BMD = Büyük Menderes Detachment; CMM = Central Menderes Massif; EG = Eğrigöz granite; NMM = Northern Menderes Massif; KMB = Küçük Menderes Basin; LN = Lycian Nappes; SF = Simav high-angle normal fault; SMM = Southern Menderes Massif.

history of western Turkey, the controversies outlined above are not relevant, and at this stage, I have no preferred interpretation.

3.2. Neogene extensional detachments and exhumation

Neogene extension divided the Menderes massif into three sub-Massifs, the Northern (or Gördes), Central, and Southern (or Çine) Menderes Massif (N-, C- and SMM), respectively (Figs. 2 and 3). The NMM and SMM record late Oligocene to latest Early Miocene zircon and apatite fission track ages (~27–16 Ma, with zircon fission track ages generally 2–3 Ma older than the apatite fission track ages), and the CMM exhumed mainly between ~16 and 5 Ma (Gessner et al., 2001a; Ring et al., 2003; Thomson and Ring, 2006).

The exhumation history of the CMM is well-constrained along multiple lines of evidence. Two ductile-to-brittle extensional detachments were identified along the northern (Alaşehir detachment) and southern (Büyük Menderes detachment) margins of the CMM (Figs. 2 and 3). In the hanging walls of both detachments, supra-detachment basin sediments with early middle Miocene ages (16.6–14.6 Ma along the Alaşehir detachment and 16.0–14.9 Ma along the Büyük Menderes detachment (Sen and Seyitoğlu, 2009)) are dipping into the detachment surface (Işık et al., 2003). In the footwall of the Alaşehir detachment, granite intrusions with crystallisation ages of 16–13 Ma (Catlos and Çemen, 2005; Glodny and Hetzel, 2007; Catlos et al., 2010) are deformed by the detachment, and were interpreted to be syn-kinematic (Hetzel et al., 1995b; Gessner et al., 2001b). Andalusite in the contact metamorphic aureoles around those granites was used to interpret an intrusion depth not exceeding 12 km (Ring et al., 2003). This is in line with the $^{40}Ar/^{39}Ar$ white mica age of 36 ± 2 Ma from rocks below the Büyük Menderes detachment, showing that the metamorphic rocks of the southern CMM were already cooled below the closure temperature for argon diffusion in white mica when middle Miocene detachment faulting started (Lips et al., 2001). The comparable ages between the granite and the supra-detachment basin sediments on either side of the Alaşehir detachment shows extensional exhumation from ~12 km depth since ~16 Ma.

The central axis of the CMM exhibits the Küçük Menderes Graben, containing volcanic deposits with ages of 14.6–13.9 Ma (Bozkurt et al., 2009). Together with early Miocene fission track ages in the central axis of the CMM, which young to Pliocene and younger ages to the north and south (Ring et al., 2003), this shows that the post-16 Ma extension only exhumed the flanks of the CMM.

Finally, the CMM pinches out eastward, and the Alaşehir and Büyük Menderes detachments converge to a point around Denizli (Fig. 2), defining an angle of ~30°. A comparable angle is found between the stretching lineations of the Alaşehir and Büyük Menderes detachments (Hetzel et al., 1995a; Bozkurt and Satır, 2000; Lips et al., 2001; Okay, 2001; Ring et al., 2003; Fig. 2). Paleomagnetic results from lower Miocene volcanics on the NMM, and upper Oligocene to uppermost lower Miocene sediments in the Lycian Nappes and on Bey Dağları demonstrate a rotation difference of ~25–30° since ~16 Ma, and have been used to argue for a syn-exhumational vertical axis rotation episode between the SMM and NMM during exhumation of the CMM around a pole north of Denizli (van Hinsbergen et al., 2010a, b; Fig. 2). It should be noted that the southern rotating domain runs further east than the pole near Denizli, and is likely delimited in the east along the middle to upper Miocene Aksu thrust and Kırkkavak right-lateral strike slip fault (van Hinsbergen et al., 2010a, b a, b; Fig. 2). Rotation east of the pole near Denizli is likely associated with contraction within the Lycian Nappes (such as for instance documented by post-early Miocene thrusting in the Denizli basin (Sözbilir, 2005).

This lateral variation in total amount of middle Miocene and younger extension in the CMM is a common phenomenon in most other extensional complexes that exhumed crystalline rocks in the Aegean region, including the Southern Rhodope core complex (Dinter and Royden, 1993; Sokoutis et al., 1993; Brun and Sokoutis, 2007; Georgiev et al., 2010), the culmination of Cycladic core complexes (Tirel et al., 2009; Jolivet et al., 2010a, b; Jolivet et al., 2010a), and the South Aegean crystalline complex of Crete and the Peloponnesos (van Hinsbergen et al., 2005c; Jolivet et al., 2010a, b). These all have lenticular shapes (Fig. 1), which have been associated with vertical axis rotation of the west-Aegean region in the course of the Neogene (van Hinsbergen et al., 2005b; 2008; Brun and Sokoutis, 2007; Georgiev et al., 2010).

The early Miocene configuration of the Menderes Massif, however, is markedly different. In contrast to the extensional systems mentioned above, the Menderes massif at the onset of the CMM exhumation and rotation episode, some 16 Ma ago, was deliniated along sharp boundaries on all sides, defining a NE–SW trending rectangle.

