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A fore-arc origin for the Thrace Basin, NW Turkey

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Abstract Hydrocarbon-bearing Thrace Basin occupies much of the European part of Turkey. The Middle Eocene to Oligocene sequence in the centre of the basin exceeds 9 km in thickness. Based on the stratigraphy, structure and the regional context of this basin, we propose that it developed as a fore-arc basin between the medial Eocene and the Oligocene, above the northward subducting Intra-Pontide Ocean. Its post-Miocene history has been dominated mainly by wrench tectonics resulting from the activity of the now-deactivated northwestern strand of the present-day North Anatolian fault zone.

Key words Fore-arc · Medial Eocene · Strike-slip · Thrace Basin · Turkey

Introduction

The Thrace Basin is a large Tertiary depression located between the Strandja massif in the north and the Biga Peninsula in the south (Fig. 1). It constitutes one of the most important hydrocarbon provinces of Turkey, with an area of more than 15 000 sq km, favourable particularly for natural-gas exploration. Its infill comprises sedimentary rocks of Middle Eocene (Lutetian) to Recent age, the cumulative thickness of which has been estimated to be more than 9 km (Turgut et al. 1991). Most of this infill is Middle Eocene to Upper Oligocene in age (Fig. 2). The basin is flanked predominantly by Triassic to Lower Tertiary rocks and for the most part overlies the Triassic to Jurassic metamorphic rocks of the Strandja massif in the north (Turgut et al. 1991; Perinçek 1991). At the surface, outcrop of the infill and its basement is limited, due to extensive Pliocene to Recent cover, but numerous boreholes throughout the basin (see Fig. 3) provide insight into the subsurface relationships of the Tertiary strata.

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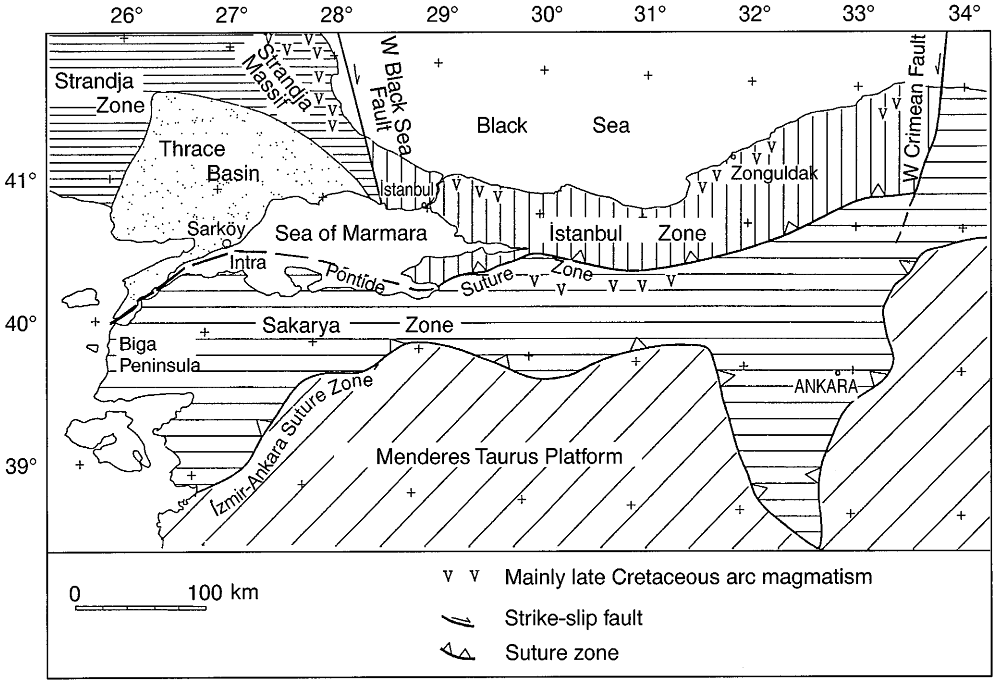
Because of its hydrocarbon content, the Thrace Basin has attracted interest particularly since the late 1950s. Due to the economical emphasis, in most of the published studies, particular attention was concentrated on stratigraphy, palaeontology, structure and reservoir characteristics (e.g. Doust and Arıkan 1974; Burke and Uğurtaş 1974; Keskin 1974; Ketin et al. 1981; Turgut et al. 1983, 1991; Perinçek 1991), but little has been published on the tectonic setting and possible evolutionary models for the basin. To date, no satisfactory model for the origin and evolution of the basin has been proposed. Because it is located mostly within the Strandja zone (Fig. 1) and formed during or soon after the final elimination of most of the Neo-Tethyan oceanic domains in Turkey, most researchers have mistakenly considered this basin to have been "intermontane" or "collision-related" in nature (Keskin 1974; Turgut and Atalık 1988; Yılmaz 1988; Perinçek 1991; Turgut et al. 1991). Contrary to this common opinion, this paper provides evidence that it evolved during the Palaeogene as a fore-arc basin, linked to the subduction of the Intra-Pontide Ocean (Fig. 1).

