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EDITORIAL

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# Menderes Massif (Western Turkey): structural, metamorphic and magmatic evolution – a synthesis

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**Abstract** The Menderes Massif covers large areas in western Turkey. The better understanding of its tectono-metamorphic history would provide insight for the Alpine evolution of western Turkey and the entire eastern Mediterranean region. This paper summarizes the available literature on the metamorphic rocks of western Turkey and that of the Menderes Massif with special reference and emphasis to the papers presented in the special issue.

## Tectonic units of western Turkey

Western Turkey comprises several continental fragments with distinctive stratigraphy, structural and metamorphic features. The amalgamation of the fragments occurred during the Early Tertiary continent–continent collision across the Neotethys. Major sutures define the boundaries of these fragments. They include the Intra–Pontide Suture, the İzmir–Ankara Suture and the Inner–Tauride Suture (Fig. 1; Okay and Tüysüz 1999).

The İstanbul Zone (Okay 1989a) consists of a Cadomian basement and an unmetamorphosed Lower Ordovician to Lower Carboniferous sedimentary sequence (Kaya 1973; Abdülselamoğlu 1977). It has been named the İstanbul Palaeozoic (Abdülselamoğlu 1977), the İstanbul Nappe (Şengör et al. 1984a) and the İstanbul Fragment (Ustömer and Robertson 1993). It is unconformably overlain by Mesozoic to Tertiary cover rocks. The non-metamorphic Palaeozoic sequence in the İstanbul Zone starts with Lower Ordovician–Lower Silurian quartzites and laminated shales. It continues with Upper Silurian–Lower Devonian limestone and fossiliferous shales with intercalated calci-turbutides. The sediments are conformably overlain by Upper Devonian nodular limestones and, in turn, by Lower Carboniferous radiolarian cherts and shallow marine flysch. The İstanbul Zone is distinguished from other continental fragments by its stratigraphy, absence of metamorphism and lack of deformation. Recently, the İstanbul Zone was divided, based on stratigraphy and deformation, into two tectonic units: namely the İstanbul and Zonguldak terranes (Göncüoğlu et al. 1996–1997; Göncüoğlu and Kozur 1998). The former is represented by a continuous Palaeozoic sequence affected by only Hercynian deformation, whereas the latter is represented by a Palaeozoic sequence, with a Devonian unconformity and Carboniferous coals, and underwent Caledonian deformation. The İstanbul Zone is separated from the Sakarya Zone by the Intra–Pontide Suture (Figs. 1 and 2). The belt has a complex history, which is not yet fully understood. The Sakarya Zone is composed of numerous tectonic units with clastic and volcanic rocks deformed and metamorphosed during the Late Palaeozoic and the latest Triassic and an unconformably overlying Jurassic–Cretaceous sedimentary sequence. The zone has also been termed the Sakarya Composite Terrane (Göncüoğlu et al. 1996–1997, 2000). Immediately south of the İzmir–Ankara Neotethyan Suture lies the Anatolide–Tauride platform made up of several tectonic units bounded by major faults (Fig. 1). These units include a blueschist belt (Tavşanlı Zone), the Bornova Flysch Zone – consisting of large Mesozoic limestone blocks within a matrix of Maastrichtian–Palaeocene greywacke-shale (Okay and Siyako 1993), the Afyon Zone (a Palaeozoic–Mesozoic sedimentary sequence metamorphosed in to the greenschist facies; Okay 1984a; Özcan et al. 1988; Göncüoğlu et al. 1992), the Menderes Massif (a Pre-

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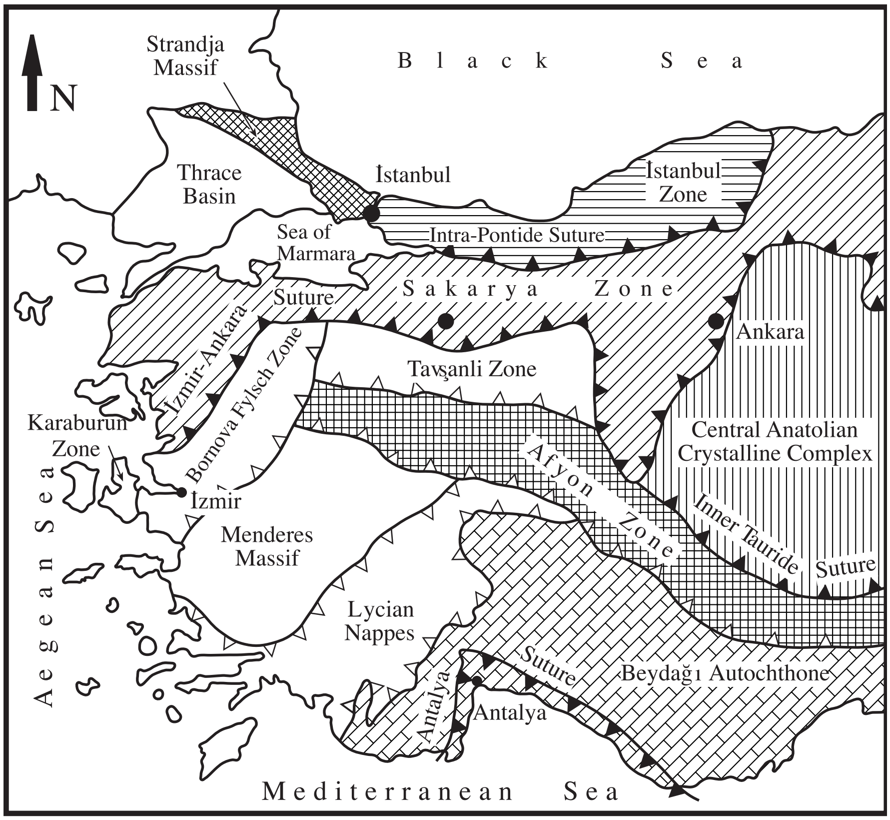
Fig. 1 Simplified tectonic map showing the location of the main Tethyan sutures and neighbouring tectonic units in western Turkey (from Okay and Tüysüz 1999). *Heavy lines with filled triangles* show sutures, and tips of the triangles indicate polarity. *Heavy lines with open triangles* indicate thrust belts, with triangles pointing in the direction of vergence




cambrian gneissic basement and the structurally overlying Palaeozoic–Palaeocene sediments metamorphosed at greenschist- to amphibolite-facies conditions) and the Lycian Nappes, composed of Mesozoic sedimentary sequences and a peridotite thrust sheet (Graciansky 1972; Ricou et al. 1975; Collins and Robertson 1997, 1998, 1999).

## Metamorphic rocks of western Turkey

The metamorphic rocks in western Turkey are widely distributed. They form distinct belts, based on their ages and degree of main metamorphism (latest event), and seven belts have been distinguished: (1) Precambrian greenschist- to amphibolite-facies and eclogite- and granulite-facies rocks; (2) Triassic greenschist-, blueschist- and eclogite-facies rocks; (3) Late Jurassic greenschist-facies rocks; (4) Cretaceous blueschist- to eclogite-facies rocks, (5) Cretaceous greenschist- to amphibolite-facies rocks; (6) Palaeocene–Eocene greenschist- to amphibolite-facies rocks and (7) Oligocene amphibolite-facies rocks (Fig. 2).

### Precambrian greenschist- to amphibolite-facies rocks (Bolu Massif)

The Bolu Massif forms the basement to the Palaeozoic rocks of the İstanbul Zone. This massif lies beneath Lower Ordovician continental clastics made up of conglomerate and coarse sandstone (Dean et al. 1997). The Massif is divided into three tectonic units (Ustömer and Rogers 1999): (1) a north-dipping thrust sheet made up of high-grade metamorphic rocks (Sünnice Group). The unit is represented by a migmatitic assemblage consisting of amphibolites with quartzo-feldspathic layers and thick gneisses with minor amphibolite alternations, cut by metagranitoids. The characteristic mineral assemblage is plagioclase, amphibole, feldspar and biotite. The replacement of amphibole and plagioclase by actinolite, chlorite and epidote minerals suggest that the Massif underwent an initial amphibolite-facies metamorphism, followed by a retrograde greenschist-facies overprinting. (2) Two calc-alkaline, volcanic-arc metagranitoids (Bolu Granitoid Complex) intruded the overlying (3) sequence of metavolcanics and volcanoclastic rocks composed of andesitic and minor rhyolitic lavas with metaignimbrites (Çaşurtepe Formation). The characteristic mineral assemblage of albite, chlorite, actinolite and quartz suggest that the volcanic sequence was metamorphosed at greenschist-facies conditions. The presence of metamorphic fragments in the unconformably overlying Ordovician clastics suggests a pre-Early Ordovician age for the rocks and the metamorphism of the Bolu Massif (see Dean et al. 1997 and Ustaömer and Rogers 1999 for details). More recently, Satır et al. (2000) documented $^{207}Pb/^{206}Pb$ single zircon ages of 590–570 Ma and 1,860–710 Ma from the gran-

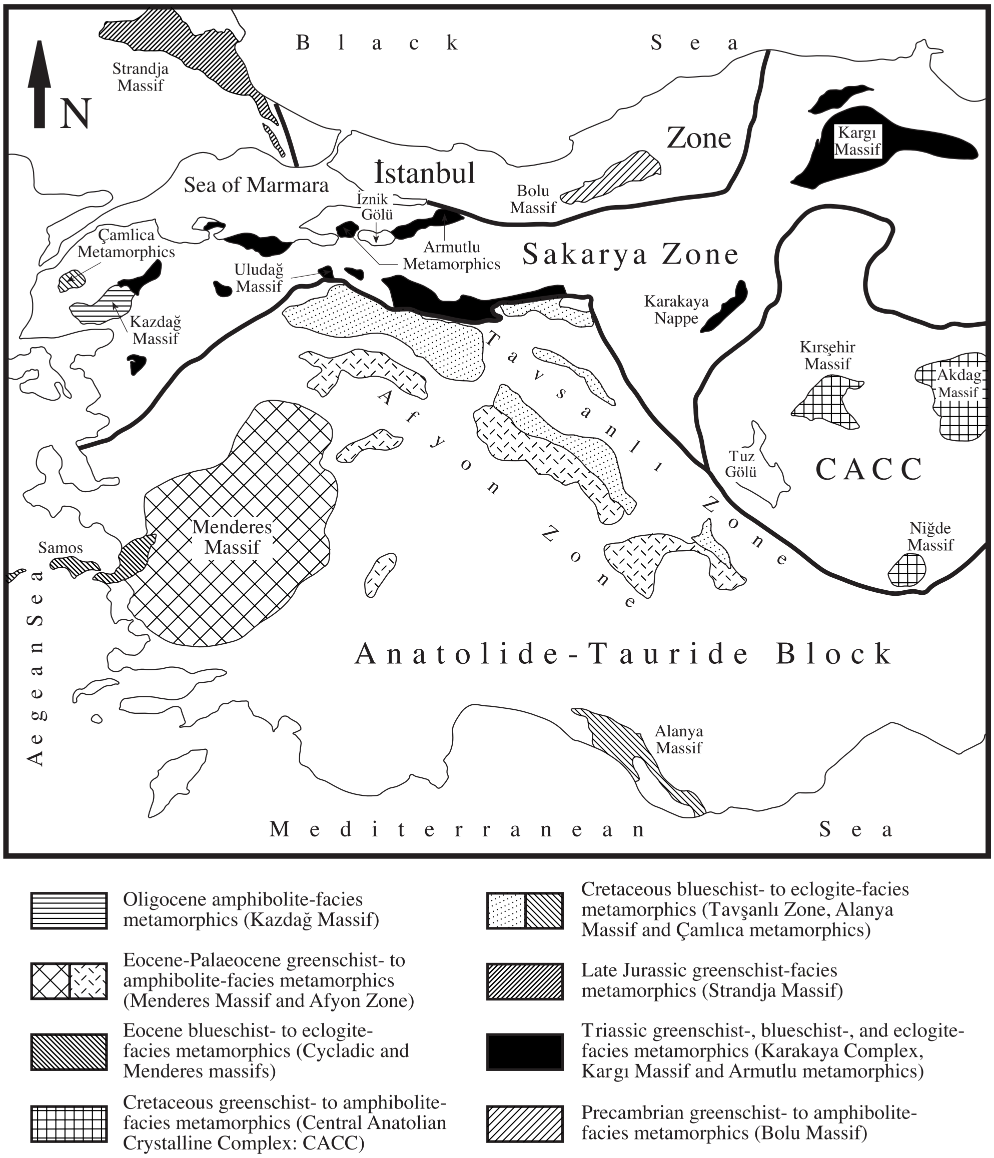



**Fig. 2** Simplified map showing the distribution of metamorphic rocks in western Turkey. *Heavy lines* show the main Tethyan sutures bounding the major tectonic units. See Fig. 1 for the names of the sutures

itoids and paragneisses, respectively. Similarly, biotite samples of granitoids and paragneisses yielded Rb/Sr ages of ca. 545±5 Ma. These ages suggest that Precambrian ages for the magmatism and metamorphism in the Bolu Massif (Satır et al. 2000).

There is also evidence for a pre-550 Ma eclogite- and granulite-facies metamorphism in the so-called 'core' rocks of the Menderes Massif. This topic will be discussed in the section on the Menderes Massif (see below).

## Triassic greenschist-, blueschist- and eclogite-facies rocks

Triassic metamorphic rocks cropping out in the Sakarya Zone cover large areas. Their exposures can be grouped into three belts: (1) the southern belt to the north of the İzmir–Ankara Suture comprises the Karakaya Complex and some inliers of high-grade Palaeozoic metamorphics (the Uludağ Massif; Okay et al. 1996; Göncüoğlu et al. 2000); (2) the northern belt exposed on the Armutlu Peninsula (Pamukova and İznik metamorphics); and (3) the eastern belt, namely the Kargı Massif, in the central Pontides (Fig. 2).

### Karakaya Complex

The Karakaya Complex represents an orogeny caused by the latest Triassic northward obduction of subduction-accretion units of Palaeotethys (Tekeli 1981; Okay et al. 1996; Picket and Robertson 1996). It was initially named the Karakaya Formation (Bingöl et al. 1973). The Karakaya Complex consists of two units (Okay et al. 1996): (1) a thick (>5 km) metamorphic sequence made up predominantly of metabasites with intercalated minor marbles and phyllites metamorphosed at greenschist-facies conditions (Nilüfer Formation); and (2) the tectonically overlying deformed Triassic greywacke with intercalated minor shales, siltstones, conglomerates and thinly bedded black cherts, and exotic Late Permian limestone blocks (Orhanlar Greywacke). Eclogite relics occur as small tectonic slices in the Nilüfer unit (Okay et al. 1996). These relics yielded $^{40}Ar/^{39}Ar$ phengite ages of 203–208 Ma, suggesting HP/LT metamorphism during latest Triassic-earliest Jurassic time (Okay and Monié 1997). The eclogites contain a greenschist-facies overprint. The undeformed Lower Jurassic shallow marine clastic rocks (conglomerates and sandstones) lie unconformably above rocks of the Karakaya Complex (Altıner et al. 1991). Exposures of the Karakaya Complex in the Ankara region have been called the Karakaya Nappe (Koçyiğit 1987, 1991).

There are also tectonic slices of Carboniferous metamorphics that form basement to the Karakaya Complex (Okay et al. 1990).

### Uludağ Massif

A tectonic window, wherein high-grade metamorphic rocks are exposed, has been termed the Uludağ Massif. The massif is made up of high-grade gneiss-amphibolite alternations covered by thick marbles and cherty marbles of unknown age (Ketin 1983). The metamorphics are intruded by Oligocene granitoids (Öztunalı 1973; Bingöl et al. 1982). Recent $^{40}Ar/^{39}Ar$ ages of 26.8±0.8 to 24.7±0.7 Ma confirm the Oligocene ages (Delaloye and Bingöl 2000).

