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Obduction, subduction and collision as reflected in the Upper Cretaceous–Lower Eocene sedimentary record of western Turkey

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Abstract – Late Cretaceous–Early Eocene Tethyan evolution of western Turkey is characterized by ophiolite obduction, high-pressure/low-temperature metamorphism, subduction, arc magmatism and continent–continent collision. The imprints of these events in the Upper Cretaceous–Lower Eocene sedimentary record of western Anatolia are studied in thirty-eight well-described stratigraphic sections. During the Late Cretaceous period, western Turkey consisted of two continents, the Pontides in the north and the Anatolide-Taurides in the south. These continental masses were separated by the İzmir-Ankara Neo-Tethyan ocean. During the convergence the Pontides formed the upper plate, the Anatolide-Taurides the lower plate. The arc magmatism in the Pontides along the Black Sea coast is biostratigraphically tightly constrained in time between the late Turonian and latest Campanian. Ophiolite obduction over the passive margin of the Anatolide-Tauride Block started in the Santonian soon after the inception of subduction in the Turonian. As a result, large areas of the Anatolide-Tauride Block subsided and became a region of pelagic carbonate sedimentation during the Campanian. The leading margin of the Anatolide-Tauride Block was buried deeply and was deformed and metamorphosed to blueschist facies during Campanian times. The Campanian arc volcanic rocks in the Pontides are conformably overlain by shaley limestone of Maastrichtian–Palaeocene age. However, Maastrichtian sedimentary sequences north of the Tethyan suture are of fore-arc type suggesting that although arc magmatism ceased by the end of the Campanian age, continent–continent collision was delayed until Palaeocene time, when there was a change from marine to continental sedimentation in the fore-arc basins. The interval between the end of the arc magmatism and continent–continent collision may have been related to a northward jump of the subduction zone at the end of Campanian time, or to continued obduction during the Maastrichtian.

1. Introduction

The Late Cretaceous–Early Eocene was a tectonically active period for the large region between the Aegean Sea and the Indian Ocean. Ophiolite obduction, subduction and high-pressure/low-temperature metamorphism as well as the initial phases of Alpide compression in the eastern Tethyan realm took place during the Late Cretaceous, while continent–continent collisions were initiated in many places during Palaeocene or Early Eocene times. These compressive events are linked to a change in the movement of Africa with respect to Europe in Late Cretaceous times from strike-slip to north–south convergence (Patriat *et al.* 1982; Livermore & Smith, 1984). These orogenic events are all inferred to have occurred within a relatively restricted area in western Turkey (e.g. Şengör & Yılmaz, 1981; Robertson *et al.* 1996). Our aim in this paper is to study the Upper Cretaceous–Lower Eocene sedimentary sequences in western Anatolia, with a view of understanding how these various orogenic events are reflected in the sedimentary record. In describing the Upper Cretaceous–Lower Eocene sedimentary sequences in western Turkey, we have tried to be precise by choosing, as far as possible, biostratigraphically controlled, measured stratigraphic sections. In the absence of these, synthetic stratigraphic sections from small and well-defined areas (less than 200 km^2) were used. The Upper Cretaceous–Lower Eocene sections from western Turkey are shown in 38 stratigraphic columns, which provide information on the rock type, thickness, stratigraphic position and fossil content for each formation.

Dating the Upper Cretaceous–Lower Eocene formations in western Turkey is commonly done using foraminifera. However, in some cases rudists are used for dating Senonian shallow marine limestones. For the pelagic Senonian sequences, *Globotruncana* species generally provide a precise age control, whereas for neritic Senonian rocks a biostratigraphic age is generally established using large foraminifera such as *Orbitoides*. Biostratigraphic control for the neritic Lower Tertiary sections are provided by large foraminifera, such as *Laffitteina*, *Nummulites*, *Discocyclina* and *Assilina*, and that for the pelagic sections by *Globorotalia* and *Globigerina* species.

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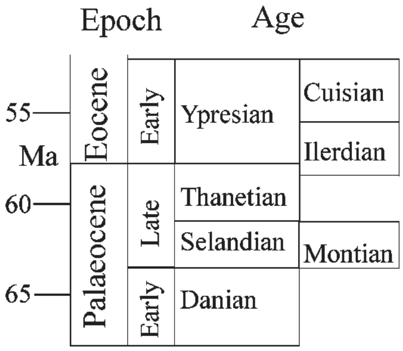

Figure 1. Palaeocene–Early Eocene geochronology (Berggren *et al.* 1995; Serra-Kiel *et al.* 1998) with local age names used in the stratigraphic columns.

In the stratigraphic columns (Figs 6–12) we indicate the ages proposed by the original palaeontological studies. Subdivision of Campanian and Maastrichtian sequences into lower, middle and upper, is based on the standard planktic foraminifera zones (Robaszynski *et al.* 1984). The Palaeocene is subdivided into three stages: Danian, Selandian and Thanetian (Berggren & Norris, 1997). 'Montian' is commonly used in palaeontological studies in Turkey for the Middle Palaeocene (e.g. Karaköse, Dinçel & Kıral, 1987). It is approximately equivalent to the Selandian of Berggren & Norris (1997). For benthic Lower Tertiary biozones the stage names 'Ilerdian' and 'Cuisian' are commonly used in Turkey. 'Ilerdian' overlaps with the late Thanetian and the early Ypresian (Early Eocene), and 'Cuisian' corresponds to the late Ypresian (Fig. 1) (cf. Serra-Kiel *et al.* 1998). For Cretaceous geochronology we use the scheme of Kent & Gradstein (1985), and for Tertiary geochronology that of Berggren *et al.* (1995).

2. The tectonic setting of western Turkey

Northwest Turkey comprises three first-order tectonic units, which were separated by the Intra-Pontide and the İzmir-Ankara oceans during most of the Mesozoic (Fig. 2) (Şengör & Yılmaz, 1981; Okay & Tüysüz, 1999). In the north, the Strandja and İstanbul zones can be broadly considered a part of the southern margin of Eurasia. The Sakarya Zone was largely an independent continental fragment between the Intra-Pontide and İzmir-Ankara Neo-Tethyan oceans. The Strandja, İstanbul and Sakarya zones together constitute the Pontides. The Anatolide-Tauride Block is separated from the Sakarya Zone by the İzmir-Ankara-Erzincan suture and from Gondwana by the Pamphylian/Assyrian/Zagros suture (Fig. 2). The Anatolide-Tauride Block was the site of a large carbonate platform during most of the Mesozoic era (Görür, 1998). During the subduction and collision events in Late Cretaceous–Early Eocene times, it formed the lower plate and therefore was strongly deformed and partially metamorphosed. This leads to the subdivision of the Anatolide-Tauride Block into secondary zones characterized by their metamorphic and structural features (Fig. 2). In contrast, the Pontides formed the upper plate during these events, and therefore were less intensely deformed than the Anatolide-Tauride Block, and were also largely free of Late Cretaceous–Tertiary regional metamorphism.

Four main tectonic events affected western Turkey in Late Cretaceous–Early Eocene times. These are subduction, ophiolite obduction, high-pressure/low-temperature metamorphism and continent–continent collision. For the first three closely spaced events there are independent isotopic data to constrain the timing, while for the continent–continent collision we have to rely solely on the stratigraphic and tectonic record. Some constraints on these four tectonic events can be placed by the age of the Tethyan oceanic lithosphere, as deduced from obducted accretionary complexes and ophiolites.

3. Age span of the Tethyan oceanic lithosphere in western Turkey as deduced from the accretionary complexes and ophiolite

Oceanic accretionary complexes cover wide areas south of the İzmir-Ankara suture (Figs 3, 4). They are made up dominantly of mafic lava, pyroclastic flow, radiolarian chert and pelagic shale with minor pelagic limestone, greywacke and serpentinite. Some accretionary complexes also comprise continental margin sequences (e.g. the Lycian melange: Collins & Robertson, 1997), probably incorporated during their emplacement over the continental crust. These Tethyan accretionary complexes differ from their circum-Pacific counterparts by the scarcity of the siliciclastic sediments, which reflects the subdued topography adjacent to the trenches. The radiolarian cherts in the accretionary complexes have major potential for establishing the age-span of the Neo-Tethyan ocean, and for constraining the age of subduction. However, to date, radiolaria from only one outcrop of cherts in a west Turkish accretionary complex have been properly dated (locality A in Fig. 3). Various blocks of radiolarian chert from this small outcrop northwest of Ankara have yielded radiolaria of late Norian, Early Jurassic, Kimmeridgian–Tithonian, Early Cretaceous and Albian–Turonian ages (Bragin & Tekin, 1996). Apart from this detailed study, Late Jurassic (late Kimmeridgean–late Tithonian) and Early Cretaceous radiolaria have been reported from the accretionary complexes from the Biga Peninsula (Beccaletto & Stampfli, 2000) and from north of Eskişehir (Göncüoğlu *et al.* 2000). These limited data imply that the İzmir-Ankara Neo-Tethyan ocean had an age span of at least late Norian to Albian.

Isolated outcrops of ophiolite, largely represented by their ultramafic portions, occur widely in the Anatolide-Tauride Block. In the southern part of the

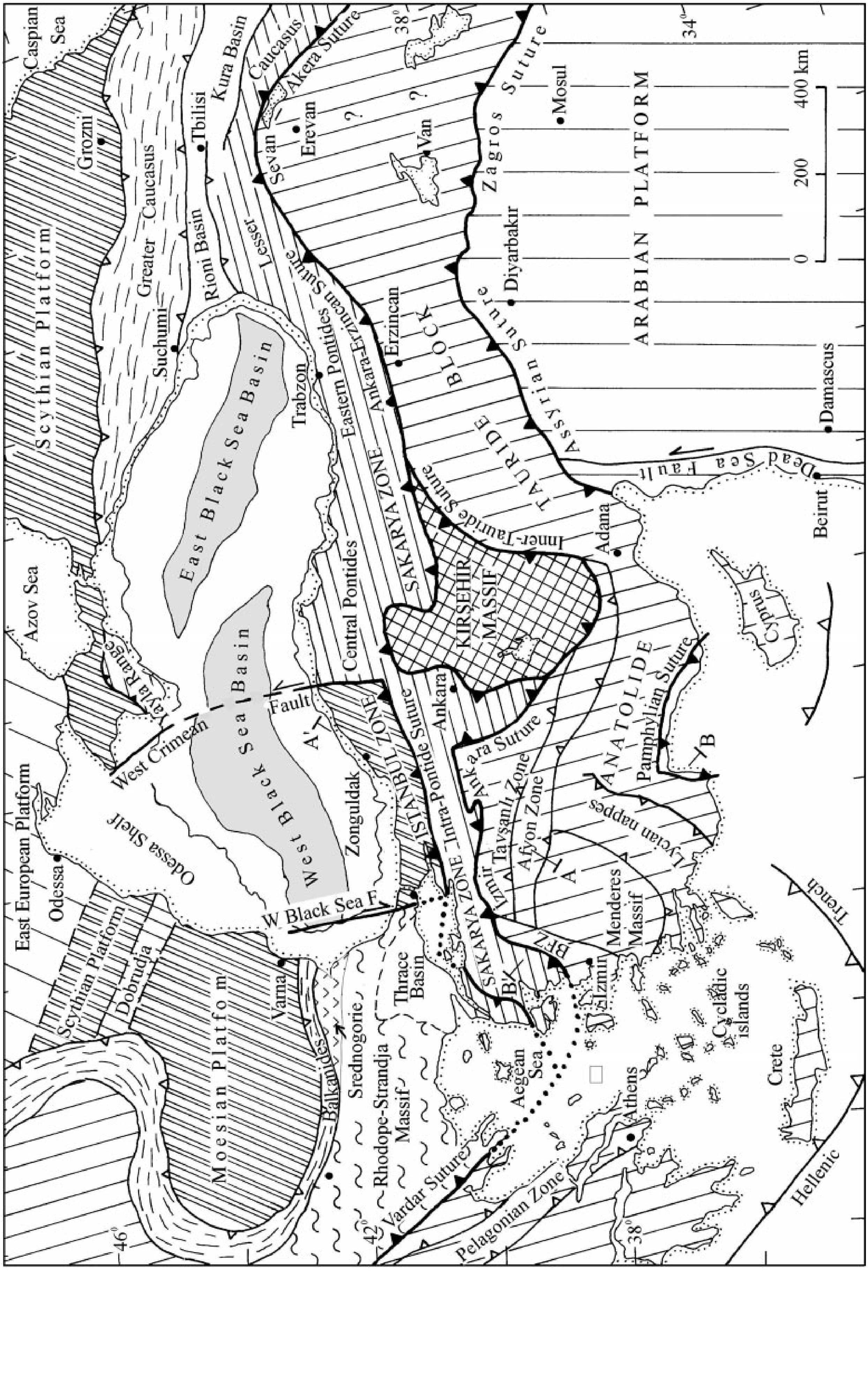

Figure 2. Tectonic map of the Aegean and Black Sea region (modified from Okay & Tüysüz, 1999).

