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The Black Sea basin: tectonic history and Neogene–Quaternary rapid subsidence modelling

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Abstract

The Black Sea basin originated as a back-arc basin during Cretaceous times. Continental rifting took place during the Aptian to Albian with large-scale crustal thinning and separation occurring since the Cenomanian, mainly along a former Albian volcanic arc. Both western and eastern Black Sea basins opened almost simultaneously during Cenomanian to Coniacian times. However, during the Santonian to Palaeocene, the Black Sea region was affected by compressional deformation. Apart from a tensional event that took place in eastern part of the region during the Eocene, deepening of the basin has been induced by compressional deformation from latest Eocene to recent times. Kinematic and dynamic modelling of the subsidence history of the Black Sea basin shows that downward bending of the lithosphere beneath the basin due to compressional deformation could be the cause of this rapid additional subsidence.

Keywords: Black Sea basin; Pliocene–Quaternary; Syn-compressional bending; Subsidence modelling

1. Introduction

The Black Sea is located between Ukraine, Russia, Georgia, Turkey, Bulgaria and Romania. It has a surface area of 423 000 km^2 and a maximum water depth of up to 2–2.2 km. The structure of the basin is known mainly through the acquisition and interpretation of seismic data (Tugolesov et al., 1985; Finetti et al., 1988; Beloussov and Volvovsky, 1989), as deep wells having penetrated only the upper Pliocene–Quaternary part of the sedimentary cover.

The Black Sea basin, in terms of crustal structure, is composed of two deep basins (Figs. 1 and 2): the western Black Sea basin, which is underlain by oceanic to suboceanic crust and contains a sedimentary cover of up to 19 km thick, and the eastern Black Sea basin, which is underlain by thinned continental crust approximately 10 km in thickness and up to 12 km thickness of sediments. These basins are separated by the Andrusov Ridge that is formed from continental crust and overlain by 5–6 km thickness of sedimentary cover (Tugolesov et al., 1985; Finetti et al., 1988; Beloussov and Volvovsky, 1989; Robinson, 1997). The Black Sea basin is surrounded by Late Cainozoic mountain belts (the Great Caucasus, Pontides, Southern Crimea and Balkanides) that are

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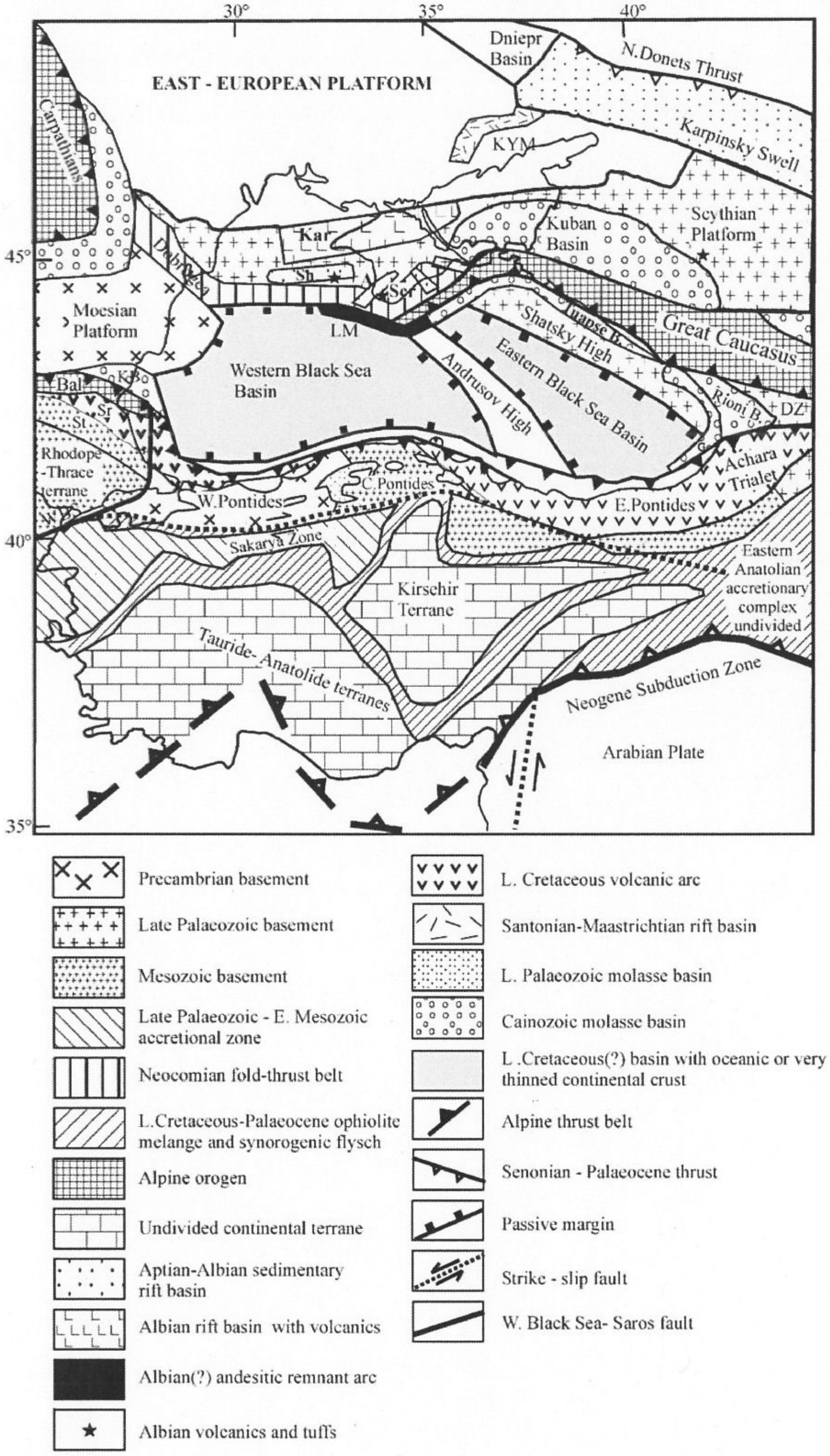