### Armutlu metamorphics

The Armutlu metamorphics outcrop in the Armutlu Peninsula (Fig. 2) and are differentiated into several tectonic units (Göncüoğlu and Erendil 1990; Yılmaz 1991–1993). The metamorphics are differentiated into four tectonic units: (1) metamorphic basement (Pamukova metamorphics; Göncüoğlu and Erendil 1990) made up of pre-Triassic mica- and graphite-schists with subordinate amounts of metabasites and recrystallized limestone lenses metamorphosed at greenschist-facies conditions (correlated with the İstanbul Palaeozoic); (2) Triassic low-grade metamorphics (İznik metamorphics; Göncüoğlu and Erendil 1990) comprise basic volcanics and volcanoclastics, dolomites and recrystallized limestone blocks and bands within a matrix of shales and sandstones (Karakaya Formation of Bingöl et al. 1973); (3) an intermediate metamorphic olistostrome of Cretaceous age, a chaotic mixture of recrystallized limestone, ophiolites (gabbro, serpentinite, metadiabase and chert) and shale blocks; (4) the structurally highest nappe is composed of low-grade Palaeozoic metamorphics composed of quartzites, phyllites intercalated with recrystallized limestones representing the Palaeozoic of İstanbul and unconformably overlying crystalline basement with schists (garnet–biotite–staurolite schists), amphibolitic metavolcanic rocks and metagreywackes cut by gneissic metagranites, resembling the Precambrian basement of the former (Göncüoğlu et al. 1987). This unit is thought to represent the Palaeozoic basement of the western Pontides. The oldest rocks unconformably above the three are of Late Campanian–Maastrichtian age.

### Kargı Massif

The Kargı Massif is located within the Sakarya Zone in the central Pontides (Fig. 2). The massif is differentiated into three superimposed tectonic units: (1) highly sheared and folded metamorphic rocks, made up of gneisses at the base, followed by black amphibolites, quartzo-feldspathic gneisses, quartzite and thin marble alternations (Devrakani Metamorphics). Both Precambrian (Yılmaz 1980) and Early Palaeozoic ages have been suggested for the precursors of these metamorphic rocks. They are unconformably overlain by Upper Jurassic limestones characterized by a basal conglomerate; (2) an imbricated pile of metavolcanics and volcanoclastics, metamorphosed at greenschist-facies conditions and structurally overlying metaophiolites made up of a sheeted dyke complex and basaltic lavas (Çangaldağ Complex). This sequence is tectonically overlain by a sequence of imbricated basic lavas, sheeted dykes, mafic- and ultramafic-cumulates and serpentinized peridotites (Elekdağ metaophiolites; Şengör et al. 1984a; Ustaömer and Robertson 1993, 1994). The rocks of the Kargı Massif are interpreted as the remnants of Palaeotethys subducted northward during Late Palaeozoic–Mesozoic time (Ustaömer and Robertson 1993, 1994); (3) a wedge of siliciclastic sediments intercalated with tectonic slices of a dismembered ophiolite (Küre Complex). The Küre ophiolite, from bottom to top, is composed of serpentinized peridotite, layered cumulate gabbro, isotropic microgabbro cut by diabase dykes, a sheeted dyke complex, alternations of pillow lava, massif lava and lava breccias, and overlying sediments (shales alternating with sandstones; Ustaömer and Robertson 1993, 1994). The ophiolites are interpreted as having originated from the oceanic basement of a marginal basin (Küre basin), opened above a subduction zone during the latest Palaeozoic–Early Mesozoic (Ustaömer and Robertson 1993, 1994). The Elekdağ and Küre ophiolites contain evidence for blueschist- and eclogite-facies metamorphism, which was probably associated with their emplacement during the Late Triassic. The HP/LT metamorphics are unconformably overlain by Malm–Upper Cretaceous sedimentary sequences, followed by the emplacement of unmetamorphosed ophiolites during the Turonian–late Campanian, which induced the second phase of metamorphism in the Kargı Massif (Yılmaz 1991–1993).

### Late Jurassic greenschist-facies rocks (Strandja Massif)

The Strandja Massif forms a northwest-trending metamorphic belt straddling the Turkey–Bulgaria border and covering large areas along the southwestern margin of the Black Sea (Fig. 2). The Strandja Massif is a composite orogenic belt deformed and regionally metamorphosed during the Late Variscan and Late Jurassic–Early Cretaceous orogenies (Okay et al. 2000). In Turkey, the Massif consists of a Late Variscan crystalline basement of granites and gneisses unconformably overlain by a Lower Mesozoic metasedimentary sequence. This relative autochthon is tectonically overlain by allochthonous Triassic units, mainly preserved in the Bulgarian sector of the Massif. Cenomanian sandy limestones unconformably overlie the metamorphic rocks. The field and geochronological evidence confirm that the Strandja Massif is a Late Variscan unit, which underwent regional metamorphism and compressive deformation during the Late Jurassic–Early Cretaceous, and that the mid-Mesozoic deformation was thick-skinned and involved thrusting of basement gneisses over the Lower Mesozoic series. The Early Permian orogeny in the Strandja Massif involved amphibolite-facies regional metamorphism, crustal anatexis, and generation and emplacement of crustal-melt granites. New single-zircon evaporation ages from both the deep level granites and the surrounding gneisses and migmatites indicate that the high-grade metamorphism and plutonism were Early Permian (~271 Ma; Okay et al. 2000). Zircon geochronological data indicate that the metamorphism of the gneissic basement of the Strandja Massif, generally thought to be of Precambrian age, was most probably Early Permian and coeval with the granite intrusions. An Early Triassic–Middle Jurassic continental to shallow marine sequence of fluviatile coarse-grained sediments with abundant detritus derived from the basement granites and gneisses unconformably overlies the Late-Variscan basement. The Late Jurassic–Early Cretaceous (Oxfordian–Barremian) corresponds to a time of thick-skinned tectonics and involved northward thrusting of the basement granites over the mid-Mesozoic cover series, emplacement of the Triassic allochthons including basic volcanic rocks and deep marine turbiditic sequences over the epicontinental Jurassic cover series of the Strandja Massif, and consequent penetrative deformation and regional greenschist-facies metamorphism of the massif area. A Rb–Sr biotite-whole rock age from a metagranite dates the regional metamorphism as Late Jurassic (155 Ma; Okay et al. 2000). The mid-Mesozoic orogeny in the Strandja Massif ceased by the Cenomanian as suggested by coeval shallow marine sandstones that unconformably overlie the metamorphic rocks. During the Senonian, the northern half of the Strandja Massif formed a basement to an intra-arc basin and to a magmatic arc generated above the northward-subducting Tethyan oceanic lithosphere, as suggested by the occurrence of a thick sequence of calc-alkaline volcanic and volcanoclastic rocks over, and several granodiorite intrusions into, the metamorphic rocks of the Strandja Massif. The magmatic arc ceased by the Maastrichtian, and the intra-arc basin closed during the Early Tertiary by renewed north-vergent thrusting of the Massif during which the Eocene–Oligocene Balkanide mountain chain formed (see Okay et al. 2000 and references therein for details).

### Cretaceous blueschist- to eclogite-facies metamorphics

These rocks form a distinct belt in western Turkey and are distinguished into three groups, namely the Tavşanlı Zone, Alanya Massif and Çamlıca metamorphics, each of which possess remarkable differences in tectono-metamorphic evolution.

### The Tavşanlı zone

In the extreme north, just to the south of the İzmir-Ankara suture, is a an exposed belt of deformed rocks affected by high pressure/low temperature (HP/LT) metamorphism (Okay 1984a, 1984b). To the east it follows the southern boundary of the Central Anatolian Crystalline Complex (Figs. 1 and 2). These rocks commonly dip under non-metamorphic ophiolitic nappes (primarily of peridotite and, rarely, gabbros) of the Neotethyan suture in the north (Okay 1984a) whereas, in the south, HP/LT rocks are partly thrust onto the metasediments of the Afyon Zone (Okay 1986). Okay (1986) has differentiated two units within the Tavşanlı Zone: (1) A non-metamorphic, imbricated and strongly tectonized volcano-sedimentary complex, consisting of closely intercalated spilites, agglomerates, radiolarian cherts, red and green pelagic shales, pelagic limestones and greywackes (Ovacık unit; Kaya 1972; Okay 1984a, 1986). It contains abundant serpentine and talc lenses with lawsonite and aragonite growth in the amygdales and veins of spilites, suggesting an incipient blueschist metamorphism (Okay 1986). (2) A metamorphosed, pervasively recrystallized and ductilely deformed metavolcano-sedimentary rocks including (in ascending order) graphite schists, phyllites, marbles and a thick sequence of intercalated metabasites, metacherts and metapelites (Orhaneli unit; Okay 1980a, 1986). The minerals lawsonite, glaucophane, jadeite, aragonite and garnet indicate blueschist metamorphism (Çoğullu 1967; Okay 1980a, 1980b, 1986; Servais 1982) at 20 kbar and 430 °C (Okay et al. 1998). The timing of deformation and HP/LT metamorphism within the Tavşanlı Zone is constrained between Cenomanian (the youngest known metamorphosed sediments) and Middle Palaeocene (the oldest known unconformable sediment; Şengör et al. 1984a). It has been argued that exhumation of the blueschists was achieved in a compressive regime (see Okay et al. 1998 for detailed discussion). The age of HP/LT metamorphism, based on Rb/Sr and Ar–Ar data, is Campanian (83–80 Ma; Sherlock et al. 1999).

Göncüoğlu et al. (1996–1997) reported that the Tavşanlı Zone is not a continuous HP belt but contains thin slivers of very low-grade (chlorite zone) metamorphic rocks (Kınık metamorphics).

## Alanya Massif

The Alanya Massif (Blumenthal 1951) is the structurally highest unit in the Tauride Mountains, made up of a stack of nappes emplaced over the Mesozoic carbonate platform during Late Cretaceous and Tertiary times (Brunn et al. 1971; Şengör and Yılmaz 1981). The Alanya Massif is differentiated into three structurally conformable superimposed nappes (Okay and Özgül 1984): (1) The structurally lowest nappe consists of metapelites, metapsammites, metadolomites, recrystallized limestones and quartzites with minor amounts of metadolerites and recrystallized radiolarian cherts (Mahmutlar nappe). Fossil assemblages preserved within the recrystallized limestones suggest that part of the sequence is of Permian age (Özgül 1985). The lower parts of the sequence are Late Cambrian–Early Ordovician in age (Göncüoğlu and Kozur 1999). (2) The intermediate unit (Sugözü nappe) is mainly composed of mica–garnet schists with thin intercalations and/or boundinaged lenses of eclogite and blueschist metabasites with a mineral assemblage of garnet, omphacite, glaucophane, paragonite and barroisite. The eclogites grade into blueschists (Okay 1989b). The presence of sodic amphibole inclusions in garnet porphyroblasts in garnet–mica schists suggests that the HP/LT metamorphism was overprinted by Barroviantype greenschist-facies metamorphism. (3) The structurally highest nappe (Yumrudağ nappe) consists of schists (pelites, psammites calc-schists, metadolomites and recrystallized limestone bands) and conformably overlying recrystallized limestones of Late Permian age (Blumenthal 1951). The structural and metamorphic evidence suggest that the initial HP/LT metamorphism of the Sugözü nappe was followed by nappe stacking and consequent regional greenschist-facies metamorphism (Okay and Özgül 1984; Okay 1989b).

## Çamlıca metamorphics

The Çamlıca metamorphics comprise well-foliated quartz–mica schists with subordinate amounts of calcschists, black marbles, quartzites, amphibolites and albite–chlorite schists. Sporadic occurrences of thin garnet-bearing metabasic rocks are the characteristic features of these metamorphics (Okay and Satır 2000a). The amphibolites have textures typical of retrograded eclogites where the occurrences of sodic amphiboles as inclusions in garnets suggest an early high-pressure metamorphism with assemblages of garnet–omphacite–glaucophane. It has been suggested that the Çamlıca metamorphics underwent an initial eclogite-facies metamorphism during the Maastrichtian (Rb/Sr phengite ages of 69–65 Ma; Okay and Satır 2000a) and subsequently were overprinted by a greenschist-facies metamorphism. Evidence for the subsequent greenschist-facies overprint includes the replacement of omphacite and glaucophane by calcic amphibole and albite (Okay and Satır 2000a). The authors also suggested that, based on similar lithostratigraphy and metamorphic history, the Çamlıca metamorphics form part of the Rhodope metamorphic complex (see Okay and Satır 2000a for details).

## Cretaceous greenschist- to amphibolite-facies rocks (Central Anatolian Crystalline Complex)

In central Anatolia, medium- to high-grade metamorphic rocks with ophiolitic, plutonic and extrusive rocks have been collectively termed the Central Anatolian Crystalline Complex (CACC; Göncüoğlu et al. 1991; Figs. 1 and 2). The local exposures have been named the Akdağmadeni Massif in the north-east (Vache 1963), the Niğde Massif in the south (Göncüoğlu 1977), and the Kırşehir Massif in-between (Seymen 1982; Fig. 2). The complex has been given different names, such as the Kırşehir Massif (Seymen 1982), the Kırşehir Crystalline Massif (Bailey and McCallien 1950; Egeran and Lahn 1951), the Central Anatolian Massif or the Kızılırmak Massif (Erkan and Ataman 1981).

The metamorphic rocks comprise a metamorphosed platform-type sequence, made up of orthogneisses and paragneisses, metasediments intercalated with metabasic rocks, marbles alternating with metamorphosed sandstones–shales–marls and marly limestones, and metaophiolites (metagabbro, serpentinite, amphibolite and various metamorphosed ultramafic and mafic rocks) (Göncüoğlu et al. 1991). U–Pb dating of zircons from the gneisses of the same area yields an age of 2,059±77 Ma for the protolith of the metamorphics (Göncüoğlu 1986). The main event that affected the whole Massif is a progressive medium pressure/medium- to high-temperature event (Göncüoğlu 1977; Seymen 1982). The first event is characterized by kyanite–biotite–garnet paragenesis and accompanying multiphase deformation (Göncüoğlu 1977; Göncüoğlu et al. 1993). The second phase is the extensive recrystallization and low-pressure/high-temperature metamorphism with andalusite–sillimanite–cordierite paragenesis around extensive plutons intruded into the system following partial melting and crustal extension (Göncüoğlu et al. 1991; Whitney and Dilek 1997). Based on available isotopic data (71.4±3.2 to 77.8±1.2 Ma; Erkan and Ataman 1981; Göncüoğlu 1986), a pre-Cenomanian age has been assigned for the main metamorphism (Göncüoğlu et al. 1991).

In addition to metamorphic rocks, the CACC also contains supra-subduction zone ophiolites (ultramafic rocks, isotropic gabbro, plagiogranites, diabase, pillow lava and epi-ophiolitic sediments) as remnants of an extensive thrust sheet (Yalınız et al. 1995, 1996). The post-collisional granitoids intrude the metamorphic rocks and the ophiolites during and after the southward obduction of the ophiolitic rocks from Neotethys onto the Tauride–Anatolide Block during the Late Cretaceous (Erler et al. 1991; Göncüoğlu et al. 1991; Akıman et al. 1993), but before the Late Maastrichtian (Yalınız et al. 1996). The isotopic data on the age of granitoids are scarce and range from 54 to 110±14 Ma (Ayan 1963; Ataman 1972; Göncüoğlu 1986; Güleç 1994; Yalınız et al. 1999).

The Kırşehir Massif is made up of metamorphic rocks ranging from greenschist- to upper-amphibolite-facies with local granulite-facies conditions. The Akdağ Massif contains abundant high-grade (staurolite–kyanite–sillimanite) metapelitic rocks. On the other hand, the Niğde Massif has a more complex tectono-metamorphic history involving shortening, internal imbrication and consequent crustal thickening, erosion and extensional exhumation followed by magmatism. It has been suggested that, following the high-temperature metamorphism during contractional deformation associated with the closure of Neotethys in the Late Mesozoic–Early Cenozoic times, the Niğde Massif underwent bivergent extensional deformation during the Miocene (U–Pb monazite ages of 23.7–20 Ma from the late syn- to post-tectonic Üçkapılı granite), and that the Niğde Massif is a typical core complex (Whitney and Dilek 1997).