Geological setting

Structural relationships between the Thrace Basin and the surrounding tectonic provinces suggest that this basin was in a fore-arc position during the medial Eocene to Oligocene time. It was located between a magmatic arc (the Pontide magmatic arc; Letouzey et al. 1977; Şengör and Yılmaz 1981) to the north and its subduction–accretion complex to the south. Part of this subduction–accretion complex now crops out along the Intra-Pontide suture zone and constitutes its main marker (Figs. 1 and 4; Şengör and Yılmaz 1981; Görür et al. 1984). The magmatic arc in the Strandja massif is represented by granodiorites (the Demirköy pluton; Aykol 1979), andesites, tuffs, agglomerates and pyroclastic rocks of mainly Upper Cretaceous age (Pamir and Baykal 1947; O. Sungurlu, unpublished; Taner and Çağatay

Fig. 1 Tectonic setting of the Thrace Basin



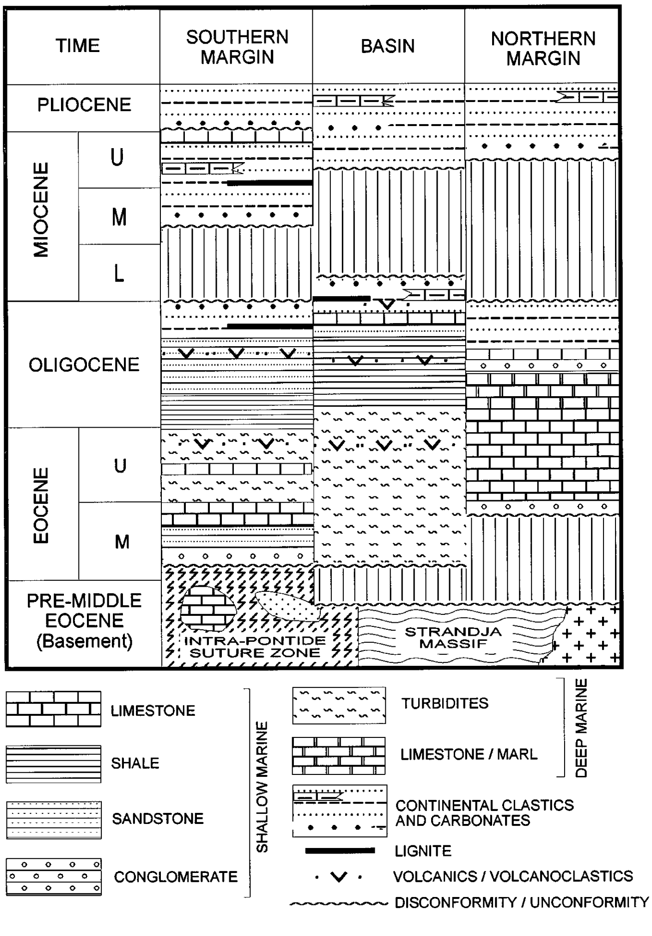
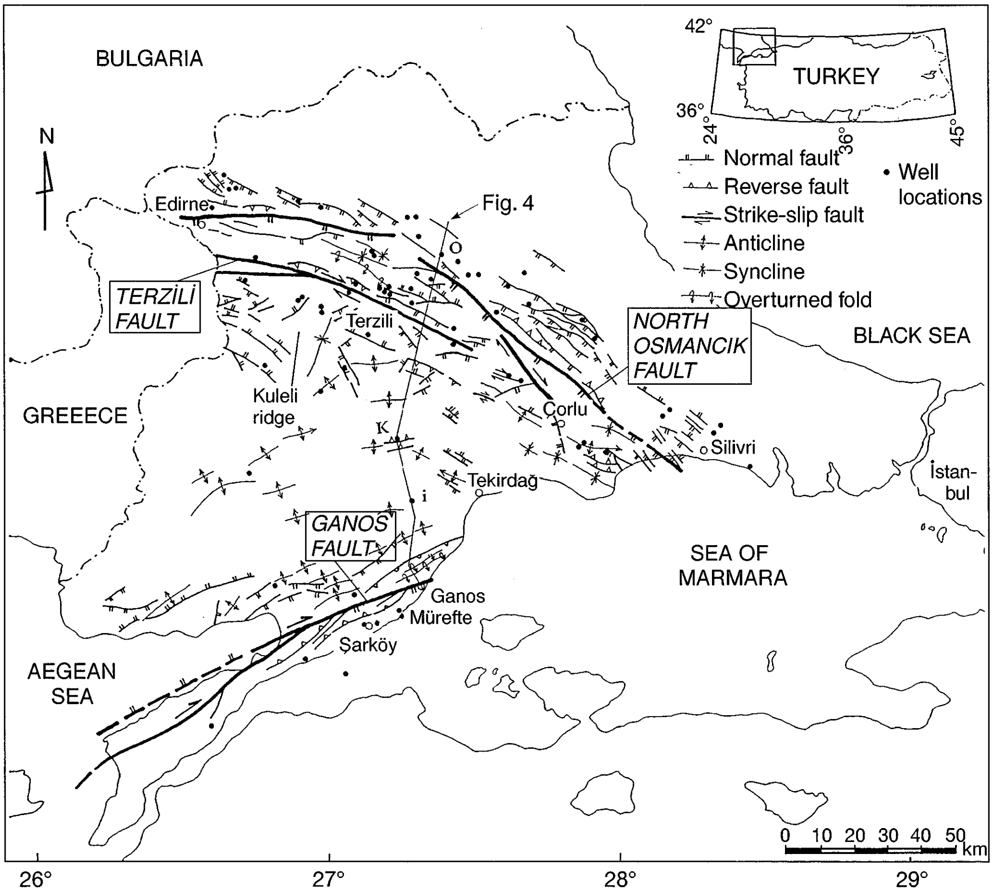


Fig. 2 Generalized stratigraphy of the Thrace Basin (compiled from Doust and Arıkan 1974; Keskin 1974; Turgut et al. 1991)