A large set of apatite fission track ages from the NMM and SMM show a range in ages from late Oligocene (~27 Ma) to early Miocene (~16 Ma; Ring et al., 2003; Thomson and Ring, 2006), indicating cooling below ~120–60 °C in this period of time. Along the northern margin of the NMM, the ductile-to-brittle Simav extensional detachment was identified by Işık and Tekeli (2001), which can be traced from the northern contact of the NMM with the overlying rocks of the Afyon zone and İzmir–Ankara ophiolites to the massifs southwest and southeast of Simav, over a distance of some 50 km parallel to the consistent NE–SW stretching lineations (Verge, 1993; Işık and Tekeli, 2001; Ring et al., 2003; Fig. 2). Thomson and Ring (2006) concluded that the detachment was active from ~25 to 19–18 Ma, and syn-kinematically with detachment activity, the large Eğrigöz granite intruded, which has crystallisation and $^{40}Ar/^{39}Ar$ cooling ages of 23–20 Ma (Işık et al., 2004; Ring and Collins, 2005; Figs. 2 and 3). The Eğrigöz granite is unconformably overlain by volcanics of ~15 Ma (Ercan et al., 1997; Işık et al., 2004). If all exhumation of the footwall can be attributed to the Simav detachment, and erosion played no role, then the modern 50 km width of the exposed Simav detachment corresponds to its displacement. If there was significant erosion after activity (as suggested by Thomson and Ring, 2006 and Hasözbek et al., 2010), then 50 km should be considered a maximum value. In the reconstructions in this paper I will use the maximum value of 50 km of NE–SW extension along the Simav detachment. To the northwest of the Menderes Massif, contemporaneous extensional exhumation was identified in the Kazdağ Massif (Bonev et al., 2009; Cavazza et al., 2009; Fig. 2), which will be accounted for in the restoration in this paper.

Apatite fission track ages of the SMM show that it cooled in the same period as the NMM, starting perhaps slightly earlier, but locating extensional structures that may account for its exhumation has proven to be problematic (Ring et al., 2003). As noted above, northern SMM exposes predominantly augen gneisses of the Çine unit with Pan-African granitic protoliths, separated from the Selimiye unit by a top-to-the-south ductile shear zone (the South Menderes shear zone, Bozkurt (2007); Fig. 2), which has been interpreted as an extensional shear zone (Bozkurt and Park, 1994; 1997) or as a thrust (Ring et al., 2003), or both (Bozkurt, 2007). Either way, this shear zone has no brittle component and can therefore not be held accountable for the cooling of the SMM recorded by the fission track ages of (Ring et al., 2003). The base of the Ören unit has also been proposed to be a candidate to have exhumed the SMM (Collins and Robertson, 1998, 2003; Güngör and Erdoğan, 2001; Régnier et al., 2003; Ring et al., 2003). A particular problem with this contact is that shearing in the metamorphosed Lycian nappes is associated with top-to-the-NE to E kinematics (Collins and Robertson, 2003; Rimmelé et al., 2003a). Therefore, Bozkurt and Park (1999) and Rimmelé et al. (2003a) interpreted these kinematics as indication of north-eastward back-thrusting after southward emplacement of the Lycian Nappes over the Menderes Massif. On the other hand, Régnier et al. (2003) interpreted these E-ward stretching lineations as rotated top-to-the-south kinematic indicators. These contradicting interpretations illustrate the notion of Ring et al. (2003) that unequivocal evidence for extensional deformation at the southern margin of the SMM is lacking.

The NMM and SMM are overlain by a series of grabens that trend parallel to the stretching lineations in their underlying basement (Bozkurt, 2003; Purvis and Robertson, 2004; Figs. 2 and 3). Especially the basins overlying the NMM are well-described, and contain volcano-sedimentary records with ages of ~21 Ma and younger (Bingöl, 1977; Ercan et al., 1996; Seyitoğlu, 1997; Purvis and Robertson, 2005; Ersoy et al., 2008) that rework leucogranite pebbles with crystallisation ages of 24–21 Ma (Seyitoğlu and Scott, 1992), in line with sedimentation during exhumation along the Simav detachment to the north and erosion of the footwall. The NW–SE extension direction of these basins has puzzled workers for a long time, and Bozkurt (2003) suggested that they form cross-grabens that formed in the hanging wall of the Alaşehir detachment. Although this scenario may valid since ~16 Ma, i.e. the onset of activity of the Alaşehir detachment, this cannot explain the first ~5 Ma if their evolution.

In summary, three unequivocal extensional detachments have been identified in the Menderes Massif, which have exhumed the entire CMM, and maximum 50 km of the NMM (Ring et al., 2003), with NE–SW stretching directions that rotated into a N–S direction in the south after 16 Ma (van Hinsbergen et al., 2010a). The exhumation of the remaining southern part of the NMM, and of the entire SMM to the surface in the early Miocene remains open for discussion. The timing of this exhumation, however, is well-constrained and occurred between ~27 and 16 Ma, and since ~21 Ma, sedimentary basins formed on the NMM (and SMM), apparently in part as a result of NW–SE extension (i.e. orthogonal to the general stretching direction preserved in the underlying metamorphic rocks).

4. The Lycian Nappes and the Ören unit

The Lycian Nappes comprise a stack of sedimentary thrust slices, below an ophiolitic mélange and ophiolite unit, which were accreted and thrusted during consumption of the northern margin of the Anatolide–Tauride platform in the İzmir–Ankara subduction zone, between ~90 Ma (the age of the metamorphic soles below the ophiolite (Celik et al., 2006)) and the Eocene (de Graciansky, 1972; Dürr et al., 1978; Gutnic et al., 1979; Okay, 1989; Collins and Robertson, 1997, 1998, 1999). These thrust slices have a consistent NE–SW strike, and Cretaceous to Eocene thrusting occurred in a SE direction (Okay, 1989; Collins and Robertson, 1997, 1998, 1999). The non-metamorphic and ultramafic Lycian Nappes overlie the above mentioned Ören unit.