## Palaeocene–Eocene greenschist- to amphibolite-facies rocks

These metamorphic rocks crop out in two distinct belts, namely the Afyon Zone and the Menderes Massif. Because the Menderes Massif forms the main part of this paper, the characteristics of the Massif will be given separately in detail, whereas those of the Afyon Zone are given briefly below.

### The Afyon Zone

The Afyon Zone is composed of a regionally metamorphosed, shelf-type Palaeozoic–Mesozoic sequence and covers large areas of central Anatolia (Figs. 1 and 2). It is tectonically overlain either by the HP/LT metamorphics of the Tavşanlı Zone or by peridotite nappes. Because of its low metamorphic grade and imbricated structure, the Afyon Zone has been interpreted as a structurally higher unit than the rest of the Menderes Massif to the south (Şengör et al. 1984b). The Afyon Zone comprises the following sequence: (1) Devonian metaclastics (sandstones, siltstones, and quartzites) with rare metabasites and recrystallized limestone horizons; (2) Carboniferous to Late Permian dark grey recrystallized limestones; (3) a Mesozoic (Triassic–Maastrichtian) sequence made up of metaclastics (basal conglomerates, sandstones, siltstones) intercalated with algal limestones and conformable overlying thick platform carbonates (Özcan et al. 1988). The carbonates are locally overlain by Late Maastrichtian to Palaeocene 'wildflysch' with Permian, Jurassic and Cretaceous limestone blocks (Akdeniz and Konak 1979; Okay 1984a). The upper carbonate units and the wildflysch are little affected by metamorphism whereas the lower levels of the Afyon Zone (Palaeozoic metaclastics) have the characteristic regional greenschist-facies mineral assemblage of quartz–albite–phengite–chlorite–biotite, suggesting a progressive increase in grade towards the deeper levels of the sequence (Okay 1984a).

Although the Afyon Zone and the Tavşanlı Zone are described here separately, there are also claims that both zones forms a single belt, namely the Kütahya–Bolkardağ Belt (Özcan et al. 1988; Göncüoğlu et al. (1996–1997).

## Oligocene amphibolite-facies metamorphic rocks (Kazdağ Massif)

The Kazdağ Massif is a structural and topographic dome in which high-grade metamorphic rocks are exposed. The Massif is overlain tectonically by low-grade metamorphic rocks of the Karakaya Complex (Fig. 2). Analogous to the Menderes Massif, the Kazdağ Massif comprises core and cover lithologies. The core is made up of gneisses, marbles, amphibolites and metagabbros and is enveloped by a marble-dominated sequence and the overlying felsic gneisses, intercalated with calc-silicate gneisses, amphibolites and marbles with minor migmatites and metaserpentinites (Bingöl 1969; Okay and Satır 2000b). The characteristic mineral assemblage is quartz–plagioclase–biotite–muscovite with the sporadic occurrence of garnet and sillimanite. Rb/Sr mica ages from the gneisses range from 18 to 24 Ma. The Massif experienced an amphibolite-facies regional metamorphism at 640±50 °C and 51 kbar during the extensional exhumation (first stage) from a depth of 14 to 7 km along a north-dipping ductile shear zone in the latest Oligocene (24 Ma) (Okay and Satır 2000b). There is no noticeable change in metamorphic grade on the scale of the Kazdağ Massif. The second-stage exhumation to the surface occurred along brittle faults of the North Anatolian Fault Zone during the Pliocene–Quaternary (after 5 Ma; Okay and Satır 2000b).

## Menderes Massif

The Menderes Massif (Paréjas 1940) forms a large part of the Alpide orogen in western Turkey. It is a large, elongate (300×200 km), crustal-scale metamorphic culmination with its long axis trending NE–SW. It forms the westernmost part of the Anatolide tectonic unit of Ketin (1966) within the Turkish section of the Alpine–Himalayan Belt (Figs. 1 and 2). The Menderes Massif was originally named the 'Lydisch–Karische Masse' (Philippson 1910–1915), 'Saruhan–Menteşe Massif' (Akyol 1924 in Pamir and Erentöz 1974) and the 'West Anatolia Massif' (Ketin 1966). The commonly used term 'Menderes Massif' was introduced into the literature by Pamir (1928 in Pamir and Erentöz 1974). The Menderes Massif has been the subject of many works and it is still the apple of many active researchers' eyes. Some of the very early workers were Hamilton and Strickland (1841) and Philippson (1910–1915).

The Massif is tectonically overlain by nappes of İzmir-Ankara Neotethyan Suture Zone (Şengör and Yılmaz 1981), including the Bornova Flysch, in the north-west, by the Afyon Zone in the northeast, and by the Lycian Nappes (Graciansky 1972; Collins and Robertson 1997, 1998, 1999) in the south (Fig. 2). The Massif is dissected into three submassifs, namely the northern (Gördes and Eğrigöz submassifs), central (Ödemiş–Kiraz submassif) and the southern (Çine submassif), along the active, E–W trending, Gediz and Büyük Menderes grabens, respectively (Fig. 3).

The Menderes Massif has traditionally been interpreted as the eastern lateral continuation of the Cycladic Massif in the Aegean (Dürr et al. 1978; Oberhänsli et al. 1998). The long-standing assumption is based on the lithological aspects of both massifs where an older crystalline core is structurally overlain by a Palaeozoic–Mesozoic cover series (see below) with metamorphic grade increasing toward the 'core'. On the other hand, it recently has been claimed that these two massifs cannot be correlated (Ring et al. 1999; Gessner et al. 2000; Okay 2000).

The Menderes Massif was generated by compressional deformation. It lies within western Anatolia, one of the most seismically active and rapidly deforming regions in the world (Jackson and McKenzie 1984; Eyidoğan and Jackson 1985; Ambraseys 1988; Le Pichon et al. 1995; Reilinger et al. 1997; Ambraseys and Jackson 1998; Altunel 1999 and references therein), and is currently experiencing a large-scale, approximately N–S basin-and-range type continental extension (Fig. 1).

The Menderes Massif also forms a perfect laboratory to study polydeformation, where contractional and extensional fabrics are superimposed, and provides valuable information about the processes involved in exhuming deeply buried metamorphic rocks. An analysis of its history would therefore yield clues leading to a better understanding of metamorphic core-complex formation.

### Rock units

The Menderes Massif consists typically of a thick, tripartite lithological succession, interpreted by Dürr (1975) in terms of a simple 'onion-shaped' structure, with a 'gneiss core' at the base, a Palaeozoic 'schist envelope' covering the gneiss core, and a Mesozoic–Cenozoic 'marble envelope' overlying both, with metamorphism increasing towards the core (Dürr 1975; Şengör et al. 1984b). The latter two units are known as 'cover rocks'. This sequence is well preserved and is exposed in the southern submassif. The lithological aspects and contact relations of 'core' and 'cover' rocks will be summarized below.

### Core

The core rocks are composed of augen gneisses (typical orthogneisses), metagranites, high-grade schists, paragneisses and metagabbros with eclogite relics (Şengör et al. 1984b; Satır and Friedrichsen 1986; Candan 1995, 1996; Candan et al. 1998, 2000; Oberhänsli et al. 1997, 1998).

Augen gneisses are the most dominant and widespread lithology of the so-called 'core' rocks. The augen gneisses are typical blastomylonites characterized by a well-developed mylonitic foliation and a N–S-trending mineral-stretching lineation. The rocks are made up of asymmetric large feldspar porphyroclasts (reaching up to 8 cm along their long axes) within a matrix of primarily quartz, mica (both biotite and muscovite) and feldspars. Locally garnet porphyroclasts are also present. The matrix foliation wraps around the porphyroclasts.

The origin of the augen gneisses has been the subject of controversy for years. Although most previous workers have suggested a sedimentary protolith for the augen gneisses (Schuiling 1962; Başarır 1970, 1975; Öztürk and Koçyiğit 1983; Şengör et al. 1984b; Satır and Friedrichsen 1986), a granitic protolith has also been suggested (Graciansky 1965; Konak 1985; Konak et al. 1987). Based on a Rb/Sr whole rock isochron, Satır and Friedrichsen (1986) suggested that the sedimentation of the protoliths of the augen gneisses began ~750 Ma ago. The protoliths of the augen gneisses were calc-alkaline, peraluminous, S-type, late-to post-tectonic, tourmaline-bearing two-mica leucogranites (Bozkurt et al. 1995).

The age of protoliths is also controversial. Most previous workers have suggested that the augen gneisses are Precambrian in age (Schuiling 1962; Gra-

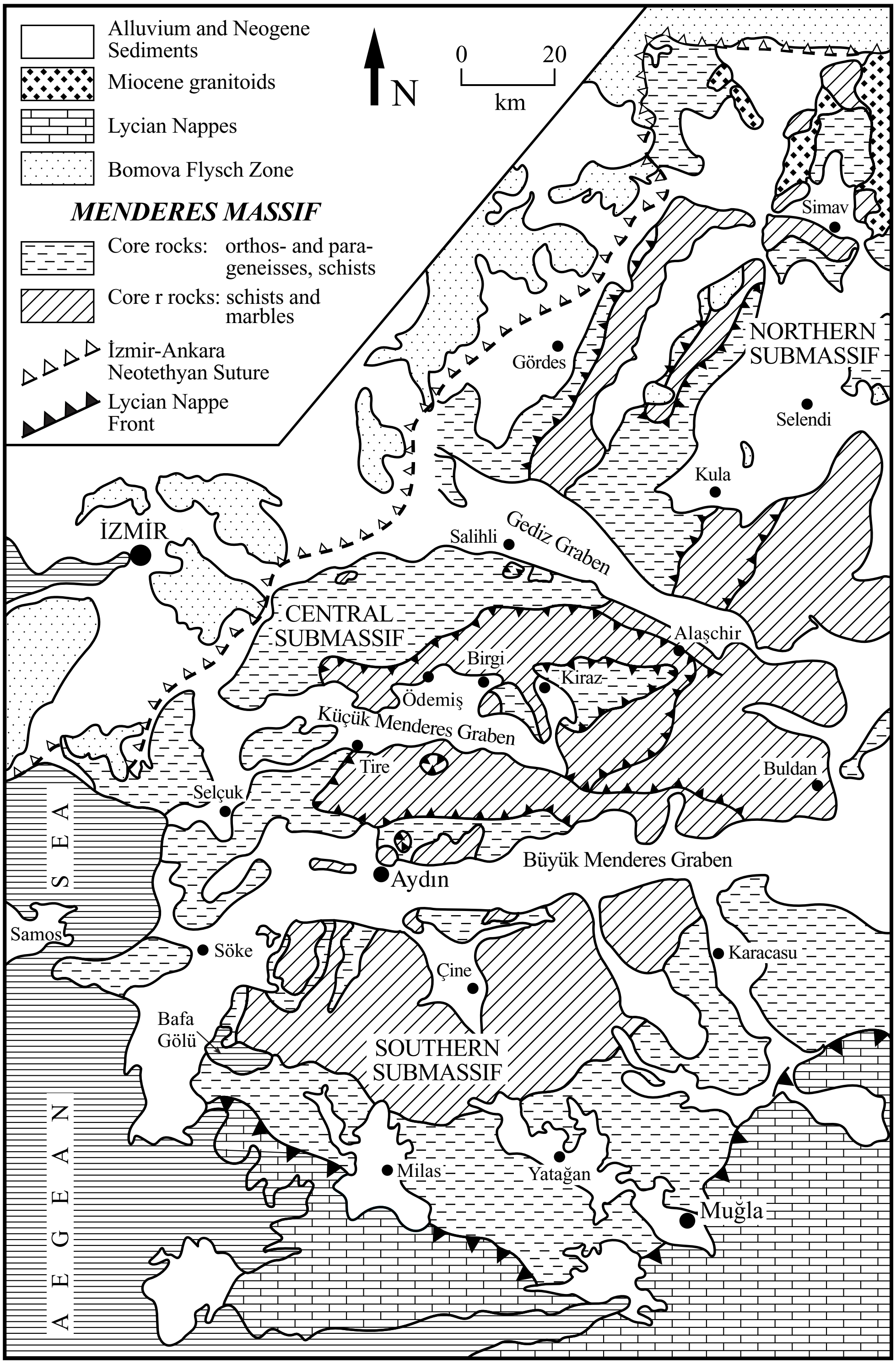



**Fig. 3** Simplified geological map of the Menderes Massif showing the subdivision of the Massif into submassifs and the distribution of 'core' and 'cover' rocks in each submassif (from Candan et al. 2000). *Lines with filled triangles* indicate thrust belts and point in the direction of vergence

ciansky 1965; Başarır 1970, 1975; Öztürk and Koçyiğit 1983; Şengör et al. 1984b; Konak 1985; Satır and Friedrichsen 1986; Konak et al. 1987; Reischmann et al. 1991). Şengör et al. (1984b) suggested that augen gneisses represent Pan-African basement in western Turkey. More recent Pb–Pb single zircon ages led to the conclusion that the granitic precursors were intruded during Late Precambrian–Early Cambrian times (521–572 Ma, averaging 550 Ma: Hetzel and Reischmann 1996; Dannat 1997; Hetzel et al. 1998; Loos and Reischmann 1999). On the other hand, it has been suggested that, based on (1) local cross-cutting relations of augen gneisses with the structurally overlying schists where the former are intrusive into the latter and (2) the age of oldest unconformable sediments (21±0.4 Ma; Becker-Platen et al. 1977), the protoliths of the augen gneisses are Tertiary (Oligocene: Bozkurt et al. 1992, 1993, 1995; Erdoğan 1992, 1993; Bozkurt and Park 1994). Bozkurt et al. (1995) argued that, because of the wide scatter of ages (2,555–1,740; Reischmann et al. 1991), zircons might have been inherited from a variety of sources. The other evidence for a Tertiary age comes from the field relations along the eastern margin of Lake Bafa where augen gneisses are intrusive into the schists and affected carbonates of the 'marble envelope' (Late Triassic to Early Eocene) with contact metamorphic assemblages thus suggesting that any intrusive body must post-date Early Eocene. It has also been argued that if the Precambrian ages are correct, then the intrusive relations indicate remobilization of the Pan-African basement during the Eocene main Menderes metamorphism (Bozkurt and Park 1997b). It is very clear that single zircon ages and the field data are contradictory. This situation arises from the fact that, in the southern submassif, any rock unit structurally beneath the so-called 'cover schists' is regarded as 'core' without paying attention to their field relations, age, structural features, etc. In fact, the core rocks in the southern submassif can broadly be divided into two as orthogneisses (true augen gneisses) and metagranites (Bozkurt 1998; Bozkurt and Park 2001; Gessner et al. 2000) where the later is intrusive into the augen gneisses and schists with a narrow contact metamorphic zone (Bozkurt 1996). The two types of rocks display contrasting kinematics and formed under different metamorphic conditions (see section on 'Deformation').

The field relations, petrological and radiometric studies suggest that the protoliths of the augen gneisses are several different intrusions that record several phases of magmatic activity. A detailed account will be given in the section on 'Magmatism'.

Metagabbros contain relics of eclogites overprinted by regional amphibolite-facies Barrovian-type metamorphism. The eclogites were originally gabbroic and noritic stocks and veins that intruded the basement rocks. They were previously considered to be Miocene (Dora et al. 1988). However, the occurrence of metagabbro xenoliths within the augen gneisses suggests that intrusion of the gabbroic protolith and subsequent eclogite-facies metamorphism occurred prior to 550 Ma (Candan et al. 2000).