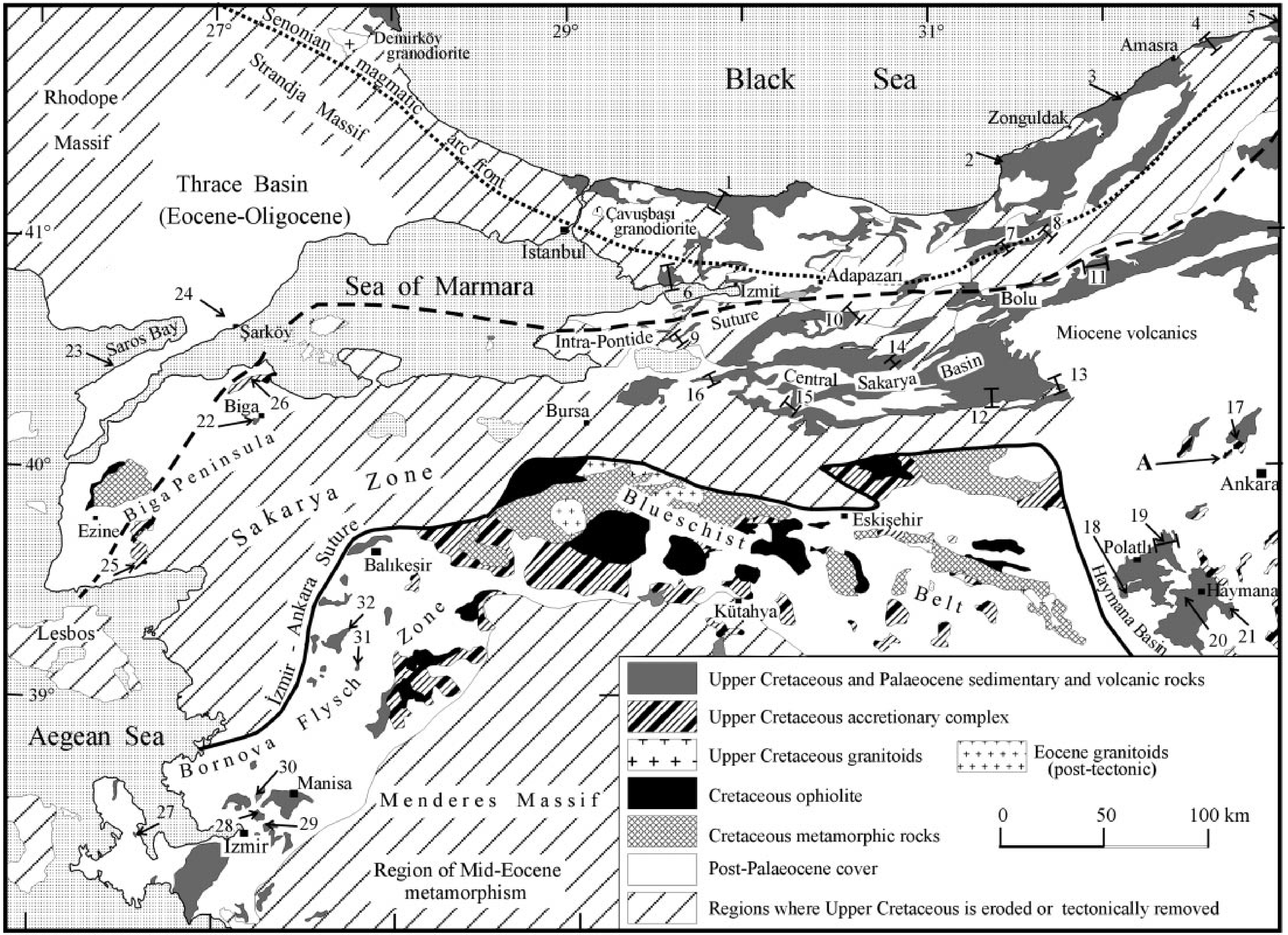

Figure 3. Outcrops of Upper Cretaceous and Palaeocene rocks in western Turkey. Locations of the stratigraphic sections portrayed in Figures 6 to 11 and discussed in the text are numbered.

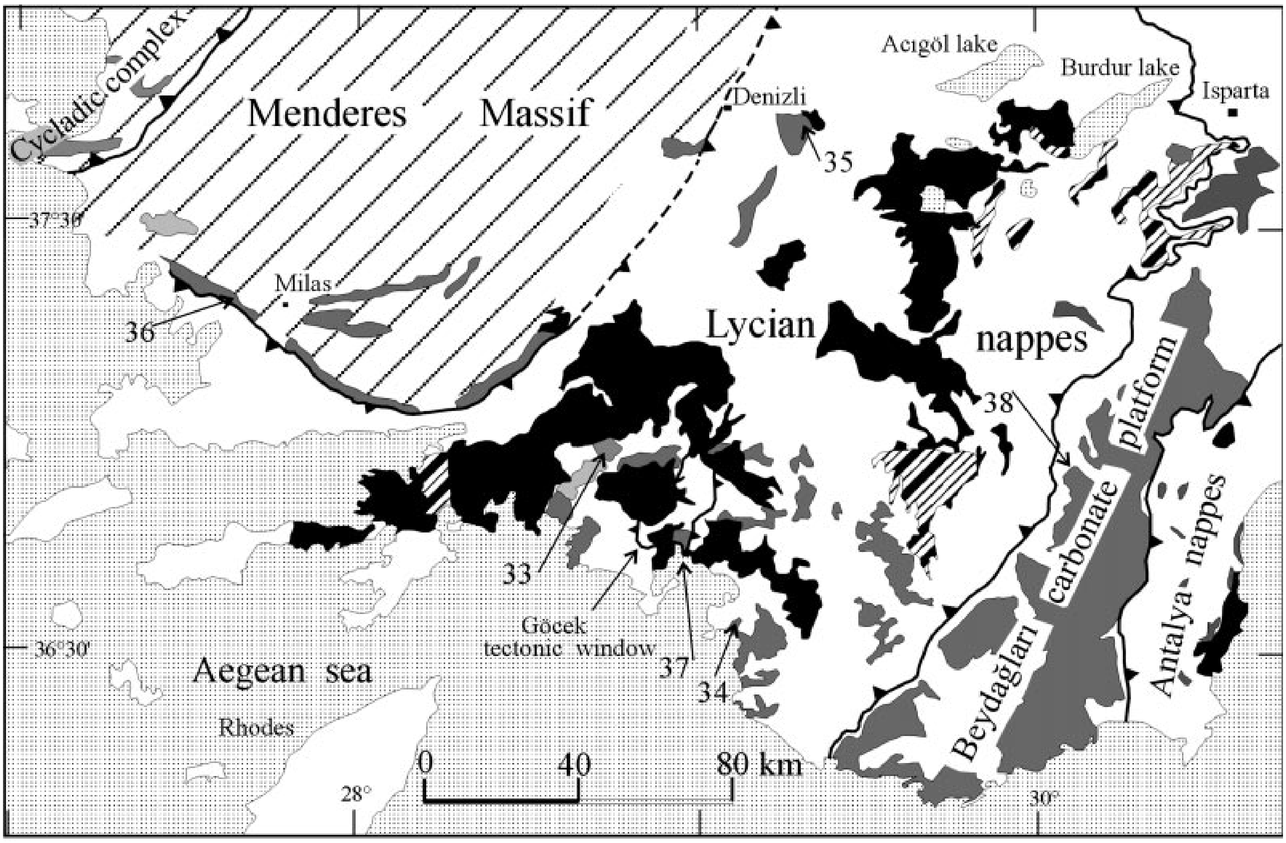

Figure 4. Outcrops of Upper Cretaceous and Palaeocene rocks in the western Taurides. Locations of the stratigraphic sections portrayed in Figure 12 and discussed in the text are numbered. The legend is the same as that for Figure 3, except the white areas in the Lycian nappes represent pre-Cretaceous and post-Eocene rocks.

Anatolide-Tauride Block, in the Taurides *sensu stricto*, ophiolite nappes lie tectonically over Mesozoic carbonates separated only by discontinuous melange slivers (Figs 4, 5) (Bernoulli, Graciansky & Monod, 1974; Tekeli *et al.* 1984; Özgül, 1997; Collins & Robertson, 1998, 1999). In the northern part of the Anatolide-Tauride Block, in the Tavşanlı Zone, the ultramafic rocks lie, either directly or separated by a tectonic layer of accretionary complex, over the blueschists (Okay, 1984).

The lack of upper, extrusive parts of the ophiolite in western Turkey means that there are no direct data on the age of formation of the ophiolite. However, in the Tethyan realm the age of metamorphic soles of the ophiolite is generally close to the age of the ophiolite generation. Ar/Ar and K/Ar isotopic data from the metamorphic soles of the ophiolite nappes in western Turkey yield a scatter of ages between 101 to 88 Ma but with an age concentration around 95–90 Ma (Thuizat *et al.* 1981; Harris, Kelley & Okay, 1994; Parlak & Delaloye, 1999; Dilek *et al.* 1999), which contrasts with the Early to Middle Jurassic ages from the metamorphic soles of the Hellenic and Dinaric ophiolites (e.g. Spray *et al.* 1984). The western Anatolian ophiolites also comprise diabase dykes, 91 to 85 Ma old (Parlak & Delaloye, 1999; Dilek *et al.* 1999). These dykes are different from those in a sheeted dyke complex, in that they are isolated and cut the ultramafic part of the ophiolite. The diabase dykes are also locally intrusive into the metamorphic sole but do not cut the underlying melange or the underlying carbonate platform, indicating that they have formed prior to the emplacement of the ophiolite over the carbonate platform (Whitechurch, Juteau & Montigny, 1984). The geochemistry of the dykes suggests that they formed above a subduction zone (Collins & Robertson, 1998; Dilek *et al.* 1999). The isotopic data from the sub-ophiolite metamorphic rocks and diabase dykes indicate that the western Anatolian ophiolites formed before the Cenomanian. The short interval between the generation of the ophiolite and that of its metamorphic sole, observed in the Tethyan ophiolites, suggests that the western Anatolian ophiolites are of Early Cretaceous age. The common lithological and isotopic features of the Anatolian ophiolites led Dilek *et al.* (1999) to suggest that they once formed part of a very large ophiolite body comparable in size to the Semail ophiolite in Oman.

The Lower Cretaceous ophiolite nappes in the Anatolide-Tauride Block are, therefore, considerably younger than the Triassic to Cretaceous Neo-Tethyan oceanic lithosphere, and possibly have formed above a subduction zone, similar to the case invoked for the Semail ophiolite in Oman (e.g. Collins & Robertson, 1998). Alternatively they may represent the last

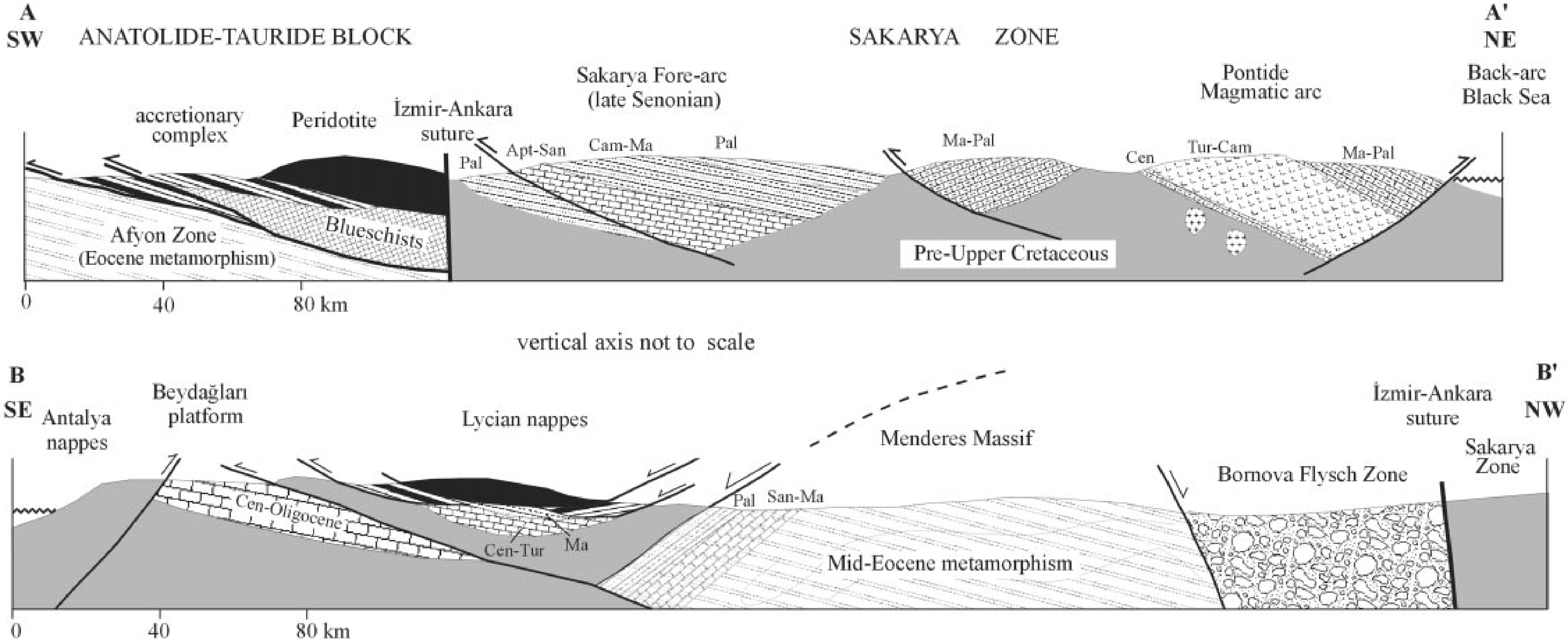

Figure 5. Schematic cross-sections from western Anatolia showing the general structure and disposition of the various Cretaceous–Eocene belts. For the location of the cross-sections see Figure 1. Apt, Aptian; Cen, Cenomanian; Tur, Turonian; San, Santonian; Cam, Campanian, Ma, Maastrichtian; Pal, Palaeocene.

oceanic lithosphere generated before the ridge subduction (Lytwyn & Casey, 1995; Polat, Casey & Kerrich, 1996). The latter model is more appealing, because arc magmatism in the Pontides started in late Turonian–Coniacian times, soon after the inferred formation of the ophiolites.