In places, klippen of the Ören unit have been identified on top of the Menderes Massif (Rimmelé et al., 2006). These preserve magnesiocarpholite-bearing, HP–LT metamorphic mineral parageneses, without evidence for a HT overprint, similar to the rest of the Ören and Afyon units(Rimmelé et al., 2005), and in places also contain SE-directed lineations, although such isolated blocks may not have preserved their original position (Rimmelé et al., 2006; Fig. 2).

In the central Aegean region and as far east as Samos (Fig. 2), the Cycladic Blueschist unit is thrust upon the Lowermost Unit, generally considered to correlate to the Tripolitza unit of mainland Greece and Crete (Ring et al., 1999a, 2001b), which underthrusted the Aegean nappe pile in the Oligocene (Ring et al., 2001b; van Hinsbergen et al., 2005c), i.e. well after the collision and underthrusting of the Menderes Massif. The structural evolution and architecture of the transition of the Menderes Massif to the Aegean region remains poorly understood, and may include significant influence of strike-slip tectonics accommodating a westward increasing amount of extension (Ring et al., 1999a; van Hinsbergen et al., 2010a, Erkül, 2010; Kondopoulou et al., in revision). In the early Miocene, the Eocene and older Lycian nappe stack moved SE-wards, overthrusting the Bey Dağları parauthochtonous carbonate platform, as indicated by an Aquitanian to lowermost Langhian (~23–16 Ma) foreland basin succession below the Lycian thrustfront (Hayward and Robertson, 1982; Hayward, 1984b; van Hinsbergen et al., 2010b). This episode of thrusting ended in the lowermost Langhian, as evidenced by the presence of proximal, Lycian Nappes-sourced fan delta conglomerates in the top of the stratigraphy (Poisson, 1977; Gutnic et al., 1979; Hayward et al., 1996; Karabıyıkoğlu et al., 2005) and Serravallian sediments covering the thrust (Flecker et al., 1995; Poisson et al., 2003).

In a window near Göçek (Fig. 2), Brunn et al. (1970) and Hayward (1984a) found Burdigalian marls and turbidites, comparable in age and facies to the foreland basin deposits of Bey Dağları, and these authors therefore concluded a minimum of ~75 km SE-ward thrust displacement of the Lycian Nappes in early Miocene times.

The Lycian nappes are overlain by post-accretionary extensional sedimentary basins with late Oligocene and younger ages. The oldest sedimentary basins are found along the contact of the Ören unit and the Lycian nappes, and constitute the upper Oligocene–lower Miocene Kale–Tavaş, Denizli and Çardak–Dazkırı basins (Akgün and Sözbilir, 2001; Seyitoğlu et al., 2004; Sözbilir, 2005; Alçiçek et al., 2007). These basins form a NE–SW trending belt, and are limited in the SE by NW-verging normal faults. The contact of the basin fill with the underlying metamorphosed Lycian Nappes has not been described so far, and is covered by younger sediments (Fig. 2). Within the Lycian Nappes, early Miocene sedimentary basins (e.g. near Acıpayam, Fig. 2) formed as a result of NW–SE extension, comparable in direction with the Oligocene basin belt, and were interpreted as piggy-back basins that formed during SE-ward emplacement of the Lycian Nappes over the Bey Dağları platform in the early Miocene (Alçiçek and ten Veen, 2008; Alçiçek, 2010).

Paleomagnetic data showed that Bey Dağları underwent a counterclockwise rotation of ~25° (with respect to the Eurasian reference frame of Torsvik et al. (2008)) between 16 and ~5 Ma (Kissel and Poisson, 1986, 1987; Morris and Robertson, 1993; Tatar et al., 2002; van Hinsbergen et al., 2010b). Importantly, there is no evidence that this region experienced any significant rotations between late Cretaceous and middle Miocene times (van Hinsbergen et al., 2010b). From the upper Oligocene of the Kale basin, van Hinsbergen et al. (2010a) obtained a similar rotation amount, which suggests that the Lycian Nappes shared the Bey Dağları rotation history, and that early Miocene SE-ward emplacement did not involve significant (i.e. >~5°) vertical axis rotations.

In summary, the Lycian Nappes display evidence for SE-ward displacement over a distance of at least 75 km between ~23 and 16 Ma ago, i.e. contemporaneous with activity of the Simav detachment, and the cooling of the NMM and SMM.

5. Analysis: a restoration

Fig. 4 shows a series of snapshots of the reconstruction in 5 Ma intervals from the Present back to 25 Ma. For these reconstructions *GPlates* software (Boyden et al., in press), freely available at http://www.gplates.org/ was used. *GPlates* subdivides the surface into undeformable *polygons* and deformable *topologies*. For instance extending basins, core complexes, or folding nappes can be represented by topologies during their deformation. Moreover, *GPlates* interpolates plate motions with constant rates between known situations, leading to continuous deforming-plate reconstructions, allowing testing whether configurations can evolve from one to another with reasonable rates and processes. Polygons and topologies are defined strictly on geological criteria, and their definition, shape files and rotation files used to create Fig. 4a are given in the Online Appendix. The reconstructions in Fig. 4 are shown in a Eurasia-fixed reference frame.