## Cover

Cover rocks comprise both the 'schist' and 'marble' envelopes. The 'schist envelope' consists of quartzofeldspathic gneisses, pelitic and psammitic gneisses, brown or buff mica schists, quartzites, garnet amphibolites, 'augen' schists, phyllites, garnet-, staurolite-, kyanite-, chloritoid-mica schists, and limestone intercalations with these schists (Dürr 1975; Başarır 1970, 1975; Evirgen and Ataman 1982; Akkök 1983; Öztürk and Koçyiğit 1983; Şengör et al. 1984b; Ashworth and Evirgen 1985a, 1985b; Satır and Friedrichsen 1986; Konak et al. 1987; Kun et al. 1988; Bozkurt 1996; Hetzel et al. 1998). These units show rapid along- and across-strike passages into one another and are locally isoclinally folded.

At structurally lower parts of the cover schists, the psammites (muscovite–quartz schists) are intercalated with garnet–mica schists. The lower-grade and/or preserved parts of the garnet–mica schists exposed in the Babadağ (Denizli) area have been dated as Late Devonian–Early Carboniferous (Çağlayan et al. 1980; Konak et al. 1987) on the basis of poorly preserved Bryzoan assemblages. Similarly, zircons from the psammites around Lake Bafa yielded minimum zircon ages of ~526 Ma (Loos and Reischmann 1999). Thus, an Ordovician–Devonian age is suggested for this section of the cover schists. The stratigraphically and structurally upper parts of the 'schist envelope' are represented by a sequence of low-grade schists made up of graphite-bearing mica schists, phyllites and grey-black dolomitic marbles. The fossiliferous marbles (named the 'Göktepe marbles' with minor amounts of quartzites and calc-schists; Önay 1949) contain well-preserved Fusulinids, and some other assemblages yielded a Carboniferous (Visean)–Permian age (Önay 1949). Recently, Bryozoa (Fenestellidae) were found in the grey marbles intercalated with low-grade graphitic phyllites (north of Aydın). They yielded Permian ages (Özer 1998).

Dürr (1975) incorporated the Göktepe marbles into the 'schist envelope'. It has also been claimed that the rocks of the 'marble envelope' start with basal metaconglomerate (with quartzite and marble pebbles) resting unconformably on the rocks of the 'schist envelope' (Çağlayan et al. 1980; Öztürk and Koçyiğit 1983; Konak et al. 1987; Dora et al. 1995). The metaconglomerates between the Palaeozoic and Mesozoic rocks of the Menderes Massif are attributed to Cimmerian orogenesis during the closure of the Karakaya Basin (Koralay et al. 2001). In contrast, Şengör et al. (1984b) argued that these rocks represent a transition from the 'schist envelope' to the 'marble envelope' and that the Göktepe marbles are the earliest representatives of the 'marble envelope'. This view is shared by Özer et al. (2000) who have done detailed palaeontological work.

Above the basal conglomerates, the 'marble envelope' consists of the following rock types: (1) Upper Triassic–Liassic marbles intercalated with schists and metavolcanics; (2) a Jurassic–Lower Cretaceous thick, massive marble unit with metabauxite lenses (Konak et al. 1987); (3) Upper Cretaceous rudist-bearing marbles intercalated locally with thin mica schists (Dürr 1975; Konak et al. 1987; Özer 1998). The thickness of the platform-type marbles reaches up to 2 km (Dürr et al. 1978; Güngör 1998). Metabauxites in the high-grade parts contain corundum whereas those in the lower-grade sections contain diaspore (Yalçın 1987).

Although Yalçın (1987) suggested that the age of the bauxite-bearing marbles is Jurassic or Early Cretaceous; Özer (1998) indicated, based on fossil evidence, that these marbles lie at the boundary between the Lower and Upper Cretaceous; (4) Upper Cretaceous–Palaeocene brecciated pinkish-reddish pelagic marbles (Konak et al. 1987). More recently, Özer (1998) documented fossil assemblages of Campanian–Maastrichtian age; and (5) an Upper Palaeocene–Lower Eocene metaolistostrome with metaserpentines, metagabbros and metabauxite-bearing marble blocks within a schist matrix (Dürr 1975; Gutnic et al. 1979; Çağlayan et al. 1980; Konak et al. 1987). The metagabbro and metaserpentine blocks are characterized by eclogitic rims suggesting HP metamorphism (Dürr 1975; Dora et al. 1995; Candan et al. 1997; Candan and Dora 1998). The metaolistostrome is thought to have formed in front of advancing Lycian ophiolitic nappes (Erdoğan and Güngör 1992). It has also been compared with the Cycladic olistostromes (Candan et al. 1997). These observations indicate that the 'marble envelope' represents a time interval from Late Triassic to Early Eocene.

The rocks of the 'marble envelope' in the central and northern submassifs form small outcrop areas in comparison with the southern submassif and usually occur at topographically higher levels. The marbles of the central submassifs are correlated with the Göktepe marbles of Önay (1949) from the southern submassif (İzdar 1971; Akkök 1981, 1983).

It is now commonly accepted that the pre-Ordovician to Early Eocene record in the Menderes Massif, particularly in the southern submassif, shows a generally conformable sequence with a number of erosional disconformities.

More recently, eclogite relics strongly overprinted by regional greenschist-facies Barrovian metamorphism have been documented from the cover rocks of the central submassif (Candan 1994, 1995, 1996; Candan et al. 1997; Oberhänsli et al. 1998; Gessner et al. 2000; Okay 2000).

A key horizon with garnet, kyanite and sillimanite assemblage (inappropriately called 'leptites') occurs between the schists and the gneisses (Dora et al. 1988, 1990; Kun et al. 1988; Candan and Kun 1991; Candan 1994). These high-grade metamorphic rocks have been variously described as feldspar-enriched quartz–schist and biotite quartz–schists. The leptites characteristically have cataclastic texture indicating localized shearing during the main Menderes metamorphism (MMM; see section on 'Metamorphism') and contributed to differential deformation of the core and cover rocks (Öztürk and Koçyiğit 1983; Şengör et al. 1984b). According to Şengör et al. (1984b), these rocks postdate the first metamorphism of the Massif and were metamorphosed and deformed at almandine-amphibolite-facies conditions, probably during the MMM. The leptites have been interpreted as originally island-arc type rhyolitic-dacitic lavas of dominantly calc-alkaline affinity (Kun and Dora 1984; Dora et al. 1988, 1990; Kun et al. 1988). The rocks were probably originally tuffs and ignimbrites. However, recent works (Dora et al. 1998, 2000) suggest that the protoliths of the leptites were predominantly clastic sediments of litharenitic composition. The metaclastic sequence (up to 7–8 km thick) consists predominantly of paragneisses, which pass upward into mica schists with almost no carbonates. Although in almost all previous works (Kun and Dora 1984; Şengör et al. 1984a; Dora et al. 1988, 1990; Kun et al. 1988) these rocks were included in the cover schists, the metasediments are intruded by granites, which have the character of granitic-augen gneisses. They also occur as partially assimilated xenoliths within the orthogneisses. Because the magmatic rocks are 550 Ma or older, these clastic metasediments are Late Proterozoic in age (Dora et al. 2000).

## Core-cover boundary relationships

The relationship between the core augen gneisses and cover schists has been largely obscured by the polymetamorphic and structural complex history of the Menderes Massif and is almost everywhere structural. It has been the subject of debate among geoscientists for many years.

Several workers have interpreted the so-called 'core-cover' contact (augen gneiss–schist contact) as a major unconformity (Schuiling 1962; Graciansky 1965; Başarır 1970, 1975; Öztürk and Koçyiğit 1983; Şengör et al. 1984b; Konak 1985; Satır and Friedrichsen 1986; Konak et al. 1987), later termed the 'main supra-Pan African unconformity' by Şengör et al. (1984b). A metaconglomerate (Gökçay metaconglomerate), consisting of deformed-elongated granite, tourmaline-rich and quartzite pebbles at structurally lower parts of the 'schist envelope' close to the contact with the augen gneisses near the town of Kavaklıdere (Muğla) has been interpreted as a basal conglomerate, thus supporting the view that the 'core-cover' contact is a major unconformity (Konak et al. 1987; Dora et al. 1998). It has also been argued that older zircon ages (546.2±1.2 Ma) from the augen gneisses and their intrusive relations to the schists suggests the presence of Precambrian schists in the southern submassif and that the inferred major unconformity may be located within the schists, not along the augen gneiss–schist contact (Hetzel and Reischmann 1996). The authors suggested that because of subsequent Alpine events. Such an unconformity may be difficult to recognize and has not yet been found.

Others, however, argue that there is no unconformity and that the 'core-cover' contact is essentially intrusive (Bozkurt et al. 1992, 1993, 1995; Erdoğan 1992, 1993). The authors indicate that the protoliths of the augen gneisses are Tertiary granitoid rocks and display locally preserved intrusive relationships with the structurally overlying 'cover' schists. This contact later being used as a major shear zone (Bozkurt and Park 1994, 1997a, 1997b).

The third and more realistic possibility is that the 'augen gneiss–schist' contact is tectonic. There are two different interpretations: (1) this contact is a south-facing high-angle Alpine extensional shear zone along which the so-called 'core' rocks in the footwall were progressively exhumed during top-to-the S–SSW shearing (Bozkurt and Park 1994, 1997a, 1997b; Hetzel and Reischmann 1996; Bozkurt and Satır 2000; Lips et al. 2000). The shear zone accompanied late orogenic collapse in western Turkey. (2) The 'core-cover boundary' is a south-facing thrust fault along which the 'cover' schists were emplaced onto the 'core' rocks during top-to-the south Alpine nappe stacking during the collision of the Anatolide–Tauride platform and the Sakarya continent across the Neotethys (Ring et al. 1999; Gessner et al. 2000). On the other hand, it is suggested that the contact may have operated as a thrust fault during a top-to-the N–NNE Alpine contractional deformational but was later reactivated during a subsequent top-to-the S–SSW Alpine extensional deformation (Lips et al. 2000).

Another possibility is that the schists above the augen gneisses are Precambrian in age, and the Pan-African granites intruded into these Precambrian sediments (Hetzel and Reischmann 1996).

## Metamorphism

Although the number and timing of different episodes are still being debated, it is now commonly agreed that the Menderes Massif has a complex, polyphase metamorphic history. Five phases of metamorphism have been distinguished (Table 1). The first two pre-Alpine phases ($M_{1-2}$) affected only the 'core' rocks. Because the traces of earlier metamorphic phases have been erased by subsequent Alpine events, evidence for pre-Alpine metamorphism is scarce. These phases are based mainly on relic assemblages.

The first event ($M_1$) is evidenced by the occurrence of metagabbros with eclogite relics (Candan 1995, 1996; Oberhänsli et al. 1997; Candan et al. 1998, 2000). The high-pressure relics (omphacite–garnet–clinozoisite–rutile and omphacite–garnet–rutile–kyanite assemblages) suggest a HP/LT metamorphism at a pressure of ~15 kbar (Candan et al. 2000). Orthopyroxene relics in the gneisses and charnockites (Oberhänsli et al. 1997; Candan and Dora 1998) indicate a possible granulite-facies metamorphism prior to eclogite formation (Candan et al. 2000). The presence of amphibolite-facies inclusions in garnets reveals a pre-eclogitic stage. The eclogites were later retrograded to garnet amphibolite during a Barrovian-type metamorphism. The absence of similar relics in augen gneisses and the presence of metagabbros as xenoliths in them suggest that the eclogite-facies metamorphism must have occurred prior to the emplacement of the precursors of the augen gneisses (pre-550 Ma: Candan et al. 2000). Candan et al. (2000) relates this phase to overthrusting and consequent crustal thickening during Pan-African orogeny.

The second event ($M_2$) occurred at 50,210 Ma (Rb–Sr whole rock ages; Satır and Friedrichsen 1986) with intense deformation at amphibolite-facies conditions associated with widespread migmatization and local anatexis (Akkök 1983; Şengör et al. 1984b; Satır and Friedrichsen 1986). Deformed tonalitic-granitic intrusions dated at 470±9 Ma (Rb/Sr whole rock ages; Satır and Friedrichsen 1986) mark the end of this episode. Şengör et al. (1984b) related this event to the last Pan-African collision and associated post-collisional convergence.

Evidence for the third phase ($M_3$) is scarce and is from Derbent area of the central submassif (Akkök 1983). Akkök believed that this phase, which post-dated Triassic (250–227 Ma: Dannat and Reischmann

**Table 1** Metamorphism in the Menderes Massif

| Metamorphic phase | Grade | Evidence | Age | Reference |
|---|---|---|---|---|
| $M_1$ | Granulite, eclogite, amphibolite | Relict eclogites within the core rocks of the southern submassif | Pre-550 Ma | Candan et al. 2000 |
| $M_2$ | Greenschist | From the central submassif, Derbent | Pre-230 Ma | Akkök 1983 |
| $M_3$ | Blueschist | Relict eclogites from the central submassif, Dilek Peninsula | 40 Ma | Oberhänsli et al. 1998 |
| $M_4$ (MMM) | Amphibolite–greenschist | Rb/Sr mica ages from the southern submassif | 35±5 Ma | Şengör et al. 1984a; Satır and Friedrichsen 1986 |
| | | $^{40}Ar/^{39}Ar$ fabric age from central submassif, northern margin of Büyük Menderes Graben | 36±2 Ma | Lips et al. 2000 |
| $M_5$ | Greenschist | $^{40}Ar/^{39}Ar$ biotite plateau and amphibole isochron ages, respectively | 12.2±0.4–19.5±1.4 Ma | Hetzel et al. 1995a |

1998; Koralay et al. 1998, 2000) granitic intrusions, occurred at greenschist-facies conditions. It was related to the closure of the Karakaya marginal basin of Palaeotethyan ocean during the Late Triassic (Akkök 1983).

Blueschists and eclogites with strong greenschist-facies overprints have been documented from the central submassif (Candan 1995; Candan et al. 1997, 1998; Oberhänsli et al. 1997, 1998). They suggest a HP metamorphism ($M_4$). The eclogites are represented by metabasite and metaserpentinite lenses within the quartzitic metapelites. The metabasites contain mineral assemblages of glaucophane–phengite–epidote–albite, typical of HP metamorphism and similar to those on Samos Island in the Aegean Sea (Okrusch et al. 1985; Chen et al. 1995; Candan et al. 1997; Oberhänsli et al. 1998). The phengites yielded $^{40}Ar/^{39}Ar$-cooling ages of 40 Ma; this is attributed to a subduction-related HP metamorphism ($M_4$) during the closure of Neotethys (Oberhänsli et al. 1998). Based on geochronological data and mineral assemblages, the high-pressure unit of the Dilek Peninsula is interpreted as the eastward lateral continuation of the HP unit of Samos Island (Candan et al. 1997; Oberhänsli et al. 1998). Oberhänsli et al. (1998) and Okay (2000) suggested that this unit has no relation to the Menderes Massif but belongs to the Cycladic Massif and should be omitted in the definition of the Menderes Massif. According to this view, the HP unit of the Dilek Peninsula was thrust onto the Menderes Massif during the Alpine orogeny (Oberhänsli et al. 1998; Gessner et al. 2000; Okay 2000). This view suggests that the Menderes and Cycladic massifs cannot be correlated.