4. Subduction and the development of the Pontide magmatic arc

Probably the best evidence of subduction in the geological record is a magmatic arc, which forms at depths in excess of 80 km above the descending oceanic lithosphere, usually 150–200 km away from the trench axis. The magmatic arc, constructed from volcanic and plutonic rocks, also provides a minimum age range for the subduction process. The Neo-Tethyan oceans in Turkey are generally believed to have subducted northward during the Late Cretaceous (e.g. Şengör & Yılmaz, 1981). This is mainly based on data from the northeastern Turkey, where there is a well-developed magmatic arc in the eastern Pontides in the eastern part of the Sakarya Zone (Akın, 1978; Akıncı, 1984; Robinson *et al.* 1995; Okay & Şahintürk, 1997). Palaeontological data from the limestones in the volcanic sequence and the isotopic ages of the plutonic rocks in the Eastern Pontides indicate that the subduction was active from Turonian to the end of Maastrichtian or in some cases even to Danian time (Okay & Şahintürk, 1997). This magmatic arc can be traced westward along the southern margin of the Black Sea coast to the Srednogorie Zone in Bulgaria. Figure 6 shows a series of stratigraphic sections along the western Black Sea coast of Turkey. Most show andesitic lavas, agglomerates and tuffs accompanied by pelagic limestones in the Coniacian–Campanian interval. Geochemically the volcanic rocks are calc-alkaline, similar to the island arc volcanics (Yeniyol & Ercan, 1990). In the Kurucaşile (profile 4 in Fig. 6) the base of the volcanic sequence is dated as middle Turonian by pelagic limestones (Akyol *et al.* 1974; Tüysüz, Kırıcı & Sunal, 1997). This Coniacian–Campanian magmatic arc sequence rests with a regional unconformity on rocks as old as Triassic (profile 1 in Fig. 6), indicating a period of uplift and erosion before the initiation of the magmatic arc. The arc sequence is conformably overlain by pelagic marl and shaley limestone, Maastrichtian to Late Palaeocene in age, locally continuing up to the Lower Eocene sequence (profiles 1 and 2 in Fig. 6). Continuous carbonate sedimentation during Maastrichtian and Palaeocene times characterizes the western Black Sea region (profiles 1, 2, 3 and 5 in Fig. 6; Dizer & Meriç, 1983; İnan, 1995; Derman & İztan, 1997; Matsumaru *et al.* 1996). All ten planktonic foraminifera zones from Coniacian to Danian have been recognized along the western Black Sea coast of Turkey (Özkan-Altıner & Özcan, 1999), indicating uninterrupted submarine sedimentation during and subsequent to arc volcanism. In the Cide section (profile 5 in Fig. 6) carbonate deposition continued into Early Eocene time, and was then succeeded by siliciclastic deposition during Middle Eocene time (Akyol *et al.* 1974; Tüysüz, Kırıcı & Sunal, 1997). The Upper Cretaceous stratigraphy of the western Black Sea coast is similar to that of the Srednogorie Zone, where a volcanic-dominated Coniacian to Campanian sequence, more than 3 km thick, is overlain by Maastrichtian shallow marine limestones (Aiello *et al.* 1997; Boccaletti *et al.* 1978; Nachev, 1993; Nachev & Dimitrova, 1995).

Structurally the western Black Sea coastal region is characterized by east–northeast trending and north-vergent thrusts, which emplace Triassic sandstones on Cretaceous sequences. The thrusts are associated with open to tight folds with an east–northeast trending fold axis. Upper Eocene to Pliocene sequences are missing in the Black Sea coastal region, which places a broad post-Middle Eocene age on the foreland type deformation (Yiğitbaş & Elmas, 1997*a*).

The Coniacian–Campanian volcanism is accompanied by minor calc-alkaline plutonism along the western Black Sea coast. The Demirköy granodiorite in the Strandja Massif has yielded biotite and hornblende K/Ar ages of 78 ± 2 and 79 ± 2 Ma, respectively (Moore, McKee & Akıncı, 1980), whereas the Çavuşbaşı granodiorite near İstanbul gave biotite Rb/Sr ages of 65 ± 10 Ma (Öztunalı & Satır, 1975). The Palaeozoic rocks in the İstanbul region and the metamorphic rocks of the Strandja Massif are also cut by numerous andesitic dykes and sills, related to the Senonian magmatism. In most subduction zones the magmatic-arc front is a well-defined line. The Pontide magmatic arc is no exception, and Figure 3 shows the Coniacian–Campanian magmatic-arc front defined by the southernmost occurrence of andesitic dykes, sills and shallow intrusions.

Subduction of either the Intra-Pontide or the İzmir-Ankara ocean could have generated the Coniacian–Campanian Pontide magmatic arc. The closure age of the Intra-Pontide ocean has been variously estimated as Late Cretaceous (Yılmaz *et al.* 1995; Yiğitbaş, Elmas & Yılmaz, 1999) or Early Eocene (Şengör & Yılmaz, 1981; Okay, Şengör & Görür, 1994). However, recent work in the central Pontides, where the İstanbul and Sakarya zones are in tectonic contact, has indicated that both zones have a common Senonian cover but differ in their pre-Senonian stratigraphy and evolution (Tüysüz, 1999). Thus, the Intra-Pontide ocean appears to have closed during Cenomanian time, concomitant with the opening of the western Black Sea basin (Görür, 1997; Tüysüz, 1999). A second argument favouring a source of Pontide arc magmatism in the İzmir-Ankara ocean, is the position of the Intra-Pontide suture. The distance between the Intra-Pontide suture and the magmatic arc-front is only 5 to 40 km, and there is no evidence for significant Cretaceous north–south shortening in the suture–arc

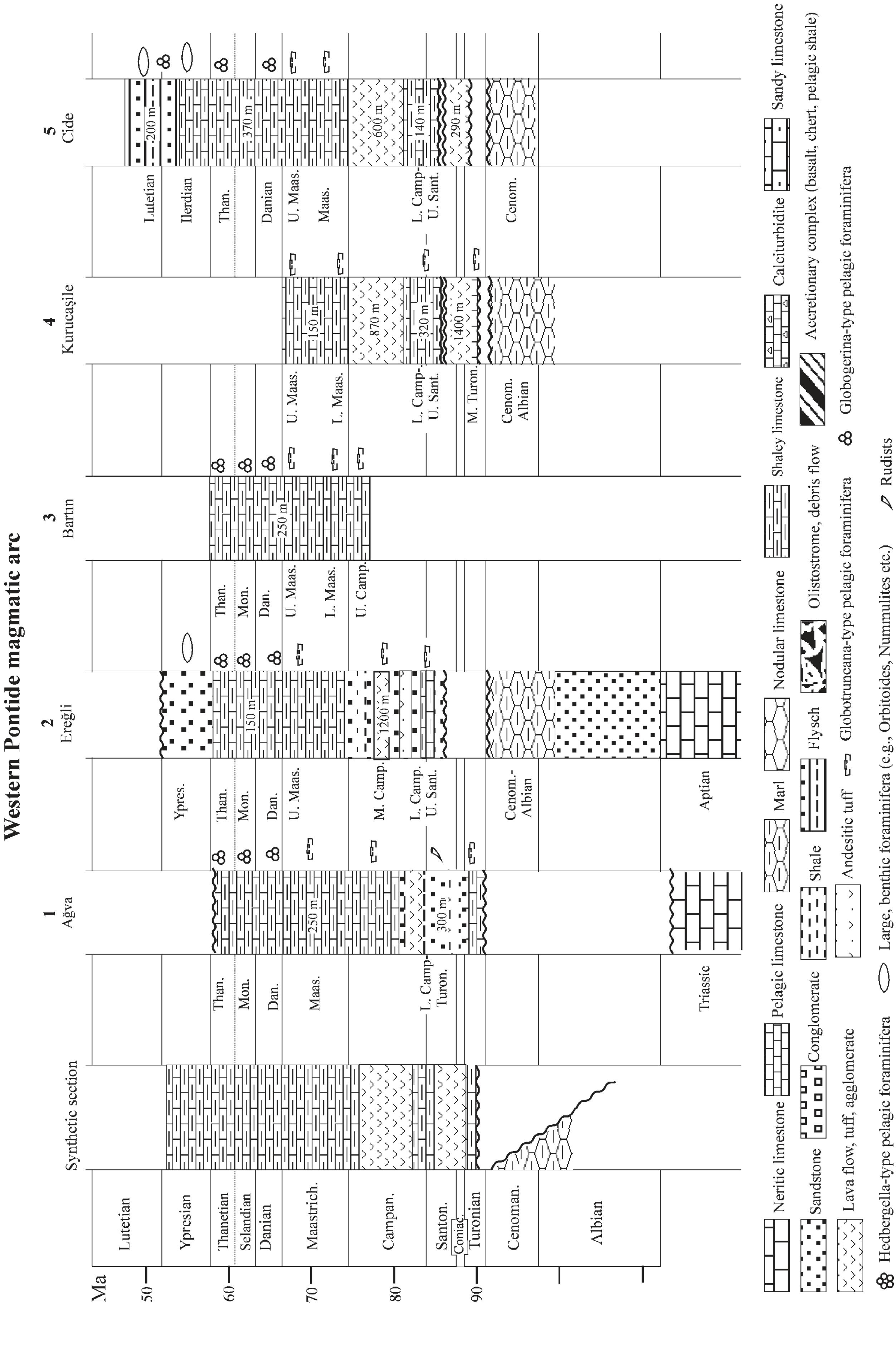

Figure 6. For legend see facing page.

interval (Figs 2, 4). This distance is much less than the 150–200 km observed in the modern subduction zones between the arc front and the trench axis. In contrast, the İzmir-Ankara suture east of Bursa is situated about 150 km south of the Pontide magmatic front, about the right distance for the generation of a magmatic arc.

5. Senonian south of the Pontide magmatic arc front

South of the Pontide magmatic arc is a belt about 20 km wide where the Senonian stratigraphy is quite varied. In many regions Maastrichtian shallow marine limestones lie unconformably over a Palaeozoic to Triassic basement (Fig. 7). The base of the transgressive sequence ranges from uppermost Campanian (profile 6 in Fig. 7) through Maastrichtian (profile 9) to uppermost Maastrichtian to Palaeocene (profile 8). These shallow marine limestones pass upward to pelagic shaley limestone and marl, similar to those observed along the Black Sea coastal region. In several sections (e.g. south of İstanbul, profile 6), all the biozones of the Maastrichtian and Palaeocene are recognized, indicating continuous pelagic deposition throughout Maastrichtian and Palaeocene times. In contrast in other localities near the arc front (profiles 7, 10 and 11 in Fig. 7) lower Senonian deep marine sequences are preserved beneath the Maastrichtian unconformity. Based on the general absence of lower Senonian deposits, Yiğitbaş, Elmas & Yılmaz (1999) suggest that there was an erosional ridge south of the western Pontide magmatic arc during Coniacian–Campanian times.

