Pre-Alpide Palaeozoic and Mesozoic orogenic events in the Eastern Mediterranean region

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Abstract: We review the Palaeozoic–Early Mesozoic evolution of the Eastern Mediterranean–Balkan region with special reference to Anatolia, and provide new isotopic data on the Palaeozoic magmatic and metamorphic rocks. The pre-Alpide evolution of the region involves episodic growth of Laurussia by accretion of oceanic terranes and Gondwana-derived microcontinents. Terrane accretion, associated with deformation, magmatism and regional metamorphism, took place in the Late Ordovician–Early Silurian, Carboniferous, Late Triassic–Early Jurassic and Mid-Jurassic. The Late Ordovician–Early Silurian accretion is inferred from stratigraphic and faunal records in the Pontides; other evidence for it is buried under young cover on the northern margin of the Black Sea. The Carboniferous orogeny is related to southward subduction and continental collision on the southern margin of Laurussia. It is marked in the Pontides by high-grade regional metamorphism, north-vergent deformation and post-orogenic latest Carboniferous–Early Permian plutonism. The latest Triassic–Early Jurassic Cimmeride orogeny involved the collision and amalgamation of an oceanic plateau to the southern margin of Laurasia. It is represented by voluminous accretionary complexes with Late Triassic blueschists and eclogites. Late Jurassic regional metamorphism and deformation is confined to the Balkans, and is the result of continental collision between the Rhodope–Serbo-Macedonian and Strandja blocks in the Late Jurassic. The Palaeozoic geological history of the Balkans and the Pontides resembles that of Central Europe, although the similarities end with the Mesozoic, as a consequence of the formation of Pangaea.

Orogenic belts and Mesozoic oceanic basins occupy the Eastern Mediterranean region between the stable areas of the East European Craton in the north, and NE Africa and the Arabian Platform in the south (Fig. 1). The East European Craton, as represented by the Ukrainian Shield north of the Black Sea, is an Archaean–Palaeoproterozoic crystalline terrane. The consolidation of the southern part of the East European Craton was completed by 2300–2100 Ma (e.g. Bogdanova *et al.* 1996; Claesson *et al.* 2001). In the Early Palaeozoic, the East European Craton formed part of the Baltica plate, which collided in the west with Laurentia, Avalonia and Armorica, creating Laurussia in the Late Palaeozoic (e.g. Pharaoh 1999; Matte 2001; Warr 2002). In contrast, Africa and the Arabian Platform constituted part of Gondwana, which preserved its unity until the Early Mesozoic opening of the southern Atlantic. Very large areas in the northern margins of Gondwana are characterized by Neoproterozoic–Cambrian plutonism and metamorphism forming part of the Pan-African–Cadomian orogenic cycle (e.g. Stern 1994), and are therefore readily distinguished from the Palaeoproterozoic basement of the East European Craton.

During the Late Palaeozoic and Mesozoic, Tethyan oceanic basins separated Laurussia from NE Africa–Arabia. Parts of the present Eastern Mediterranean Sea represent a Triassic to Jurassic remnant of a Tethyan oceanic crust, whereas the Black Sea is a Late Cretaceous oceanic back-arc basin that opened during the northwards subduction of a Tethyan ocean (e.g. Şengör & Yılmaz 1981; Garfunkel 1998).

The Anatolian–Balkan region between the Eastern Mediterranean and the Black Sea consists of several small continental fragments or terranes bearing evidence of various periods of deformation, metamorphism and magmatism, the latest and strongest of which is the Alpide orogeny. The Alpide orogeny resulted in the amalgamation of the continental fragments into a single landmass in the Tertiary. Previous to this amalgamation, these continental fragments were situated on the margins of the Tethyan oceans, or formed small edifices within the ocean. The pre-Alpide orogenic history of these terranes forms the subject of this paper.

Terranes in the Eastern Mediterranean–Balkan region

As orogenic events are restricted to the plate margins, identification of former plates is important for an understanding of the orogenic evolution. Only the deformed continental parts of the former microplates would be expected to be preserved, and they would be rimmed by sutures marked by linear zones of accretionary complex, blueschist, eclogite and ophiolite, and would show distinctive stratigraphic features, especially if the intervening oceans were large. The main methods used in the differentiation of the former plates include recognition of sutures, palaeomagnetism, faunal provinciality and stratigraphy. A complication in this picture is that the number and configuration of the plates change through time. For example, the Anatolian microplate, which makes up most of the present Anatolian landmass, did not exist before the Miocene (e.g. Şengör 1979). Furthermore, late-stage strike-slip faulting may lead to the dispersal of a single palaeoplate, as happened during the Cretaceous opening of the Black Sea (Okay *et al.* 1994). Therefore, a better term for such a palaeo-plate would be 'terrane', and this would be distinguished by its distinctive stratigraphic, palaeomagnetic, faunal, structural, metamorphic and magmatic features. The following terranes are defined in the Anatolian region from south to north: the Arabian Platform comprising part of SE Anatolia, the Anatolide–Tauride Block, the Kirşehir Massif, the Sakarya Zone, the Istanbul Zone and the Strandja Massif (Fig. 1; Okay & Tüysüz 1999). Of these, the last three are grouped together as the Pontides. Pre-Alpide orogenic events are especially strong and well documented in the Pontides (Fig. 2). With the possible exception of its northwestern margin, the Anatolide–Tauride Block was largely free of Palaeozoic and early Mesozoic deformation. Therefore, most of this review concerns the pre-Alpide geological history of the Pontides.

In addition to the Anatolian terranes listed above, several other terranes have been defined in the Balkans (e.g. Burchfiel 1980; Stampfli 2000; Stampfli *et al.* 2001). The major ones include the Moesian Platform, the Rhodope–Serbo-Macedonian Massif, Pelagonia (which comprises the Pelagonian Zone and the

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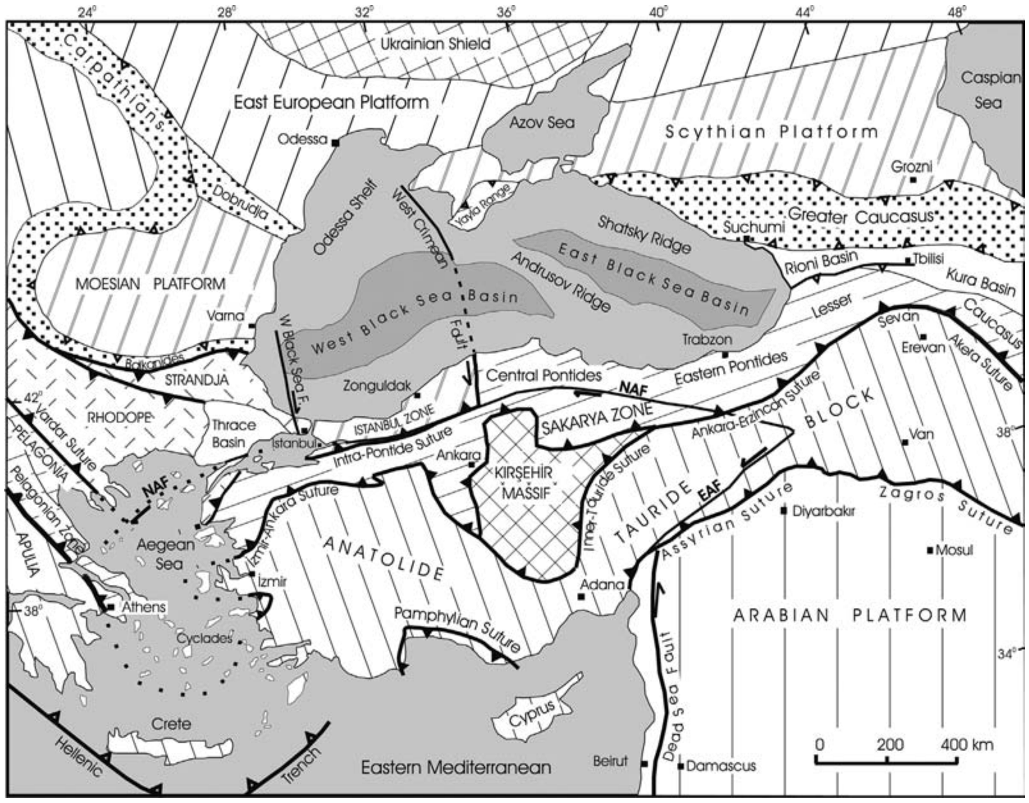

Fig. 1. Tectonic map of the Eastern Mediterranean region showing the major terranes and the bounding sutures. The filled triangles indicate the polarity of the subduction (modified from Okay & Tüysüz 1999). NAF, North Anatolian Fault; EAF, East Anatolian Fault.

Cyclades) and Apulia, which includes Greece south of the Pindos suture (Fig. 1; Stampfli *et al.* 2001). Palaeozoic rocks are not exposed on Apulia, so it is not known whether this region was affected by the Variscan orogeny. Assuming that it was not, then Apulia probably forms a single terrane jointly with the Anatolide–Tauride Block. Isotopic data indicate Carboniferous (325–295 Ma) plutonism and regional metamorphism in the Pelagonian Zone and in the Cyclades (see the discussion by Vavassis *et al.* 2000), which contrasts with the Neoproterozoic magmatic and metamorphic basement ages for the Anatolide–Tauride Block. Their Mesozoic histories are also different, with Late Jurassic–Early Cretaceous ophiolite obduction on the northern margin of Pelagonia contrasting with the Late Cretaceous ophiolite obduction on the Anatolide–Tauride Block. The contact between Pelagonia and the Anatolide–Tauride Block is represented by an Eocene thrust, where the cover sequence of the Cycladic Massif, metamorphosed at blueschist-facies conditions in the Eocene, is thrust on the Menderes Massif (Fig. 3; Okay 2001). This contact, which may represent the extension of the Pindos suture, may link up with the İzmir–Ankara suture, in which case Pelagonia will be correlated with the Sakarya Zone (Fig. 1). Such a correlation is supported by the similar Variscan magmatic and metamorphic ages from Pelagonia and the Sakarya Zone (see below), although their Mesozoic histories are separate.