1983; Aykol and Tokel 1991). Magmatic rocks of Eocene age consist mainly of andesitic and dacitic tuffs (Doust and Arıkan 1974; Turgut et al. 1991; Perinçek 1991) and are mostly encountered within the basin, suggesting that by the Eocene time the locus of the magmatism may have migrated to the south. The subduction complex, forming the southern frame of the basin, is represented by an ophiolitic mélange with included blueschist blocks (Şentürk and Okay 1984). It crops out north and east of Şarköy in the cores of small post-Miocene anticlines and is encountered below the Middle Miocene limestones in the petroleum wells drilled in the Sea of Marmara (Figs. 1–4; Kopp et al. 1969; Şentürk and Okay 1984; Siyako et al. 1989; Okay and Tansel 1992). Şengör and Yılmaz (1981) interpreted the ophiolitic mélange as marking the suture of an ocean (the Intra-Pontide Ocean) located between the Strandja zone in the north and the Sakarya zone in the south (Fig. 1). The main evidence for this suture comes from the east of the Sea of Marmara where it separates the İstanbul and the Sakarya zones with entirely different basement and cover stratigraphies along a zone of intensively sheared ophiolites, showing evidence of an original southerly vergence (Şengör and Yılmaz 1981; Yılmaz 1990). The İstanbul zone carries a late Cretaceous through early Cenozoic island-arc association betraying the duration of ocean elimination along the suture zone. There are very few reliable observations on the age of opening and closing of this ocean. On the basis of the sparse data on early Jurassic facies changes, Şengör and Yılmaz (1981) and Görür et al. (1984) suggested that it began opening in the Lias and closed during late Cretaceous to Palaeocene time. However, the discovery of Middle Palaeocene pelagic limestone blocks along with blocks of serpentinite and radiolarian

Fig. 3 Structural map of the Thrace Basin (compiled from Doust and Arıkan 1974, Turgut et al. 1991; Yaltırak 1995). *O* Osmancık-1 well; *K* Kandamış-1 well; *İ* İncik-1 well



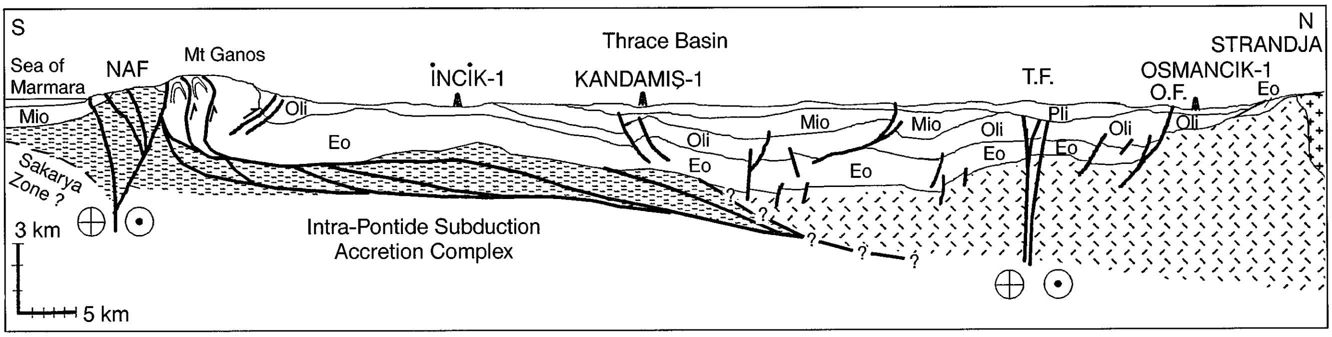


Fig. 4 A geological cross section showing structural relationships between the Thrace Basin and the surrounding tectonic provinces. For location see Fig. 3. *O.F.* North Osmancık fault; *T.F.* Terzili fault; *NAF* North Anatolian fault; *Eo* Eocene; *Oli* Oligocene; *Mio* Miocene; *Pli* Pliocene. (Modified from Doust and Arıkan 1974)

chert in the infill of the Thrace Basin, north of Şarköy, indicates that this ocean persisted at least into the Palaeocene time (Okay and Tansel 1992).

Stratigraphy

Due to numerous exploratory wells and large amounts of both geological and geophysical studies, the stratigraphy of the Thrace Basin is well known (Fig. 2). Therefore, it is described only briefly herein. The interested reader is referred to Doust and Arıkan (1974), Keskin (1974), Ataman and Gökçen (1981) and Turgut et al. (1991) for more detailed descriptions.