6. Upper Cretaceous fore-arc sequences in the Sakarya Zone

6.a. Central Sakarya basin

North of the İzmir-Ankara suture there is a well-developed Jurassic to Eocene trough called the Central Sakarya basin (Fig. 2). The Mesozoic stratigraphy in the Central Sakarya basin is known in detail (Altınlı, 1975; Altıner *et al.* 1991; Saner, 1978*a*, 1980) and representative Cretaceous to Lower Eocene stratigraphic profiles from the Central Sakarya basin are shown in Figure 8. The eastern part of the Central Sakarya basin shows uninterrupted sedimentation between Late Jurassic and Palaeocene time (profile 12 in Fig. 8). In the Nallıhan section, the Tithonian–Early Campanian interval is represented by pelagic limestone deposition (Tansel, 1980). The first sandstone interbeds in the carbonate sequence start in the lower Campanian, and by the lower Maastrichtian the sequence is transformed into siliciclastic turbidites with rare marl intervals. Further east in the Beypazarı region, the siliciclastic deposition apparently began earlier during Santonian time (profile 13 in Fig. 8). The Campanian to Lower Maastrichtian sandstones often contain grains of glaucophane (Varol, 1979; Saner, 1980; Tunç, 1984). The turbidite sequence shows a regressive development and in the upper Maastrichtian sequence, shallow marine sandstones with large benthic foraminifera (*Orbitoides* sp.) have been deposited. These pass up into continental red sandstones and conglomerates presumably of Early Palaeocene age (Tansel, 1980). The northern margin of the Central Sakarya basin in the Mudurnu-Göynük area (profile 14 in Fig. 8) shows a similar stratigraphic development but with two important differences. The Turonian–Santonian pelagic micrites comprise tuff horizons, and the pelagic sequence continues into the Middle Palaeocene before passing up into continental red sandstones of presumably of Late Palaeocene age (Saner, 1978*b*; Altıner *et al.* 1991; Meriç & Şengüler, 1986).

Upper Cretaceous sequences are less continuous in the western part of the Central Sakarya basin; in particular, there is a major unconformity of late Campanian–early Maastrichtian age (profiles 15 and 16 in Fig. 8). The Maastrichtian sequence above the unconformity is represented by a heterogeneous series of turbidites, limestones, debris flows, olistostromes and tuffs, which, in many localities, lie unconformably over Triassic and older rocks. This east–west facies difference in the Central Sakarya basin stretches back to the Late Jurassic–Early Cretaceous, when the western part of the basin was characterized by neritic carbonate sedimentation and the eastern part by pelagic biomicrite deposition (Saner, 1980; Altıner *et al.* 1991).

Structurally the Central Sakarya basin is a fold and thrust belt with east–west trending fold axes and generally south-vergent thrusts. The deformation is constrained to the Late Palaeocene and Middle Eocene by an unconformable cover of Lutetian limestones (e.g. profile 15 of Fig. 8). The Central Sakarya basin is situated only 20 km north of the İzmir-Ankara suture. It is tempting therefore to regard the Upper Campanian–Maastrichtian sequence of the Central Sakarya basin as a fore-arc sequence. The presence of blueschist and serpentinite detritus in the Maastrichtian sandstones is in keeping with a fore-arc setting, assuming that the fore-arc basin was fed by the uplifted wedges of the accretionary complex. A fore-arc setting has already been suggested for the Maastrichtian–Tertiary sequences in the Haymana basin further east in the Sakarya Zone.

Figure 6. Stratigraphic sections along the western Black Sea coast. The locations of the sections are shown in Figure 3. The stage names on the left of the stratigraphic columns indicate those characterized by the fossil group on the right of the column. The data sources are: 1. Baykal, 1943; Tansel, 1989; 2. Kaya *et al.* 1984; 3. Dizer & Meriç, 1983; 4. Akyol *et al.* 1974; Tüysüz, Kırıcı & Sunal, 1997; 5. Akyol *et al.* 1974; Sirel, 1998; Kırcı & Özkar, 1999.

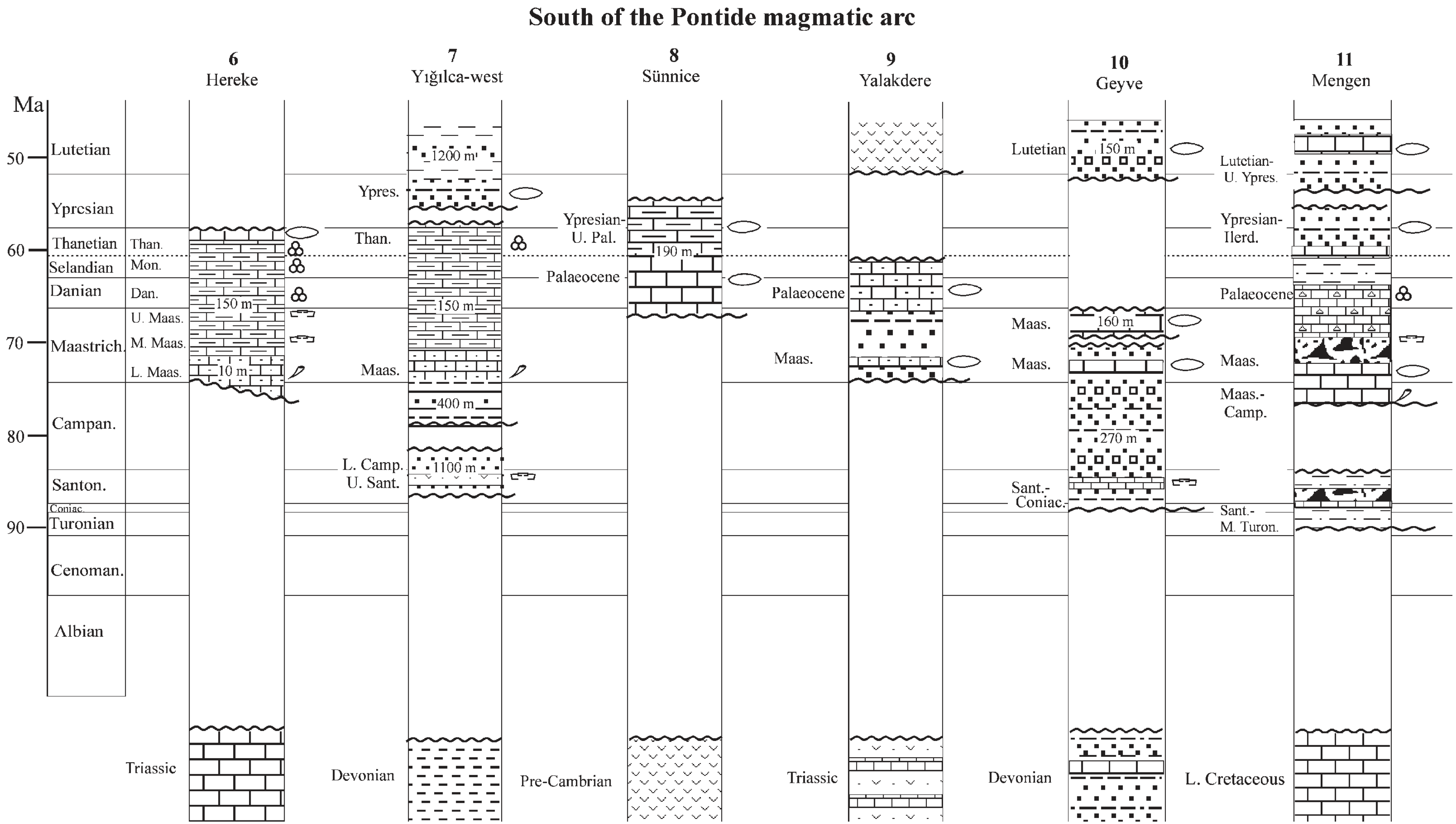

Figure 7. Stratigraphic sections south of the Pontide magmatic arc. For locations of the profiles see Figure 3, and for legend see Figure 6. The data sources are: 6. Özer, Tansel & Meriç, 1990; 7. Kaya *et al.* 1987; 8. Yiğitbaş & Elmas, 1997*b*; 9. Akartuna, 1968; Bargu & Sakınç, 1987; 10. O. R. Atan, unpub. Thesis, İstanbul Univ., 1973; Akartuna & Atan, 1981; 11. Yiğitbaş & Elmas, 1997*b*.

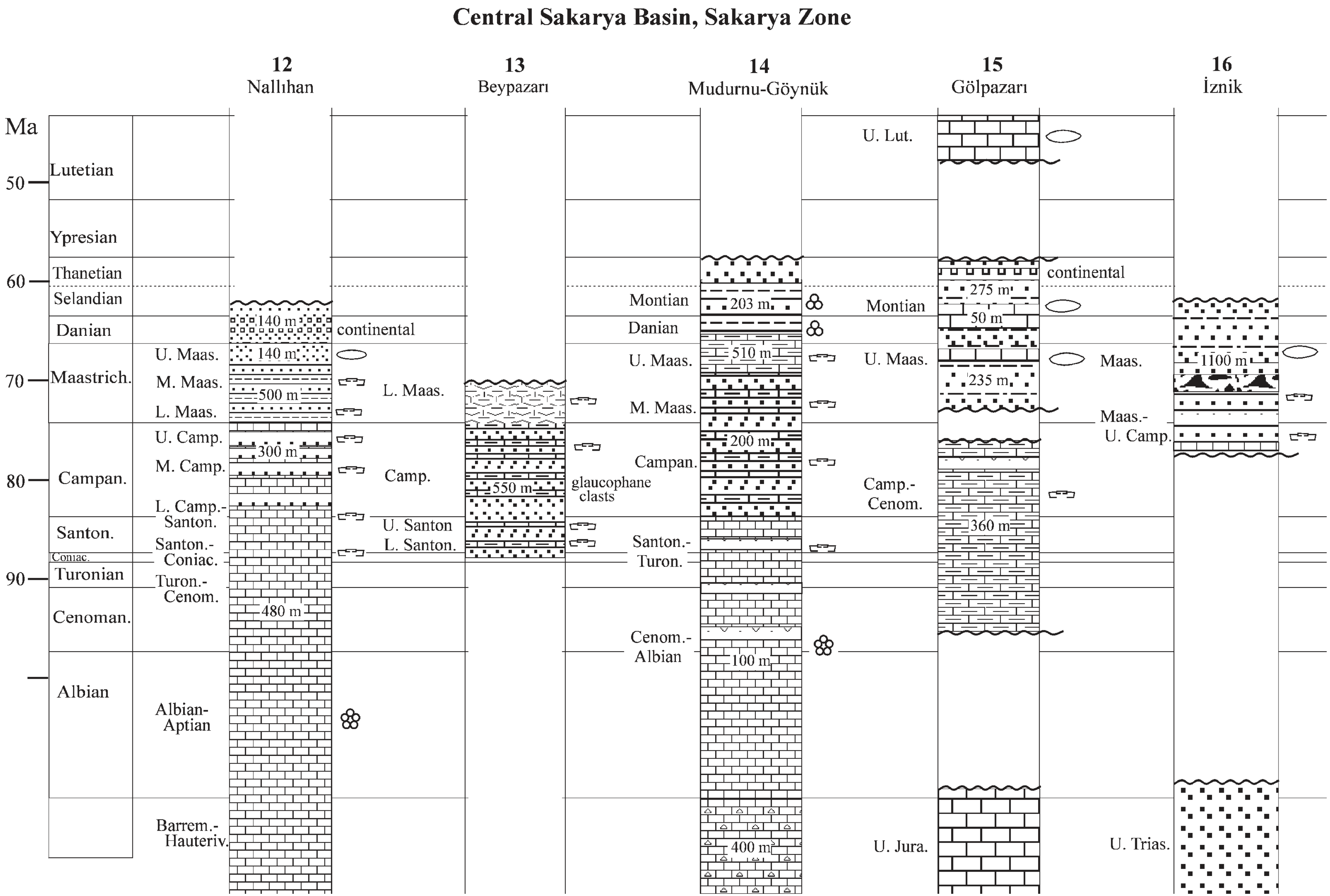

Figure 8. Stratigraphic sections from the Central Sakarya basin and further west. For locations of the profiles see Figure 3, and for legend see Figure 6. The data sources are: 12. Tansel, 1980; 13. Tunç, 1984; 14. Meriç & Şengüler, 1986; Altıner *et al.* 1991; 15. Eroskay, 1965; Dizer & Meriç, 1983; 16. Bargu, 1982; C. Genç, unpub. Ph.D. Thesis, İstanbul Technical Univ., 1993.

6.b. Haymana basin

The Haymana basin comprises a continuous Maastrichtian to Eocene sedimentary sequence, over 5 km in total thickness. The sedimentology and biostratigraphy of the Haymana sequence are well known (Sirel, 1975; Ünalan *et al.* 1976; Görür *et al.* 1984; Görür, Tüysüz & Şengör, 1998). The Palaeocene and Eocene sequences in the Haymana basin constitute reference sections for the Palaeocene–Eocene shallow benthic foraminiferal biozones (e.g. Serra-Kiel *et al.* 1998). Furthermore, all three biozones of the Maastrichtian have been recognized in the Haymana basin (Toker, 1979). The Upper Cretaceous sequence in the Haymana basin lies unconformably over an accretionary complex or over the Upper Jurassic–Lower Cretaceous pelagic carbonates of the Sakarya Zone (profiles 17 and 18 in Fig. 9). The accretionary complex is made up of basalt, radiolarian chert and serpentinite with minor turbidite. The geochemistry of some of the basalts is similar to that of the ensimatic arc volcanic rocks (Tüysüz, Dellaloğlu & Terzioğlu, 1995). Biostratigraphic data from the accretionary complex include radiolaria as young as Albian–Turonian (Bragin & Tekin, 1996). The accretionary complex was emplaced northward over the carbonates of the Sakarya Zone during Cenomanian–Turonian times. This poorly understood compressive event can be traced eastward to the Eastern Pontides and even to the Sevan-Akera Zone in the Lesser Caucasus (Okay & Şahintürk, 1997) but is not observed further west, for example in the Central Sakarya basin.