The Alpide orogeny in the Mediterranean area started with the convergence between the Africa–Arabian and Eurasian plates during the Late Mesozoic. The relative movement between these two plates was sinistral strike-slip from Early Jurassic to mid-Cretaceous time (e.g. Savostin *et al.* 1986; Dewey *et al.* 1989). Starting with the Cenomanian–Albian (100–90 Ma) the Africa–Arabian and Eurasian plates started to converge, presumably with the initiation of subduction. In the geological record the Albian flysch of the Central Pontides (Tüysüz 1999), the Turonian (*c.* 91 Ma) high-temperature–medium-pressure metamorphism in the Kırşehir Massif (Whitney *et al.* 2003), and the Campanian (*c.* 80 Ma) high-pressure–low-temperature metamorphism in the Anatolide Tauride Block (Sherlock *et al.* 1999) are the first recognized events of the Alpide orogeny in Turkey. Therefore, the period discussed in this review extends to the Early Cretaceous. We also do not discuss the Neoproterozoic–Cambrian-aged Pan-African orogenic events in Anatolia, which, although important (e.g. Kröner & Şengör 1990; Yiğitbaş *et al.* 2004), are poorly preserved and documented.

The Anatolide–Tauride Block

The Anatolide–Tauride Block has a Neoproterozoic crystalline basement overlain by a sedimentary succession ranging from Mid-Cambrian to Miocene in age (e.g. Gutnic *et al.* 1979; Özgül 1984, 1997). It was strongly deformed and partly metamorphosed during the Alpide orogeny, and now consists of metamorphic regions in the north (the Anatolides) and a south-vergent Eocene nappe stack in the south (the Taurides). Stratigraphy of several nappe units in the Taurides reveals Palaeozoic to Mesozoic sedimentary sequences with no evidence of pre-Cretaceous deformation or metamorphism (e.g. Gutnic *et al.* 1979; Özgül 1984, 1997). Rare reports of Late Triassic deformation in the Anatolide–Tauride Block (e.g. Monod & Akay 1984) have been questioned and need confirmation (Göncüoğlu *et al.* 2003). The Anatolide–Tauride Block is here considered as not affected by significant pre-Alpide Phanerozoic contractional deformation, which contrasts with the regions to the west and north that were deformed and metamorphosed during the Variscan orogeny.

The largest outcrops of the Precambrian basement in the Anatolide–Tauride Block are found in the Menderes and Bitlis massifs (Figs 3 and 4). In the Menderes Massif, the metagranitoids, which make up most of the crystalline basement, have been dated as to *c.* 550 Ma using a stepwise Pb evaporation method on zircons (Hetzel & Reischmann 1996; Loos & Reischmann 1999); this is a similar age to those found in the Arabian Platform and in NE Africa. The eclogite-facies metamorphic rocks in the basement of the Menderes Massif are also Neoproterozoic in age (Candan *et al.* 2001). The Palaeozoic stratigraphy in the Anatolide–Tauride Block is also similar to

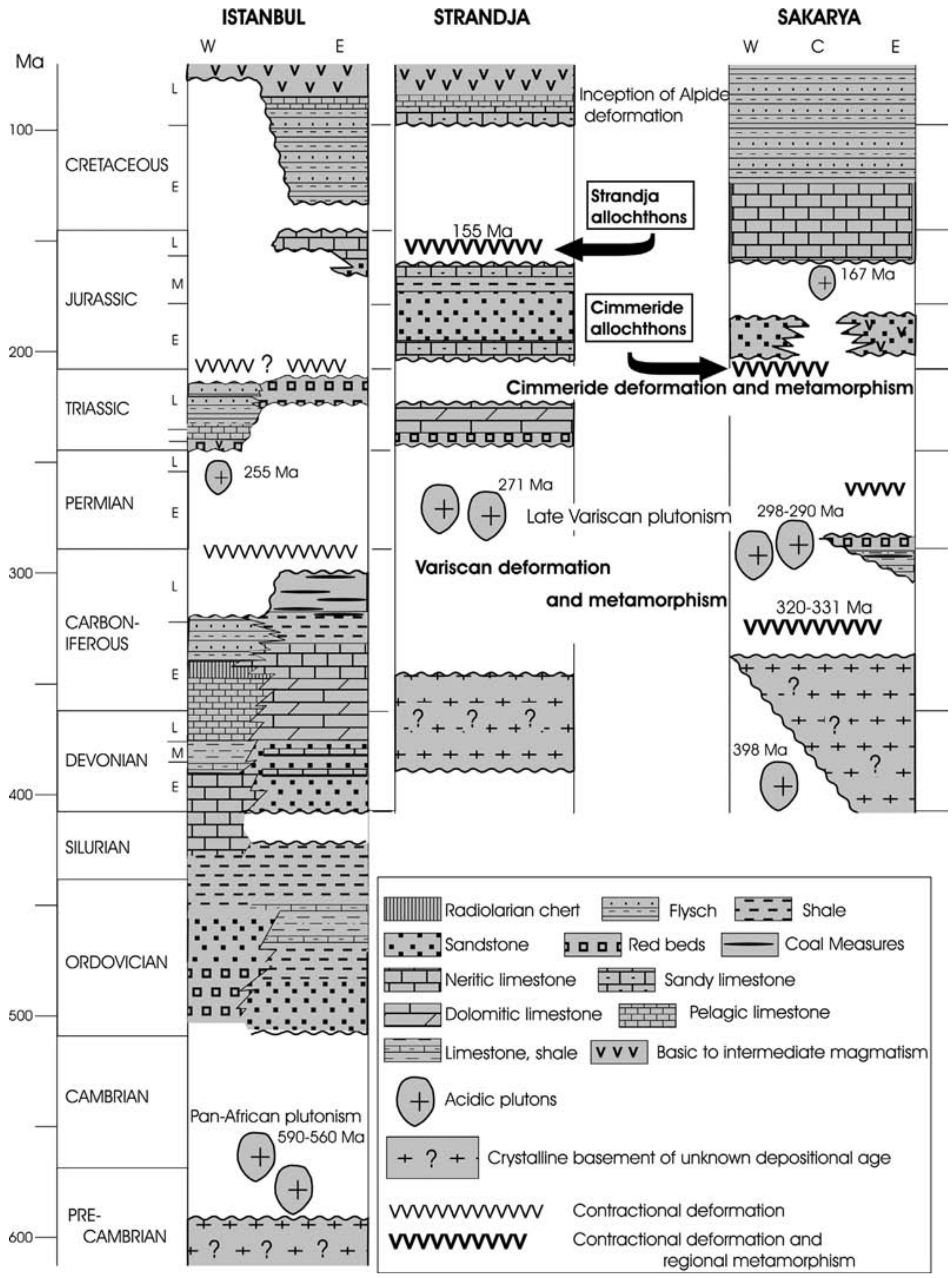

Fig. 2. A chronostratigraphic chart showing generalized geological relationships of the Pontic terranes. The main sources of the data are: for the Istanbul Zone, Gedik (1975), Alişan & Derman (1995), Görür *et al.* (1997) and Dean *et al.* (2000); for the Strandja Zone, Chatalov (1988), and Okay *et al.* (2001); for the Sakarya Zone, Altıner *et al.* (1991) and Okay & Leven (1996).

that of the northern margin of the Arabian Platform. Hence, since Smith (1971), all palaeogeographical reconstructions place the Anatolide–Tauride Block into the Eastern Mediterranean Sea between the Levant and Egyptian margins. Recent evidence for the latest Ordovician (Hirnantian) glaciation in the Anatolide–Tauride Block, including the presence of striated pebbles and striated basement (Monod *et al.* 2003), supports this pre-drift position. Stratigraphic evidence from the Levantine (e.g. Bein & Gvirtzman 1977; Garfunkel & Derin 1984) and Gondwana margins in SE Anatolia (Fontaine *et al.* 1989) indicates that the Anatolide–Tauride Block rifted away from Gondwana during the Triassic or Early Jurassic with the opening of the Tethyan ocean. However, it was never far away from Gondwana, and drifted with Gondwana to the north, as shown by its Jurassic palaeomagnetic record (Piper *et al.* 2002) and by its Jurassic–Cretaceous stratigraphy, which resembles that of the SE Anatolia.

The Istanbul Zone

The Istanbul Zone consists of a Neoproterozoic crystalline basement overlain by Lower Ordovician to Eocene sedimentary rocks (Fig. 2; e.g. Haas 1968; Görür *et al.* 1997). Before the Late Cretaceous opening of the Black Sea, the Istanbul Zone was situated east of the Moesian Platform and adjacent to the Scythian Platform (Okay *et al.* 1994). These three tectonostratigraphic units have a similar basement and show a similar Palaeozoic–Mesozoic stratigraphic development (e.g. Tari *et al.* 1997; Nikishin *et al.* 1998), and formed a single late Proterozoic–Early Palaeozoic terrane, named here the MOIS (Moesian–Istanbul–Scythian) terrane.

The crystalline basement of the Istanbul Zone is characterized by voluminous granitoids, which intrude low- to medium-grade metasediments and metavolcanic rocks (Ustaömer & Rogers 1999; Yiğitbaş *et al.* 1999, 2004). The granitoids have yielded U–Pb and Pb/Pb

Fig. 3. Tectonic map of the western Anatolia illustrating the geological features discussed in the text. The cross-hatched area shows the extent of the metamorphosed Palaeozoic and Mesozoic rocks of the Menderes Massif.

zircon ages of 590–560 Ma, and the surrounding metasediments have provided similar Rb–Sr mica ages (Chen *et al.* 2002). The geochemistry of the granitoids and the metavolcanic rocks indicates a subduction-zone setting in the Neoproterozoic. In terms of age, lithology and geochemistry, the basement of the Istanbul Zone is similar to the Pan-African basement of northern Gondwana, and unlike the East European Craton. Therefore, the Istanbul Zone is generally regarded as a Peri-Gondwana terrane (e.g. Stampfli 2000).