5.1. Reconstruction back to ~15 Ma

During the last 5 Ma, western Turkey was affected by brittle faulting, and formation of several basins (see e.g. Seyitoğlu et al., 1992; Bozkurt, 2000, 2001; Sözbilir, 2001; Alçiçek et al., 2005, 2006; Alçiçek, 2007; Çiftçi and Bozkurt, 2009; ten Veen et al., 2009; Alçiçek, 2010). However, the displacement amounts associated with these faults are relatively minor, and have little influence on the scale of this reconstruction. The reconstruction at 5 Ma ago thus shows a similar geometry of the main units as today, corrected for

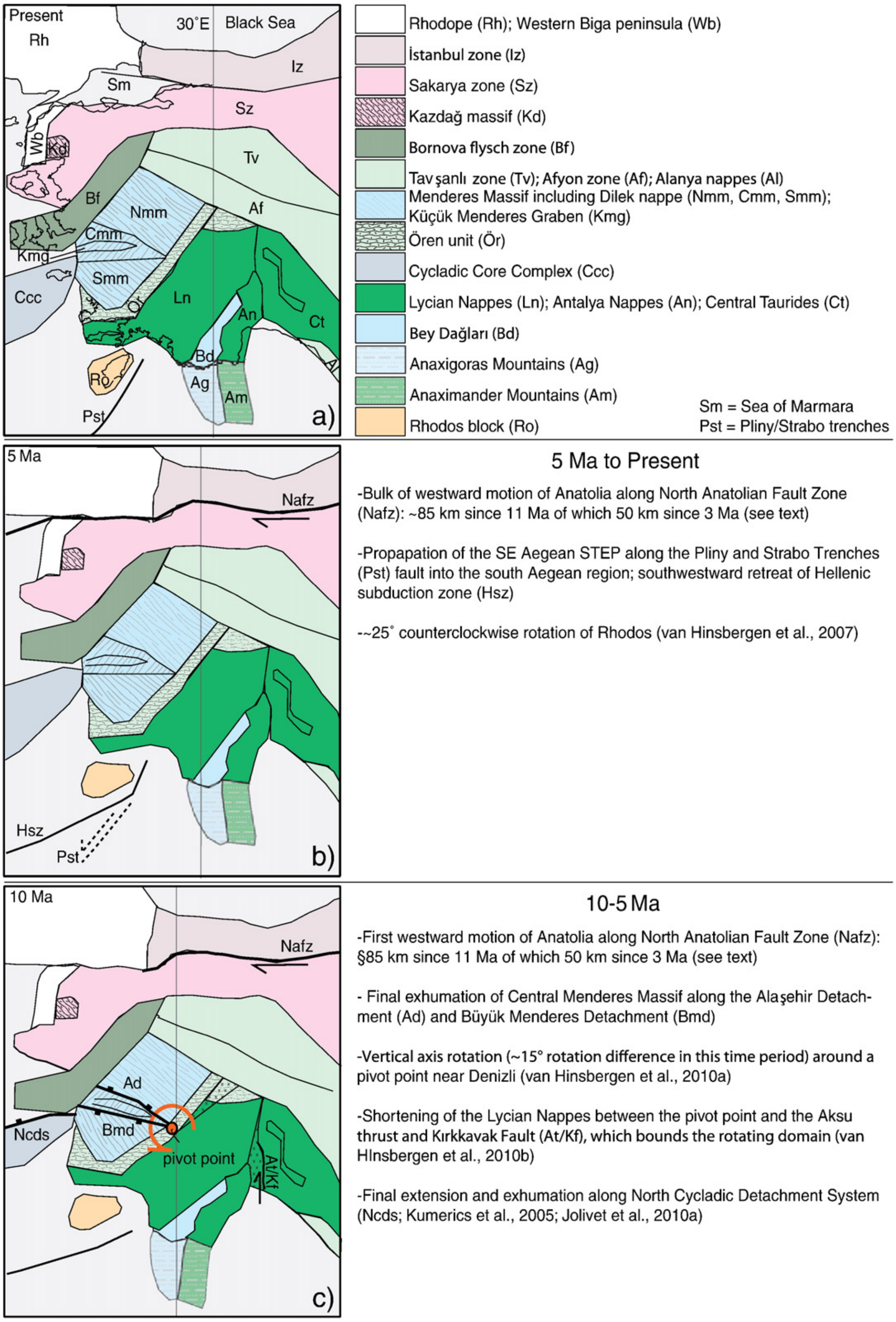

Fig. 4. Restoration of western Turkey since 25 Ma, in 5 Ma time slices. Basic geological map with main units based on Okay et al. (1996). Sea of Marmara is closed based on interpretations of Armijo et al. (1999). Western part of the Biga peninsula is included in the Rhodope block, following Beccaletto and Jenny (2004). Evolution of the Cycladic core complex (Ccc) is tentative, and will be subject of future restorations. A post-20 Ma opening of the eastern part of the Ccc is in line with geochronological data of Kumerics et al. (2005). Rotation of Rhodos is according to van Hinsbergen et al. (2007). Positions and affinity to onshore equivalents of the Anaximander and Anaxigoras mountains is based on Zitter et al. (2003) and ten Veen et al. (2004). Areas indicated with x-symbols at 10 and 15 Ma in the Lycian Nappes area indicated the area that is inferred to be consumed by convergence upon rotation of the SW-Anatolian domain (see van Hinsbergen et al., 2010a, b). The connection of the Hellenic subduction zone to the Cyprus subduction zone is enigmatic and will be subject of future reconstructions. For now, the reader is referred to ten Veen et al. (2004).