The base of the transgressive Upper Cretaceous sequence is dated by the *Globotruncana* species as Early Maastrichtian in the Haymana basin (Toker, 1979). The early to late Maastrichtian is represented by siliciclastic turbidites, more than 1000 m thick (Fig. 9). The turbidites comprise ophiolite and blueschist detritus, including pebbles of glaucophane–lawsonite schists (Batman, 1978). The overlying Lower Palaeocene sequence shows facies variation and ranges from continental clastics in the northwest (profile 19 in Fig. 9) through shallow marine limestones (profile 20) to pelagic shales in the southeast (profile 21). This and the absence of the Maastrichtian sequence in the western part of the Haymana basin (profile 18 in Fig. 9) indicate uplift in the region around the İzmir-Ankara suture during Early Palaeocene times (Ünalan *et al.* 1976; Görür *et al.* 1984; Koçyiğit, 1991). In the Haymana basin a second transgressive cycle starts with Late Palaeocene (Thanetian) shallow marine limestone and marl, continues with Thanetian–Ilerdian turbidites and ends with Ilerdian–Cuisian shales. The shales are conformably overlain by shallow marine to fluviatile sandstones and conglomerates of Cuisian–Lutetian age (profiles 19 and 21 in Fig. 9).

The location of the Haymana basin adjacent to the İzmir-Ankara suture, the stratigraphic position of the basinal sediments over an accretionary complex led to the interpretation of the Haymana basin as a fore-arc basin (Görür *et al.* 1984; Koçyiğit, 1991). The Upper Cretaceous–Palaeocene sequences in the eastern Central Sakarya basin and northwestern Haymana basin are similar, and reflect the formation of a fore-arc basin during the Campanian–Maastrichtian followed by the uplift in the Palaeocene. As discussed in Section 9, this uplift may be related to the initiation of the continent–continent collision.

6.c. Cretaceous–Palaeocene in the western Sakarya Zone

Cretaceous–Palaeocene deposits are absent over a large area between Bursa and the Aegean Sea mainly due to Oligocene erosion. In this region Miocene continental sediments are frequently observed to lie unconformably over Triassic and older rocks. A recent discovery of a small Palaeocene inlier around the town of Biga in the Biga peninsula is therefore of considerable interest (M. B. Yıkılmaz, unpub. Thesis, İstanbul Technical Univ., 1999). The sequence consists of pelagic red micrite intercalated with calciturbidite, thick debris and grain flow deposits made up of Jurassic limestone clasts in a red sandy or micrite matrix (profile 22 in Fig. 10). A rich fauna of pelagic foraminifera indicates an Early Palaeocene age. The Palaeocene sequence is unconformably overlain by shallow marine limestones with Lutetian nummulites. Another small outcrop of Palaeocene pelagic sediments is found in the southern margin of the Saros Bay in the Aegean Sea (Fig. 3). Here, a thick pelagic micrite sequence of Maastrichtian to Palaeocene age is unconformably overlain by Lower Eocene shallow marine sediments (profile 23 in Fig. 10) (Önal, 1986). The bases of the Biga and Saros pelagic sequences are not exposed. Blocks of Lower and Middle Palaeocene pelagic limestone associated with blocks of serpentinite, blueschist, diabase, Senonian (Campanian, Maastrichtian) red pelagic micrite and Lower Eocene reefal limestone, occur in the Eocene turbidites north of Şarköy in Thrace (profile 24 in Fig. 10) (Okay & Tansel, 1994). Thus, sporadic evidence suggests that the region between Bursa and the Aegean was characterized by pelagic carbonate deposition in the late Senonian–Middle Palaeocene interval.

Ophiolitic melanges, marking the northwestern limit of the Sakarya Zone, crop out in the central Biga peninsula (Fig. 3). Like the accretionary complexes south of the İzmir-Ankara suture, these are mainly composed of basalt, radiolarian chert, pelagic shale but also include blocks of Middle (Anisian) and Upper Triassic (Norian), Upper Jurassic–Lower Cretaceous and Upper Cretaceous limestone (profile 25 in Fig. 10) (Okay, Siyako & Bürkan, 1991; Okay *et al.* 1996; Beccaletto & Stampfli, 2000). The Upper Triassic blocks are commonly several hundred metres in size, whereas the Turonian–Santonian red marly

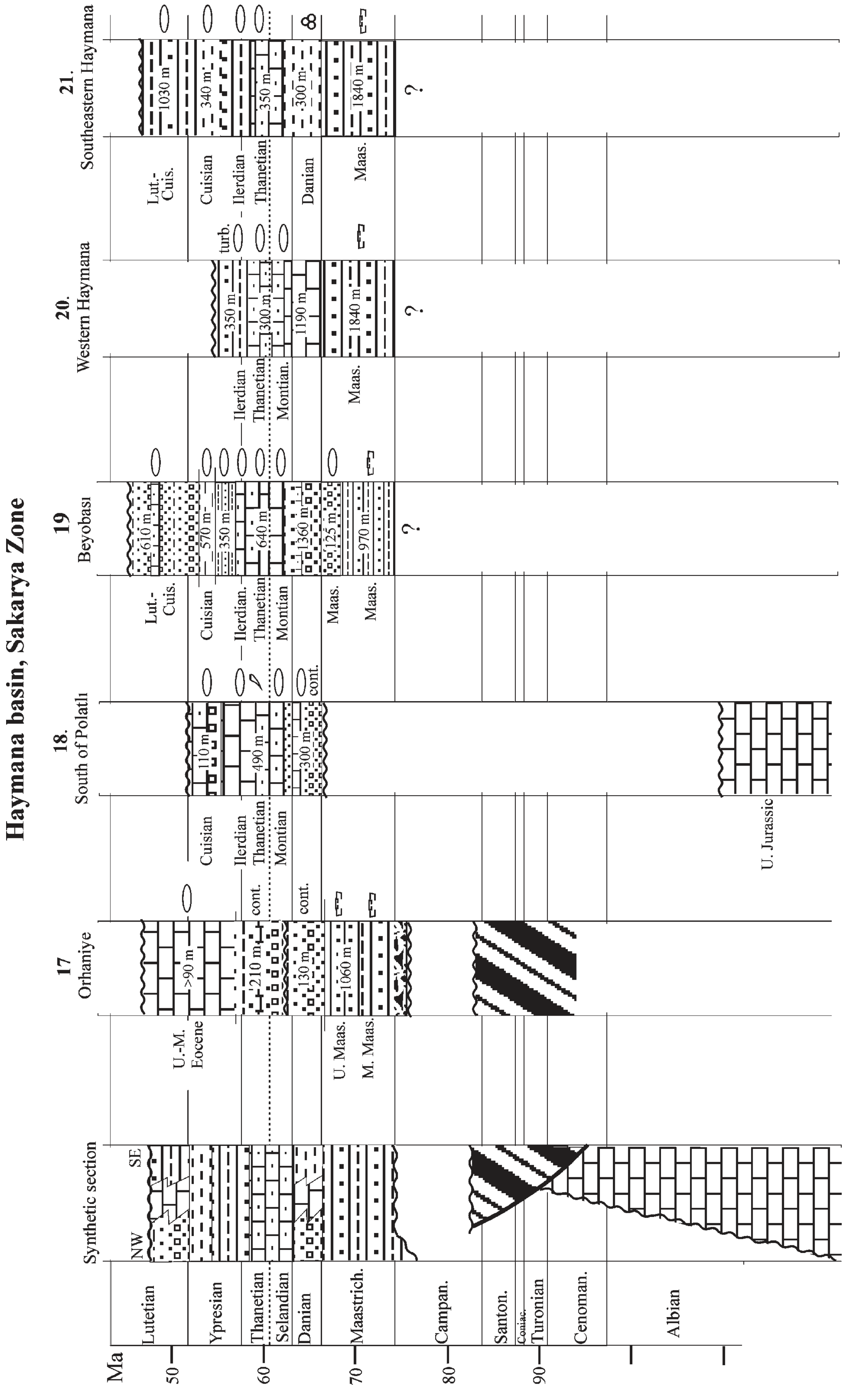

Figure 9. Stratigraphic sections from the Haymana basin of the Sakarya Zone. For locations of the profiles see Figure 3, and for legend see Figure 6. The data sources are: 17. Gökten, Kazancı & Acar, 1988; Sagular & Toker, 1990; Koçyiğit, 1991; 18. Sirel, 1975; 19–21. Ünalan *et al.* 1976; Toker, 1979.

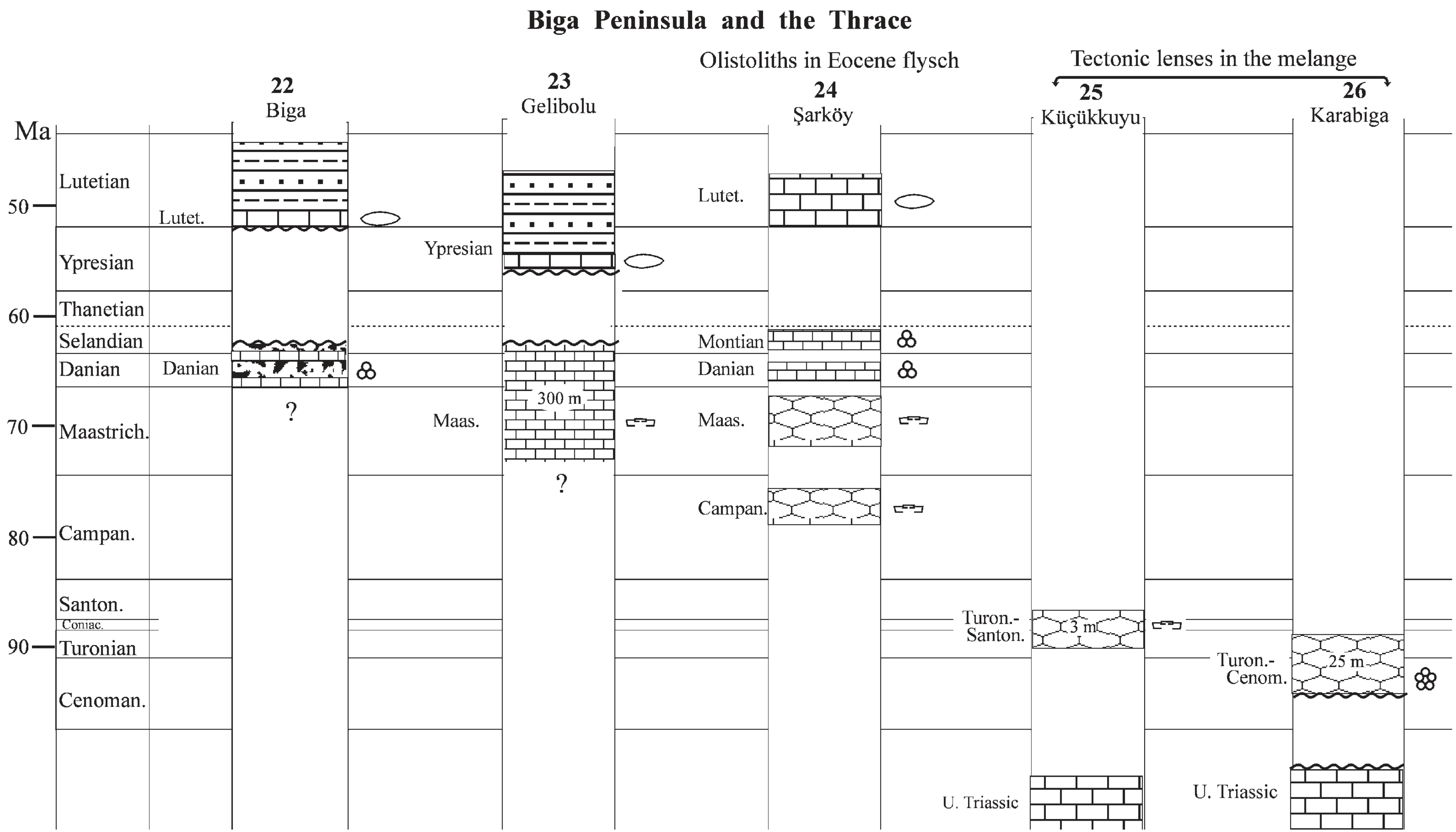

Figure 10. Stratigraphic sections from the western Biga Peninsula and Thrace. For locations of the profiles see Figure 3, and for legend see Figure 6. The data sources are: 22. M. B. Yıkılmaz, unpub. Thesis, İstanbul Technical Univ., 1999; 23. Önal, 1986; 24. Okay & Tansel, 1994; 25. Brinkmann *et al.* 1977; Okay, Siyako & Bürkan, 1991; 26. Okay, Siyako & Bürkan, 1991.

limestones occur as small blocks. In one large block north of Biga, the Upper Triassic limestones are unconformably overlain by Turonian–Cenomanian pelagic micrites (profile 26 in Fig. 10) (Okay, Siyako & Bürkan, 1991). The limestone blocks in the Biga melanges are stratigraphically akin to the blocks in the Bornova Flysch Zone.