The Neoproterozoic basement of the Istanbul Zone is overlain by a thick Palaeozoic sedimentary succession extending from the Ordovician to the Carboniferous (Fig. 2). There are significant stratigraphic differences between the western and eastern parts of the Istanbul Zone, which led to a suggestion that the Istanbul Zone consists of two terranes, the Istanbul terrane in the west and the Zonguldak terrane in the east (Kozur & Göncüoğlu 1998; Stampfli *et al.* 2002; von Raumer *et al.* 2002). The most important difference is in the Carboniferous system, which in the west is represented by Viséan radiolarian cherts overlain by siliciclastic turbidites, but in the east, in the Zonguldak region, is represented by Viséan neritic carbonates overlain by Namurian to Westphalian coal measures (Figs 2 and 5). However, evidence for a Phanerozoic ocean, in terms of pelagic sedimentary rock, mélange, ophiolite or blueschist, is missing between the western (Istanbul *sensu stricto*) and eastern parts (Zonguldak) of the Istanbul Zone (see Fig. 5), and the stratigraphic differences are a result of facies changes. A similar situation has been reported in the Moesian Platform, where neritic carbonate deposition in the Tournaisian in the north is replaced by radiolarian chert sedimentation in the south in the Elovitza region (Haydoutov & Yanev 1997). During the Carboniferous, the MOIS terrane was part of the southern continental margin of Laurussia. Turbidite deposition in a continental slope setting took place in the western part of the Istanbul Zone and on the southern margin of the Moesian Platform, whereas coal deposition occurred in swamps in the north (Fig. 6). The palaeogeographical situation was similar to that of SW Britain at the same period, when coal measures were being deposited in Wales and siliciclastic turbidites (Culm facies) in Cornwall and Devon (Fig. 6; e.g. Guion *et al.* 2002).

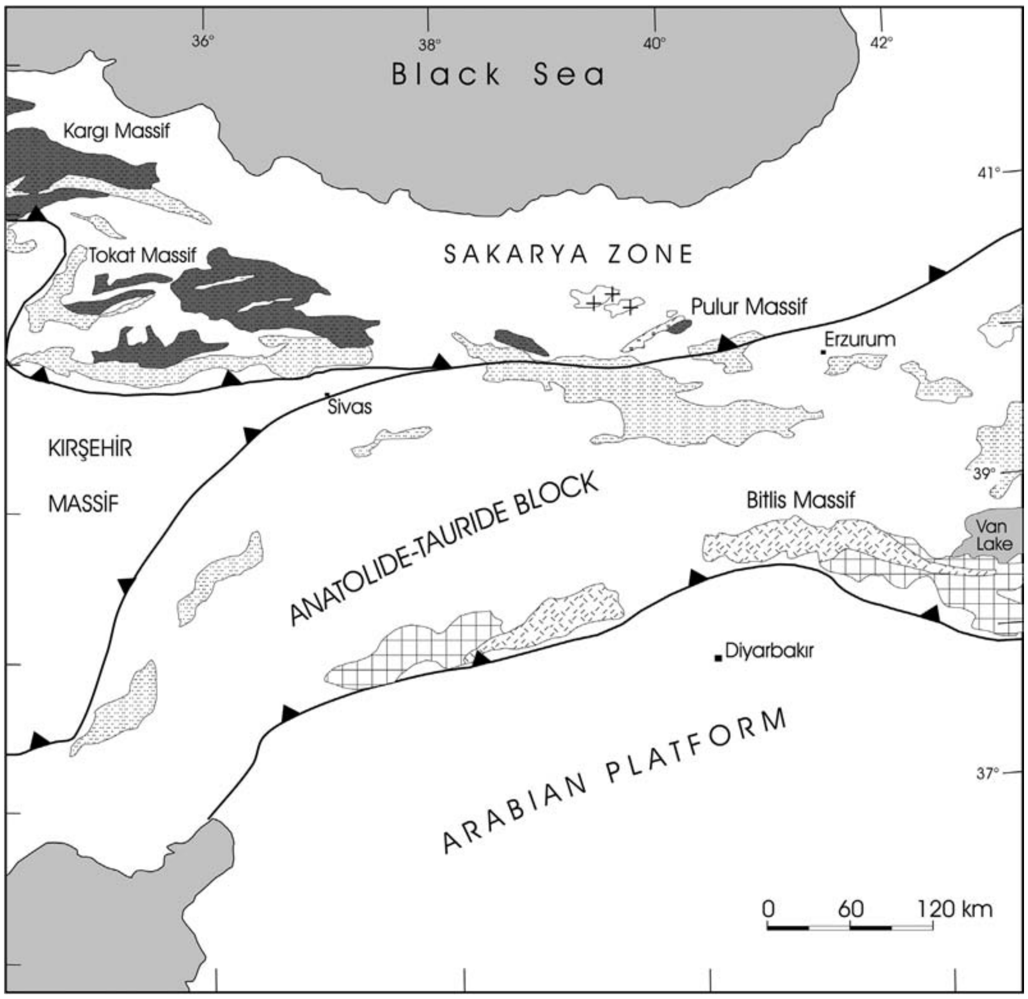

Fig. 4. Tectonic map of eastern Anatolia illustrating the geological features discussed in the text. (For legend see Fig. 3.) The cross-hatched area shows the extent of the metamorphosed Palaeozoic and Mesozoic rocks of the Bitlis Massif.

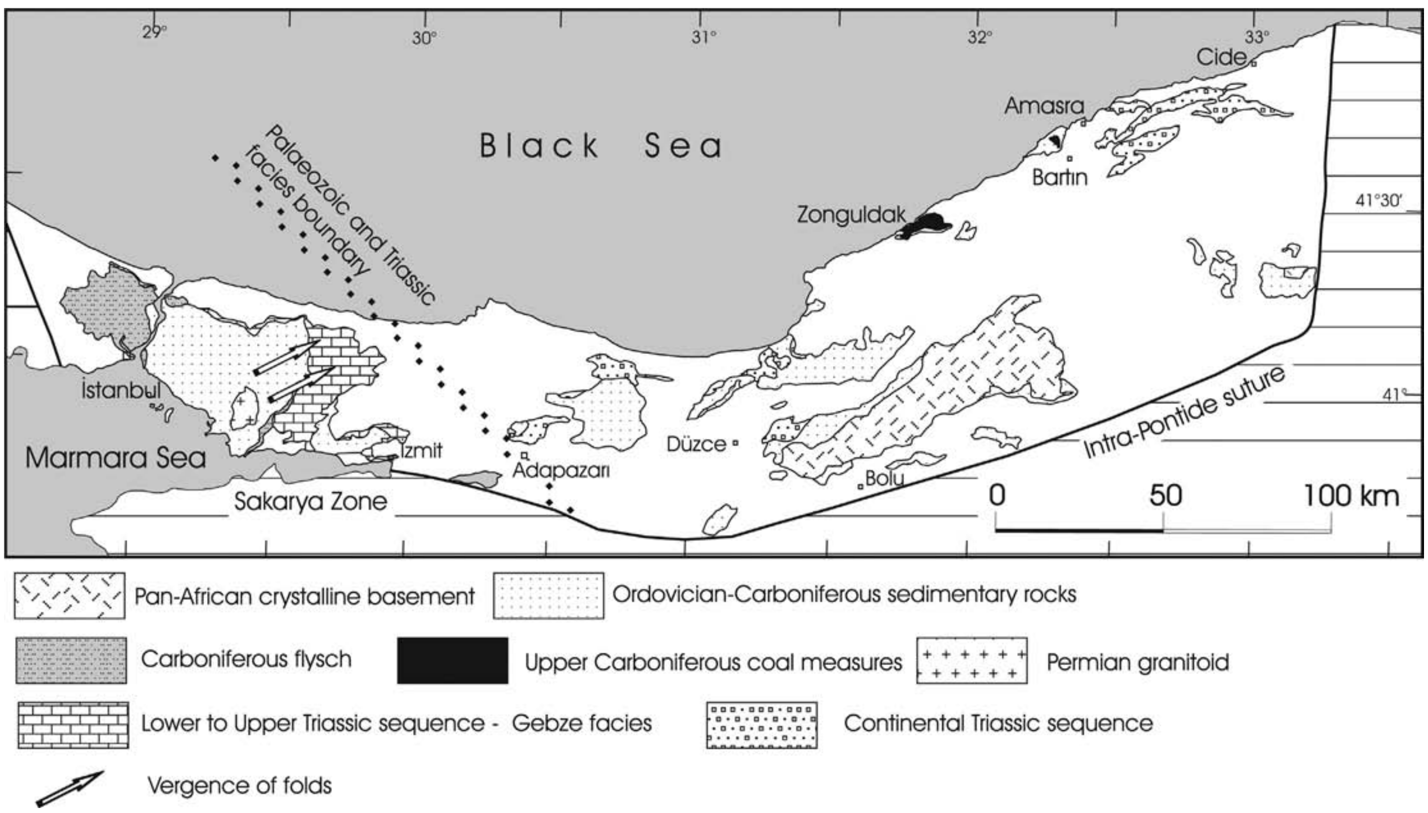

Fig. 5. The distribution of pre-Jurassic rocks in the Istanbul Zone (simplified from Aksay *et al.* 2002; Türkecan & Yurtsever 2002). Noteworthy features are the different Carboniferous and Triassic facies in the west and east, and the trend of the facies boundary, which is highly oblique to the Intra-Pontide suture.