Fig. 1. Tectonic setting of the Black Sea basin. Kar—Karkinit trough, Sh—Shtormovaya graben, A—Alma basin, Scr—Southern Crimea Orogen, Bal—Balkanides, K—Kamchia foreland basin, Sr—Srednogorie, St—Strandzha, WBS—West Black Sea–Saros Fault, Dz—Dzirula, G—Guriy basin, KYM—Konka–Yaly–Molochnaya graben.

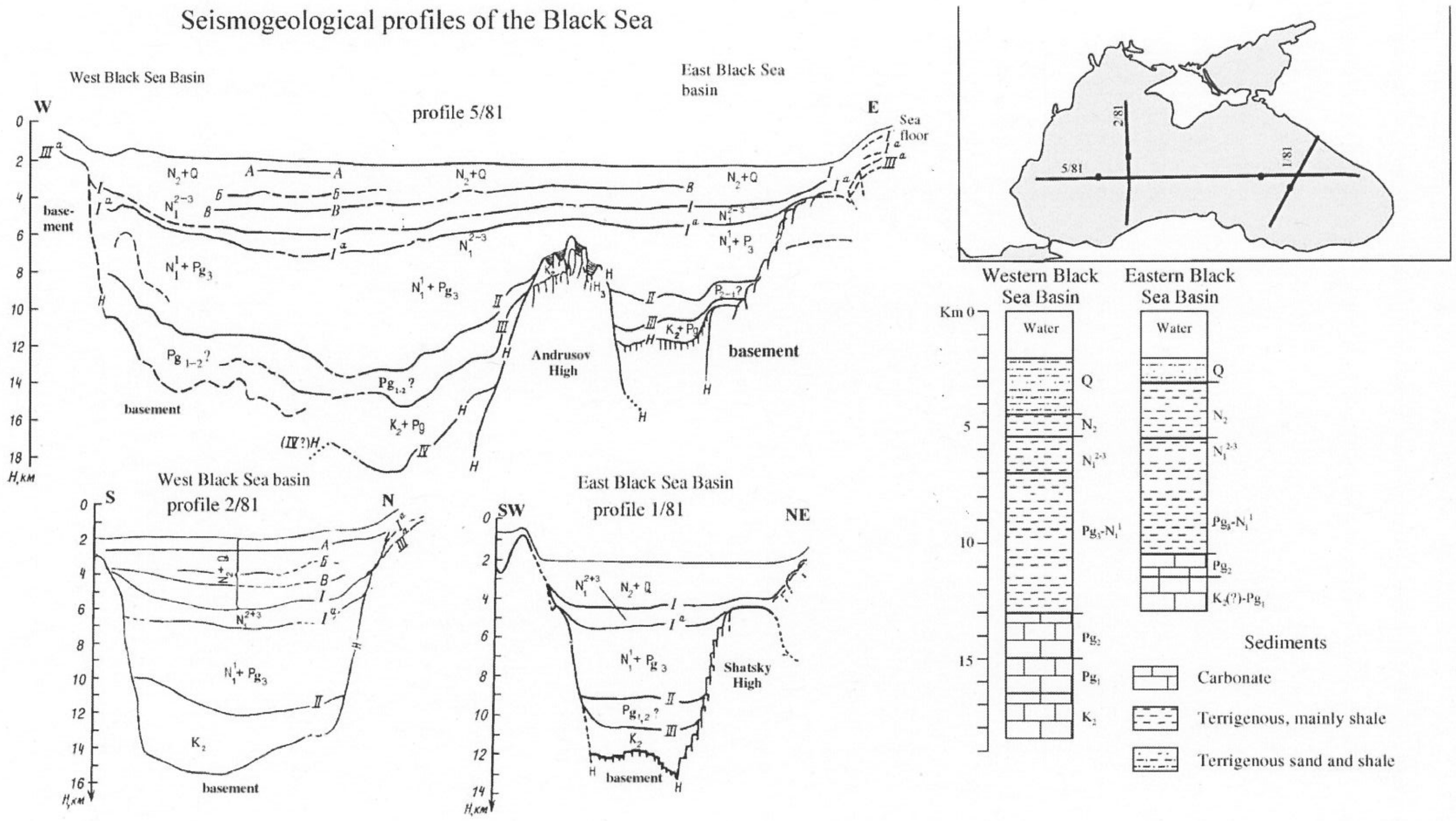

Fig. 2. Cross-sections for the Black Sea region (after Beloussov and Volvovsky, 1989) and location map of profiles and points of pseudo-wells used for the modelling of burial history.

represented by a thickened continental crust of some 40–50 km (Beloussov and Volvovsky, 1989; Volvovsky and Starostenko, 1996).

Both western and eastern Black Sea basins are marked by positive irregular Bouguer gravity anomalies (Goncharov et al., 1972; Bur'yanova et al., 1989). In addition, the axial zones of these basins are marked by positive magnetic anomalies of more than 200 mgal (Bur'yanova et al., 1989). The presence of a thick sedimentary cover considerably reduces the heat flow such that its value in central part of the basins is 30 mW/m^2, while a maximum of 70 mW/m^2 is found in the Crimean and Caucasus regions (Duchkov, 1996; Kutas et al., 1996).

Earthquake foci are situated on the borders of the deep-water basin. The average depth of these earthquakes is 5–20 km, they have a magnitude of up to 6–8 and are induced mainly by compression stresses (Volvovsky, 1989). Indeed, modern stress fields obtained from structural data (thrusts in the Pontides and the Crimean part of the Black Sea), earthquake data (Barka and Reilinger, 1997; Reilinger et al., 1997), stress field measurements in Crimean and the Caucasus regions (Rastsvetaev, 1987) and GPS data (Barka and Reilinger, 1997; Reilinger et al., 1997) show a dominantly compressional environment. The general source of compression is the collision between the Eurasian and Arabian plates. However, no data is presently available on the stress field of the inner part of the Black Sea.