Quartz-micaschists with rare lenses and layers of marble and metabasite crop out in the western Biga peninsula (Fig. 3). Some of the metabasites retain eclogitic textures and minerals, indicating an early high-pressure/low-temperature metamorphism which was strongly overprinted by greenschist facies metamorphism. Three muscovite Rb/Sr ages from the quartz-micaschists from the Çamlıca metamorphic rocks range from 69 to 65 Ma, indicating a Maastrichtian age for the regional metamorphism (Okay & Satır, 2000). The western Biga peninsula forms part of a large metamorphic area of the Rhodope-Serbo-Macedonian Massif, and its geological evolution is linked to this massif rather than to the Sakarya Zone.

7. Ophiolite obduction over the Anatolide-Tauride Block

The term obduction refers to the emplacement of oceanic lithosphere above a passive continental margin. In the Tethyan realm ophiolite obduction is generally marked in the sedimentary record by a transition from neritic to pelagic carbonate sedimentation followed by the deposition of siliciclastics. The siliciclastic deposition is related to internal thrust slicing, as well as to the arrival of the ophiolite thrust sheet. The change from neritic to pelagic carbonate sedimentation and the inception of the siliciclastic deposition provide a temporal record of ophiolite obduction. Most of the northern margin of the western Anatolide-Tauride Block is too strongly metamorphosed for any meaningful biostratigraphy. However, two tectonic belts in the Anatolide-Tauride Block, free from regional metamorphism, namely the Bornova Flysch Zone and the Lycian nappes (Fig. 2), provide detailed information on the timing and direction of ophiolite obduction.

7.a. The Bornova Flysch Zone

The Bornova Flysch Zone is a 50–90 km wide zone of chaotically deformed Maastrichtian–Lower Palaeocene greywacke and shale with Mesozoic neritic limestone, and rare mafic volcanic rock, chert, pelagic shale and peridotite blocks (Okay & Siyako, 1993; Okay *et al.* 1996). It forms a northeast trending belt between the Menderes Massif and the İzmir-Ankara suture (Fig. 3). The pre-Palaeocene stratigraphy of the Bornova Flysch Zone consists of Norian to Lower Cretaceous neritic carbonates overlain unconformably by Senonian pelagic limestones (Fig. 11). The Senonian pelagic carbonates must have passed up into a thick flysch sequence, however, most of the present day contacts between the pelagic Senonian limestones and the Bornova flysch are sheared, and the Bornova flysch forms a highly tectonized matrix to the Mesozoic carbonate blocks. Most of the blocks in the Bornova flysch are made up of Norian to Lower Cretaceous neritic carbonates. Continuous neritic carbonate sections, ranging from Norian to Cenomanian and even to Santonian, exist in the Karaburun and Bornova regions (profile 29 in Fig. 11) (Erdoğan, 1990; Erdoğan *et al.* 1990; Robertson & Pickett, 2000). Senonian red biomicrite, calciturbidite and red marly micrite rest unconformably over the neritic limestones. In many carbonate blocks Senonian pelagic limestones lie unconformably over the Upper Triassic carbonates marking the deep erosion of the Mesozoic carbonate platform before the deposition of the pelagic sediments (profiles 27 and 31 in Fig. 11). Abundant *Globotruncana* species in the Senonian biomicrites allow a precise age range to be determined, and the characterization of the unconformity. In the Balıkesir-Soma region the pelagic biomicrites are late Santonian to late Campanian in age (profiles 30 and 31 in Fig. 11), while in the Karaburun-Bornova area they are Campanian–middle Maastrichtian (profile 27) (Brinkmann *et al.* 1977; Şahinci, 1976; Poisson & Şahinci, 1988; Erdoğan, 1990, Erdoğan *et al.* 1990; Okay & Siyako, 1993). The inception of the siliciclastic sedimentation is more difficult to date, as most of the carbonate–siliciclastic contacts are tectonic, and the matrix of the Bornova flysch is generally devoid of fossils. Nevertheless, in the Kocaçay section near Bornova, the sandstones above the Upper Maastrichtian red marls contain Palaeocene limestone blocks, indicating that the Bornova flysch is Palaeocene in age in this locality (Konuk, 1977) (profile 28 in Fig. 11). In the northeast, in the Balıkesir-Soma region, where pelagic limestones appear to be Campanian or older, the Bornova flysch is possibly of Maastrichtian age. Between Balıkesir and Manisa, the Bornova Flysch is unconformably overlain by undeformed shallow marine carbonates of late Early Eocene (late Cuisian) age (profile 32 in Fig. 11), which places an upper limit for deposition and deformation in the Bornova Flysch Zone (Akdeniz, 1980; Okay & Siyako, 1993; Önoğlu, 2000).

7.b. The Lycian nappes

The Lycian nappes occupy a large area between the Menderes Massif and the Beydağları carbonate platform (Fig. 4). They are generally believed to have been derived north of the Menderes Massif from the region of the İzmir-Ankara suture (e.g. Brunn *et al.* 1976). They were emplaced on the Menderes Massif during Eocene time and were translated southeast, probably together with the Menderes Massif, over the Beydağları platform during Miocene time (Fig. 4).

Four main components can be distinguished in the Lycian nappes. These components are, from the base upwards, two thrust sheets of dominantly Mesozoic

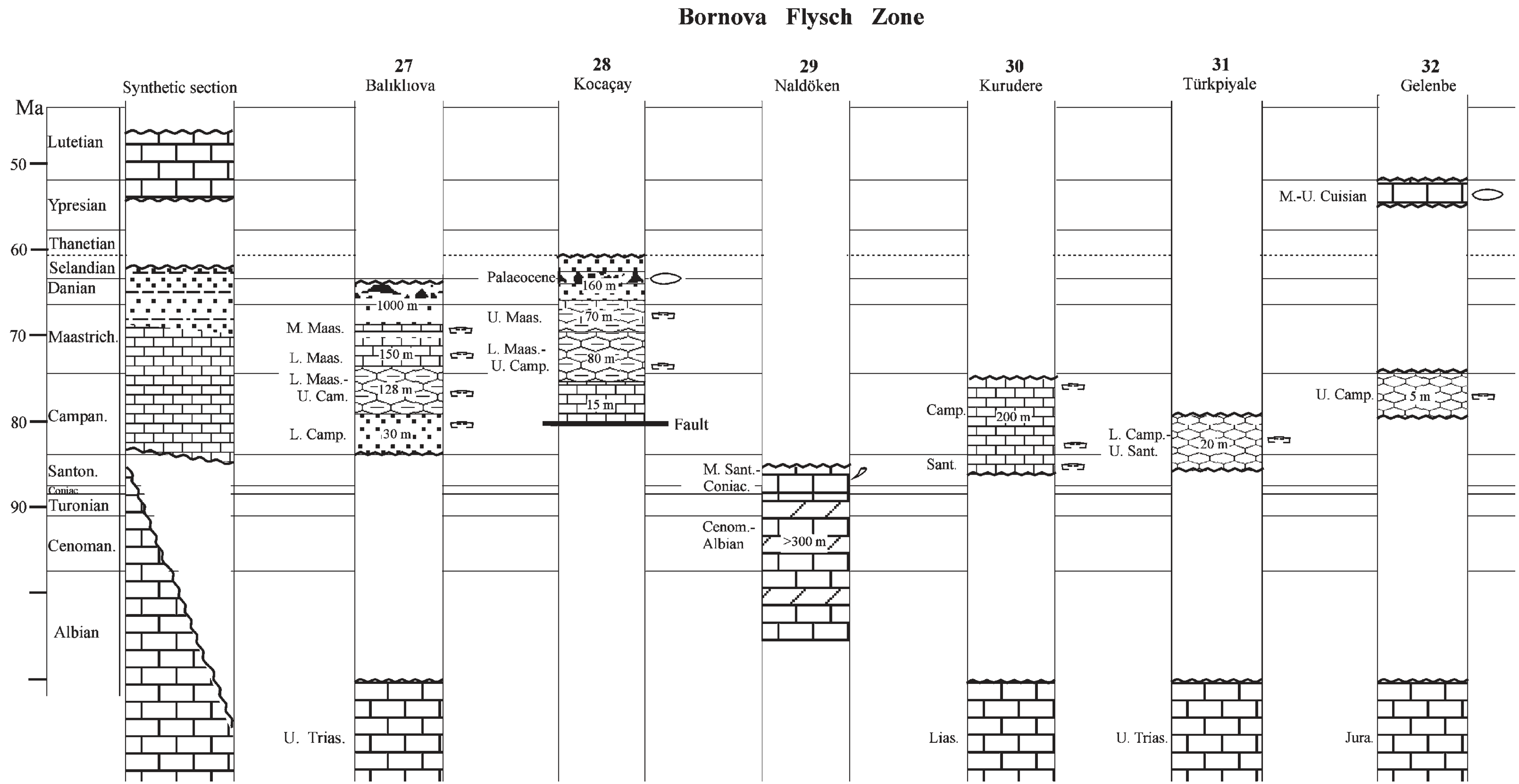

Figure 11. Stratigraphic sections from the Bornova Flysch Zone. For locations of the profiles see Figure 3, and for legend see Figure 6. The data sources are: 27. Brinkmann *et al.* 1977; Tansel, 1990; Erdoğan *et al.* 1990; 28. Konuk, 1977; Erdoğan *et al.* 1990; 29, 30. Erdoğan *et al.* 1990; 31, 32. Okay & Siyako, 1993; Önoğlu, 2000.

continental margin sediments, a melange-type accretionary complex and a peridotite thrust sheet (Fig. 5) (Collins & Robertson, 1998, 1999; Robertson, 2000). The Köyceğiz thrust sheet below the accretionary complex has a Middle Jurassic–Lower Cretaceous sequence, over 1000 m thick, of pelagic micrite and calcareous turbidite. The pelagic limestone sequence extends up to the lowermost Turonian, and is overlain, through a carbonate breccia horizon, by siliciclastic turbidites (profile 33 in Fig. 12) (Graciansky, 1972; Bernoulli, Graciansky & Monod, 1974; Collins & Robertson, 1999). The turbidites contain reworked late Turonian to early Senonian *Globotruncana* and are possibly of Santonian to Campanian age. They pass up into debris flows and olistostromes with limestone, diabase and radiolarian chert blocks. The olistostromes are tectonically overlain by the accretionary complex or by the peridotite thrust sheet. The age of the youngest block in the olistostromes is Late Campanian–Early Maastrichtian suggesting that the emplacement of the ophiolite occurred during the Maastrichtian (Graciansky, 1972; Bernoulli, Graciansky & Monod, 1974).

The siliciclastic sedimentation, recording the tectonic slicing of the Anatolide-Tauride platform, started earlier in the Köyceğiz thrust sheet (?Santonian–Campanian) than in the Bornova Flysch Zone (Maastrichtian–Palaeocene), suggesting a depositional position closer to the Tethys ocean for the Köyceğiz thrust sheet. This is also consistent with the nature of the respective Middle Jurassic–Lower Cretaceous sequences, which are represented by pelagic micrites and calciturbidites in the Köyceğiz thrust sheet and by neritic carbonates in the Bornova Flysch Zone. The Köyceğiz thrust sheet represents sediments of the north-facing passive continental margin of the Anatolide-Tauride Block. In terms of its tectonic setting and initial palaeogeographic position, it shows similarities to the Hawasina nappes in Oman.

The first indication of deformation of the Anatolide-Tauride carbonate platform, most probably related to the ophiolite obduction, was in the Santonian (~86 Ma). This event is recorded by the inception of siliciclastic turbidite deposition in the Köyceğiz thrust sheet, and by the foundering of the carbonate platform in the Bornova Flysch Zone. This foundering of the carbonate platform appears to have been diachronous, ranging from Santonian in the northeast to early Campanian in the southwest. The main deformation, leading to the almost complete dislocation of the Mesozoic carbonate platform in the Bornova Flysch Zone, is constrained to the Palaeocene (66–55 Ma), and in the south possibly to the Late Palaeocene, through the unconformable cover of undeformed Lower Eocene carbonates. Tectonism in the Anatolide-Tauride Block related to ophiolite obduction appears to have been a prolonged event comprising the whole of Senonian and Early Palaeocene times (89–63 Ma) and merges with that associated with continent–continent collision.

The lowermost Lycian nappes have a Middle Jurassic–Cretaceous sequence of pelagic micrite and calciturbidite, similar to that of the Köyceğiz thrust sheet (Collins & Robertson, 1999). However, the pelagic limestones in the lower Lycian nappes continue into the Lower Palaeocene (Şenel, 1991; Özkaya, 1991) and in places into the Lower Eocene section (Okay (1989), and are followed by the Palaeocene–Middle Eocene turbidites, basalts and debris flows (profile 34 and 35 in Fig. 12). The depositional setting of the lower Lycian nappes, whether north of the Menderes Massif (Collins & Robertson, 1998; 1999), or south in a basin between the Menderes Massif and the Beydağları carbonate platform (Poisson, 1984; Okay, 1989; Özkaya, 1991) is controversial.