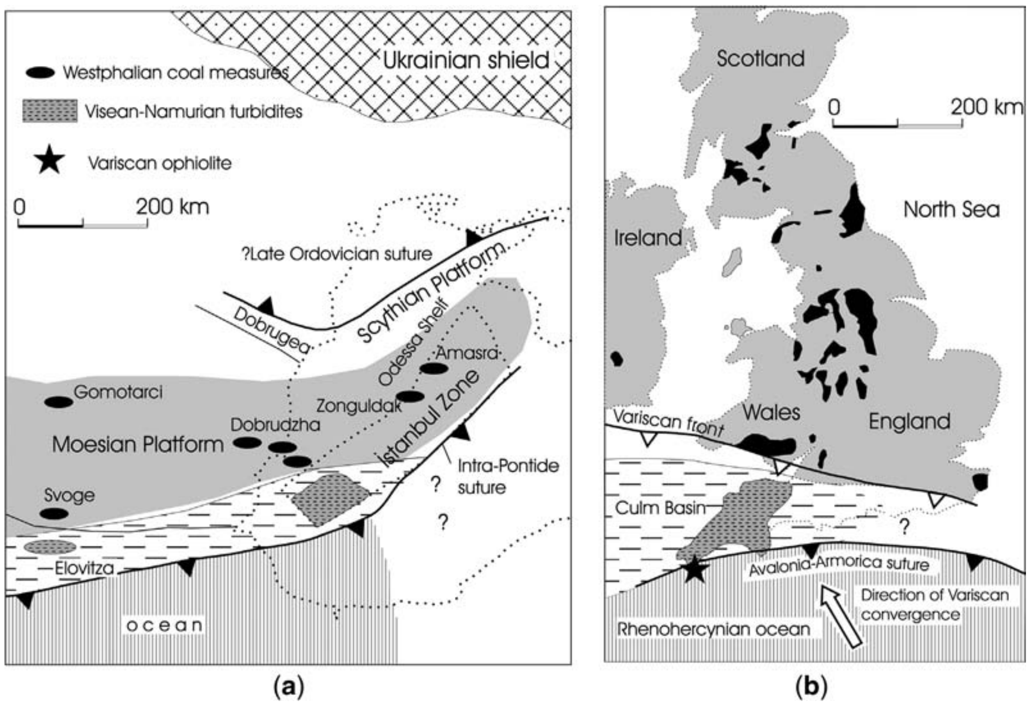

Fig. 6. Carboniferous palaeogeography in the southern margin of Laurussia (**a**) compared with that of Britain at the same period (**b**). Both maps are of the same scale. In (a) the Istanbul Zone is restored to its predrift position before the Cretaceous opening of the Western Black Sea basin (Okay *et al.* 1994). In the Tournaisian–Viséan, neritic carbonate deposition took place in Moesia and in the eastern Istanbul Zone; this was succeeded by the accumulation of coal during the Namurian and Westphalian. In the same period radiolarian chert sedimentation gave way to siliciclastic turbidite deposition in a continental slope setting in the western Istanbul Zone and the southern margin of Moesia. A similar picture exists in Britain, where, in addition, the Early Devonian Lizard ophiolite in Cornwall provides another indication of the Rheno-Hercynian ocean in the south. It should be noted that the Intra-Pontide suture truncates the facies belts. The Moesia data are from Dachev *et al.* (1988), Popova *et al.* (1992), Tenchov (1993), Haydoutov & Yanev (1997) and Tari *et al.* (1997); the data for Britain are from Guion *et al.* (2002) and Warr (2002).

The Intra-Pontide suture, which marks the southern boundary of the Istanbul Zone, truncates the Palaeozoic and Triassic facies boundary between the western and eastern parts of the Istanbul Zone (Figs 5 and 6). This suggests removal of a major section of the Istanbul Zone, possibly by post-Triassic strike-slip faulting.

Late Carboniferous deformation and plutonism

The Palaeozoic sequence in the Istanbul region ends with Viséan to Namurian siliciclastic turbidites, whereas in the east, in the Zonguldak region, it extends into the Westphalian coal measures (Fig. 2; Görür *et al.* 1997). The Palaeozoic rocks in the Istanbul region are deformed in a contractional mode, with the generation of recumbent folding, local cleavage and minor thrusting, whereas deformation is less intense in the Zonguldak region. The minor folds generally show an east to NE vergence (Seymen 1995; Zapcı *et al.* 2003), although the timing of deformation, whether Variscan or later, is difficult to constrain. Nevertheless, the observation that the lowermost Triassic red beds step down from Carboniferous to Ordovician (Türkecan & Yurtsever 2002) indicates significant deformation and erosion in the Late Carboniferous–Permian interval. The deformed Palaeozoic rocks are intruded by a Permian granite east of Istanbul, which has biotite K–Ar and whole-rock Rb–Sr ages of *c.* 255 Ma (early Late Permian; Figs 3 and 5; Yılmaz 1977). The age of the undeformed pluton constrains the Variscan deformation in the Istanbul Zone to the Late Carboniferous–Early Permian.

The Triassic restoration

The Palaeozoic rocks in the Istanbul Zone are unconformably overlain by Triassic continental clastic rocks with basaltic flows (Fig. 2). In the Istanbul region, the Triassic sequence continues with neritic to pelagic carbonates, capped by Carnian or Norian siliciclastic turbidites, showing a typical transgressive passive margin type of development (e.g. Gedik 1975; Yurttaş-Özdemir 1971), whereas in the east the Triassic is represented mainly by continental clastic rocks and lacustrine limestones (Figs 2 and 5). The change in the Triassic facies closely follows that of the Palaeozoic, suggesting a long-term hinge, possibly controlled by a deep-seated fault (Fig. 5). The termination of deposition in the Carnian or Norian in the Istanbul Zone probably reflects the Cimmeride orogeny, which is particularly strong in the Sakarya Zone farther south.

Palaeogeographical affinity

The Infra-Cambrian to Cambrian granitoids (590–560 Ma) and Neoproterozoic metamorphism in the basement of the Istanbul Zone suggest a location on the Gondwana margin in the latest Precambrian. This is supported by the Ordovician trilobite faunas, which are similar to those from Central European and Anglo-Welsh successions, and differ from those of Baltica, as well as from those of typical Gondwana realms of the Anatolide–Tauride Block and the Arabian Platform (Dean *et al.* 2000). Therefore, a location of the MOIS terrane on the western margin of Baltica during the Early Ordovician, as shown in some reconstructions (von Raumer *et al.* 2002) is not possible. The absence of the latest Ordovician (Hirnantian) glaciation in the Istanbul Zone provides another constraint on its location on the Gondwana margin. However, from the Late Silurian onwards the Istanbul Zone became part of Laurussia, as indicated by its palaeomagnetic record from sediments of Late Silurian, Devonian, Carboniferous and Triassic age (Sarıbudak *et al.*

1989; Evans *et al.* 1991), and by the Devonian–Carboniferous foraminiferal assemblages (Kalvoda 2003; Kalvoda *et al.* 2003). These data imply that the MOIS terrane separated from Gondwana during the Ordovician, and docked with Baltica in the Late Ordovician–Early Silurian; however, there is little evidence for Ordovician–Silurian collision in the geological record of the Istanbul Zone. Apparently, the zone of collision is hidden under young cover on the northern margin of the Black Sea. The Early Palaeozoic history of the MOIS terrane appears to be remarkably similar to that of Avalonia (Stampfli *et al.* 2002; Winchester & the PACE TMR Network Team 2002).

The Strandja Massif

The Strandja Massif forms part of large metamorphic region in the Balkans, which includes the Rhodope, Serbo-Macedonian and Peri-Rhodope zones (Fig. 1). The relationship between these metamorphic units, and their ages of regional metamorphism are poorly known. The Strandja Massif crops out both in Turkey and in Bulgaria, and is bordered in the west by the Rhodope Massif. It consists of a metamorphic basement of unknown age, intruded by Permian granitoids, and overlain by continental to shallow marine sedimentary rocks of Triassic to Mid-Jurassic age (Fig. 2). During the Late Jurassic, the cover and the basement of the Strandja Massif underwent contractional deformation and regional metamorphism, and Triassic allochthons were emplaced on the Mid-Jurassic metasediments. Cenomanian and younger sediments lie unconformably over the metamorphic rocks (Chatalov 1988; Okay *et al.* 2001).

The basement of the Strandja Massif consists of gneisses and micaschists intruded by voluminous plutonic rocks, several of which have been dated as Early Permian (*c.* 271 Ma) using stepwise Pb evaporation method on single zircon grains (Okay *et al.* 2001). The overlying Triassic sequence of the Strandja Massif shows affinities to the Central European Germanic Triassic facies, with a basal continental clastic series overlain by Middle Triassic shallow-marine carbonates (Chatalov 1988, 1991). A hiatus between the Late Triassic and Early Jurassic is probably a distant echo of the Cimmeride deformations farther south (Fig. 2). The shallow marine sedimentation continued into the Mid-Jurassic (Bathonian), and was terminated by the Late Jurassic Balkan orogeny.

Late Jurassic deformation and metamorphism in the Strandja Zone

The Triassic to Jurassic sedimentary cover sequence of the Strandja Massif, together with its crystalline basement, underwent deformation and greenschist-facies metamorphism during the Late Jurassic. The age of regional metamorphism is constrained to the Late Jurassic–Early Cretaceous (Callovian–Albian) by the Bathonian age of the youngest metamorphosed strata (Chatalov 1988), and by the Cenomanian post-metamorphic cover (Fig. 2). Rb–Sr and K–Ar biotite ages from the deformed and metamorphosed Permian granites of the Strandja Massif fall in the range of 155–149 Ma (Aydın 1988; Okay *et al.* 2001), indicating a Late Jurassic age for the regional metamorphism.

The Late Jurassic metamorphism in the Strandja Massif was associated with north-vergent thrusting, folding, and the generation of foliation and lineation (Okay *et al.* 2001). Permian granitoids were penetratively deformed and thrust north over the Triassic to Jurassic mylonitic metasediments and marbles. Large allochthons, composed of Triassic deep-sea metasediments and metavolcanic rocks, were thrust northwards over the epicontinental Triassic–Jurassic rocks of the Strandja Massif (Chatalov 1985, 1988; Dabovski & Savov 1988). A foreland basin, called the Nij–Trojan trough, developed in the Oxfordian between the Strandja–Rhodope massifs and the Moesian Platform. The Nij–Trojan trough migrated northward and persisted until the Early Cretaceous (Barremian; Tchoumatchenko *et al.* 1990; Harbury & Cohen 1997).

The Sakarya Zone

The Sakarya Zone forms a continental sliver, over 1500 km long, south of the Istanbul Zone and the eastern Black Sea (Fig. 1). It consists mainly of Jurassic and younger sedimentary and volcanic rocks, which unconformably overlie a heterogeneous basement. The only sign of the Late Jurassic–Early Cretaceous deformation and metamorphism that is so intense in the Strandja Massif is a parallel unconformity at the base of Callovian–Oxfordian limestones (Fig. 2; Altıner *et al.* 1991).