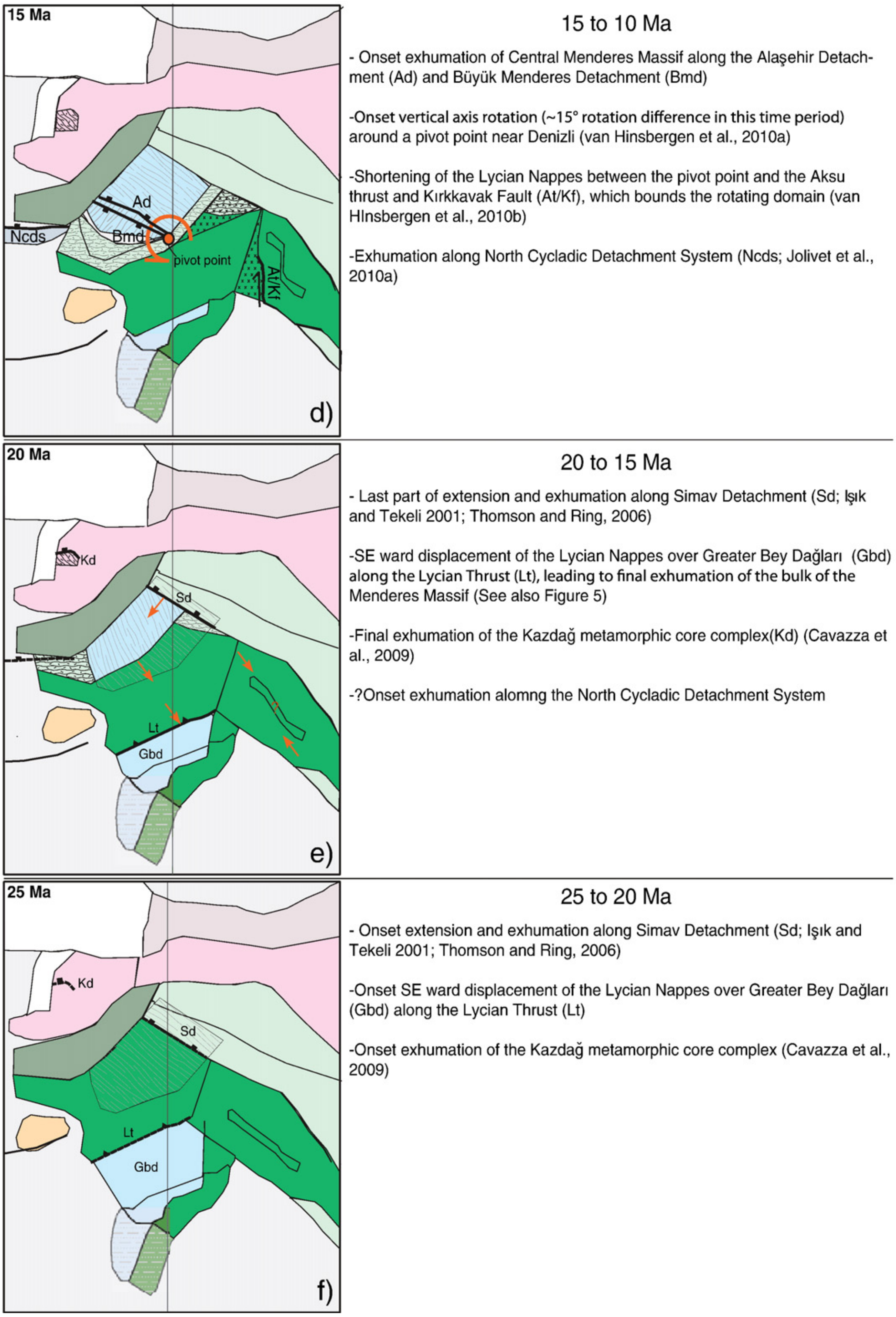

Fig. 4 (*continued*).

the motion along the North Anatolian Fault. For this motion, I adopted a total displacement of 85 km since 11 Ma following Şengör et al. (2005), of which 50 km was accommodated in the last 3 Ma, following Hubert-Ferrari et al. (2009). The rates are intrapolated between these ages. Paleomagnetic data of van Hinsbergen et al. (2010a) in the region north of the CMM show no evidence for post-early Miocene vertical axis rotation with respect to Eurasia, and I therefore do not adopt the conclusion of Kissel et al. (2003) for a Turkey-wide counterclockwise rotation history of ~20° since the late Miocene. The reconstructed 85 km of motion along the curved North Anatolian Fault Zone would lead to a counterclockwise rotation of ~2°, well within the typical error margins of paleomagnetism.

The reconstruction of Fig. 4d shows the west-Anatolian situation prior to the counterclockwise rotation of the SMM, Lycian Nappes and Bey Dağları. I assumed a constant rotation rate between 15 and 5 Ma, and the 10 Ma reconstruction of Fig. 4c is an intrapolation of the situations at 15 and 5 Ma.

The vertical axis rotation is, as outlined above, accommodated by the exhumation of the CMM, which consequently is closed at 15 Ma. The position of the central axis of the CMM and the Küçük Menderes basin between the CMM and NMM is outlined in Fig. 4c. Post-12 Ma N–S extensional exhumation on the island of Kos (van Hinsbergen and Boekhout, 2009) was not specifically corrected for due to its local nature.

East of the rotation pole near Denizli, the vertical axis rotation episode must have been accommodated by compressional deformation of the Lycian nappes. Two triangles in Fig. 4d indicate the total area that must have been consumed during this rotation. The eastern one was likely consumed by dextral transpression partitioned along the Aksu thrust and the Kırkkavak fault, which have been shown to be active in middle to late Miocene times (Poisson et al., 2003; Çiner et al., 2008). Because the exact boundary of the rotating domain in the north is not known, and few data are available on the structural evolution of the northeastern Lycian nappes in the Neogene, I have not specified the exact locations of accommodation of compression here, and let the Lycian Nappes topology deform as a whole. Post-Oligocene thrusting in the Denizli basin (Sözbilir, 2005), and significant clockwise rotations in the volcanic fields of Afyon along the northeastern margins of the Lycian nappes (Gürsoy et al., 2003), are in line with the prediction of (compressional) middle–late Miocene deformation in this region.