The Beydağları carbonate platform occurs at the base of the Lycian nappe pile and is regarded as the relative autochthon (Fig. 5). It also crops out underneath the Lycian nappes in the Göcek tectonic window (Fig. 4). Its stratigraphy is known in great detail (Poisson, 1977; Gutnic *et al.* 1979) and essentially consists of Cenomanian neritic limestones overlain by Coniacian to Palaeocene pelagic carbonates. The sequence continues into the Lower Miocene with neritic or pelagic carbonate deposition (profile 38 in Fig. 12). In the Göcek tectonic window the sequence extends without any major breaks from Cenomanian to Lower Miocene (profile 37) and terminates with Burdigalian olistostromes marking the emplacement of the Lycian nappes (Graciansky, 1972; Bernoulli, Graciansky & Monod, 1974).

7.c. Menderes Massif

The Menderes Massif and its northern extension, the Afyon Zone, were regionally metamorphosed during Middle Eocene times. Furthermore, the Mesozoic sequence in northern and central Menderes is absent either due to the Oligocene erosion or through thrust sheet translation during the Middle Eocene. Stratigraphic data for the Cretaceous are therefore scarce and patchy for the Menderes Massif and the Afyon Zone with the Cretaceous–Palaeocene strata mainly preserved along the southern rim of the Massif (Fig. 4). Özer (1998) gives the most precise Cretaceous–Palaeocene stratigraphy of the southern Menderes Massif, on the basis of rudists and foraminifera. The thick carbonate sequence in the Milas region contains Santonian–Campanian rudists and is overlain by red pelagic marbles with Upper Campanian–Maastrichtian foraminifera (profile 36 in Fig. 12). These, in turn, pass up into siliciclastics with possibly reworked Middle Palaeocene pelagic foraminifera. Gutnic *et al.* (1979) report, without giving any details, Early Eocene *Nummulites* from the topmost part of the Menderes Massif. Ar–Ar mica

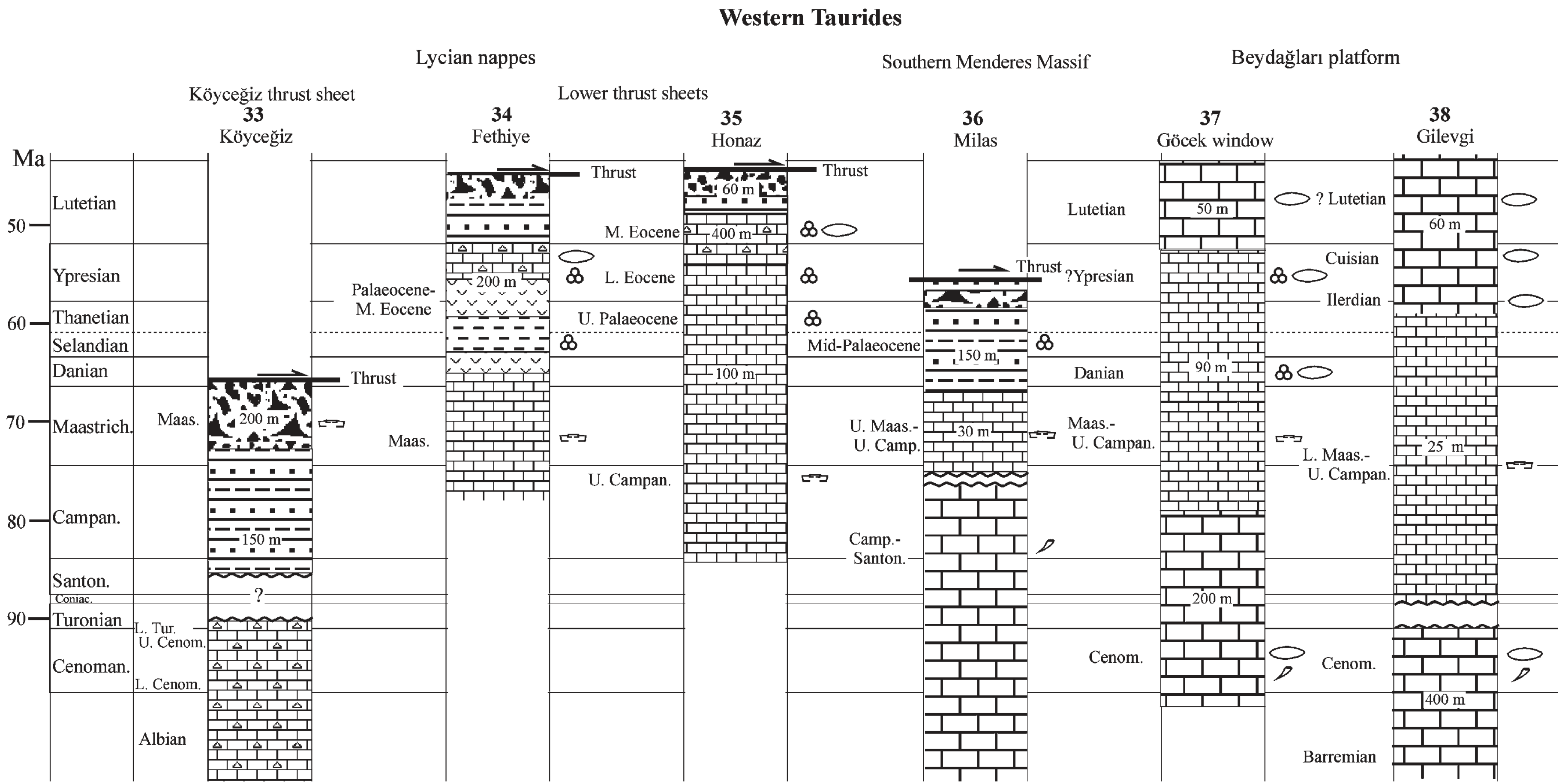

Figure 12. Stratigraphic sections from the western Taurides. For locations of the profiles see Figure 4, and for legend see Figure 6. The data sources are: 33. Graciansky, 1972; Bernoulli, Graciansky & Monod, 1974; 34. Şenel, 1991; Özkaya, 1991; 35. Okay, 1989; 36. Özer, 1998; 37. Graciansky, 1972; Bernoulli, Graciansky & Monod, 1974; 38. Poisson, 1977; Gutnic *et al.* 1979.

isotopic ages from the southern Menderes Massif are 43 to 37 Ma, indicating Middle Eocene regional metamorphism (Hetzel & Reischmann, 1996). Blueschists and eclogites with Middle Eocene Ar/Ar phengite ages of 40 Ma (Oberhänsli *et al.* 1998) occur as a thrust sheet above the Menderes Massif (Fig. 4). They form part of the Cycladic metamorphic complex and their stratigraphy appears to be similar to that of the Menderes Massif (Özer, 1998).

8. Blueschist metamorphism and exhumation

Blueschists form a large coherent belt, about 250 km long and 50 km wide, along the northern margin of the Anatolide-Tauride Block south of the İzmir-Ankara suture (Fig. 3). The protoliths of the blueschists are sandstone, siltstone, shale and limestone, as well as mafic volcanic rocks, and are believed to represent a Palaeozoic to Mesozoic sequence of the Anatolide-Tauride Block. The blueschists are tectonically overlain by an accretionary complex or directly by large slabs of ultramafic rock (Okay, 1984). The widespread occurrence of glaucophane, lawsonite and jadeite in the blueschists allows a precise estimation of the peak metamorphic conditions as 20 kbar and 440 °C, indicating that the northern margin of the Anatolide-Tauride block was subducted to depths of about 60 km during the blueschist metamorphism. Rb–Sr and Ar–Ar isotopic data constrain the age of the peak metamorphism to 80 ± 5 Ma (Campanian: Okay & Kelley, 1994; Sherlock *et al.* 1999). The blueschists and the tectonically overlying accretionary complex and ophiolite are intruded by several granodiorite plutons with Early to Middle Eocene Ar–Ar isotopic ages (53 to 48 Ma: Harris, Kelley & Okay, 1994; Sherlock *et al.* 1999).

9. Continent–continent collision

Continent–continent collision is initiated when the oceanic lithosphere between two opposing continents is eliminated, and the opposing continental crusts come into contact. During progressive convergence, the deep oceanic basin between the two opposing continents is gradually filled by sediments and is transformed from a pelagic to a continental basin followed by strong shortening. In most orogenic belts continent–continent collisions are diachronous and prolonged events. In the Himalaya, shortening related to the Eocene collision of India with Asia is still continuing. In the geological record the most reliable criteria for dating the initiation of the continent–continent collision will be the termination of the subduction-related magmatism, and the start of the contractional deformation of the lower slab. However, in the Tethyan realm deformation of the continental lower slab is not always a good criterion for dating the collision, since in most cases the lower continental slab already has been strongly deformed and metamorphosed during the earlier ophiolite obduction, as is the case for the Campanian deformation and metamorphism in the Anatolide-Tauride block. The upper plate can, however, also be affected by compressive deformation prior to continental collision, as is the case with the Cenomanian–Turonian deformation in the central and eastern Pontides. Only an evaluation of the magmatic, metamorphic, structural and stratigraphic data from both the lower and upper plate can place constraints on the timing of the continental collision.

An upper bound on the continental collision can be inferred from the lower Lycian nappes, where the change from deep shelf to foredeep sedimentation, and arrival of allochthonous detrital clasts took place in the Middle Eocene. The Eocene emplacement of the Lycian nappes over the Menderes Massif and the subsequent Middle Eocene Barrovian metamorphism and deformation of the Menderes Massif are most probably a direct consequence of the continental collision between the Anatolide-Tauride Block and the Pontides. In the more internal parts of the orogen, the deformation related to the continental collision could have started earlier, however, it is difficult to separate this deformation from that caused by the Campanian–Maastrichtian ophiolite obduction. In the Bornova Flysch Zone the deformation occurred during Palaeocene time, however, it is not clear whether this deformation is a result of the last stages of ophiolite obduction or due to continent–continent collision. In the most internal part of the orogen, granodiorite plutons were emplaced into the blueschists at depths of about 10 km during Early to Middle Eocene times (48 to 53 Ma: Harris, Kelley & Okay 1994; Sherlock *et al.* 1999). These post-tectonic plutons occur immediately south of the İzmir-Ankara suture and are spread over a distance of 250 km (Fig. 3). Their restriction to the blueschist belt suggests that they were generated through the upwelling of the asthenosphere following the rapid isostatic rebound of the subducted continental slab (Okay, Harris & Kelley, 1998). Their undeformed post-tectonic nature suggests that deformation related to the continent–continent collision most probably occurred prior to Early Eocene time in the internal parts of the orogen.

During Early Palaeocene time there was a transition from deep sea turbidites to redbeds in the fore-arc basins of Haymana and Central Sakarya of the Pontides, especially in the regions near the İzmir-Ankara suture (Fig. 9). In the Sakarya basin the Palaeocene continental sedimentation migrated northward and was followed by south-vergent folding and thrusting during Palaeocene–Early Eocene times. However, sedimentation continued in the Haymana basin without a major break until early Middle Eocene time.

The stratigraphic and structural evidence constrains the continental collision between the Anatolide-Tauride Block and the Pontides to the Palaeocene–Early Eocene interval. One line of evidence supporting initiation of the collision during the Early Palaeocene, comes from the relative motion of Africa and Laurasia in the vicinity of Turkey. This motion shows a sudden decrease between the latest Cretaceous and Early Eocene (70 to 48 Ma) (Livermore & Smith, 1984), which can be attributed to the jamming of the subduction zone by the continental crust. However, there is no distinct break between the deformation associated with the ophiolite obduction and continent–continent collision.

The Pontide arc magmatism, which can potentially constrain the timing of continental collision, ceased by the end of Campanian time, and evidence for arc magmatism during the Maastrichtian and Palaeocene is missing (Fig. 6). However, as discussed above, there is nothing in the sedimentary record of the Anatolide-Tauride Block or that of the Pontides to suggest continent–continent collision during the Maastrichtian age. The end of arc magmatism coincides with that of the blueschist metamorphism in the northern margin of the Anatolide-Tauride Block. Thus, the termination of the arc magmatism is probably related to the choking of the subduction zone by the continental crust (Fig. 14d).

Initiation of the arc magmatism (Turonian, 91–88 Ma), sub-ophiolite metamorphism (95–90 Ma), and the inferred age of the ophiolite in western Anatolia (Early Cretaceous–Cenomanian) are closely spaced. This may be explained if the subduction in the İzmir-Ankara ocean was initiated through the ridge subduction during the Cenomanian (Fig. 14a,b). This has already been suggested for the origin of the sub-ophiolite metamorphic rocks and diabase dykes in the ultramafic portions of the ophiolite (Lytwyn & Casey, 1995; Polat, Casey & Kerrich, 1996; Dilek *et al.* 1999). The sub-ophiolite metamorphic rocks would have been produced during the early stages of ridge subduction closely followed by arc magmatism when the subducting oceanic lithosphere reached a depth of 80 km. The arc magmatism was terminated by the end of Campanian time, following the complete subduction of the oceanic lithosphere south of the former ridge (Fig. 14d). The last vestige of the oceanic lithosphere would have taken the northern margin of the Anatolide-Tauride Block into the subduction zone leading to continental subduction and high-pressure/low-temperature metamorphism during the Campanian.