Five main seismic sedimentary units within the Black Sea basin are recognised: the Upper Cretaceous, Palaeocene–Eocene, Oligocene–Lower Miocene, Upper Miocene and Pliocene–Quaternary (Tugolesov et al., 1985; Beloussov and Volvovsky, 1989). The proposed thickness of the Cretaceous sedimentary unit is 5–6 km in the west and 3–4 km in the east and a carbonate composition is supported by seismic evidence (Tugolesov et al., 1985). The thickness of the Palaeocene–Eocene unit is 5 km in the western Black Sea basin and 3 km in the eastern Black Sea basin, and consists of terrigenous and carbonate rocks (Tugolesov et al., 1985). The Oligocene–Lower Miocene (Maikopian) complex has a maximum thickness of 5 km in the western Black Sea basin and 4 km in the eastern Black Sea basin, and is composed of clay sediments (Tugolesov et al., 1985; Beloussov and Volvovsky, 1989; Ivanov, 1999). The Middle–Upper Miocene complex also has a terrigenous composition and a thickness varying from hundreds of metres (Shatsky Ridge) to 3 km (Guriy foredeep). The Pliocene–Quaternary unit consists mostly of clays (Tugolesov et al., 1985; Beloussov and Volvovsky, 1989) and has a thickness of 2–3.5 km in the inner parts of the basin. All Cretaceous to recent sequences are nearly horizontal and there is no indication of marked deformation within the basin (Tugolesov et al., 1985). The stratigraphy of the lower part of the sedimentary cover is not well constrained and there has been some variation in its description in the literature (Tugolesov et al., 1985; Finetti et al., 1988; Beloussov and Volvovsky, 1989; Robinson, 1997). The presence of Aptian–Albian rift units is possible below the Upper Cretaceous cover. Seismic data show that the basement beneath the basin is sometimes disrupted by normal faults (Tugolesov et al., 1985; Finetti et al., 1988; Robinson et al., 1996), supporting an extensional origin for the basin (Zonenshain and Le Pichon, 1986; Görür, 1988; Dercourt et al., 1993; Okay et al., 1994; Robinson, 1997; Nikishin et al., 1998, 2001). A volcanic calc-alkaline belt of Cretaceous age is located just to the south of the Black Sea basin along the Pontides. This combination of extensional basin and volcanic arc has led to a widely accepted conclusion that the Black Sea basin has a back-arc origin (Zonenshain and Le Pichon, 1986; Görür, 1988; Dercourt et al., 1993).

In this paper, the following key and unsolved problems for the Black Sea region will be discussed:

1. The time of the Black Sea basin origin (Cretaceous to Eocene ages have been proposed in the literature);
2. The kinematics of the basin opening (i.e., the direction of opening, fault control and the timing of opening of the two sub-basins);
3. The tectonic history of the basin after the time of opening;
4. Why is the thickness of Neogene to recent sediments so large?

2. Tectonic setting of the Black Sea basin

The tectonic evolution of the Black Sea region has been discussed by many authors (Tugolesov et al., 1985; Görür, 1988; Finetti et al., 1988; Beloussov and

Volvovsky, 1989; Ziegler, 1990; Dercourt et al., 1993; Okay et al., 1994; Dixon and Robertson, 1996; Robinson et al., 1996; Robinson, 1997; Banks and Robinson, 1997; Yilmaz et al., 1997; Ustaömer and Robertson, 1997; Nikishin et al., 1998, 2001; Stampfli et al., 2001). In this section, we use the review of the tectonics of the Black Sea region presented by Nikishin et al. (2001) along with newly published data. Fig. 1 shows the tectonic setting of the Black Sea basin and surrounding regions. The tectonic units of different ages surrounding the basin are as follows.

(1) The Scythian Platform, which consists of Late Palaeozoic basement deformed at the time of the Triassic/Jurassic boundary (Muratov, 1969; Milanovsky, 1991; Nikishin et al., 1998, 2001).

(2) The southern Crimea orogen with pre-Bajocian, pre-Callovian and intra-Berriasian orogenic phases (Nikishin et al., 2001).

(3) The Great Caucasus Alpine orogen resulting from the shortening and closing of a former Jurassic–Eocene back-arc basin during Late Eocene to recent times (Milanovsky, 1991; Ershov et al., 1998, 1999, 2003; Nikishin et al., 1998, 2001).

(4) The northern Dobrogea orogen, which consists of a former Permo-Triassic rift basin that underwent thrusting and folding from the time of the Jurassic/Cretaceous boundary up to Neocomian times (Banks, 1997; Seghedi, 2001; Nikishin et al., 2001).