Infill of the Thrace Basin is dominated by a shallowing-upward, dominantly clastic succession of Middle Eocene to Oligocene age (Fig. 2). The medial Eocene to early Oligocene sedimentation within the basin was characterized by locally tuffaceous turbidites, whereas continental to shallow-marine clastics and carbonates with subordinate volcanoclastics were laid down both along the basin margins and on the bathymetric highs (e.g. the Kuleli ridge) projecting into the basin interior (Figs. 2 and 3; Keskin 1974; Doust and Arıkan 1974; Ataman and Gökçen 1981; V. Ediger, unpublished;

Turgut et al. 1991; Perinçek 1991). Post-early-Oligocene sedimentation was mainly marginal marine to terrestrial represented by coal-bearing clastics and carbonates in part with some tuffaceous material (Doust and Arıkan 1974; Turgut et al. 1991).

The succession in the central part of the basin starts with turbidites of Middle Eocene to Lower Oligocene age (the Keşan formation; Doust and Arıkan 1974) deposited unconformably on the metamorphic and crystalline rocks of the Strandja massif in the north and the ophiolitic mélange of the Intra-Pontide Suture in the south (Figs. 1 and 2). Both of these features formed the principal provenance, contributing detritus to the basin throughout its depositional history. The turbidites are composed mainly of more than 4000-m-thick interbedded sandstones and shales with subordinate pelagic carbonates and conglomeratic intervals, largely confined to proximal sites near the basin margins. The sandstones are mostly fine-grained and poorly sorted arenites to wackes with abundant angular to subrounded grains of metamorphic and magmatic rocks, quartz, feldspars, micas and various minerals of volcanic and metamorphic origin (Doust and Arıkan 1974; Ataman and Gökçen 1981). The interbedded shales consist mainly of grey to black calcareous siltstones and claystones, rich in mica and carbonaceous material. Upward in the section, the turbidites contain pyroclastic material and ashfall tuffs of andesitic, dacitic and rhyolitic composition and pass into 450- to 1000-m-thick Upper Oligocene shales with subordinate rippled sandstones and argillaceous limestones (the Muhacir formation; Holmes 1961). The shales are grey to green, calcareous and micaceous with well-developed lamination and common shallow-water fossils including fish remains, gastropods, bivalves, ostracods and diatoms (Doust and Arıkan 1974). These marginal marine sediments are overlain by the uppermost Oligocene non-marine strata of fining-upward, fine- to medium-grained, and poorly sorted sandstones and calcareous shales with conglomeratic and tuffaceous interlayers (the Danişment formation, NV Turkse Shell, unpublished). Lignite seams and thin dolomite bands are encountered in the calcareous shales. Ostracods, fish and plant remains form the main fossil assemblage of these non-marine sediments. Unconformably above them are the fluvio-lacustrine deposits of the Upper Miocene to Quaternary age (the Ergene formation; A. W. Holmes, unpublished) These deposits consist mainly of loose to poorly cemented pebbles, sands, pebbly sands, shales, marls and chalky limestones. Their thickness is variable, ranging from 500 to 1700 m (Doust and Arıkan 1974).

The basin margin succession is a thinner one (rarely exceeding 2500 m) and is characterized by shallow- to deep-marine clastics and reefal carbonates (Doust and Arıkan 1974; Turgut et al. 1991). Along the southern periphery of the basin, the succession seems to have developed on the outer non-volcanic arc associated with the north-dipping Intra-Pontide subduction zone (Doust and Arıkan 1974; Şentürk and Okay 1984; Okay and Tansel 1992). Here, the succession begins with Middle Eocene coarse conglomerates, sandstones and shales overlain by coralgal reef limestones (Fig. 2). Towards the top, the reef limestones pass, through the Upper Eocene to Lower Oligocene calcareous shales and turbiditic sandstones, into coal-bearing marginal marine to continental clastic deposits of the Upper Oligocene age. The succession continues upward across an angular unconformity with Middle to Upper Miocene terrigeneous to shallow-marine rocks (the Gazhanedere and the Kirazlı formations; Turgut et al. 1991). On the northern flank of the basin, the marginal sequence (the Saray group; Doust and Arıkan 1974) displays similar stratigraphic and lithological characteristics with abundant carbonate facies, although deposition of the sediments here seems to have started a bit later in the Eocene (Fig. 2; Doust and Arıkan 1974).