A major event affecting the whole massif is the 'main Menderes metamorphism (MMM)′ ($M_5$), which was associated with intense deformation. It reached upper-amphibolite-facies conditions in the central and northern submassifs (Akkök 1981, 1983; Ashworth and Evirgen 1985a, 1985b; Kun et al. 1988), almandine–amphibolite grade associated with local anatectic melting in the structurally lower parts of the southern submassif (Başarır 1970, 1975; Dürr 1975; Akkök 1981, 1983; Öztürk and Koçyiğit 1983; Ashworth and Evirgen 1984; Şengör et al. 1984b; Konak 1985; Satır and Friedrichsen 1986; Konak et al. 1987), but only greenschist grade in the structurally upper parts of the 'cover' rocks. This metamorphism is considered to be the product of latest Palaeogene collision across the Neotethys and the consequent internal imbrication of the Menderes–Tauride block that resulted in the burial and intense shearing of the massif area within a broad zone along the base of the southward advancing Lycian Nappes, and must have been generated during and after their emplacement onto the Massif (Şengör and Yılmaz 1981; Şengör et al. 1984b). The subsequent cooling occurred in the Early Tertiary as suggested by mean Rb/Sr ages of 35±5 Ma (63–48 Ma muscovite and 50–27 Ma biotite ages; Satır and Friedrichsen 1986). More recently, Bozkurt and Satır (2000) provided new Rb/Sr geochronological data (62–43 Ma) supporting the previous considerations. It is commonly believed that the MMM affected the whole Massif and indeed gave it its 'massif' character. The estimated conditions are 450–660 °C and 5–8 kbar in the central submassif (Akkök 1981, 1983; Ashworth and Evirgen 1985b; Okay 2001) and 330–530 °C and 5 kbar in the southern submassif (Ashworth and Evirgen 1984).

Mineral paragenesis in the central submassif range from chlorite- to sillimanite-bearing (Akkök 1981, 1983; Evirgen and Ataman 1982; Evirgen and Ashworth 1984; Ashworth and Evirgen 1985a, 1985b; Hetzel et al. 1998). The sporadic occurrence of sillimanite is attributed to local disturbances of a regional P–T path passing close to the kyanite/sillimanite phase boundary (Ashworth and Evirgen 1985b). The inverted metamorphism in the central submassif along the northern margin of Küçük Menderes graben was first documented by Evirgen and Ataman (1982) who mapped the metamorphic rocks according to their mineral paragenesis (see their Fig. 2), and was later interpreted by Hetzel et al. (1998) who suggested that it is related to Alpine contractional events (see the section on 'Deformation'). Okay (2001) documented inverted metamorphism south of the Küçük Menderes Graben and related it to the whole scale inversion of the Menderes massif sequence.

The inverted metamorphic sequence documented from the central submassif (Hetzel et al. 1998; Okay 2001). The sequence is interpreted as the result of numerous north-directed thrusts, or to the whole scale inversion of the sequence of the Menderes Massif.

The last phase ($M_6$) in the Massif was a retrogressive greenschist-facies metamorphism associated with exhumation of metamorphic rocks in the footwall of presently low-angle normal faults that accommodated latest Oligocene–Early Miocene orogenic collapse in western Turkey. This phase was associated with synkinematic Miocene granitoid intrusions (see Table 3 for detail information about age relations).

## Magmatism

The geochronological and field evidence suggest that there have been four distinct pulses of magmatic activity in the Menderes Massif: Proterozoic, Cambrian, Triassic and Tertiary (Table 2).

Evidence for Proterozoic magmatic activity comes from the entire Menderes Massif where orthogneisses and metagranites yield $^{207}Pb/^{206}Pb$ single zircon ages that range from 2,555–1,740 Ma (Reischmann et al. 1991). This event was followed by major magmatic activity during the Cambrian. The orthogneisses, commonly known as augen gneisses, have yielded $^{207}Pb/^{206}Pb$ single zircon ages of: (1) 546.2±1.2 Ma (Hetzel and Reischmann 1996) and 521±8 to 572±7 (mean 550 Ma; Loos and Reischmann 1999) from the southern submassif; (2) 528±4.3 to 541.4±2.5 Ma (Dannat

1997) from the entire Menderes Massif; and (3) 551±1.4 Ma (Hetzel et al. 1998) from the central submassif.

In central submassif (the Alaşehir and Derbent areas), NNE-directed leucocratic metagranitoids are intrusive into so-called 'core' schists (Akkök 1983; Candan 1994; Koralay et al. 1998). The metagranites display a distinct penetrative Alpine foliation parallel to the schistosity of the country rocks. The single crystal zircon ages of ~230–240 Ma were obtained from leucocratic metagranitoids (Reischmann et al. 1991; Dannat 1997; Dannat and Reischmann 1998; Koralay et al. 1998, 2000). The intrusion of these granitoids was attributed to the closure of the Karakaya basin, a remnant basin of Palaeotethys (Koralay et al. 2000).

In the central submassif, the metamorphic rocks in the footwall of a presently low-angle normal fault – which forms the contact between the metamorphics and structurally overlying Neogene sediments – are intruded by two Miocene granitoids, namely the Turgutlu and Salihli granitoids. The Turgutlu granitoid is characterized by a wide contact metamorphic aureole with andalusite (Hetzel et al. 1995a; Koçyiğit et al. 1999). The age of intrusion is Burdigalian (a $^{40}Ar/39Ar$ hornblende age of 19.5±1.4 Ma; Hetzel et al. 1995a). The Burdigalian age of the Turgutlu granitoid has been confirmed by the most recent K–Ar age of 16.3±0.3 Ma (Delaloye and Bingöl 2000). Similarly, the metamorphic rocks in the northern submassif were intruded by the Late Oligocene–Early Miocene Eğrigöz granite (20–29 Ma: K–Ar and $^{40}Ar$–$^{39}Ar$ mica and $^{207}Pb/^{206}Pb$ single zircon ages; Bingöl et al. 1973; Öztunalı 1973; Reischmann et al. 1991; Işık and Tekeli 2000). The intrusion of these granitoids is interpreted as syntectonic with respect to the exhumation of the massif (see section on 'Deformation').

## Deformation

More recent studies have shown that the structure of the Menderes Massif is much more complicated than previously thought, and dominated by extensive imbrication and shearing with the development of enormous cataclastic zones, asymmetric folds and steep-gentle compressional and extensional shear zones. The complex internal structure is the manifestation of a non-coaxial, multiphase deformation history and repeated metamorphism.

For many years, a simple dome structure, with a Precambrian core enveloped by a Palaeozoic–Early Tertiary cover, has been accepted for the Menderes Massif (Schuiling 1962). However, studies in recent years suggest that the Menderes Massif is a complex nappe pile, a product of Alpine compressional tectonics (Konak et al. 1987, 1994; Dora et al. 1995; Partzsch et al. 1998; Yusufoğlu 1998; Gessner et al. 2000). In the period following this compression, the Massif was exhumed along shear zones and low-angle normal faults that accompanied Early Miocene extensional tectonics, and thus the Massif took on much of its present-day form (Hetzel et al. 1995a, 1995b). A still active, Late Miocene–Pliocene E–W-trending graben system and normal faults continue to deform the Massif. It is now commonly agreed that the Massif has undergone polyphase deformation although the number and timing of phases are still controversial.

**Table 2** Data concerning with the magmatic activity in the Menderes Massif

| Magmatic Phase | Lithology | Location | Age (Ma) | Method | Reference |
|---|---|---|---|---|---|
| 1 | Metagranites and orthogneisses | Entire Menderes Massif | 2555–1740 | $^{207}Pb/^{206}Pb$ single zircon evaporation | Reischmann et al. 1991 |
| 2 | Weakly deformed granite | Southern Menderes Massif | 546.2±1.2<br>521±8 to 572±7 (mean ~550) | $^{207}Pb/^{206}Pb$ single zircon evaporation | Hetzel and Reischmann 1996<br>Loos and Reischmann 1999 |
| | Metagranites and orthogneisses | Entire Menderes Massif | 528±4.3 to 541.4±2.5 | | Dannat 1997 |
| | Weakly deformed granite | Central Submassif | 551±1.4 | | Hetzel et al. 1998 |
| 3 | Metagranites | Central Submassif | 240.3±2.2<br>226.5±6.8 | $^{207}Pb$–$^{206}Pb$ single zircon evaporation | Dannat 1997<br>Koralay et al. 1998 |
| 4 | Egrigöz Granite | Northern Submassif | ~20 | $^{207}Pb/^{206}Pb$ single zircon evaporation | Reischmann et al. 1991 |
| | | | 24.6±1.4– 21.2±1.8<br>20.4±0.6– 20.0±0.7 | K–Ar Orthoclase age<br>K–Ar biotite age | Delaloye and Bingöl 2000 |
| | | | 20–29 | K–Ar and $^{40}Ar/^{39}Ar$ mica ages | Öztunalı 1973; Bingöl et al. 1982; Işık and Tekeli 2000 |
| | Turgutlu and Salihli granitoids | Central Submassif | 19.5±1.4 | $^{40}Ar/^{39}Ar$–amphibole isochron | Hetzel et al. 1995a |
| | | | 12.2±0.4– 13.1±0.2 | $^{40}Ar/^{39}Ar$–biotite plateau age | |
| | | | 16.3±0.3 | K–Ar biotite age | Delaloye and Bingöl 2000 |

**Table 3** Deformation in the Menderes Massif

| Deformation Phase | | Kinematics | Age | Reference |
|---|---|---|---|---|
| Pre-Alpine ($D_1$) | | | ? | |
| Alpine | Contractional ($D_2$) | Top to the N–NNE | Eocene | Bozkurt 1995; Hetzel et al. 1998; Bozkurt and Park 1999 |
| | Extensional ($D_3$) | Top to the N–NNE in the northern submassif | Early Miocene | Verge 1995; Işık and Tekeli 2000 |
| | | Bivergent in the Central Submassif | | Hetzel et al. 1995a, 1995b, 1998; Emre 1996; Koçyiğit et al. 1999; Bozkurt 2000b; Sözbilir and Emre 1998; Gessner et al. 2000; Gökten et al. 2000; Lips et al. 2000 |
| | | Top to the S–SSW in the southern submassif | | Bozkurt and Park 1994, 1997a, 1997b; Hetzel and Reischmann 1996; Lips et al. 2000 |
| Neotectonic ($D_4$) | | Approximately N–S extension | Pliocene or younger (~5 Ma onwards) | Koçyiğit et al. 1999; Bozkurt 2000a; Sarıca 2000; Yılmaz et al. 2000 |

The deformation needs to be examined as two groups: pre-Alpine and Alpine (Table 3).

## Pre-Alpine ($D_1$)

Since the traces of earlier deformational phases (pre-Alpine fabrics) have been erased, and primary contact relationships of various lithologies have been obliterated by subsequent Alpine events (see below) – during which the massif has acquired a complex nappe structure and was exhumed to the surface – evidence for a pre-Alpine phase, here referred to as $D_1$ for simplicity, is scarce.

The first evidence for $D_1$ deformation is documented from the so-called Pan-African core rocks (Precambrian–Cambrian basement), particularly those exposed in the southern submassif, which contain widespread relic textures and paragenesis (typical of granulite- and eclogite-facies metamorphism) in metagabbro lenses (Oberhänsli et al. 1997; Candan and Dora 1998; Candan et al. 2000). Similar relics in metagabbro lenses within the Birgi augen gneiss are also documented from the central submassif (Hetzel et al. 1998). Following granulite- and eclogite-facies metamorphism, the metagabbros have been overprinted by amphibolite-facies metamorphism (Candan et al. 2000).

Evidence for the age of this deformation is scarce. Pb–Pb dating from the augen gneisses in the southern submassif yielded zircon ages of 550 Ma (Hetzel and Reischmann 1996; Dannat 1997; Hetzel et al. 1998; Loos and Reischmann 1999). Similarly, magmatic zircons from Birgi metagranite are dated at 551.5±1.4 Ma (U–Pb ages) where the metagranite is intrusive into the augen gneisses in the central submassif (Hetzel et al. 1998). The presence of metagabbros, with relics of granulite- and eclogite-facies textures and paragenesis, as xenoliths within the augen gneisses and the lack of such relics in the augen gneisses themselves, strongly suggest that the age of the $D_1$ deformation must predate the age of augen gneiss precursor, i.e. pre-550 Ma. The metamorphic monazite age of 660±62 Ma from granulite-facies orthopyroxene-bearing Birgi augen gneisses (central submassif) is interpreted as the age of granulite-facies metamorphism in the Massif (Warkus et al. 1998). It has been suggested that the granulite- and eclogite-facies metamorphism is the manifestation of the Pan-African orogeny (Oberhänsli et al. 1997; Candan and Dora 1998; Hetzel et al. 1998; Candan et al. 2000).

Other evidence for the presence of pre-Alpine deformation ($D_1$) comes from the central submassif (Derbent area) with poorly documented Triassic greenschist-facies metamorphism (Akkök 1983). Triassic post-tectonic granite(s) are interpreted to mark the end of this episode (see section on ‘Magmatism’ for details).

It is important to note that, although the Menderes Massif has experienced a complex pre-Alpine tectonometamorphic history with polyphase deformation and metamorphism, the number and timing of the different phases are still unknown. However, the available data suggest that the pre-Alpine development of the Menderes Massif includes a Pan-African Orogeny and a (?) Permo-Triassic Cimmerian event during the closure of the Karakaya basin.

## Alpine

Alpine deformations in the Menderes Massif are distinguished, based on contrasts in kinematics and metamorphic grade, into two distinct phases: older contractional and younger extensional (Table 3). They are referred to as $D_2$ and $D_3$, respectively, for simplicity although it is possible to distinguish many sub-stages within each phase.

### Contractional ($D_2$)

A $D_2$ contractional deformation has affected the entire Menderes Massif and in fact gave it its ‘massif’ character. The $D_2$ fabrics are characterized by a penetrative $S_2$ regional foliation and an invariably associated, approximately N–S-trending mineral lineation ($L_2$). It was synchronous with the main Menderes metamorphism (MMM; see section on 'Metamorphism'). The matrix foliation-porphyroblast microstructural relationships show syn-tectonic grain growth of albite, garnet, staurolite, kyanite and sillimanite at greenschist- to upper-amphibolite-facies conditions (Bozkurt 1995; Hetzel et al. 1998; Bozkurt and Park 1999). On the scale of the Menderes Massif, this regional foliation has a typical dome-shaped pattern. On the scale of the central submassif, this foliation defines two antiforms and a synform where the latter corresponds to the present-day location of the Küçük Menderes graben (Gessner et al. 2000). N- to NNE-vergent folds, deforming the $S_2$ regional foliation, at micro-, meso- and mapable-scales are the other common structural elements of this deformation. Based on the regional correlation of stratigraphy and structures, the Menderes Massif is interpreted as a major recumbent fold (Okay 2000).

It has been suggested that the 'MMM' and associated deformation are the results of the burial of the Menderes Massif beneath the southward-advancing Lycian Nappes during the Late Palaeogene collision of the Sakarya Continent in the north and Anatolide–Tauride block in the south across Neotethys (Şengör and Yılmaz 1981; Şengör et al. 1984b). Structural analysis of $D_2$ fabrics, however, demonstrates a top-to-the N–NNE shear sense (southern submassif: Bozkurt 1995; Bozkurt and Park 1999; central submassif: Hetzel et al. 1998), thus suggesting N-directed thrusting. The kinematics during $D_2$ deformation was opposite to that which has been proposed as the cause of metamorphism and coeval deformation. The top-N deformation is therefore attributed to a northward backthrusting and consequent internal imbrication of the Menderes Massif at depth whereas southward thrusting of the Lycian Nappes took place at uppermost crustal levels (Bozkurt and Park 1999).

Ductile thrusting during $D_2$ deformation has resulted in the internal imbrication of the so-called 'core' and 'cover' rocks of the Menderes Massif. The evidence includes: (1) the presence of augen-gneiss klippen above the presently low-angle normal faults along the northern margin of Büyük Menderes graben and along the southern margin of the Gediz graben in the central submassif (Hetzel et al. 1995a, 1995b, 1998; Emre 1996; Emre and Sözbilir 1997; Bozkurt 2000b; Gessner et al. 2000; Lips et al. 2000; Okay 2000). The klippen strongly contrast with the underlying greenschist- facies schists, containing typical upper-amphibolite-facies microstructures. Moreover, klippen in the central submassif made up of the Birgi augen gneiss contain eclogite relics (Dora et al. 1995). (2) The inverted metamorphic sequence documented from the central submassif (Evirgen and Ataman 1982; Hetzel et al. 1998). The sequence is interpreted as the result of numerous north-directed thrusts that post-date the peak of the 'MMM' and place higher-grade rocks onto lower-grade rocks (Hetzel et al. 1998).