The pre-Jurassic basement of the Sakarya Zone includes Devonian plutonic rocks, Carboniferous plutonic and metamorphic rocks, and Triassic accretionary complexes with blueschists and eclogites (Figs 3 and 4). The pre-Jurassic relation between these basement units is strongly overprinted by Alpide deformations. The Devonian and Carboniferous units, and the Triassic accretionary complexes, are described below.

Early Devonian plutonism in the Sakarya Zone

The Devonian was a period of widespread granitoid plutonism in the Caledonides in NW Europe (e.g. Woodcock & Strachan 2002), whereas granitoids of this age were unknown in the Eastern Mediterranean region. Therefore, it was a surprise when a single sample from a granitoid in NW Turkey was dated as Early Devonian (Okay *et al.* 1996). The Çamlık granodiorite in the Biga peninsula (Fig. 3) forms a 20 km long and 3–4 km thick thrust sheet in an Alpide thrust stack. It is a leucocratic granodiorite consisting mainly of quartz, plagioclase and chloritized biotite, and is unconformably overlain by Upper Triassic arkosic sandstones. As the age of the granite is tectonically significant, zircons from a second sample from the Çamlık granodiorite were dated using the stepwise Pb-evaporation method. The details of the dating method have been given by Okay *et al.* (1996). Two zircon grains from the Çamlık granodiorite gave an Early Devonian age of 397.5 ± 1.4 Ma (Fig. 7, Table 1), confirming the earlier less precise age of 399 ± 13 Ma obtained by Okay *et al.* (1996). The relationship between the Çamlik Granodiorite and the other pre-Jurassic basement units of the Sakarya Zone are not known. However, the proximity of the high-grade Carboniferous metamorphic rocks of the Kazdağ and the essentially unmetamorphosed Devonian Çamlık granodiorite in NW Turkey (see Fig. 3) suggest major pre-Jurassic shortening between these two units.

Carboniferous deformation and metamorphism in the Sakarya Zone

The high-grade Variscan metamorphic basement of the Sakarya Zone is exposed in only a few areas throughout its 1500 km length. These include the Kazdağ and Uludağ massifs in the west, the Devrekani Massif in the Central Pontides, and the Pulur Massif in the Eastern Pontides (Figs 3 and 4). These metamorphic regions are composed of gneiss, amphibolite and marble metamorphosed at amphibolite- to granulite-facies conditions, and in the Kazdağ and Pulur massifs there are also meta-ultramafic rocks within the sequence (Okay 1996; Okay *et al.* 1996; Duru *et al.* 2004; Topuz *et al.* 2004*a*). Isotopic age data exist only for the Pulur and Kazdağ massifs. Monazite Pb ages from a Pulur gneiss are late Early Carboniferous

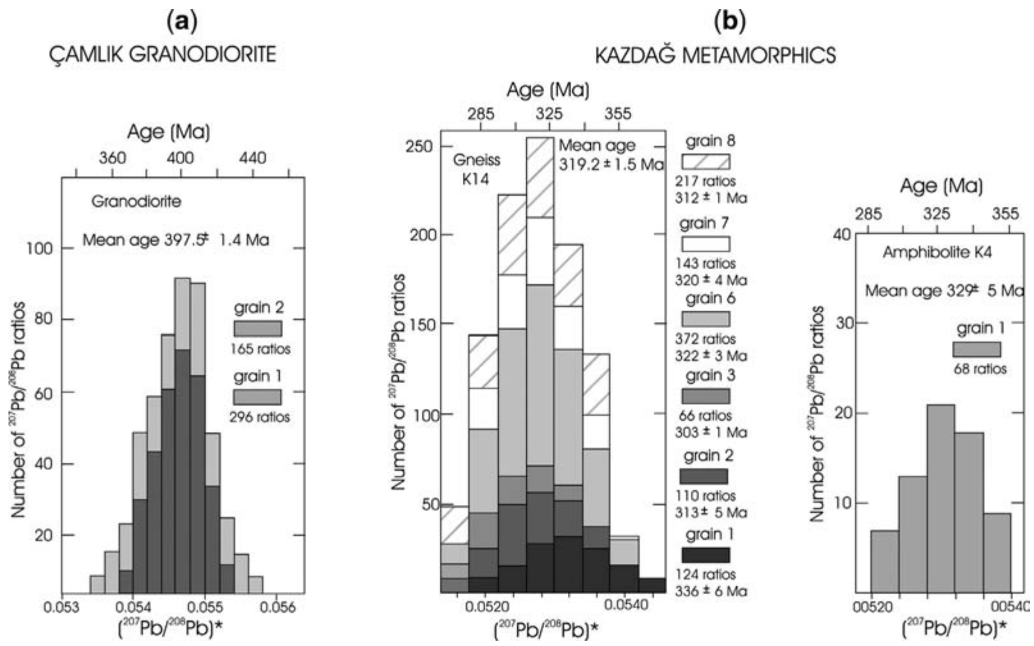

Fig. 7. Histograms showing the distribution of radiogenic Pb isotope ratios derived from the evaporation of two zircon grains from the Çamlık granodiorite (**a**) and from a gneiss and an amphibolite of the Kazdağ Group (**b**) in the Sakarya Zone, NW Turkey.

(331–327 Ma, Namurian), considered as the age of high-grade metamorphism (Topuz *et al.* 2004*a*). The 315–310 Ma (Westphalian) Nd–Sm, Rb–Sr and Ar–Ar ages from the Pulur gneisses are regarded as cooling ages. Zircons from two gneiss samples from the Kazdağ Massif, dated by the stepwise Pb-evaporation method, gave an age of 308 ± 16 Ma (Okay *et al.* 1996). To further refine the age of high-grade metamorphism in the Kazdağ Massif, we have dated a gneiss and an amphibolite from the Kazdağ Massif using the same method (Okay *et al.* 1996). Six zircon grains from the gneiss sample produced a relatively precise age of 319.2 ± 1.5 Ma (early Late Carboniferous, latest Namurian), and one zircon grain from the amphibolite gave an age of 329 ± 5 Ma (Fig. 7, Table 1). The isotopic data indicate high-grade metamorphism and associated deformation in the mid-Carboniferous (Namurian) in the Sakarya Zone.

Permo-Carboniferous plutonism in the Sakarya Zone

Pre-Jurassic granitoids are common in the Sakarya Zone, although few are dated. The Söğüt granite in the western Sakarya Zone gave an Ar–Ar biotite plateau age of 290 ± 5 Ma (Carboniferous–Permian boundary, Okay *et al.* 2002) confirming earlier U–Pb zircon and K–Ar biotite ages (Çoğulu *et al.* 1965; Çoğulu & Krummenacher 1967). K–Ar biotite ages from the Gönen and Karacabey granites, east and west of Bandırma, respectively, are in the range 286–298 Ma (Delaloye & Bingöl 2000). These data indicate late orogenic acidic plutonism in the Sakarya Zone in the latest Carboniferous to early Permian period. The Variscan granites and high-grade metamorphic rocks in the Sakarya Zone were exhumed and unconformably overlain by the latest Carboniferous continental to shallow marine sedimentary rocks in the Eastern Pontides (Okay & Leven 1996; Okay & Şahintürk 1997; Çapkınoğlu 2003) and in the Caucasus (e.g. Khain 1975).

Accretionary complexes in Anatolia: data on the spatial and temporal aspects of the Tethyan oceans

There are widely differing views on the number, location, age span and name of the Tethyan oceans that existed during the Phanerozoic (e.g. Şengör & Yılmaz 1981; Robertson & Dixon 1984; Şengör *et al.* 1984; Dercourt *et al.* 1986; Ricou 1994; Robertson *et al.* 1996; Stampfli *et al.* 2001, 2002). One way to approach this problem is through a biostratigraphic and isotopic study of the accretionary complexes. Because of their relatively low density the accretionary complexes have a wide preservation potential, and crop out widely in orogenic belts. The accretionary complexes may comprise three types of constituents: (1) pelagic sedimentary and basic magmatic rocks scraped at subduction

Table 1. *Isotopic data from single-grain $^{207}Pb/^{206}Pb$ evaporation analyses of zircons from the basement of the Sakarya Zone, NW Turkey*

Lithology and sample number	Grain	Number of scans	$^{204}Pb/^{206}Pb$	$^{208}Pb/^{206}Pb$	Mean value of $^{207}Pb/^{206}Pb$ ratios	$^{207}Pb/^{206}Pb$ age (Ma)
Kazdağ gneiss K14	1	124	0.00210	8.2	0.053178 ± 102	336.4 ± 4.4
	2	110	0.00115	8.4	0.052621 ± 104	312.5 ± 4.5
	3	66	0.000255	10.6	0.052407 ± 130	303.2 ± 5.7
	4	372	0.000179	5.8	0.052832 ± 59	321.6 ± 2.5
	5	143	0.000115	8.2	0.052792 ± 90	319.9 ± 3.9
	6	217	0.000095	8.5	0.052609 ± 88	312.0 ± 3.8
	mean					319.2 ± 1.5
Kazdağ amphibolite K4	1	68	0.000071	7.7	0.053009 ± 112	329.2 ± 4.8
Çamlık metagranite CL1	1	296	0.000074	12.0	0.054640 ± 37	397.6 ± 1.5
	2	165	0.000336	12.1	0.054631 ± 87	397.2 ± 3.6
	mean					397.5 ± 1.4

Errors are given at 95% confidence level and refer to the last digits.

zones from the downgoing oceanic crust; (2) greywacke and shale, which represent the trench infill; (3) blueschists and eclogites brought up along the subduction channel. The pelagic sedimentary rocks in the accretionary complexes provide an age range for the subducted ocean, whereas the greywackes and the isotopic age of the blueschists give an indication of the duration of subduction. The structural position of the accretionary complexes provides clues to the location of the associated oceans. Accretion ends by collision, or when the subduction zone is clogged by large oceanic or continental edifices, such as oceanic islands, oceanic plateaux, or isolated continental slivers (e.g. Cloos 1993). Termination of subduction in the Eastern Mediterranean south of Cyprus by the collision of the Eratosthenes Seamount provides a present-day example (e.g. Robertson 1998). The age of the accretionary complex can be defined as the age of subduction. At least four distinct accretionary complexes can be defined in the Balkan–Anatolian region.