(5) The Moesian Platform, consisting of Late Precambrian basement affected by Late Variscan deformation (Okay et al., 1994; Banks, 1997).

(6) The Balkanide Alpine thrust–fold belt (Banks, 1997).

(7) The Strandzha Mesozoic polyphase orogenic belt of main Neocomian age (Okay et al., 1994; Banks, 1997).

(8) The Rhodope Massif consisting of Palaeozoic and/or Precambrian basement, that experienced Variscan deformation and metamorphism as well as further deformation during the Mesozoic, mainly in pre-Cenomanian or even pre-Maastrichtian times (Burg et al., 1996; Banks, 1997).

(9) The western Pontides (or Istanbul Zone) whose basement is similar to the Moesian Platform (Okay et al., 1994; Banks and Robinson, 1997).

(10) The Central Pontides consisting of an Early Triassic(?) ophiolite complex covered by Triassic to Mid-Jurassic flysch. The main orogeny in this region occurred in pre-Callovian times (Ustaömer and Robertson, 1997) similar to southern Crimea.

(11) The eastern Pontides have a Late Palaeozoic basement (Okay and Sahintürk, 1997) and experienced a major orogenic event in the Aalenian–pre-Bajocian (Robinson et al., 1995) similar to Southern Crimea and the Great Caucasus (Nikishin et al., 2001).

(12) The Achara–Trialet Zone is characterised by a Cretaceous magmatic arc superimposed on a Late Palaeozoic(?) basement affected by Eocene rifting and post-Eocene closing (Lordkipanidze, 1980; Karyakin, 1989; Nikishin et al., 2001).

(13) The Dzirula Massif consists of Precambrian basement affected by Variscan deformation accompanied by Late Palaeozoic granitoids. The Shatsky High within the Black Sea is a possible prolongation of the Dzirula Massif. In addition, the Andrusov High has a basement that is proposed to be similar to the Shatsky High. Late Jurassic conglomerates observed in Southern Crimea in the Demerdzhi Mountain were transported from the south. It implies at this place the existence of large boulders of Palaeozoic and Precambrian granitoids and metamorphosed rocks, Triassic limestones and Mid-Jurassic volcanics (Chernov, 1971).

The Black Sea basin is bounded to the south by the Srednogorie–Pontide–Achara–Trialet–Karabakh Cretaceous magmatic belt (Lordkipanidze, 1980; Zonenshain and Le Pichon, 1986; Dercourt et al., 1993; Okay et al., 1994; Banks and Robinson, 1997; Nikishin et al., 1998, 2001; Stampfli et al., 2001). However, the time of formation of this magmatic belt is not well constrained. The Achara–Trialet magmatic arc in Georgia was active from the Aptian to Turonian, and, to a lesser extent, to the Campanian (Lordkipanidze, 1980; Karyakin, 1989; Nikishin et al., 2001). A small-scale Aptian–Albian volcanism was also proposed in the Pontide magmatic belt in Turkey. This belt was active from Cenomanian(?)–Coniacian to Campanian, and, to a lesser extent, to the Maastrichtian (Görür et al., 1993; Okay and Sahintürk, 1997; Yilmaz et al., 1997). The Srednogorie magmatic arc in Bulgaria was active during the Cenomanian(?) to Maastrichtian with maximum volcanism occurring during the Senonian (Beloussov and Volvovsky, 1989; Okay et al., 1994).

There is a remnant Cretaceous volcanic arc along the northern margin of the western Black Sea basin

(Shnyukov et al., 1997; Nikishin et al., 2001). Submarine investigations discovered a belt of Cretaceous magmatic rocks along the continental slope to the south and southwest of the town of Sevastopol in Crimea (Zhigunov, 1983; Shnyukov et al., 1997). This so-called Lomonosov Massif consists of basalts, andesites, shoshonitic basalts and dacites overlapped by Late Cretaceous clays and marls. The geochemistry of the volcanic rocks shows that they are probably arc-related. Their age, which is constrained by K/Ar dates, ranges from 150 and 100 Ma up to 65 Ma (Shnyukov et al., 1997). Coarse-grained Late Albian tuffs with lapilli are well documented in Crimea close to the towns of Balaklava–Sevastopol (Muratov, 1969; our data), while the latest Albian sandstones of the Southern Crimea are enriched by andesitic tuff material (Mazarovich and Mileev, 1989; our data). These data may support the proposal that arc volcanism within the Lomonosov Massif has an Albian age (Nikishin et al., 2001). We propose that the origins of the western Black Sea basin are to be found in, at least, a partial splitting of an Albian volcanic arc (Nikishin et al., 2001).