The $D_2$ contractional deformation has resulted in a complicated nappe structure in the Massif (see Ring et al. 1999 for review). It is important to note that, in spite of recent claims, the nappe concept in the massif is not new and has been proposed by many workers (Konak et al. 1987, 1994; Dora et al. 1995; Partzsch et al. 1998; Yusufoğlu 1998). Particularly, Konak et al. (1994), based on regional lithostratigraphic correlations, pointed out that the Menderes Massif comprises many nappe slices.

Evidence for an Alpine age of top-to-the N–NNE-thrusting is correlated with that of the main Menderes metamorphism (MMM) and is given in the section on 'Metamorphism'. On the other hand, there are also claims that this deformation was pre-Alpine (Ring et al. 1999; Gessner et al. 2001), but no evidence has been provided. More recently, Lips et al. (2000) provided a new $^{40}Ar/^{39}Ar$ fabric age of 36±2 Ma from augen gneisses exposed along the northern margin of Büyük Menderes graben. This data, once more, confirms that top-to-the N–NNE deformation in the Menderes Massif is Alpine.

## Extensional ($D_3$)

Orogenic collapse and consequent thinning of the previously over-thickened crust followed crustal thickening during Late Palaeogene collision across Neotethys. The collapse was accompanied by extensional exhumation of the metamorphic rocks of the Menderes Massif along presently low-angle normal faults, attesting to the core-complex nature of the Massif (Bozkurt and Park 1994, 1997a, 1997b; Hetzel et al. 1995a, 1995b, 1998; Verge 1995; Emre 1996; Emre and Sözbilir 1997; Koçyiğit et al. 1999; Bozkurt 2000b; Gessner et al. 2000; Gökten et al. 2000; Işık and Tekeli 2000; Lips et al. 2000). Extensional deformation that accompanied orogenic collapse is here referred to as $D_3$ deformation.

$D_3$ fabrics are characterized by an $S_3$ shear-band foliation, strongly overprinting the $S_2$ regional foliation where the latter is preserved in microlithons, and in an approximately NNE–SSW-trending mineral-stretching lineation ($L_3$). In most cases, the relationship between the $D_2$ and $D_3$ fabrics is hampered by close parallelism between the $S_2$ and $S_3$ foliations and between the $L_2$ and $L_3$ lineations. That is why the metamorphic rocks of the Massif apparently show one foliation and one lineation.

$D_3$ fabrics formed during retrograde metamorphic evolution at greenschist-facies conditions. Structural analysis demonstrates that $D_3$ fabrics are the result of bivergent NNE–SSW-directed extension. The present-day subdivision of the Menderes Massif into southern, central and northern submassifs has important implications for the kinematics of $D_3$ deformation: (1) top-to-the-S–SSW fabrics in the southern submassif (Bozkurt and Park 1994, 1997a, 1997b; Hetzel and Reischmann 1996; Lips et al. 2000); (2) bivergent extension in the central submassif being top-to-the N–NNE along the southern margin of the Gediz Graben in the north (Hetzel et al. 1995a, 1995b; Emre 1996; Koçyiğit et al. 1999; Gessner et al. 2000; Gökten et al. 2000) and top-to-the S–SSW along the northern margin of the Büyük Menderes graben in the south (Emre and Sözbilir 1997; Bozkurt 2000b; Gessner et al. 2000; Lips et al. 2000); and (3) top-to-the N–NNE fabrics in the northern submassif (Verge 1995; Işık and Tekeli 2000; Lips et al. 2000). This suggests that the Menderes Massif may be composed of at least three (or even more) small core complexes.

The exhumation of the Massif occurred in the footwall of presently low-angle normal faults. In almost all parts of the Massif, the footwall rocks possess a structural sequence typical of extensional shear zones, made up of (in ascending order) mylonite, brecciated mylonite, and with cataclasites at the contact. Above the low-angle normal faults lie back-tilted Neogene sediments, exhumed in the footwall of active graben-margin-bounding high-angle normal faults. The sediments contain evidence that the deformation in the footwall and sedimentation in the hanging wall were synchronous. The original geometry of the presently low-angle normal faults is controversial and three different views have been proposed: (1) they were originally low-angle structures (Hetzel et al. 1995a); (2) they were high-angle structures but back-rotated to lower angles during the course of extension (Bozkurt 2000b); and (3) these structures were originally thrust faults but reactivated during $D_3$ extension as normal faults (Bozkurt 2000b; Lips et al. 2000).

It is important to note that the cataclastic zone of up to 10 km width, composed of mylonites, protomylonites, blastomylonites, pseudotachylites, brecciated mylonites and cataclasites, exposed along the southern margin of the Gediz graben beneath the Neogene sediments, was first recognized and described in detail by Evirgen (1979) and Evirgen and Ataman (1982). However, the authors did not comment on the tectonic significance of the cataclastic zone simply because the core-complex phenomenon, detachment faulting and associated progressive deformation in the footwall and the role of low-angle normal faults in exhuming deep-seated metamorphic rocks were not yet known. Philippson (1910–1915) first described these cataclastic rocks as 'Tumulos-schutt'.

The age of the $D_3$ deformation is constrained by the age of the oldest unmetamorphosed Neogene sediments (20–14 Ma; Seyitoğlu and Scott 1996 and references therein) and by the syn-extensional Turgutlu granitoid (19.5±1.4 Ma hornblende and 12.2±0.4 to 13.1±0.2 Ma $^{40}Ar/^{39}Ar$ mica cooling ages; Hetzel et al. 1995a; Table 2). The evidence suggests that the $D_3$ deformation commenced 20 Ma and subsequent exhumation and cooling occurred by 12 Ma. This, in turn, suggests that core-complex formation and thus $D_3$ extensional deformation started in the Early Miocene (Seyitoğlu et al. 1992; Hetzel et al. 1995a, 1995b, 1998; Emre 1996; Emre and Sözbilir 1997; Koçyiğit et al. 1999; Bozkurt 2000a, 2000b; Gessner et al. 2000; Lips et al. 2000). Recent documentation of $^{40}Ar/^{39}Ar$ mica ages of 7 Ma from the footwall rocks of the presently low-angle normal fault along the southern margin of the Gediz graben (Lips et al. 2000) suggests either that this fault was reactivated during the neotectonic period or that core-complex formation continued until late Tortonian (Bozkurt 2000b; Sarıca 2000).

On the other hand, the age of the extensional shear zone in the southern submassif is still debated. Hetzel and Reischmann (1996) argued, on the basis of $^{40}Ar/^{39}Ar$ ages of 43–37 Ma, that the shear-zone deformation and subsequent cooling occurred during the Eocene and that unroofing of the submassif is erosional. It is has also been claimed, based on field relations and available literature, that the age of the shear zone is constrained between the age of top-to-the N–NNE contractional fabrics (Eocene: 36±2 Ma; Lips et al. 2000) and the age of the oldest unconformable, unmetamorphosed sediments (~ 21 Ma; Becker-Platten et al. 1977; Bozkurt and Park 1994, 1997a, 1997b; Bozkurt and Satır 2000) and that, based on calculated denudation rates, the unroofing of the submassif was tectonic (and associated with normal faulting) rather than erosional (Bozkurt and Satır 2000).

### Neotectonic ($D_4$)

The last deformation ($D_4$) that affected the Menderes Massif was high-angle normal faulting and consequent graben formation. During this phase, the Massif was dissected into northern, central and southern submassifs, with the Gediz and Büyük Menderes grabens taken as dividing lines.

It has been demonstrated that the high-angle faults clearly cut and displace the low-angle normal faults (Paton 1992; Cohen et al. 1995; Hetzel et al. 1995a; Koçyiğit et al. 1999; Bozkurt 2000a; Gessner et al. 2000; Gökten et al. 2000; Lips et al. 2000), attesting to younger ages for the grabens. Along the graben-margin-bounding normal faults, the metamorphic rocks in the footwall are dissected and elevated.

The age of the grabens is controversial. There are claims that the formation of the grabens, and thus the neotectonic period in western Turkey, commenced by the Early Miocene (Seyitoğlu and Scott 1992, 1996 and references therein) whereas others argue that the grabens are much younger structures (Pliocene–Pleistocene: Koçyiğit et al. 1999; Bozkurt 2000a; Sarıca 2000; Yılmaz et al. 2000). Extension in western Turkey evolved in two separate stages (the episodic two-stage extensional model of Koçyiğit et al. 1999) – a core-complex mode (Early Miocene) and a subsequent wide-rift mode (Bozkurt 2000a). The Miocene

(20–14 Ma: Eskihisar sporomorph association and isotopic ages of the intercalated volcanic rocks) sediments along the exhumed margins of the Gediz and Büyük Menderes grabens are regarded as the earliest sediments of these grabens. The two-stage graben model acknowledges the presence of tilted and folded Miocene sediments but identifies a regional unconformity above which lies undeformed horizontal Pliocene or younger (Koçyiğit et al. 1999; Bozkurt 2000a; Sarıca 2000; Yılmaz et al. 2000). The age of the younger sediments have been determined from mammalian fossils (Ünay and De Bruijn 1998; Ünay et al. 1998; Sarıca 2000). Even Sarıca (2000) argued that some of the coal-bearing sediments dated by Seyitoğlu and Scott (1996) are reworked and transported fragments lying within Pleistocene conglomerates.

## The volume

This special issue of the *International Journal of Earth Sciences* incorporates selected scientific contributions presented at the Third International Turkish Geology Symposium that was held at Middle East Technical University (METU) in Ankara from 31 August–4 September 1998. These contributions describe recent research into the metamorphic belts of western Turkey, in particular the Menderes Massif.

The volume contains 13 papers from a range of international contributors and includes both major tectonic overviews and more local studies with significant regional implications. The meeting at METU provided the opportunity for geoscientists to present up-to-date results and ideas from a wide range of current research activity in this region. About 200 presentations, both oral and poster, covered a wide range of geological topics.

The volume is divided into four sections. The first section is concerned with the deformation and structure of the Menderes Massif. The section opens with paper by Okay. This is followed by the contributions of **Bozkurt**, **Gökten et al.**, **Işık and Tekeli**, **Gessner et al.** and **Lips et al.**, which describe structural aspects of different parts of the Massif. The papers also comment on the metamorphism coeval with the different phases of deformation. In the succeeding section on metamorphism and magmatism, contributions by **Candan et al.**, **Satır and Taubald** and **Koralay et al.** present chemical, petrographic and geochronological evidence, particularly from the southern and central submassifs, on the metamorphic and magmatic evolution of the Menderes Massif. The penultimate section is devoted to the stratigraphy of the 'core' (mainly leptites; **Dora et al.**) and 'cover' (particularly 'marble envelope') rocks (**Özer et al.**). The final section on the Lycian Nappes includes papers describing the metamorphic history of the Lycian Nappes (**Oberhänsli et al.**) and their relation to the Menderes Massif (**Güngör and Erdoğan**).

### Deformation and structure

The first paper by Okay presents evidence from the Aydın Mountains where the central submassif sequence exhibits stratigraphic and metamorphic inversions with older and higher metamorphic-grade rocks exposed in structurally higher parts of the sequence. The submassif is interpreted, based on the similar observations from the Bozdağ horst, as a major synformal anticline. Okay also presents a detailed stratigraphy of the southern submassif in the Babadağ region (west of Denizli). The correlation of structures and stratigraphy of both the southern and central submassifs led the author to interpret the whole Menderes Massif as a major southward closing recumbent fold structure where the southern submassif represents the downward plunging nose of this fold while the central submassif corresponds to the inverted limb.

Okay relates the ubiquitous N–S-trending stretching lineation and top-to-the south ductile shear zones along the southern part of the southern submassif and the northern part of the central submassif as coeval with the folding. The model by Okay therefore predicts that late orogenic collapse-accommodated extension, which began by the latest Oligocene–Early Miocene (Seyitoğlu and Scott 1992), has played an ignorable role in the exhumation of the metamorphic rocks of the central submassif in the Aydın Mountains – dominated by contractional structures with quite subordinate extensional structures. This, in turn, suggests that the southern and central submassifs are not metamorphic core complexes, contrary to some other researchers who have suggested that the ductile shear zones in the Menderes Massif are extensional and have controlled the exhumation of the submassifs (Bozkurt and Park 1994, 1997a, 1997b; Hetzel et al. 1995a, 1995b, 1998; Verge 1995; Emre 1996; Hetzel and Reischmann 1996; Emre and Sözbilir 1997; Koçyiğit et al. 1999; Bozkurt 2000a, 2000b; Bozkurt and Satır 2000; Gessner et al. 2000; Gökten et al. 2000; Işık and Tekeli 2000; Lips et al. 2000). Okay interprets the augen gneiss klippens above the schists along the northern margin of the Büyük Menderes graben as representing the sheared and southward-translated core of the recumbent fold.

A thrust sheet with serpentinite and metabasites lenses was emplaced over the Menderes Massif following the formation of the recumbent fold. Okay interprets this sheet as belonging to the Cycladic metamorphic complex, which is characterized by eclogite relics strongly overprinted by greenschist-facies metamorphism. Okay suggests the age of these contractional deformation phases is constrained between 43 and 20 Ma.

One of the other most important conclusions of Okay's paper supports the previous contention of Candan et al. (1997) and Oberhänsli et al. (1998) that metamorphic rocks with eclogite relics described from the Massif have stratigraphic and metamorphic affinities that are different from the typical Menderes Massif sequence known from the southern submassif, have undergone Eocene Barrovian regional metamorphism, resemble the Cycladic crystalline complex, and therefore should be excluded from the definition of Menderes Massif.

Bozkurt describes the structures observed and reports kinematic data collected along the southern flank of the central submassif at the northern margin of the Büyük Menderes Graben, between Germencik in the west and Umurlu in the east. There, an overall dome-shaped Alpine foliation pattern and a N–NNE-trending stretching lineation characterize the central submassif. Three major rock types crop out in the area: metasediments, orthogneisses ('augen gneisses') and Neogene and Quaternary sediments. Microstructural analysis indicates that the rocks of the central submassif have experienced two distinct phases of Alpine deformation: top-to-the N–NNE contraction and top-to-the S–SSW extension. The top-to-the N–NNE movement in the metasediments suggests deformation along north-directed thrusts and consequent internal imbrication and deformation of the Menderes Massif under greenschist- to upper amphibolite-facies conditions (main Menderes metamorphism) during the latest Palaeogene closure of Neotethys. The contractional phase was followed by extensional collapse, with spreading and thinning of a previously thickened orogen that led to the formation of top S–SSW fabrics under retrograde greenschist-facies conditions along extensional shear zones that accompanied Early Miocene orogenic collapse and continental extension in western Turkey. Progressive exhumation of the lower plate resulted in the formation of a zone of cataclastic rocks below the south-facing, presently low-angle normal fault above which Neogene sediments were tectonically emplaced. The boundary between the cataclasites and structurally overlying Neogene sediments is marked by a major south-dipping low-angle (22–34°) normal fault. Assuming a distributed vertical shear, following Westaway and Kusznir (1993), the initial dip of the fault is calculated as 49±5°. The normal fault was later back-tilted as a result of substantial extension. Similarly, the initial dip for the first generation of normal faults that form the northern boundary of the submassif in the Gediz graben is calculated as 48°. These calculations clearly demonstrate that the northern and southern boundary faults of the submassif were originally high-angle normal faults. The isolated occurrence of augen gneisses and metasediments in the hanging-wall above the south-facing low-angle fault suggests that high-grade augen gneisses were brought into contact with low-grade cover schists along south-facing thrust(s), thus supporting the contention that large-scale thrust faults exist in the submassif and were later reactivated during the extensional period (e.g. Hetzel et al. 1998; Partzsch et al. 1998; Lips et al. 2000).