Karaburun–Chios accretionary complex (Carboniferous)

Carboniferous accretionary complexes are found on the Karaburun peninsula and the adjacent island of Chios (Fig. 3; Stampfli *et al.* 1991, 2003). Both areas are situated in the Aegean on the north-western margin of the Anatolide–Tauride Block immediately south of the Neotethyan İzmir–Ankara suture. The complexes consist of strongly deformed siliciclastic turbidites, regarded as a Franciscan-type trench infill, which are unconformably overlain by Lower Triassic basinal sedimentary and volcanic rocks (Robertson & Pickett 2000; Zanchi *et al.* 2003). The Lower Triassic pelagic sediments pass up into a typical Tauride carbonate platform of Triassic to Early Cretaceous age (Erdoğan *et al.* 1990). The intensely deformed and tectonically sliced and repeated turbidites comprise exotic limestone, radiolarian chert and volcanic blocks, up to kilometre scale, and Silurian–Carboniferous in age (Fig. 8; Kozur 1995). The age of the turbidite matrix is probably Early Carboniferous (Groves *et al.* 2003; Zanchi *et al.* 2003).

Karakaya–Küre accretionary complex (Triassic–Early Jurassic)

Triassic–Early Jurassic accretionary complexes are widely exposed in the Sakarya Zone below the Jurassic unconformity (Figs 3 and 4; Tekeli 1981; Tüysüz 1990; Ustaömer & Robertson 1993, 1994; Pickett & Robertson 1996; Yılmaz *et al.* 1997; Okay 2000; Okay & Göncüoğlu 2004). In the western part of the Sakarya Zone they are attributed to the Karakaya Complex, and in the central Pontides to the Küre Complex. Some Triassic palaeogeographical reconstructions show the Karakaya and Küre accretionary complexes as belonging to different oceans separated by a continental sliver, attributed to the western part of the Istanbul Zone and to the northern parts of the Sakarya Zone (e.g. Ustaömer & Robertson 1993; Stampfli *et al.* 2001; Ziegler & Stampfli 2001). However, no coherent continental fragment can be defined

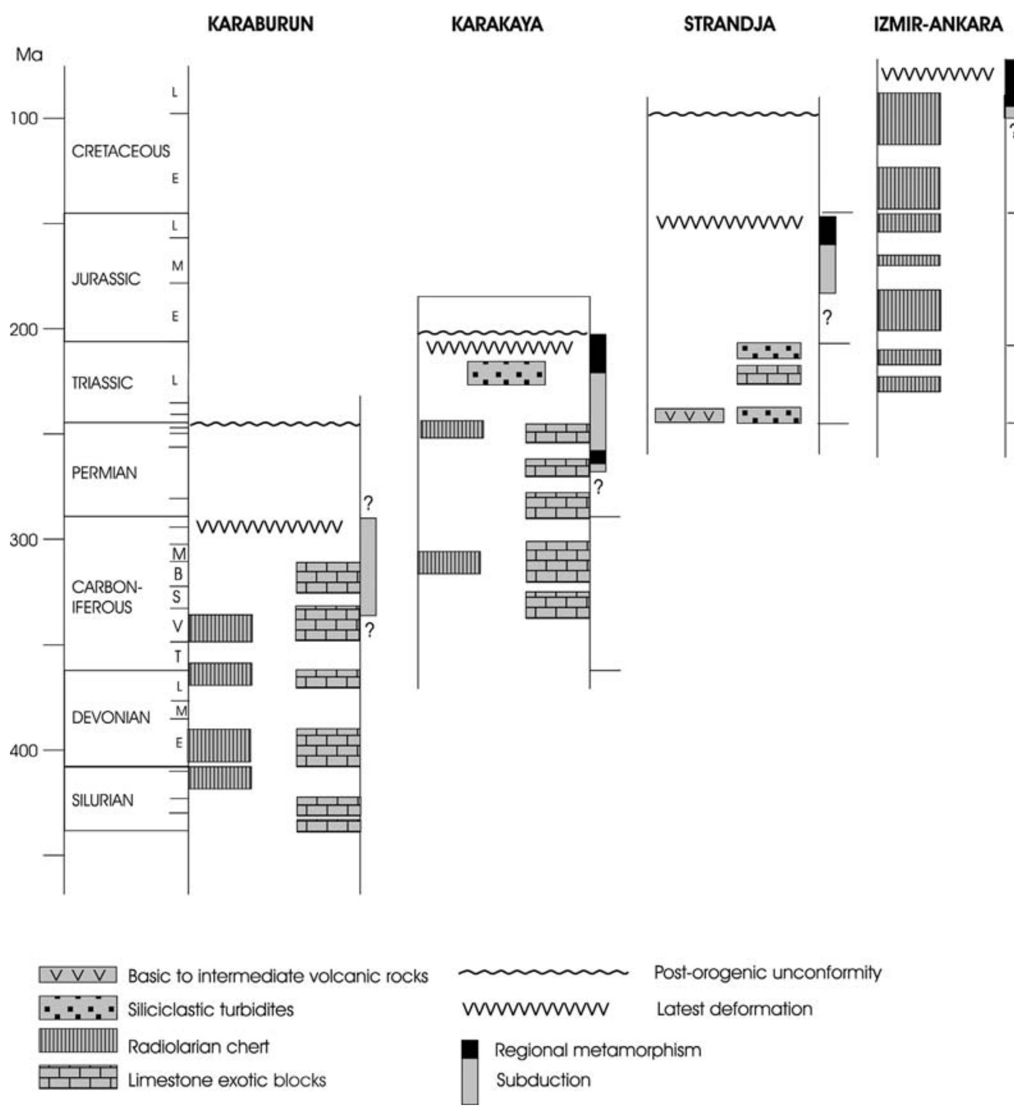

Fig. 8. A chronostratigraphic chart showing biostratigraphic and isotopic data from the Anatolian accretionary complexes. Data for the Karaburun complex are from Kozur (1995) and Groves *et al.* (2003); for the Karakaya–Küre Complex from Kozur & Kaya (1994), Okay & Mostler (1994), Kozur (1997), Okay & Monié (1997) and Okay *et al.* (2002); for the İzmir–Ankara accretionary complexes from Bragin & Tekin (1996), Sherlock *et al.* (1999) and Tekin *et al.* (2002).

between the outcrops of the Küre and Karakaya complexes (Fig. 3). Furthermore, no Phanerozoic accretionary complex exists in the Istanbul Zone (Fig. 5). The Küre and Karakaya complexes are similar in lithology, tectonostratigraphy and in structural position, but slightly differ in age, and will be treated together. The youngest palaeontological ages from the Karakaya Complex are latest Triassic (Leven & Okay 1996; Okay & Altıner 2004), whereas the age of the Küre Complex extends to Early Jurassic (Kozur *et al.* 2000), and the complex is cut by granitoids of Mid-Jurassic age (Boztuğ *et al.* 1984; Yılmaz & Boztuğ 1986). Before the Cretaceous opening of the Black Sea, the Küre Complex was contiguous with the Taurian Flysch of the Crimean Peninsula.

The Karakaya–Küre Complex consists of a lower metamorphic unit made up of a strongly deformed thrust stack of metabasite–phyllite–marble with tectonic slices of ultramafic rock, broadly referred to as the Nilüfer Unit. The depositional age of the Nilüfer Unit, based on scarce conodonts in the marbles in NW Turkey, is Early to Mid-Triassic (Kaya & Möstler 1992; Kozur *et al.* 2000). The geochemistry of the metabasites in the Nilüfer Unit suggests a within-plate tectonic setting (Genç & Yılmaz 1995; Pickett & Robertson 1996, 2004; Genç 2004). The Nilüfer Unit generally shows a high-pressure greenschist-facies metamorphism, although, in several localities in the Sakarya Zone, it also includes tectonic slices of blueschist and eclogite. The HP–LT metamorphic rocks in the Nilüfer Unit are dated in the Bandırma and Eskişehir regions of NW Turkey (Fig. 3) as latest Triassic (205–203 Ma) using Ar–Ar method on phengites (Okay & Monié 1997; Okay *et al.* 2002). The structural setting and the lithological, metamorphic and geochemical features of the Nilüfer Unit suggest an origin as an oceanic plateau or oceanic island, which was accreted to a Late Triassic active margin (Pickett & Robertson 1996, 2004; Okay 2000; Genç 2004).

Recently, Topuz *et al.* (2004*b*) reported Early Permian (263–260 Ma) Rb–Sr and Ar–Ar hornblende and muscovite ages from a metabasite–phyllite sequence from the Pulur region in the Eastern Pontides (Fig. 4). The metabasite–phyllite sequence, which is correlated with the Nilüfer Unit, is tectonically overlain by the granulite-facies gneisses of mid-Carboniferous age (Okay 1996; Topuz *et al.* 2004*b*). If these isotopic data are confirmed then the subduction–accretion represented by the Karakaya–Küre Complex will extend back to the Early Permian (Fig. 8).

In the Sakarya Zone, the Nilüfer Unit is overlain by Triassic to Lower Jurassic siliciclastic and volcanic sequences, which were strongly deformed, probably in a subduction zone setting, in the latest Triassic–earliest Jurassic (Okay 2000). In NW Turkey, the siliciclastic rocks comprise olistostromes with numerous Carboniferous and Permian shallow marine limestone blocks (Leven 1995; Leven & Okay 1996), and smaller numbers of Middle Carboniferous (Bashkirian), Permian and Triassic radiolarian chert and pelagic limestone exoticblocks (Fig. 8; Kozur & Kaya 1994; Okay & Mostler 1994; Kozur 1997; Kozur *et al.* 2000; Göncüoğlu *et al.* 2004). Olistostromes with the Carboniferous and Permian shallow marine limestone blocks form a belt, over 150 km long and 5–10 km wide, in NW Turkey immediately NW of the İzmir–Ankara suture (Fig. 3). The origin of the Permo-Carboniferous limestone blocks is controversial; the fauna in the blocks is interpreted either as Laurussian (Leven & Okay 1996) or as Gondwanan in origin (Altiner *et al.* 2000).