The reconstruction at 15 Ma outlines the configuration of the Lycian Nappes and the Menderes Massif at the end of the early Miocene. Back-rotation of the southern domains aligns the stretching lineations throughout the Menderes massif into a NE–SW orientation.

5.2. Identification of the early Miocene exhumation problem

The reconstructed situation at 15 Ma quickly defines the problems that arise from the — at first sight logical — interpretation that the early Miocene exhumation history of the Menderes Massif was only associated with NE–SW extension as suggested by the invariably NE–SW stretching lineations found throughout the Menderes Massif (e.g. Seyitoğlu et al., 2004).

Firstly, the width of the Menderes Massif around 15 Ma parallel to this dominant NE–SW stretching direction is 150 km. Around 25 Ma, most of the NMM and SMM were buried to at least several kilometers of depth as suggested by apatite and zircon fission track data (Ring et al., 2003; Thomson and Ring, 2006). It is unlikely that all exhumation can be contributed to erosion, mainly because it seems unlikely that the Menderes massif experienced many kilometers of erosion, whereas the surrounding (higher) mountain belts still preserve the highest structural units, formed by the ophiolites of the İzmir–Ankara zone. It is therefore likely that most exhumation is related to tectonic unroofing. The eastern boundary of the Menderes Massif is very abrupt, and borders with a region with no, or only a very minor record of extension. This boundary should therefore accommodate a total of 150 km of early Miocene NE–SW extension exhuming the Menderes Massif to the northwest from a region without extension to the southeast. In other words, this boundary should be a transform fault, offsetting the southwestern Lycian nappes over a distance of 150 km. Such a fault should crosscut the Lycian Nappes and separate the southwestern, offset Lycian Nappes from the eastern part. In modern coordinates, this strike-slip fault should be oriented approximately N–S, running from Denizli to Fethiye (Fig. 2). Such a fault would be hard to miss, and is evidently absent, especially illustrated by the continuous ophiolite belt that runs NE–SW over the Lycian Nappes. Even if only the proven displacement of the Simav detachment of maximum 50 km (Ring et al., 2003) is taken, this would still require a major strike-slip fault.

Secondly, the Lycian Nappes, and its thrustfront over Bey Dağları, run almost parallel to the early Miocene stretching direction in the Menderes Massif. There is no evidence for 150 km of extension in the northeastern part of the Lycian Nappes in early Miocene time, and it is therefore unlikely that the Lycian Nappes were coupled with the Menderes Massif during the early Miocene.

Thirdly, if I ignore the lack of evidence for a transform fault crosscutting the Lycian Nappes to accommodate the Menderes Massif stretching, and the Menderes Massif is restored to its pre-early Miocene history, I inevitably have to restore the island of Rhodos along with the Menderes Massif to the northeast. This would bring it into a position north of the Lycian Nappes. This is precluded by the fact that Rhodos was the location of an Oligocene foreland basin (Mutti et al., 1970), correlated across southern and western Greece linked to the northward subducting Hellenic slab. This would then have to bypass the Lycian Nappes, an impossibility.

Fourthly, closure of the Menderes Massif in a SW–NE direction would not bring the Lycian Nappes in contact with its equivalents along the NE and NW side of the Menderes Massif.

In other words, it is not possible to un-exhume the Menderes Massif with only NE–SW extension. This problem was recently also realised by ten Veen et al. (2009) who already suggested that N–S extension in the Menderes Massif was contemporaneous with southeastward translation of the Lycian Nappes.

5.3. A working hypothesis for the early Miocene unroofing history of the Menderes Massif

The foreland basin deposits on Bey Dağları of 23–16 Ma clearly show that the SE-ward motion of the Lycian nappes occurred contemporaneously with at least part of the activity of the Simav detachment (active between ~25 and 19 Ma according to Thomson and Ring, 2006). Adopting the constraint of at least 75 km of SE-ward movement of the Lycian Nappes provided by the Göçek window already places the Lycian Nappes on top of the Menderes Massif, covering half of the area exposed at 15 Ma. The similarity in shape of the NW and SE margin of the Menderes Massif at 15 Ma requires little imagination to suggest that the amount of displacement of the Lycian Nappes was approximately twice the distance from Göçek to the modern thrustfront. This would restore the Lycian Nappes back against the Karaburun peninsula and the Bornova Flysch zone in the NW, and link the carpholite bearing 'cover sequence', the overlying ophiolitic mélange and the Ören unit with their equivalents formed by the Dilek Nappe and Selçuk Mélange and the overlying Ören klippen in western Turkey. Moreover, it connects the Ören unit with the Afyon zone in the NE (Figs. 4 and 5).

In other words, during the NE–SW extension of max. 50 km along the Simav detachment, the Lycian Nappes slide southeastward, exhuming the remainder of the Massif in the early Miocene. This, however requires a decollement between the Lycian Nappes and the Menderes Massif, for the latter provides no evidence for SE-ward sliding (Fig. 2).

5.4. Kinematic requirements and predictions

The above scenario integrates the available temporal and kinematic evidence into a restorable tectonic history of the early Miocene unroofing of the Menderes Massif. However, this reconstruction also predicts deformation in regions where little kinematic information, or time-constraints on deformation are present.