On the scale of the central submassif, extensional exhumation and related deformation have resulted from non-coaxial flow with a N–NNE-trending sense of vorticity along a north-facing low-angle normal fault at the northern margin of submassif (Hetzel et al. 1995a, 1995b, 1998; Koçyiğit et al. 1999) and with top-to-the-S–SSW shearing at the southern margin (this study; Gessner et al. 2000). The opposite sense of vorticity on either side of the submassif suggests a model of symmetrical gravity collapse of the previously thickened crust and that the central Menderes Massif is a typical 'symmetric metamorphic core complex'. Bozkurt concludes that the results support the previous bivergent extension model of Hetzel et al. (1995a) for the central submassif and that of Vandenberg and Lister (1996) for the whole Aegean, including Menderes and Cycladic massifs.

Gökten et al. presents both structural and stratigraphic evidence from the northern part of the central submassif along the southern margin of the Gediz graben. There the Massif is made of augen gneisses, mica schists and marbles, the age of which ranges from Precambrian to Early Palaeocene. The metamorphic rocks are overlain by Lower–Middle Miocene fluvial and limnic sediments (Alaşehir Group) and by Pliocene–Pleistocene fluvial sediments (Gediz Group). The sedimentary units are separated by an angular unconformity. Deformation in the Massif is described in terms of six phases ($D_1$–$D_6$), of which the first three are compressional while the others are extensional. The pre-Palaeocene deformation in the Massif is ascribed to $D_1$ deformation. The evidence for this phase is completely erased by the subsequent Alpine deformations. The $D_2$ deformation was associated with a regional Barrovian-type metamorphic event during Early Eocene–Early Oligocene time. The development of pervasive regional foliation and the generation of NW–SE-trending, tight to isoclinal folds characterize the $D_2$ phase. The submassif, including both the core and cover rocks, is internally sliced and imbricated along south-facing thrust faults. During $D_2$ deformation, upper levels have moved up-to-the north, similar to movement reported from the southern submassif (Bozkurt 1995; Bozkurt and Park 1999) and central submassif (Hetzel et al. 1998). The third phase of deformation ($D_3$) is an axial-planar folding, caused by NW–SE-directed compression that resulted in the refolding of $F_2$ structures.

Subsequent to internal imbrication and consequent crustal thickening, the area of the Massif orogenically collapsed along low-angle normal faults ($D_4$). The metamorphic rocks in the footwall were progressively mylonitized and exhumed while, in the footwall, the deposition of the Alaşehir Group occurred. The kinematic indicators in the mylonites suggest a top-to-the-N shearing along the N-facing detachment fault. The progressive deformation of the metamorphic rocks produced a typical structural sequence where ductile fabrics are overprinted by brittle structures, similar to those reported in the literature (Hetzel et al. 1995a, 1995b; Emre 1996; Koçyiğit et al. 1999). The Miocene sediments are deformed along NNE–SSW-trending folds suggesting an approximately E–W-directed compression during the Late Miocene ($D_5$). The detachment fault surface also appears to have been folded in N–S direction. However, Koçyiğit et al. (1999) demonstrated the existence of approximately E–W-trending faults in the Miocene sediments and suggested that a short-term N–S-directed compression followed the detachment faulting. They have also pointed out the presence of N–S-trending folds both in the Miocene sediments and the detachment fault, but suggested that these folds indicate a minor refolding that resulted in E–W compression. However, Gökten et al. do not mention the existence or absence of E–W-trending folds in the Miocene sediments. Following the Late Miocene compressional period ($D_5$), the area was subjected to a N–S extension along high-angle normal faults ($D_6$), which has resulted in the formation of the neotectonic Gediz graben. During this phase, the Pliocene–Pleistocene Gediz Group was deposited and exhumed in the footwalls of basin-bounding normal faults.

The authors conclude that N–S extension has not been continuous since the Early Miocene (Seyitoğlu and Scott 1992) but was interrupted by a short-term compression during Late Miocene and that neotectonic extension in western Turkey commenced by the Early Pliocene. This paper supports the episodic, two-stage graben model of Koçyiğit et al. (1999) wherein a Miocene-Early Pliocene first stage (a consequence of orogenic collapse) and a Plio-Quaternary second phase (tectonic escape) is separated by an intervening phase of short-term compression.

Işık and Tekeli present evidence for late-orogenic crustal extension and metamorphic core-complex formation from the poorly known northern submassif – here named the 'Simav metamorphic core complex'- and demonstrate that a low-angle, ductile-to-brittle normal fault (Simav detachment fault) separates non-mylonitic and mylonitic 'core rocks' in the footwall from the allochthonous brittley deformed 'cover rocks' in the hanging-wall. Core rocks comprise high-grade metamorphics, including migmatitic, banded gneisses, mica schists, subordinate amphibolites and marbles. The rocks are compared with Pan-African rocks reported from the southern and central submassifs. The core rocks are intruded by granitoid plutons (Eğrigöz and Koyunoba plutons). The uppermost part of the core rocks, just beneath the detachment, is intensely deformed and mylonitized. The cover rocks contain a (?) Palaeozoic schist–marble unit, Mesozoic recrystallized limestones and an ophiolitic mélange recording the closure of Neotethys during the latest Cretaceous. The metamorphic rocks are unconformably overlain by a Neogene volcano-sedimentary succession wherein the volcanic rocks are dated at 15.2±0.3 and 15.8±0.3 Ma (Ercan et al. 1997).

Deformation in the Simav metamorphic core complex is described in terms of two phases, $D_1$–$D_2$, of which $D_1$ developed within the Precambrian core rocks while the latter ($D_2$) is the result of Tertiary crustal extension and affected the whole complex. $D_1$ is characterized by a penetratively developed regional gneissic foliation with a weak mineral lineation. $D_2$ occurred in two substages: (1) the first stage involved the formation of a variably developed ductile (mylonitic) deformation ($D_{2d}$) of metamorphic and granitic core rocks at greenschist-facies conditions. The mylonites display a well-developed mylonitic foliation and invariably associated SW-plunging mineral-stretching lineation. The kinematic indicators all are consistent in indicating a top-to-NE sense of shear which is in close agreement with the N–S regional extension in western Turkey; (2) the second stage involves brittle deformation ($D_{2b}$) that affected all core and cover rocks and the formation of cataclasites (chloritic breccia) and high-angle normal faults. The timing of $D_2$ deformation is constrained between the age of the Eğrigöz granitoid (29±3 Ma Rb–Sr and 20±0.7 Ma K–Ar biotite ages; Öztunalı 1973 and Bingöl et al. 1982, respectively) and the age of the oldest unconformable Neogene succession (15.5 Ma; Ercan et al. 1996). This indicates that the exhumation of the SMCC and extension-related deformation occurred before Early Miocene. The authors conclude that the exhumation of the southern (Bozkurt and Park 1994, 1997a, 1997b; Hetzel et al. 1998; Bozkurt and Satır 2000; ), central (Hetzel et al. 1995a, 1995b, 1998; Bozkurt 2000b) and northern (this study) massifs occurred at different times and that the Menderes Massif had several phases of core-complex formation rather than a single phase of core-complex development.

Gessner et al. point out that deformation fabrics in Proterozoic/Cambrian granitic rocks of the Çine nappe, and mid-Triassic granites of the Çine and Bozdağ nappes, constrain aspects of the tectonometamorphic evolution of the Menderes nappes of south-western Turkey. Based on intrusive contacts and structural criteria, the Proterozoic/Cambrian granitic rocks of the Çine nappe are subdivided into older orthogneisses (widely known as augen gneisses) and younger metagranites. The granitic rocks document two major deformation events in their history. An early deformation event ($D_{PA}$) during amphibolite-facies metamorphism affected only the orthogneisses and produced predominantly top-to-NE shear-sense indicators associated with a NE-trending stretching lineation. The younger metagranites are deformed both by isolated shear zones, and by a major shear zone along the southern boundary of the Çine submassif. This deformation event is referred to as $D_{A3}$. $D_{A3}$ shear zones are associated with a N-trending stretching lineation, which formed during greenschist-facies metamorphism. Kinematic indicators associated with this stretching lineation reveal a top-to-south sense of shear. The greenschist-facies shear zones cut the amphibolite-facies structures in the orthogneisses. $^{207}Pb/^{206}Pb$ dating of magmatic zircons from a metagranite, that crosscuts orthogneiss containing amphibolite-facies top-to-NE shear-sense indicators, confirms that $D_{PA}$ occurred before 547.2±1.0 Ma. Such an age is corroborated by the observation that mid-Triassic granites of the Çine and Bozdağ nappes lack $D_{PA}$ structures. This means that top-N fabrics in the orthogneisses cannot be related to Alpine thrusting although there are claims that the top-N fabrics both in the core and cover rocks are the result of Eocene Alpine deformation and associated main Menderes metamorphism (Bozkurt 1995, 2000b; Hetzel et al. 1998; Bozkurt and Park 1999; Lips et al. 2000). The younger, top-to-south fabrics formed most likely as a result of top-to-south Alpine nappe stacking that caused the final juxtaposition of the Menderes nappes with the overlying units of the Cycladic blueschist unit, the İzmir–Ankara suture zone and the Lycian Nappes during collision of Anatolia with the Sakarya continent to the north in the Eocene. This further means that the core-cover boundary in the southern Menderes Massif is a south-facing thrust fault whereas the contact has been interpreted as an extensional shear zone beneath which the orthogneisses in the footwall were progressively mylonitized and exhumed (Bozkurt and Park 1994, 1997a, 1997b; Hetzel and Reischmann 1996; Bozkurt and Satır 2000; Lips et al. 2000) during Miocene orogenic collapse.

Gessner et al. also document evidence for bivergent extension of the central submassif. They name the two detachment faults exposed at the northern and southern margin of the submassif as ‘Kuzey detachment’ and ‘Güney detachment’, respectively. The detachments are top-N and top-S cataclastic shear zones, respectively, along which the central submassif in the footwalls was progressively exhumed; Miocene sediments occur in the hanging walls of the detachments. They are cut and offset by younger high-angle, basin-bounding normal faults of the Gediz and Büyük Menderes grabens. The deformation along the detachments accompanied late orogenic extension in western Turkey and commenced by Early Miocene. N–S cross-sections, showing relationships between the detachments and high-angle graben-bounding normal faults, are also illustrated.

Lips et al. present evidence for a multistage exhumation of the Menderes Massif, based on the reassessment of existing studies combined with new $^{40}Ar/^{39}Ar$ laserprobe experiments on two syn-kinematic white mica populations and additional structural investigations. A structural analysis of the Massif, with a synthetic cross section along selected transects, is also presented.

Unlike previous considerations of a single Early Miocene phase of extensional collapse in western Anatolia (Seyitoğlu and Scott 1996), Lips et al. suggest an episodic multi-stage collapse model for the gradual exhumation of the basement series in the Menderes Massif. The authors present isotopic data from the region for three stages of exhumation events associated with the extensional collapse at ~40–35, 20–18 and 7–6 Ma. The first stage of exhumation around 40–35 Ma followed imbrication of the composite basement and northward-directed thrusting. This event is constrained by 36±2 Ma $^{40}Ar/^{39}Ar$ fabric age from the augen gneisses exposed along the northern margin of the Büyük Menderes graben (this work), 35±5 Ma Rb–Sr ages from the southern Menderes Massif (Satır and Friedrichsen 1986), and 43–37 Ma $^{40}Ar/^{39}Ar$ mica cooling ages from the southern submassif (Hetzel and Reischmann 1996). The geochronological investigations suggest that the top-N fabrics in the augen gneisses occurred in Eo–Oligocene times (36±2 Ma). This is consistent with contention that the top-N fabrics are related to Alpine thrusting (Bozkurt 1995, 2000b; Hetzel et al. 1998; Bozkurt and Park 1999), but in contradiction with that of Ring et al. (1999) and Gessner et al. (2000) who consider that the fabrics date from the Neoproterozoic. Lips et al. relate this deformation to young thrusting related to the emplacement of blueschist-facies rocks on top of the tectonostratigraphic pile in the western part of the Massif (Candan et al. 1997; Oberhänsli et al. 1998; Ring et al. 1999). The top-to-the N–NNE deformation was accompanied, or shortly followed, by extensional ductile faulting along the current southern margin of the Menderes Massif, supporting the exhumation. The second phase was the Early Miocene late orogenic extensional collapse during which the metamorphic rocks of the Massif were exhumed in the footwall of a north-dipping low-angle normal fault (Gediz detachment) along the southern margin of the Gediz graben. The deformation started to operate at ductile crustal conditions and continued its activity under semi-ductile conditions. Intrusion of syn-extensional granitoids (Salihli and Turgutlu granitoids) accompanied ductile deformation in the footwall. Synchronously, a south-dipping antithetic low-angle normal fault (Büyük Menderes Detachment) started to operate at semi-ductile to brittle crustal levels along the northern margin of the Büyük Menderes graben. This accelerated the exhumation of the Menderes Massif in the footwall of detachments and resulted in the doming of the Massif. The exhumation of the Massif was accompanied by the formation of two large half grabens (the Gediz and Büyük Menderes grabens) in the hangingwall of detachments as supradetachment basins. The age of this phase is 20–18 Ma (age of the Turgutlu granitoid, Hetzel et al. 1995a; and age of the basin fill of the supradetachment basins, Seyitoğlu et al. 1992). A 7±1 Ma $^{40}Ar/^{39}Ar$ fabric age from a syn-kinematic white mica in footwall rocks of the Gediz detachment reflects a continuous movement along the fault zone during most of the Miocene. The model of Lips et al. predicts that the parts of the Gediz and Büyük Menderes detachments may represent reactivated Alpine thrust planes that were originally low-angle structures although there are claims that these structures were originally high-angle faults and tilted to lower angles during the course of extension (Bozkurt 2000b). The last stage was the result of movement along high-angle normal faults that crosscut the low-angle normal faults. The activity of these faults resulted in the relative uplift of the graben margins and further exhumation of the detachment structures and their footwalls. Lips et al. conclude that the post-7 Ma ages of high-angle normal faults are consistent with the episodic, two-stage extension model that constrain different stages in the development of the Gediz and Büyük Menderes grabens and their basin fill (Koçyiğit et al. 1999; Bozkurt 2000a).

## Metamorphism and magmatism

The metamorphism and magmatism in the Menderes Massif form a second group of papers. Recently, relics of eclogites overprinted by regional Barrovian-type amphibolite-facies metamorphism have been documented from the so-called 'core' rocks of the Menderes Massif (Candan 1994, 1996; Candan et al. 1997; Oberhänsli et al. 1997). The eclogites were originally gabbroic and noritic stocks and veins, which intruded the basement rocks. It is certain that a better understanding of HP metamorphism in the core rocks would provide valuable information for interpreting the tectono-metamorphic evolution of the Menderes Massif.