The jamming of the subduction zone by the Anatolide-Tauride Block possibly led to the establishment of a second subduction zone further north along the continent–ocean boundary, which consumed the northern half of the İzmir-Ankara ocean (Fig. 14e). The common presence of blueschist detritus in the Maastrichtian fore-arc sequences suggests that detritus from the accretionary complexes and blueschists along the southern margin of the İzmir-Ankara ocean reached the northern margin. The narrow oceanic lithospheric area in place during Maastrichtian time may also explain the lack of arc magmatism during this period. An alternative is that the oceanic lithosphere was detached and overrode the Anatolide-Tauride Block without the creation of a second subduction zone. The latter model would provide an explanation for the continuous nature of obduction- and subduction-related deformation.

10. Conclusions

Imprints of four major orogenic events, subduction, obduction, collision and regional metamorphism, were studied in the Upper Cretaceous–Lower Eocene stratigraphic record of western Turkey. The most straightforward and internally consistent stratigraphic record is that of arc magmatism. It shows that the arc volcanism started by the end of Turonian time along the west Black Sea coast and was terminated by the end of Campanian time (Fig. 13), when it was replaced by the deposition of marly limestones from early Maastrichtian up to the end of Palaeocene times.

The sedimentary record of continental subduction, high-pressure/low-temperature metamorphism, obduction and continent–continent collision in the Anatolide-Tauride Block are closely related. Carbonate deposition at the northern continental margin of the Anatolide-Tauride Block continued at least up to Turonian time. The long-standing Triassic–Lower Cretaceous carbonate platform of the Anatolide-Tauride Block started to subside by late Santonian time, and was the site of pelagic carbonate sedimentation throughout the Campanian age (84–74 Ma). The Campanian (80 ± 5 Ma) was also the period of blueschist metamorphism of the northern margin of the Anatolide-Tauride Block (Fig. 14d). During Campanian time the northern margin of the Anatolide-Tauride Block must have been passively pulled down a subduction zone by its oceanic lithospheric root. This appears initially not to have caused any internal deformation in the Anatolide-Tauride Block, apart from regional uplift and erosion at the start of the continental subduction, and deposition of siliciclastic turbidites in the Lycian nappes. The Campanian interval in the Anatolide-Tauride carbonate platform is generally characterized by continuous pelagic carbonate deposition, continuing in many places into Maastrichtian time (Fig. 14d,e). The internal deformation of the Anatolide-Tauride Block, marked by the fragmentation of the carbonate platform and flysch deposition, started in latest Maastrichtian and Palaeocene times as a result of obduction. Thus, during the initial period of continental subduction (Campanian–Maastrichtian, ~85–70 Ma), the northern margin of the Anatolide-Tauride Block was deeply

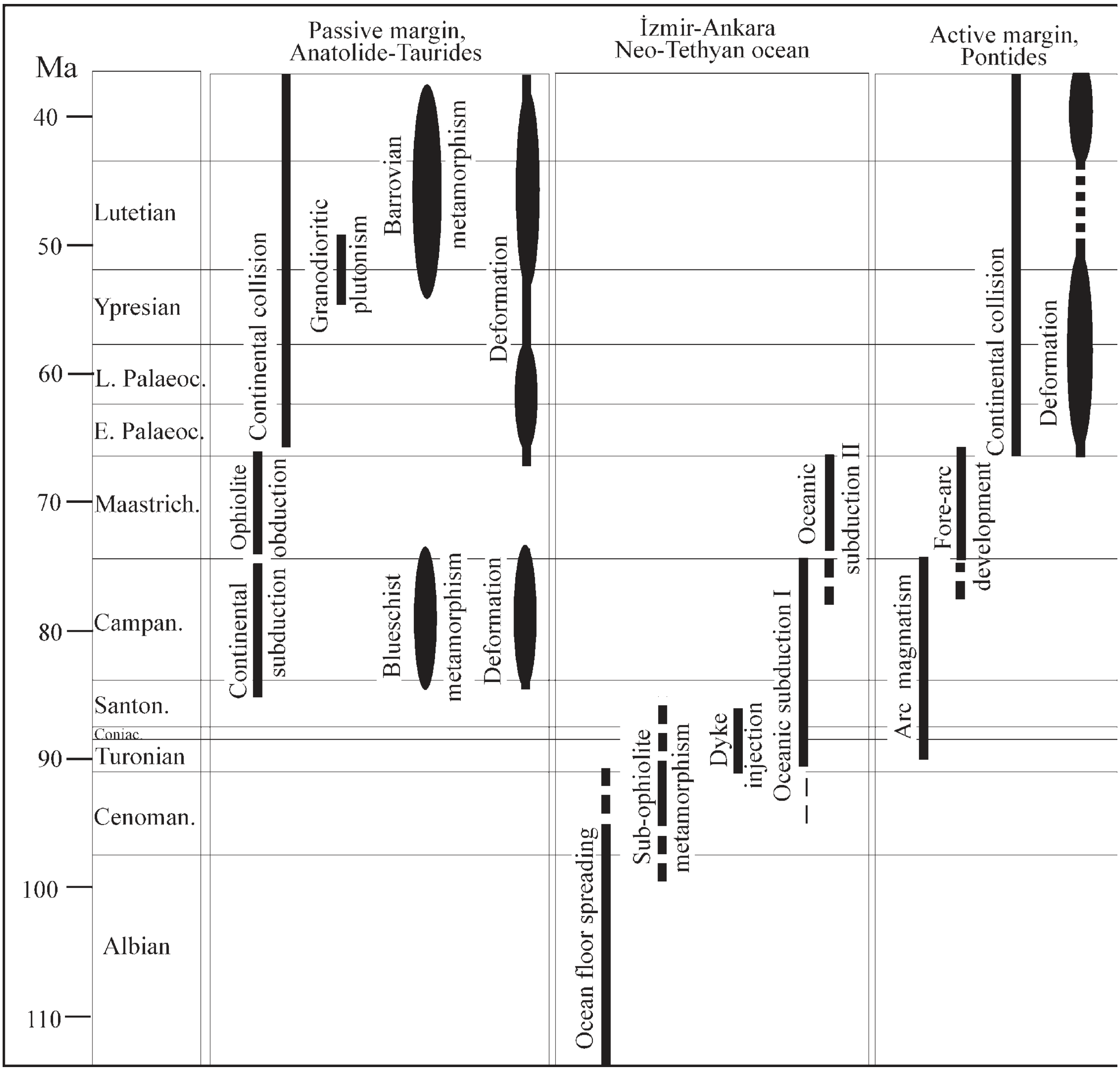

Figure 13. Timing of orogenic events in western Turkey as deduced from structural, isotopic data, and from the biostratigraphy of the Upper Cretaceous–Lower Eocene sequences.

buried, deformed and metamorphosed but with little evidence of near-surface deformation. This was followed by a second period of obduction *sensu stricto* during latest Maastrichtian–Palaeocene times (~70–60 Ma) marked by flysch deposition and large-scale deformation of the Anatolide-Tauride Block. Continental collision was probably initiated during the Palaeocene. However, the main deformation of the Anatolide-Tauride Block, including regional metamorphism of the Menderes Massif, occurred in the Middle Eocene. The deformation continued until the Early Miocene in the external parts of the Taurides. There is no clear temporal or spatial limit between deformation associated with the obduction (latest Maastrichtian–Palaeocene) and that with the continent–continent collision (Palaeocene–Early Miocene) in western Anatolia.

The relatively abundant and precise biostratigraphic data for the Late Cretaceous and Palaeocene in central northwest Turkey allows a standard plate tectonic interpretation of the orogenic events for this period. In contrast, for the region west of Bursa the Upper Cretaceous–Palaeocene stratigraphic data are scarce and difficult to interpret. In particular, the marked divergence of the Senonian arc front and the İzmir-Ankara suture in this region (Fig. 3) suggests important but poorly documented strike-slip movements after Campanian time. Finally, it is worth noting that although the Black Sea coastal region has a large number of well-exposed Maastrichtian–Danian pelagic

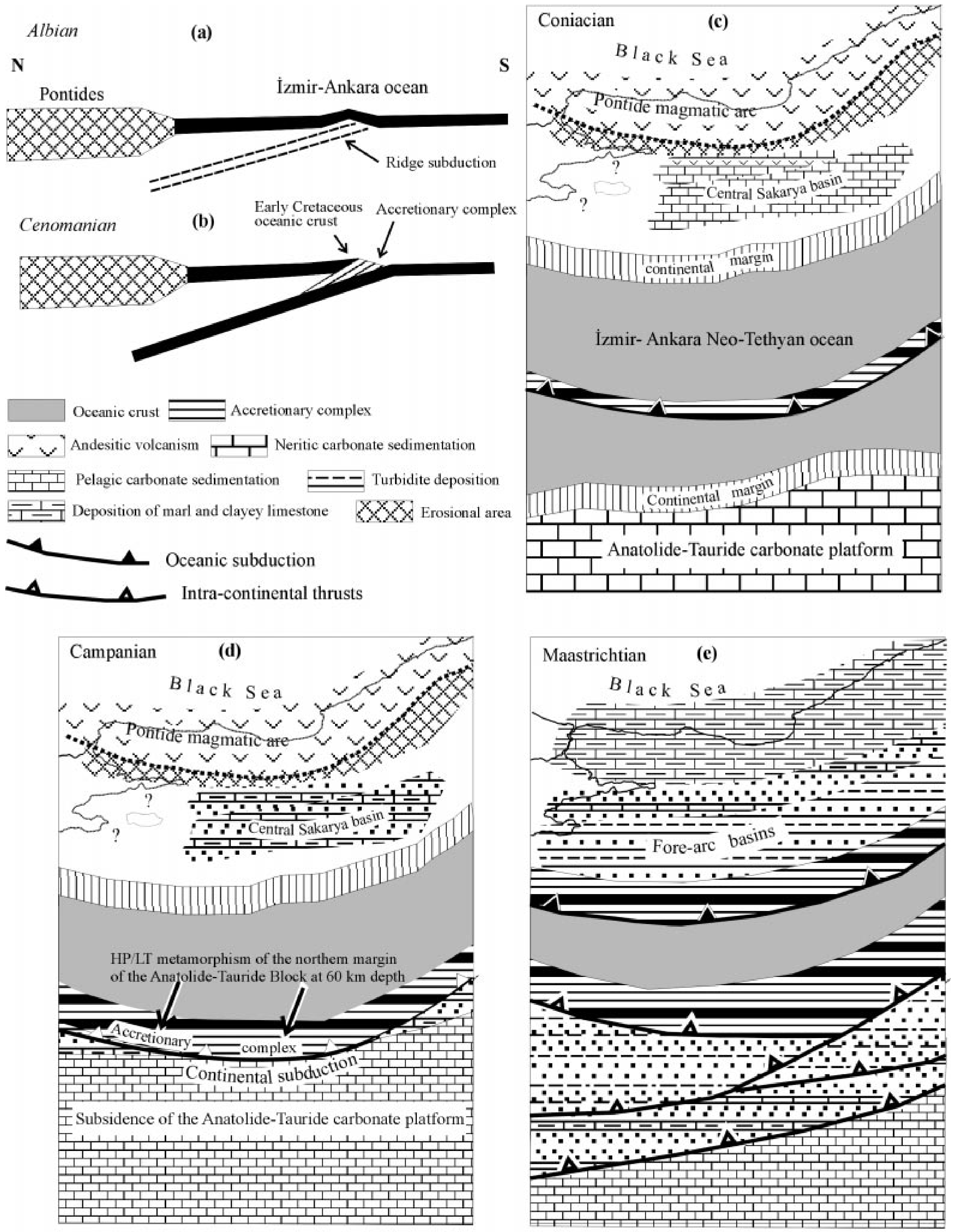

Figure 14. Palaeogeographic and tectonic evolution of northwest Turkey during the Late Cretaceous. Ridge subduction starts in the Cenomanian–Turonian (b); the subduction gives rise to a magmatic arc in the late Turonian and Coniacian (c). A possible subareal ridge south of the arc prevents the volcanic detritus reaching the Central Sakarya basin. In the Campanian the Anatolide-Tauride Block collides with the subduction zone, and its leading edge is metamorphosed in blueschist facies at a depth of 60 km (d). The growing accretionary wedge above the subduction zone supplies clastic detritus to the Central Sakarya basin. The jamming of the subduction zone by continental crust eventually leads to the establishment of a second subduction zone further north (e), which consumes the remaining Neo-Tethyan oceanic lithosphere north of the former ridge.

carbonate sections (Fig. 6), there have been very few studies dealing with the Cretaceous–Tertiary boundary event in Turkey (Matsumaru *et al.* 1997).

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