There are numerous Aptian–Albian continental rifted basins around the Black Sea (Muratov, 1969; Görür et al., 1993; Okay et al., 1994; Robinson and Kerusov, 1997; Yilmaz et al., 1997; Nikishin et al., 1998, 2001). In eastern Crimea, the fault bounded by the Belogorsk basin, which contains more than 1 km of sediments, developed during the Aptian–Albian. In southeastern Crimea, the Salgir graben also developed at the same time (Muratov, 1969; Byzova, 1981). In the Gulf of Odessa, the deep Karkinit trough, which contains more than 3 km of sediments, began to subside either during the Barremian or Late Hauterivian and remained active till Albian times. Its Middle-to-Late Albian main rifting stage was accompanied by volcanic activity, while post-rift evolution commenced during the Cenomanian, as documented by deep well and seismic data in the northern Crimea (Muratov, 1969; Chaitsky, 1984; Nikishin et al., 2001). The Shtormovaya graben, located south of the Karkinit basin, is defined by seismic reflection data (Robinson and Kerusov, 1997). The Alma basin of the southwestern Crimea is characterised by relatively rapid Valanginian–Early Aptian subsidence (Nikishin et al., 1998, 2001). The Kuban basin, located on the northwestern flank of the Great Caucasus trough, subsided rapidly during the Early Cretaceous and particularly during Aptian–Albian times. Its subsidence was accompanied by basaltic and andesitic volcanism (Bolotov, 1996). There are a few Aptian–Albian rift-like basins in the western and central Pontides (Görür et al., 1993; Robinson and Kerusov, 1997). Albian to Cenomanian tension-related basaltic magmatism took place along the Great Caucasus trough (Lomize, 1969; Lordkipanidze, 1980). Also, Albian syn-rift uplift took place within the Moesian Platform (Harbury and Cohen, 1997) and southern Crimea regions (Nikishin et al., 2001).

Ophiolitic sutures are located south of the Pontides (Parlak and Delaloye, 1999). The main Izmir–Ankara–Erzincan suture contains Cretaceous ophiolites and deep-water sediments of Norian to Senonian ages, indicating that an oceanic basin existed at least since Late Triassic (or earlier) to Senonian (Banks and Robinson, 1997; Okay, 1999, 2000; Nikishin et al., 2001). The Izmir–Ankara–Erzincan oceanic basin was closed mainly during latest Cretaceous–Palaeocene time (Yilmaz et al., 1997), or just before the Danian (Ozgul et al., 1999). The Intra-Pontide suture separates the western Pontides (or Istanbul Zone) to the north and Sakarya Zone to the south (Yilmaz et al., 1997). The interpretation of this suture is very controversial (Robinson, 1997). It could be a micro-oceanic basin that closed at the end of Cretaceous times and certainly before the Palaeocene (Elmas et al., 1999). However, new data show that the oceanic basin was closed probably during Neocomian to Mid-Cretaceous times, before the opening of the Black Sea basin (Okay and Tüysüz, 1999; Okay, 2000).

3. Timing and kinematic model of the Black Sea basin opening

Many authors consider the Black Sea as a Cretaceous to Palaeogene back-arc basin (Zonenshain and Le Pichon, 1986; Görür, 1988; Dercourt et al., 1993; Okay et al., 1994; Robinson, 1997; Nikishin et al., 1998, 2001). However, two key problems remain unsolved: the timing and the kinematics of the opening. One point of discussion is whether the western and eastern Black Sea sub-basins have the same time of origin, or, alternatively, is the eastern Black Sea basin younger? To date, no comprehensive solution has been proposed. However, we will try to constrain some possible models.