Strandja accretionary complex (Late Triassic–Early Jurassic)

The Triassic–Middle Jurassic epicontinental sediments of the Strandja Massif are tectonically overlain by a highly deformed volcano-sedimentary complex of siliciclastic turbidites, carbonates, mafic and acidic volcanic rocks of Early to Late Triassic age (Chatalov 1980, 1988). Şengör *et al.* (1984) and Ustaömer & Robertson (1993) interpreted the Strandja allochthons as an accretionary complex, although definite evidence for the oceanic origin of the Strandja allochthons (e.g. ultramafic rocks or deep-sea radiolarian cherts) is missing. Dismembered ophiolites of Late Jurassic age, including peridotite, gabbro and basalt, occur on the eastern margin of the Rhodope Massif (Fig. 3; Tsikouras & Hatzipanagiotou 1998), and are associated with a slightly metamorphosed epicontinental sequence of Triassic to Early Jurassic age (Kopp 1969). The ophiolites of this Circum-Rhodope Zone may represent the root zone of the Strandja allochthons. Although the Karakaya–Küre and the Strandja accretionary complexes are similar in age, they differ markedly in their structural setting and lithology, and, as discussed below, are ascribed to different oceans.

İzmir–Ankara accretionary complex (Late Cretaceous)

Late Cretaceous accretionary complexes cover large regions in the Anatolide–Tauride Block south of the İzmir–Ankara–Erzincan suture (Figs 3 and 4; Okay 2000). They generally form tectonic imbricates sandwiched between the ophiolites above and the Anatolide–Tauride carbonate platform below. In many regions near the İzmir–Ankara suture, such as north of Eskişehir (Okay *et al.* 2002), east of Ankara (Koçyiğit 1991) and in the Tokat Massif (Bozkurt *et al.* 1997) the İzmir–Ankara accretionary complexes are imbricated with those of the Karakaya–Küre Complex (Figs 3 and 4).

The Late Cretaceous accretionary complexes consist mainly of basalt, radiolarian chert, pelagic shale and pelagic limestone, and in the old literature were often referred to as ophiolitic mélange or coloured mélange. In the north near the İzmir–Ankara suture, the accretionary complexes have undergone low-grade blueschist-facies metamorphism dated at *c.* 80 Ma (Sherlock *et al.* 1999). Palaeontological study of radiolarian cherts and pelagic limestones in these accretionary complexes has shown the presence of Triassic, Jurassic and Cretaceous rocks (Fig. 8; Bragin & Tekin 1996; Tekin *et al.* 2002). In contrast, no Palaeozoic pelagic sedimentary rocks were described in the İzmir–Ankara accretionary complexes, the oldest ones being Late Triassic (Late Carnian) radiolarian cherts (Tekin *et al.* 2002).

Age and location of the Tethyan oceans

Biostratigraphic and isotopic data from the accretionary complexes in Anatolia indicate the presence of several Tethyan oceans. A Tethyan ocean north of the Anatolide–Tauride Block, called the İzmir–Ankara ocean, had a minimum age span from Mid–Late Triassic to the Cretaceous, suggesting an opening as early as the Early Triassic. This is in accord with the Triassic stratigraphy from the Karaburun Peninsula, which is indicative of rifting in the Early Triassic (Robertson & Pickett 2000). The widespread unconformity at the base of the Lower Triassic rocks in the northern margin of the Anatolide Tauride Block (e.g. Eren 2001; Göncüoğlu *et al.* 2003) is probably also related to shoulder uplift before the rifting.

The outcrop pattern of the Karakaya–Küre Complex indicates an ocean in a similar position to the İzmir–Ankara ocean (e.g. north of the Anatolide–Tauride Block and south of the Istanbul Zone) but older (at least Mid-Carboniferous to Early Jurassic). The Karakaya–Küre ocean must also have been located south of the Variscan basement of the Sakarya Zone, as the exotic blocks in the Karakaya–Küre Complex were most probably derived from the south rather than the north (Okay 2000). The absence of continental fragments between the Karakaya–Küre and İzmir–Ankara accretionary complexes (Figs 3 and 4) implies that the Karakaya–Küre ocean corresponds to the main Palaeotethys (Fig. 9), rather than to a small back-arc basin as

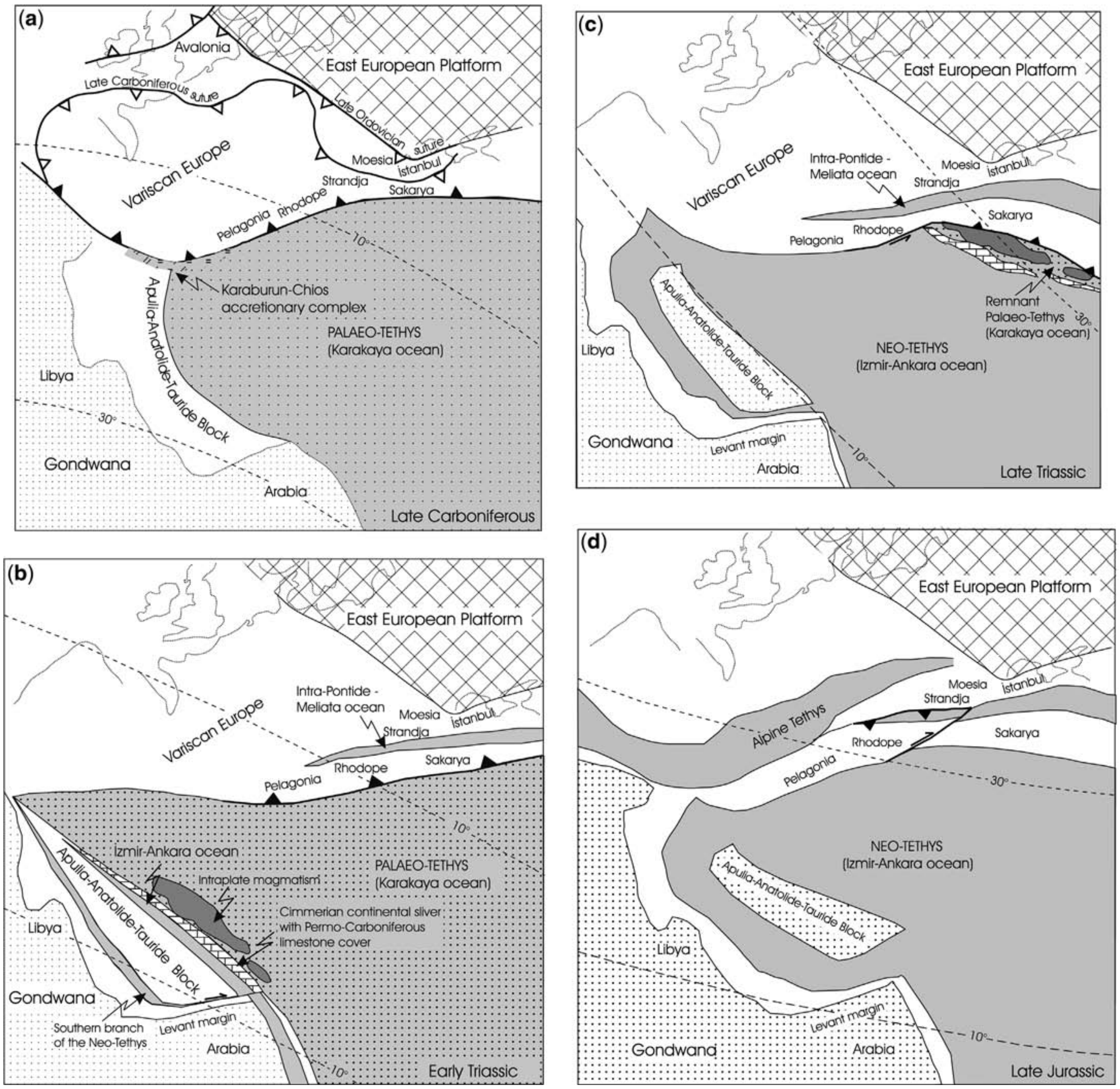

Fig. 9. Palaeogeographical reconstructions of the Tethyan realm for the Late Carboniferous (**a**), Early Triassic (**b**), Late Triassic (**c**) and Late Jurassic (**d**), showing the possible locations of the terranes and oceans discussed in the text. The general palaeogeographical framework is taken from Stampfli *et al.* (2001).

shown in most models (e.g. Şengör *et al.* 1984; Stampfli *et al.* 2001).

Biostratigraphic data from the Karaburun–Chios accretionary complexes suggest an ocean of Silurian to Carboniferous age, again situated north of the Anatolide–Tauride Block. The Karaburun–Chios accretionary complex may be related to the Variscan subduction (Zanchi *et al.* 2003), in which case it must have been displaced eastwards from its original position by strike-slip faulting (Fig. 9a).

The Strandja allochthons indicate the presence of a Triassic to Early Jurassic ocean between the Strandja Massif in the north and the Rhodope–Serbo-Macedonian massifs in the south (Fig. 3). The Triassic stratigraphy of the Istanbul Zone indicates rifting in the Early Triassic. This ocean probably formed an eastern extension of the Hallstatt–Meliata ocean, described farther west in the Eastern Alps and the Balkans (Kozur 1991; Channel & Kozur 1997). The relation between these Tethyan oceans and the surrounding continental terranes, which gave rise to the orogenic events, is discussed below.

Variscan orogeny in the Balkans and the Black Sea region

The Late Carboniferous orogeny in the Pontides forms a link between the Variscan orogen in Central Europe and the Uralides of Eastern Europe. The Variscan orogeny comprises Carboniferous to Early Permian deformation, metamorphism and magmatism linked to the collision and amalgamation of Gondwana, Laurussia and the intervening terranes (e.g. Matte 2001; Warr 2002). The eastward extension of the Variscan orogen towards the Balkans and Anatolia is obscured by the strong overprint of the Alpide orogeny, or is concealed by the younger cover.