The first major requirement for this reconstruction to be valid is that there was a decollement between the Menderes Massif and the Lycian Nappes that accommodated the SE-ward displacement of the latter. There is only one candidate for this decollement, as was already proposed by Collins and Robertson (2003), which is the contact of the Lycian Nappes with the Ören unit. The thickness of the Lycian Nappes was probably not much more than the modern maximum elevations of ~2000 m: the window of Göçek shows that in a segment near the coast the base of the Lycian nappes is still at sea level. Placing them

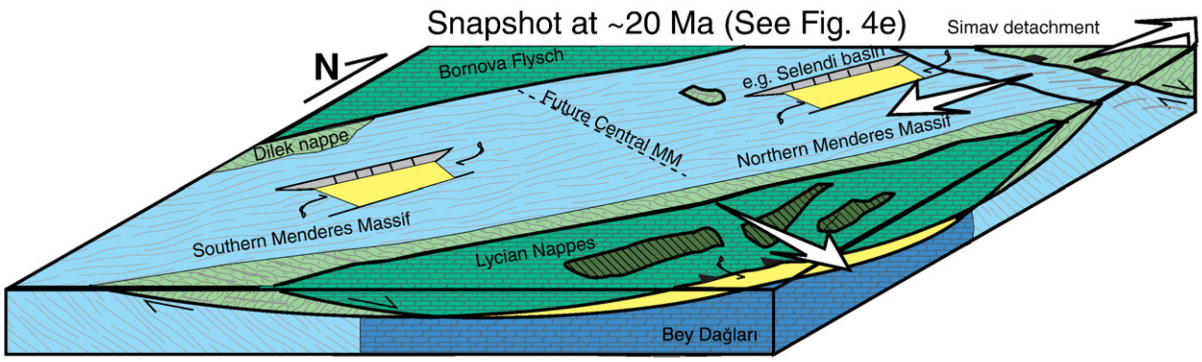

Fig. 5. Schematic block-diagram summarizing the early Miocene unroofing scenario proposed in this paper.

back to the northwest would thus bury the Menderes Massif would to a depth of approximately 2 km or, depending on the amount of thinning in Ören unit related to the SE ward sliding, and the amount of post-Oligocene erosion of the ophiolite sequence of the Lycian Nappes, more. This appears to be in line with the conclusion of Thomson and Ring (2006) that most of the northern Menderes Massif south of the Eğrigöz granite underwent 2–3 km of exhumation in early Miocene times based on zircon and apatite fission track ages. They ascribed this exhumation to erosion, I here suggest that it mainly corresponds to the SE-ward tectonic transportation of the Lycian Nappes. A limited amount of exhumation in early Miocene time of the bulk of the Menderes Massif, apart from the ~50 km wide region exhumed along the Simav detachment, as well as the younger CMM, may also explain the lack of a ductile Miocene overprint, and the preservation of Eocene and/or Pan-African NE–SW stretching lineations across the massif. In fact, most of the Menderes Massif would already have resided at upper crustal levels at the beginning of the Neogene. The brittle early Miocene history of the Menderes is entirely in line with NW–SE extension: the cross-grabens on the NMM and SMM with early Miocene basin fills recorded such extension directions.

The restoration proposed here predicts that the modern Bey Dağları platform was separated by approximately 150 km from the pre-Miocene thrustfront of the Lycian Nappes. This is in line with absence of significant deformation and ongoing carbonate deposition on the Bey Dağları platform throughout the late Cretaceous to Oligocene (Poisson, 1967, 1977; Farinacci and Köylüoğlu, 1982; Hayward and Robertson, 1982; Sarı and Özer, 2002; Sarı et al., 2004, 2009).

The Lycian Nappes continue further to the NE than the Bey Dağları platform, and because the early Miocene unroofing did probably not involve vertical axis rotation of the Lycian Nappes (van Hinsbergen et al., 2010a), an equal amount of early Miocene NW–SE shortening should have been accommodated in the Isparta Angle. Evidence for early Miocene shortening has not been reported, but cannot be excluded either: the sedimentary basins in the heart of the Isparta Angle near Yalvaç (Fig. 2) are middle Miocene and younger (Yağmurlu, 1991) and pre-middle Miocene compression is possible. Moreover, Fig. 2 highlights a curved belt of Jurassic limestones in the eastern limb of the Isparta angle, which suggests a larger amount of convergence in the north than in the south. The age of this deformation and curving is unknown, and it is tentatively ascribed to early Miocene in Fig. 4f.

Finally, the max. 50 km of extension accommodated by the Simav detachment should be represented in the northeastern Lycian Nappes, with an increasing amount of NE–SW stretching from the SE to NW, where they were overlying the Menderes Massif the longest (Fig. 5). Although the lack of detailed information does not allow more than speculation on this matter, the map pattern of the Isparta Angle may be in line with this suggestion: a triangular 'gap' exposing Ören Unit or Afyon Zone rocks is defined south of the line from the Isparta Angle to the Simav detachment. The opening of this angle is tentatively included in the reconstruction of Fig. 4.

5.5. Implications for exhumation mechanisms of metamorphic rocks

The reconstruction above argues that the bulk of exhumation of the Menderes Massif occurred prior to the activity of Neogene extensional detachments, and the formation of metamorphic core complexes of the CMM and along the Simav detachment. In fact, most of the Massif already resided in the upper crust at the inception of extensional detachment faulting, and the southeastward sliding of the Lycian Nappes.

This situation is reminiscent on the unroofing history of HP–LT rocks on the island of Crete (Fig. 1), where originally, an extensional detachment was held accountable for ~100 km of extension and exhumation from ~35 km depth or more (Jolivet et al., 1996; Ring et al., 2001b; van Hinsbergen et al., 2005c), pre-dating 15–13 Ma indicated by fission track ages (Thomson et al., 1998; Marsellos et al., 2010). Recently, however, Rahl et al. (2004) argued that the metamorphic contrast across this detachment may only explain some 6 km of exhumation, and the age of the Cretan supradetachment basin of 10.8–10.4 Ma shows that crustal extension commenced well after the bulk of exhumation occurred (Ring et al., 2001a; van Hinsbergen and Meulenkamp, 2006; Zachariasse et al., in revision). Comparable to western Turkey, the preservation of remnants of the entire nappe stack on Crete precludes a major contribution of erosion (Rahl et al., 2004; Zachariasse et al., in revision).