Internal structure

Due to the little-deformed extensive Plio-Quaternary sedimentary cover, structural outcrop data from the Thrace Basin are very limited. Most of our knowledge on basin structure comes mainly from seismic studies and borehole data. These studies reveal that the internal structure of the basin consists of numerous folds and faults, trending more or less parallel with the basin margins (Figs. 3 and 4; Doust and Arıkan 1974; Burke and Uğurtaş 1974; Turgut et al. 1991; Perinçek 1991). They trend NW–SE in the north and SW–NE in the south, thus converging to the east somewhere in the Sea of Marmara (Fig. 3).

Broadly open folds of various sizes occur throughout the basin. They are mostly broad linear features and appear to be symmetrical in most parts of the basin, although towards the margins, they become asymmetrical or overturned to the south (Figs. 3 and 4). South-vergent structures are particularly common along the southern basin margin defined by the subduction complex of the Intra-Pontide suture. These folds have affected most of the basin fill, but mostly did not involve the basement (Doust and Arıkan 1974; Burke and Uğurtaş 1974; Turgut et al. 1991).

Although faults of various kinds, sizes and ages occur throughout the Thrace Basin, they seem to be concentrated along the basin marginal areas (Figs. 3 and 4). Two types of faults are common throughout the basin: normal and strike-slip. Normal faults predominate over the others and have various amounts of vertical displacement. They may be confined entirely to the basin fill or affect the basement as well. Some of them seem to be associated with the folding and are found disrupting the crests of anticlines and the flanks of synclines. Where the folds are asymmetrical or overturned, the associated faults are dominantly reverse or thrust in nature. Reverse and thrust faults are common in the Ganos region on the southern margin and in the Çorlu-Terzili belt on the northern margin (Figs. 3 and 4;

Doust and Arıkan 1974). To the north of the Çorlu-Terzili belt, normal faults affect the entire Tertiary succession and form in places graben structures of various sizes (Doust and Arıkan 1974). The largest and the most important deep-seated normal fault of these northern areas (the North Osmancık fault; Turgut et al. 1991) is present between Silivri and Edirne where it forms a distinct zone parallel to the tectonic strike (Figs. 3 and 4). This zone probably coincides with the shelf brake and may represent, therefore, a significant boundary–fault zone in the basement.

Two major dextral strike-slip fault zones exist along both the northern and the southern margins of the Thrace Basin (Figs. 3 and 4; Erkal 1987; Perinçek 1991; Turgut et al. 1991). The northern one (the Terzili fault; Turgut et al. 1991) is located to the south of the North Osmancık boundary–fault zone. It stretches from the Greek frontier down to the Sea of Marmara with local development of positive flower structures along its course (Turgut et al. 1991). The southern strike-slip fault zone (the Ganos fault zone) is a segment of the dextral North Anatolian fault zone and is still active (Figs. 3 and 4). It has played an important role since the late Miocene to Pliocene in the neotectonic evolution of the Thrace Basin, causing intense uplift and deformation particularly on its southern margin (Şengör et al. 1985; Turgut et al. 1991).

The structures in the Thrace Basin were created during various episodes of the development of the basin. Most of the folds seem to have formed during late Oligocene to early Miocene time by syn- and post-collisional compression of the basin, because they appear to have affected most of the infill, except the post-Middle Miocene sediments. Deformation of the younger sediments seems to be related to the North Anatolian fault zone. Generation of the faults probably started along with the basin formation in the medial Eocene and therefore have a wide age span. The deep-seated normal fault zone along the northern margin of the basin perhaps formed in the medial Eocene and functioned throughout the pre-collisional Eocene period. Origin of the strike-slip faults of both the basin margins seems to be related to the present westward escape tectonic régime of Anatolia. They were probably created during the late Miocene to Pliocene when this régime initiated (Dewey and Şengör 1979; Şengör et al. 1985; Perinçek 1991; Taymaz et al. 1991). However, the southern strike-slip fault may have been a replacement structure; it may have originated during the Eocene evolution of the basin as a normal boundary fault and then may have been replaced by the North Anatolian strike-slip fault in the late Miocene to Pliocene.