Candan et al. present evidence for the eclogite-facies assemblages in these mafic rocks and subsequent amphibolite-facies conditions. The Menderes Massif is made up of Pan-African 'core' and a Palaeozoic–Early Tertiary 'cover sequence' imbricated by Late Alpine deformation. The core comprises primarily orthogneisses (widely known as augen gneisses), medium- to high-grade schists, paragneisses, migmatites, metagranites and metagabbros. The metagabbros were previously thought to be Miocene in age (Dora et al. 1988). However, recent investigations have clearly shown that the metagabbros are Precambrian in age and have experienced multiphase metamorphism. The observation of metagabbros only in the so-called Pan-African basement rocks suggests that these rocks intruded the basement before the deposition or tectonic juxtaposition of the Palaeozoic schist envelope. Moreover, metagabbros also occur as xenoliths in the orthogneisses dated at ~550 Ma (Hetzel and Reischmann 1996; Loos and Reischmann 1999). Although there is no direct data for the age of the eclogite-facies metamorphism in the basement, lack of any HP phase in the orthogneisses suggest that the eclogites must have been formed prior to their intrusion. The high-pressure relics in the core are of two types: eclogites (omphacite–garnet–clinozoisite–rutile) and eclogitic metagabbros (omphacite–garnet–rutile–kyanite) where the latter is related to Precambrian gabbroic stocks and is characterized by preserved igneous texture and relic magmatic phases. The P–T conditions of the high-P metamorphism in the eclogites are estimated to be 644 °C with a minimum pressure of ~15 kbar, corresponding to a burial depth of ~50 km. Petrographic and chemical evidence from the high-P rocks suggests a possible pre-eclogitic stage, which is mainly represented by amphibolite-facies inclusions in garnets. Under isothermal decompression conditions, the eclogites were partly to completely retrograded to garnet amphibolites during a Barrovian-type metamorphism at 7 kbar and 623 °C. Orthopyroxene relics in the gneisses and charnockites (Oberhänsli et al. 1997; Candan and Dora 1998) suggest a possible granulite-facies metamorphism during the early stages of the Pan-African evolution of the Massif. Finally, the eclogite relics in the Pan-African basement of the Menderes Massif are correlated with eclogites in the Bitlis Massif. A mechanism of crustal thickening by overthrusting during the Pan-African orogeny has been proposed for the eclogite occurrences in the Menderes Massif.

Satır and Taubald report new stable isotope data for hydrogen and oxygen and the chemical compositions of quartz, feldspar, biotite and muscovite from 11 selected samples of the crystalline 'core' and 12 from the schist 'cover' along a north–south profile from the edge to the centre of the southern submassif. They address the problems of the equilibrium or disequilibrium character of mineral assemblages, temperatures of equilibration, and characterization of relevant exchange processes and source of fluids.

The submassif comprises an inner crystalline core with gneissic rocks and an enveloping schist cover with predominantly metasedimentary rocks. Both units have a complex metamorphic history including a late Alpine overprint. A N–S traverse within the cover schists in the southern submassif shows a pronounced decrease both in the $\delta^{18}O$ values of quartz and in the fractionation between quartz and coexisting minerals. This indicates a sedimentary origin for these rocks and an increase in temperature from south to north during the Alpine, main Menderes metamorphism (MMM). More positive $\delta^{18}O$ values in all minerals from the cover schists may reflect a higher abundance of sedimentary precursor material, while biotites and muscovites from the edge and centre are indistinguishable with regard to their hydrogen-isotopic compositions. The oxygen isotope fractionation curves yield temperatures of 528–531 °C for the core rocks, whereas temperatures calculated for schist cover samples increase from 457 to 577 °C along a N–S profile. This may have been caused by either increasing influence of sedimentary material or by a 'mixing' trend between upper metasedimentary series and an overlying crystalline core, produced by retrograde fluid infiltration. The calculated average oxygen-isotopic temperatures of 450 to 600 °C indicate an increase in metamorphic grade (from the corundum isograd temperature of 420 °C to the staurolite–sillimanite isograd temperature of ~600 °C) from the 'cover' to the 'core' of the submassif; however, pre-Alpine temperatures may still be reflected in quartz-muscovite mineral pairs. The calculated temperatures are similar to those estimated by mineral stability (Dürr 1975) for Alpine assemblages. Similarly, previous analysis of mineral oxygen-isotopic compositions showed an increase from 420 °C in the 'schist cover' to ~600 °C in the 'core' (Satır and Taubald 1991). Most biotite samples have very low D values, implying that the hydrogen-isotope composition of biotite was probably disturbed by Alpine extensional deformation as this allowed infiltration of and exchange with D-depleted meteoric water, whereas muscovite retained its Alpine characteristics. This contention is also supported by the age of altered biotites, ranging between 13 and 50 Ma, where the younger age of 13 Ma is attributed to rejuvenation during extensional deformation following Alpine compressional deformation associated with the MMM. Satır and Taubald conclude that, in the southern submassif, hydrogen- and oxygen-isotopic compositions of pre-Alpine mineral assemblages were considerably disturbed during Alpine metamorphism.

Koralay et al. present geochronological and geochemical evidence from the leucocratic orthogneisses of the central submassif and focus on the Early Cimmerian tectono-metamorphic/magmatic evolution of the submassif. They report on three areas in the eastern part of the central submassif: Derbent, north of Alaşehir, and north-west of Nazilli. In Derbent, the metamorphic series is made up of four nappe units. In ascending order, the first three-nappe units were formed by the internal imbrication of the Pan-African basement series whereas the uppermost nappe unit belongs to the Early Palaeozoic cover series, comprising a phyllite–quartzite–marble intercalation. The lowest nappe unit is made up entirely of garnet–mica schist. The second nappe unit predominantly comprises garnet–mica schists that contain thin carbonate horizons. The Triassic leucocratic orthogneisses are found only in this nappe unit. Furthermore, the garnet–mica schists were intruded by 560–570 Ma Pan-African orthogneisses (Koralay et al. 1998). The third nappe slice is dominated by orthogneisses (Dora et al. 1995; Partzsch et al. 1998; Ring et al. 1999). The leucogranites intruded the metasediments of the Upper Palaeozoic 'cover schists' north-west of Nazilli and the Precambrian 'cover schists' north of Alaşehir. Typical contact metamorphic assemblages characterize the original intrusive contact relation between the leucocratic orthogneisses and the country rocks.

The protoliths of the leucocratic orthogneisses were calc-alkaline, peraluminous, S-type, syn- to post-collisional granites with a crustal affinity. Single zircon $^{207}Pb$–$^{206}Pb$ evaporation dating of three samples from the leucocratic orthogneisses yielded ages between 235 and 246 Ma (Early–Middle Triassic). These ages are interpreted as the age of intrusion. Similar ages (227–240 Ma) have also been reported from the other parts of the central submassif (Dannat and Reischmann 1998). It is suggested that the protoliths were emplaced after Early Cimmerian orogeny during which the metamorphism of the Palaeozoic series occurred. This event is attributed to the collision of Laurasia and Gondwana and consequent closure of the Palaeotethyan Ocean along a south-dipping subduction zone during Permo-Triassic time. The granites were later deformed and metamorphosed during Alpine orogeny and transformed into orthogneisses. Rb–Sr mica ages (22.5±0.2–24.2±0.2 Ma biotite and 26.2±0.3–29.8±0.2 muscovite ages; Koralay et al. 1998) demonstrate that cooling of these rocks following peak Alpine metamorphism occurred during the Early Oligocene–Early Miocene. Koralay et al. conclude that, since Triassic granites are common in the Cycladic Massif, the Menderes and Cycladic massifs have a common pre-Early Triassic history.

The authors identify three phases of magmatic activity during the evolution of the Menderes Massif: at Late Proterozoic/Cambrian boundary (550 Ma; Hetzel and Reischmann 1996; Loos and Reischmann 1999); in the Early–Middle Triassic (227–250 Ma; Dannat and Reischmann 1998; Koralay et al. 1998) and in the Early Miocene (12.2±0.4 to 19.5±1.4 Ma; Hetzel et al. 1995a). This manuscript documents, for the first time, solid evidence for Triassic magmatic activity in the Massif.

## Stratigraphy

The succeeding two papers deals with the stratigraphy of the Menderes Massif. Dora et al. present new evidence on the origin of so-called "leptites" from the 'core' rocks of the Menderes Massif. Leptite–gneisses crop out over a large area in the three submassifs of the Menderes Massif. They are partially migmatized and occur as huge xenoliths within the Pan-African orthogneisses (known as augen gneisses), and are cut by granites dated at 550 Ma and by many gabbroic and noritic stocks. The leptites are typically purple, purplish grey and grey massive rocks with poorly developed foliation and contain black to yellowish-black relic cordierite porphyroblasts, replaced completely by sillimanite and garnet. Based on the presence of porphyroblasts, the leptites are divided into two groups: (1) homogeneous paragneisses that lack porphyroblasts, and (2) spotted paragneisses with extensive ellipsoidal mineral clots. Leptites also contain widespread boundinaged lenses and bands of calc-silicates. Cordierite relics and the rare presence of orthopyroxene suggests that Pan-African metamorphism reached granulite-facies conditions and then was overprinted by an upper-amphibolite-facies metamorphism, as suggested by the presence of sillimanite replacing cordierite in the clots.

Based on detailed logs, field observation and petrography, it is suggested that the protoliths of the leptites were predominantly sandy sediments (litharenite) that show both vertical and lateral passage into clayey sediments. Zircons from the leptite gneisses are also typically rounded, supporting the contention that the leptites are sedimentary in origin. In addition, the REE patterns of the leptite–gneisses are characteristic of supra-crustal sediments derived from a cratonic provenance. Unlike the long-standing view that the protoliths of leptite-gneisses are rhyolitic-andesitic volcanic rocks that formed in a calc-alkaline island-arc (Dora et al. 1988, 1990), Dora et al. propose that leptites are of supra-crustal sedimentary origin and should be renamed sillimanite–garnet gneiss and/or paragneiss. It is also concluded that that the protoliths of the paragneisses were deposited in the Late Proterozoic because they experienced Pan-African multiphase metamorphism and were subsequently intruded by ~550-Ma-old granites.

Özer et al. presents, for the first time, a detailed biostratigraphy of the uppermost levels of the metamorphic sequence (the so-called 'marble envelope') from four reference areas in the southern and eastern Menderes Massif. It is common to all research conducted to date that the stratigraphy of the Massif is controversial and that a better understanding of this stratigraphy would provide valuable information about the tectonic evolution of not only the Menderes Massif but of south-western Turkey. The authors describe Cretaceous sequences with rudists and pelagic marbles from the Yatağan–Kavaklıdere area. The so-called 'marble envelope' of the southern Menderes Massif in Akbük–Milas and Yatağan–Kavaklıdere areas comprises four distinct units: (?) Cenomanian emery-bearing marbles, Santonian–Campanian rudist-bearing marbles, Upper Campanian–Maastrichtian red-pinkish pelagic marbles, and Middle Palaeocene flysch-like sediments. The contact between the mica schists of the so-called 'schist envelope' and the overlying emery-bearing marbles is gradational, as opposed to previous views of an unconformity (Dürr 1975; Çağlayan et al. 1980; Konak et al. 1987). Around Çal and Tavas, carbonate sedimentation ceased abruptly at the end of the Cretaceous. The Santonian–Campanian carbonates are unconformably overlain by Lower Eocene clastics in the Tavas area, and by Lower–Middle Eocene sediments in Çal area. However, at Honaz Dağı, the carbonate sedimentation continued throughout the Early Tertiary (Okay 1989b). In latest Campanian–Maastrichtian time, the deposition of pelagic carbonates and the drowning of the same carbonate platform were followed by the deposition of deepwater flysch-like clastic sediments. The break in carbonate sedimentation in the Çal and Milas areas mark the initial obduction of the Lycian ophiolite onto the passive margin to the north of the Menderes carbonate platform during the latest Cretaceous. Final overthrusting of the Lycian Nappes over the southern submassif occurred in Late Palaeocene. The nappes progressively moved south-eastward throughout the Palaeocene, as suggested by continued sedimentation until Early to Middle Eocene times at Tavas and Çal.

Özer et al. conclude that sedimentation in the southern Menderes Massif continued at least until the Middle Palaeocene, not Eocene (cf. Konak et al. 1987), and that thrusting of the Lycian Nappes onto the southern Menderes Massif occurred probably during the Middle Palaeocene – earlier than previously thought (Late Eocene; Collins and Robertson 1998, 1999).

## Lycian nappes

The volume finishes with two papers on the Lycian Nappes and their relation to the tectono-metamorphic evolution of the Menderes Massif. Oberhänsli et al. describes, for the first time, the occurrences of Fe–Mg carpholites and their breakdown products, such as chlorite, pyrophyllite, phengite and chloritoid along the basal thrust sheets of the Lycian Nappes and from klippen from the Lycian Nappes atop the Menderes Massif. The distribution of Fe–Mg carpholite and its relics suggest an extensive HP overprint in the sediments of the passive continental margin, which prevents any attempt to correlate the HP–LT event in the Lycian Nappes with the 25 Ma HP overprint in the Crete. The data presented also do not support the hypothesis of subduction in a marine basin located to the south of present-day Menderes Massif. Oberhänsli et al. then discuss the geodynamic significance of carpholite occurrences, with the possibility of two alternative hypotheses, and concluded that a geodynamic scenario producing cold exhumation paths and avoiding decompression must be envisaged because the stability field of carpholite is restricted to a narrow window of 300–420 °C at elevated pressure of >8 kb. The authors suggest a simple shear-related exhumation and nappe-transport model that resulted in cool retrograde P–T paths and allowed preservation of Fe–Mg carpholite close to a major shear zone at the base of a Lycian Nappe stack. Oberhänsli et al. suggest that occurrences of Fe–Mg carpholite and its relics at the base of the Lycian Nappe pile document a rapid tectonic transport along with intense deformation over a relatively cold continental margin, similar to an Oman-type obduction complex.

Güngör and Erdoğan present evidence for the emplacement age and direction of the Lycian Nappes over the Menderes Massif from the Söke–Selçuk region. The data presented in this paper are an important contribution to a better understanding of the polyphase metamorphic and deformational history of the Menderes Massif. In the Söke–Selçuk region, Triassic–Jurassic marbles and a calc-schist–mica schist alternation at the base, and emery- and rudist-bearing massive marbles at the top represent the Menderes Massif. A wild-flysch-type blocky mélange, characterized by meta-ophiolite and emery-bearing marble blocks within a highly sheared, garnet schist matrix, overlies the metamorphic rocks along a gradational contact. In the region, the Lycian Nappes are represented by a typical carbonate sequence and occur as tectonic slices above the Menderes Massif.

The boundary between the Lycian Nappes and the Menderes Massif is marked by low-angle fault along which thin slices of red–green phyllite are exposed. It is suggested that the deformation is localized along the contact where phyllites acted as a décollement zone. The S–C fabrics from phyllites suggest that upper levels have moved SSW during emplacement of the Lycian Nappes. This contact also marks a major metamorphic break between the Menderes metamorphics and the structurally overlying Lycian Nappes. The kinematics and the metamorphic break indicate that this fault is a low-angle normal fault. The authors also suggest that the emplacement of the Lycian Nappes was a post-metamorphic event relative to the Barrovian-type regional main Menderes metamorphism (MMM). This event occurred sometime between the Early–Middle Eocene (age of MMM) and Middle Miocene (age of oldest unconformable sediments).

The conclusion of this paper, that the contact between the Lycian Nappes and the Menderes Massif is a low-angle normal fault that post-dates the MMM, has profound tectonic significance. The MMM is attributed to burial of the Massif area beneath the base of the southward-advancing Lycian Nappes (Şengör and Yılmaz 1981; Şengör et al. 1984b). Structural analysis of coeval fabrics, however, indicates that upper levels of the Massif have moved to the N–NNE (Bozkurt 1995; Hetzel et al. 1998; Bozkurt and Park 1999; Bozkurt 2000b), thus suggesting N-directed thrusting (a direction opposite to that which has been proposed as the cause of metamorphism and coeval deformation). Bozkurt and Park (1999) suggested that (1) either the Lycian Nappes have no connection with the metamorphism and deformation of metasediments in the Menderes Massif; or (2) the previous interpretation is wrong and the tectonic transport of the Lycian Nappes was from south to north. The top-N deformation is therefore attributed to a northward back-thrusting and consequent internal imbrication of the Menderes Massif. The data in the Güngör and Erdoğan paper add more flavour to the relationship between the Lycian Nappes and the MMM.

## Conclusion

We believe that the papers presented in this volume contribute greatly not only to the understanding of the tectonic evolution of the Menderes Massif, but also of western Turkey and the entire eastern Mediterranean region. The data presented on the structure, evolution and contact relations of the Menderes Massif will be useful in elucidating the development of the Alpine orogen in western Turkey.

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