The available structural evidence in western Turkey, and the unlikelihood of a significant contribution of erosion fails to explain the exhumation of the Menderes Massif to upper crustal levels, and the analogy to Crete suggests that this is not a unique problem. As pointed out earlier in this paper, the amount of pre-Miocene exhumation of the Menderes Massif is a controversial subject, with advocates for depths equivalent to greenschist facies conditions (Gessner et al., 2001b, c; Ring et al., 2003; Régnier et al., 2007) or amphibolite-facies and perhaps even HP–LT conditions (Bozkurt, 2007). Whichever scenario one prefers, exhumation of the Menderes Massif to only a few kilometers of depth at the inception of the Miocene cannot be linked to the known extensional structures today. A solution for this problem remains open for discussion, but it seems that the available concepts cannot entirely explain exhumation of metamorphic rocks in the cases discussed here.

5.6. Geodynamic implications

An important consequence of the restoration in this paper is that the amount of early Miocene extension in the Aegean back-arc is much smaller than previously assumed. Only some 50 km or less of NE–SW extension can be accounted for in the early Miocene, as opposed to the previously assumed 150 km or more. This reconstruction therefore implies a slow rate of back-arc extension in the early Miocene, and a dramatic acceleration in the middle Miocene, associated with the curvature of the Aegean orocline (van Hinsbergen et al., 2005b, 2008). This consequence is in line with the consistently young fission track ages representing the timing of exhumation in the Cyclades of central Greece, which are invariably younger than 15 Ma (Hejl et al., 2002, 2008; Kumerics et al., 2005; Brichau et al., 2006, 2007, 2008, 2010). Testing this prediction will be subject of future restorations.

Finally, this reconstruction still allows for 50 km of NE–SW extension, with a sharp boundary east of the Menderes Massif, buried below the SE-ward slid Lycian Nappes. This boundary may be caused by roll-back of the Aegean slab, and represent transform fault accommodating SW-ward roll-back of a slab-edge (Govers and Wortel, 2005) below the eastern part of the Menderes Massif in early Miocene times.

6. Conclusions

This paper reviews the tectono-metamorphic history of western Turkey, and presents the first restoration of this region for the Neogene, contemporaneous with the exhumation of the major Menderes metamorphic massif to the surface. While making this restoration, it became evident that the general consensus on simple NE–SW unroofing of the Menderes Massif is impossible to reconstruct, and this paper provides therefore an alternative solution.

Reconstructing western Turkey back to approximately 15 Ma is relatively straightforward. Since the early middle Miocene, southwestern Turkey, including the Southern Menderes Massif, the Lycian Nappes and the Bey Dağları platform, underwent a ~25° counterclockwise rotation with respect to northwestern Turkey about a pole north of Denizli. To the west, this rotation was accommodated by the Central Menderes metamorphic core complex. To the east, the rotation must have been accommodated by compressional deformation in the Lycian Nappes. Previously documented distributed transpression in the Isparta Angle, and post-Oligocene thrusting in the Denizli basin are in line with this inference, but further work is needed to establish the exact accommodation of the predicted convergence.

The Northern and Southern Menderes Massifs were unroofed in the course of the early Miocene, and at ~15 Ma, it formed a rectangular shaped window, with abrupt and straight boundaries. Previous work has shown that only 50 km or less of this unroofing occurred along the top-to-the-northeast Simav detachment between 23 and 15 Ma. This, in combination with preserved (pre-Miocene) consistently NE–SW stretching lineations across the massif, and syn-exhumational thrusting of the Lycian Nappes over the Bey Dağları platform is usually logically interpreted to reflect simple NE–SW extensional unroofing of the Menderes Massif, associated with ~150 km of extension.

Such unroofing requires a major strike-slip fault along the eastern boundary of the Menderes Massif, that would need to cross-cut the Lycian Nappes. There is no such fault in the latter, and the generally agreed upon unroofing scenario is impossible to restore.

Instead, I here propose that simultaneously with NE–SW extensional exhumation along the Simav Detachment, over a distance of 50 km or less in the early Miocene, the Lycian Nappes slid SE-ward, i.e. orthogonal to the Simav detachment sense of shear. This sliding accounted for the final unroofing of the remainder of the Northern and Southern Menderes Massifs, and was accommodated along a decollement formed by the Ören unit, an upper Cretaceous HP–LT metamorphic metasedimentary sequence underlying the Lycian Nappes. The presence of NE–SW trending, lower Miocene extensional basins in both the Lycian Nappes and on the Menderes Massif support this hypothesis.

Two important implications of this reconstruction are that 1) early Miocene NE–SW extension in the eastern Mediterranean back-arc was limited to not more than 50 km and 2) exhumation of the southern, and most of the northern Menderes Massif south of the exposures of the Simav detachment from depths equivalent to greenschist-facies or higher-grade conditions cannot be attributed to the extensional detachment faults that are known today. The restoration in this paper, in combination with the available age constraints on the cooling history of the Menderes Massif suggests that most of it resided at upper crustal levels at the inception of extensional detachment faulting, a situation reminiscent of the exhumation history of the island of Crete. A solution for this problem remains open for discussion.

Finally, the early Miocene extension in the Menderes massif marks the onset of roll-back of the subducted Aegean slab. The sharp eastern boundary of roll-back related extension between the Menderes Massif and the rest of Turkey, now buried below the SE-ward slid Lycian Nappes, likely represents a transform fault that accommodated the SW-ward roll-back of the eastern Aegean slab-edge.

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Appendix A. Supplementary Data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.earscirev.2010.05.005.

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