Tectonic model

We suggest that the Thrace Basin developed as a fore-arc basin between medial Eocene and Oligocene time. The main evidence for this suggestion include (see Dickinson 1995):

1. The basin is flanked on one side by the Pontide magmatic arc and on the other side by the subduction complex of the Intra-Pontide suture zone (Fig. 4).
2. The basin fill typically shoals upward from a turbidite association to a marginal marine to terrestrial sequence (Fig. 2).
3. Petrographic and sedimentological characteristics of this fill record unroofing and dissection of both the adjacent arc massif and the subduction complex.

Details of the tectonic evolution of this fore-arc basin are as follows: At the beginning of Eocene time, the Intra-Pontide Ocean was still open, although its eastern part may have been reduced to a narrow canal, due to the approaching İstanbul zone from the north. According to Okay et al. (1994) the İstanbul zone was initially contiguous with the Moesian platform and moved south during late Cretaceous to Palaeocene time along two transform faults, the dextral west Black Sea and the sinistral west Crimean faults (Figs. 1 and 5). It collided during early Eocene time with the Sakarya zone in the south. This continental collision obliterated the eastern part of the Intra-Pontide Ocean to the south of the İstanbul zone. Its western part remained open and continued to subduct northward throughout the Eocene (Fig. 5; Okay and Görür 1995). The contraction associated with this oceanic subduction could not, however, be accommodated by the relative movement between the now interlocked İstanbul and the Sakarya zones. To accommodate this contraction, the Rhodope-Pontide magmatic arc developing on the overriding Strandja zone started to rift and extend, thus forming the Thrace Basin. The continuation of the calc-alkaline volcanic material-bearing marine sedimentation in this basin well into the Oligocene indicates that the consumption of the Intra-Pontide oceanic crust and the following collision continued as late as the Oligocene time (Okay and Görür 1995). After the collision, the basin-fill was folded, eroded and then unconformably overlain by the Middle Miocene to Quaternary continental to shallow marine sediments (Doust and Arıkan 1974) which cover the suture along the present Dardanelles strait. The Miocene sediments are typical molasse deposits accumulated in WNW–ESE trending troughs (Doust and Arıkan 1974) and bear no relation to the earlier fore-arc basin. The Plio-Quaternary fluvio-lacustrine sediments formed in graben structures. These structures resulted from wrench tectonics as the post-collisional compression in the basin gave way to a strike-slip régime when the North Anatolian fault zone developed during late Miocene to Pliocene time along the southern border of the basin (Şengör 1979; Barka and Hancock 1984; Şengör et al. 1985). The post-collisional deformation was not very intense in the Thrace Basin. All three major components of the arc-trench system, i.e. the subduction complex, the fore-arc basin and the magmatic arc, are exposed on land more or less in their original relationship. Two reasons may have