The East European Craton is bordered in the south by a narrow tectonic belt, called the Scythian Platform, which is generally considered as a Late Palaeozoic (Early Carboniferous) orogen (e.g. Nikishin *et al.* 1998, 2001). The Palaeozoic stratigraphy in the Scythian Platform is concealed beneath the Mesozoic and younger strata, and there are only patchy data from a few boreholes, which indicate a thick Lower Devonian continental sandstone succession overlain by Middle Devonian to Lower Carboniferous shallow marine limestones. The overlying Viséan–Namurian sequence consists of paralic and limnic deposits, and the Permian of red clastic rocks (Vaida & Seghedi 1997). During the Devonian and Carboniferous the Istanbul Zone and the Moesian Platform were adjacent to the Scythian Platform, and formed a south-facing passive continental margin (Figs 6 and 9a). The western part of the Istanbul Zone was the site of deep marine sedimentation in the Devonian and Carboniferous, in a continental slope setting, and hence was closer to the ocean compared with its eastern part and the Scythian–Moesian platforms (Fig. 6).

The Late Carboniferous deformation in the Istanbul Zone is coeval with the high-grade metamorphism in the Sakarya Zone. It is plausible to relate the deformation and regional metamorphism to collision of the Laurussia margin with an ensialic arc represented by the basement rocks of the Sakarya Zone and possibly of the Strandja Zone. Absence of Palaeozoic magmatism in the MOIS terrane suggests southward subduction, which is compatible with the general north to NE vergence of Carboniferous deformation in the Istanbul Zone (Görür *et al.* 1997) and the northward migration of the coal deposition in the Moesian Platform (Tari *et al.* 1997). The ocean between the Laurussia margin and the Sakarya–Strandja microplate probably started to close by the Early Devonian, producing a magmatic arc represented by the Çamlık Granite in the Sakarya Zone. The Late Carboniferous collision was followed by the latest Carboniferous–Early Permian plutonism in the core of the orogen in the Strandja and Sakarya zones, possibly linked to crustal thickening. The latest Carboniferous–Early Permian molasse deposition in the Eastern Pontides and the Caucasus marks the end of the Variscan orogeny in northern Turkey. The Intra-Pontide suture between the Istanbul and Sakarya zones probably links up with the Late Carboniferous Rheic suture in Central Europe (Fig. 9a; Ziegler & Stampfli 2001). The Variscan evolution of northern Turkey and the Balkans appears to be similar to that of NW Europe, with the Istanbul–Moesia–Scythian Block corresponding to Avalonia, and the Sakarya–Strandja zones to Armorica (Stampfli *et al.* 2002; Winchester & The PACE TMR Network Team 2002).

Early Triassic rifting and magmatism

The Early Triassic is characterized by widespread rifting and mafic magmatism in the Eastern Mediterranean region, possibly associated with mantle plumes (e.g. Dixon & Robertson 1999). The Istanbul Zone started to rift from the Sakarya Zone along the former Carboniferous suture, as shown by the deposition of earliest Triassic continental sandstones and conglomerates intercalated with basaltic flows (Fig. 9b). In the Mid-Triassic the Istanbul Zone became separated from the Sakarya Zone, as the rift turned into the Intra-Pontide–Meliata ocean. On the Gondwana side in the south, mafic magmatism was associated with the break-up of Permo-Carboniferous carbonate platforms, and the separation of the Anatolide–Tauride Block from Gondwana (Fig. 8b). Possibly a thin carbonate sliver, corresponding to the Cimmerian continent of Şengör *et al.* (1984), rifted away from the Anatolide–Tauride Block in the Early Triassic. Associated with this rifting, major intra-plate mafic magmatism occurred and an abnormally thick oceanic crust or oceanic plateau was created adjacent to the passive continental margin. The northward drift of this narrow continental sliver is shown to close the Palaeozoic Tethys and open up the Mesozoic Tethys in the Triassic (Fig. 9b and c), although as discussed below there is no unequivocal evidence for the Cimmerian continent in the Pontides.

Cimmeride orogeny in the Pontides

In Turkey deformation and metamorphism of latest Triassic to earliest Jurassic age is particularly marked in the Sakarya Zone. It is associated with the emplacement of large oceanic allochthons over the Variscan basement. In contrast, the Cimmeride deformation is weak, and the Cimmeride metamorphism is absent in the other Pontic zones, where this period is generally marked as an unconformity (Fig. 2).

The cause of the Cimmeride orogeny in Anatolia was generally thought to be the collision and amalgamation of a Cimmerian continent with the Laurasian margin (e.g. Şengör 1984; Şengör *et al.* 1984). However, it has not been possible to define a Cimmerian continent in the field, which would have been readily recognized by its Gondwana-type stratigraphy, free of Variscan deformation and metamorphism. In many regions along the İzmir–Ankara suture the accretionary complexes of the İzmir–Ankara and Karakaya–Küre oceans are tectonically intercalated with no evidence of an intervening continental fragment (Figs 3 and 4; Bozkurt *et al.* 1997; Okay *et al.* 2002). Apparently, the narrow Cimmerian continental sliver, responsible for the opening of the İzmir–Ankara ocean, was completely subducted, with only its Permo-Carboniferous limestone cover providing blocks to the accretionary complex. The Cimmeride orogeny in the Pontides was largely accretionary, caused by the collision and partial accretion of an oceanic plateau to the Laurasian margin during the latest Triassic (Fig. 9c; Okay 2000). This is compatible with the short duration of deformation and regional metamorphism observed in the Karakaya–Küre Complex.

Late Jurassic Balkan orogeny

Apart from the Strandja Massif, Late Jurassic deformation is strangely absent, or is marked by only a slight disconformity in the Pontic zones. The Late Jurassic was a period of opening of the Alpine Tethys in the west, where contractional deformation is also not reported. This leaves a relatively small space for the Balkan orogeny on the southern margin of Laurasia (Fig. 9d). The Balkan orogeny is possibly linked to the subduction of the Intra-Pontide–Meliata ocean between the Strandja and the Rhodope–Serbo-Macedonian massifs, and the ensuing collision (Fig. 9d). The north-vergent deformation in the Strandja Massif indicates a southward subduction under the Rhodope Massif, with the implication that the latest Jurassic–Early Cretaceous granitoids in the Serbo-Macedonian Massif were generated in a magmatic arc. The eastern part of the Intra-Pontide–Meliata ocean between the Sakarya and Istanbul zones did not close until the mid-Cretaceous, suggesting the existence of a transform fault between Pelagonia and the Sakarya Zone (Fig. 9d).

Conclusions

The Pre-Alpide geological history of the Eastern Mediterranean–Balkan region can be viewed as the episodic growth of Laurussia by the accretion of oceanic and continental terranes, interrupted by the opening of narrow back-arc basins on the southern margin of Laurussia. The continental terranes were invariably derived from Gondwana, and were accreted to Laurussia during the Late Ordovician–Early Silurian and Late Carboniferous, whereas a major accretion of oceanic crustal material occurred during the Late Triassic–Early Jurassic.

Orogenic deformation associated with the Late Ordovician–Early Silurian accretion of the Istanbul–Moesia–Scythian Platform (the MOIS terrane) is buried under young cover in the northern margins of the Black Sea. The Carboniferous accretion of the Strandja–Sakarya terrane to the Laurussian margin, along a south-dipping subduction zone, resulted in strong deformation, mid-Carboniferous metamorphism, and latest Carboniferous–Early Permian post-orogenic plutonism. The ensuing suture is probably an extension of the Rheic suture in Central Europe (Ziegler & Stampfli 2001). In contrast to these Palaeozoic continental collisions, a major accretion of oceanic crustal rocks occurred during the Late Triassic–Early Jurassic. The accretionary complexes in the Pontides comprise voluminous metabasic rocks with latest Triassic blueschist and eclogite ages. The Late Jurassic deformation and metamorphism, observed only in the Balkans, were the result of the closure of a narrow back-arc basin, the Meliata ocean between the Rhodope–Serbo-Macedonian and the Strandja massifs. In contrast to the poly-orogenic history of the Pontides and the Balkans, the Anatolide–Tauride Block south of the İzmir–Ankara suture was largely free of Palaeozoic–early Mesozoic deformations, except along its northwestern margin, where a Carboniferous accretionary complex has been recognized (Stampfli *et al.* 1991).

Biostratigraphic and isotopic data from the Anatolian accretionary complexes and their structural position indicate the presence of three oceanic realms north of the Anatolide–Tauride Block during the Phanerozoic. Two of them correspond to the mid-Palaeozoic–Early Jurassic Palaeotethys, and Early Triassic–Tertiary Neotethys, respectively, both of which were subducted along the İzmir–Ankara suture, which represents the main boundary between Laurussia and Gondwana. The third ocean, the Meliata–Intra-Pontide ocean, opened as a marginal back-arc basin on the Laurussian margin (e.g. Stampfli 2000).

A major point from this review and that is also implicit in some recent studies (e.g. Dean *et al.* 2000) is the mobility of the small plates that make up Anatolia. The assumption of conjugate margins, common in the old Tethyan reconstructions (e.g. Şengör & Yılmaz 1981; Robertson & Dixon 1984) is clearly not correct. Prior to the Tertiary, the Pontides and the Anatolide–Tauride Block never formed a single contiguous terrane, which is implicit in recent palaeogeographical reconstructions (e.g. Stampfli *et al.* 2001). Translation of continental terranes oblique to the rifted margin, and margin-parallel strike-slip faulting led to the juxtaposition of unrelated continental fragments. For example, in the Early Ordovician both the Anatolide–Tauride Block and the Istanbul Zone were probably located on the northern margin of Gondwana but separated by several thousand kilometres (see Dean *et al.* 2000).

The sutures separating the terranes in the Eastern Mediterranean–Balkan region were major zones of weaknesses, and were rejuvenated at various times. For example, the Intra-Pontide suture started as a Late Carboniferous suture, and later became the site of an Early Triassic rift, which developed into the Meliata–Intra-Pontide ocean. This ocean closed in the Late Jurassic–Early Cretaceous, generating a second suture. In the Miocene the suture was reused by the North Anatolian Fault, which at present defines the northern margin of the Anatolia microplate.

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