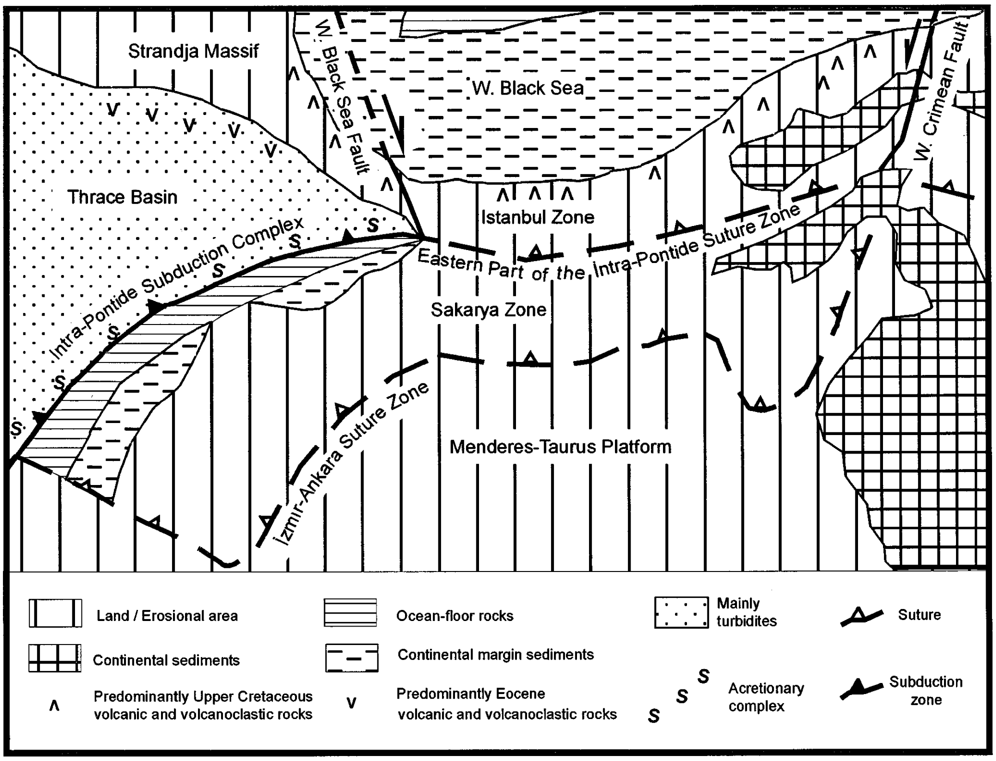


Fig. 5 Late Eocene palaeogeographic map of northwestern Turkey

been responsible for this preservation. These are the diachronous closure of the Intra-Pontide Ocean and the neotectonic strike-slip régime. The diachronous closure of the Intra-Pontide Ocean left a narrow oceanic trough between the Strandja and the Sakarya zones (Fig. 5). The accretionary mélange developed along the southern flank of the Thrace Basin probably played a cushion role between the colliding zones and thus saved the Thrace Basin from severe compression. The neotectonic strike-slip régime also helped the preservation of the original tectonic elements of this basin, because its further compression was terminated by the strike-slip movement along the North Anatolian fault zone, converting the convergent margin into a transform margin.

Summary and conclusions

Regional geology indicates that the Thrace Basin developed as a fore-arc basin during medial Eocene to Oligocene time. It was typically bounded on the arc flank by the volcano-plutonic assemblages and the associated metamorphic rocks of the Strandja massif, and on the trench flank by the accretionary mélange of the northward subducting Intra-Pontide Ocean. Both of these bounding features provided nearby sources of abundant detritus that accumulated until early Oligocene time as turbidites in the basin interiors, and clastic and associated carbonate rocks on the basin margins (Doust and Arıkan 1974; Turgut et al. 1991). Towards the end of the Oligocene, the basin had become completely choked with sediment and overflowed, losing its deep-basin character and distinct outlines. Following the closure of the Intra-Pontide Ocean to the south, limited convergence of its bounding continental blocks constricted and deformed the basin sediments which were unconformably overlain by molasse sediments of Miocene age. Post-Miocene deposits accumulated under the present strike-slip tectonic régime which replaced the previous convergent tectonics of the basin since the formation of the North Anatolian fault zone in the latest Miocene to the Pliocene (Şengör et al. 1985).

The basin fill commonly lapped unconformably upon the eroded flank of the arc massif along the arcward side of the Thrace Basin (Burke and Uğurtaş 1974; Turgut et al. 1991). However, in such places where extension occurred across the fore-arc belt, normal faults (such as the North Osmancık fault) dropped the floor of the basin downward from the elevated tract of the arc. Along the trenchward edge of the basin, a faulted contact between the basin fill and the uplifted parts of the subduction complex developed. Presently, this boundary fault occurs as an active dextral strike-slip structure, although it may have operated as a normal fault until the late Miocene when the tectonics in the area turned from a convergent into a strike-slip régime (Şengör 1979; Şengör et al. 1985).

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