According to Okay et al. (1994), the western Black Sea basin was limited during the opening phase to the west and to the east by the West Crimea and West Black Sea strike-slip faults, respectively. They proposed a large-scale closing of the Great Caucasus trough resulting in space for the opening of the new eastern Black Sea basin. Obtained data show that no compressional deformation of the Great Caucasus trough took place during the Cretaceous until the Santonian–Campanian but Albian–Cenomanian basaltic magmatism indicates a possible extensional environment. The models of Okay and Robinson (Robinson et al., 1996; Banks and Robinson, 1997; Robinson and Kerusov, 1997) present some problems with regard to the opening of the eastern Black Sea basin because they do not include the Great Caucasus trough in their kinematic restorations. Thus, their models require a space to allow the opening of the eastern Black Sea following the western Black Sea basin's formation. Our restoration of the Black Sea region for Neocomian times (before the Black Sea opening) (Fig. 3) is similar to those of proposed by Zonenshain and Le Pichon (1986), Okay et al. (1994) and Robinson and Kerusov (1997), and is a more detailed and updated version of our former restorations (Nikishin et al., 1998, 2001). The Rhodope Massif was rotated counterclockwise by more than 12°, whereas recent data show its clockwise rotation since Mid-Oligocene (Dimitriadis et al., 1998). Based on this restoration, large-scale movement along a dextral strike-slip fault between the western Pontides and Rhodope–Thrace massifs seems a more likely explanation for the opening of the Black Sea basins. Hence, such movements of the whole Pontides have lead to a nearly synchronous opening of the western and eastern Black Sea basins (or one just after the other). This strike-slip fault is located presently along the Gulf of Saros and close to the Marmara Sea in Turkey, separating the western Pontides and the Srednogorie magmatic arcs. We propose to name this large-scale fault the West Black Sea–Saros Fault. In this restoration, no splitting of the Pontide magmatic arc by hypothetical orthogonal strike-slip faults is required, contrary to restorations proposed by Okay et al. (1994) and Robinson and Kerusov (1997). Our restoration shows that the opening of the western and eastern Black Sea basins took place in a similar regional event and during a single tectonic phase (Fig. 3). However, the reconstruction raises the problem of explaining the post-opening history of the area, near the proposed West Black Sea–Saros Fault.

The timing of the opening of the Black Sea remains quite controversial. It is generally considered that the Black Sea basin originated during the Mid-Cretaceous to Palaeocene, or even in Eocene times (Lordkipanidze, 1980; Zonenshain and Le Pichon, 1986; Dercourt et al., 1993; Okay et al., 1994; Robinson, 1997; Spadini et al., 1997; Nikishin et al., 2001). The following constraints regarding this timing are presented below.

(1) The continental rifting and rift-related uplift in the Black Sea area took place during the Albian or Aptian–Albian. The axial splitting of a former Albian(?) magmatic arc represents the origin of the Black Sea basin.

(2) In general, oceanic crust back-arc spreading follows continental rifting, so oceanic crust formation (or large-scale continental crustal thinning) took place since the Cenomanian in the Black Sea (at least in the western Black Sea basin). Since that time, large areas of the Moesian and Scythian platforms were covered by marine sediments (Harbury and Cohen, 1997; Nikishin et al., 1998, 2001); Cenomanian limestones in Crimea have thin benthonitic horizons (A. Alekseev, personal communication, 1999; our data) evidencing the near proximity of an active volcanic arc.

(3) Assuming that the commonly accepted duration for the opening of a back-arc basin is estimated to be around 5–15 million years (Kaiho and Saito, 1994), it would be unlikely to consider that the Black Sea basin opening stage lasted during the whole of the Late Cretaceous to Palaeocene times.

(4) Since Late Santonian or Campanian times, the southern part of the East European Platform underwent compressional deformation. For example, the Donbass (Donets basin) was submitted to a phase of uplift and intraplate thrusting (Nikishin et al., 1999, 2001; Stovba and Stephenson, 1999), intraplate inversion tectonics took place around all of Eastern Europe (Nikishin et al., 1999), southern Crimea experienced a few small compressional events since the end of Santonian to the Palaeocene and tectonically(?)-controlled olistostrome formed along the northern slope of the Great Caucasus trough during the Santonian to

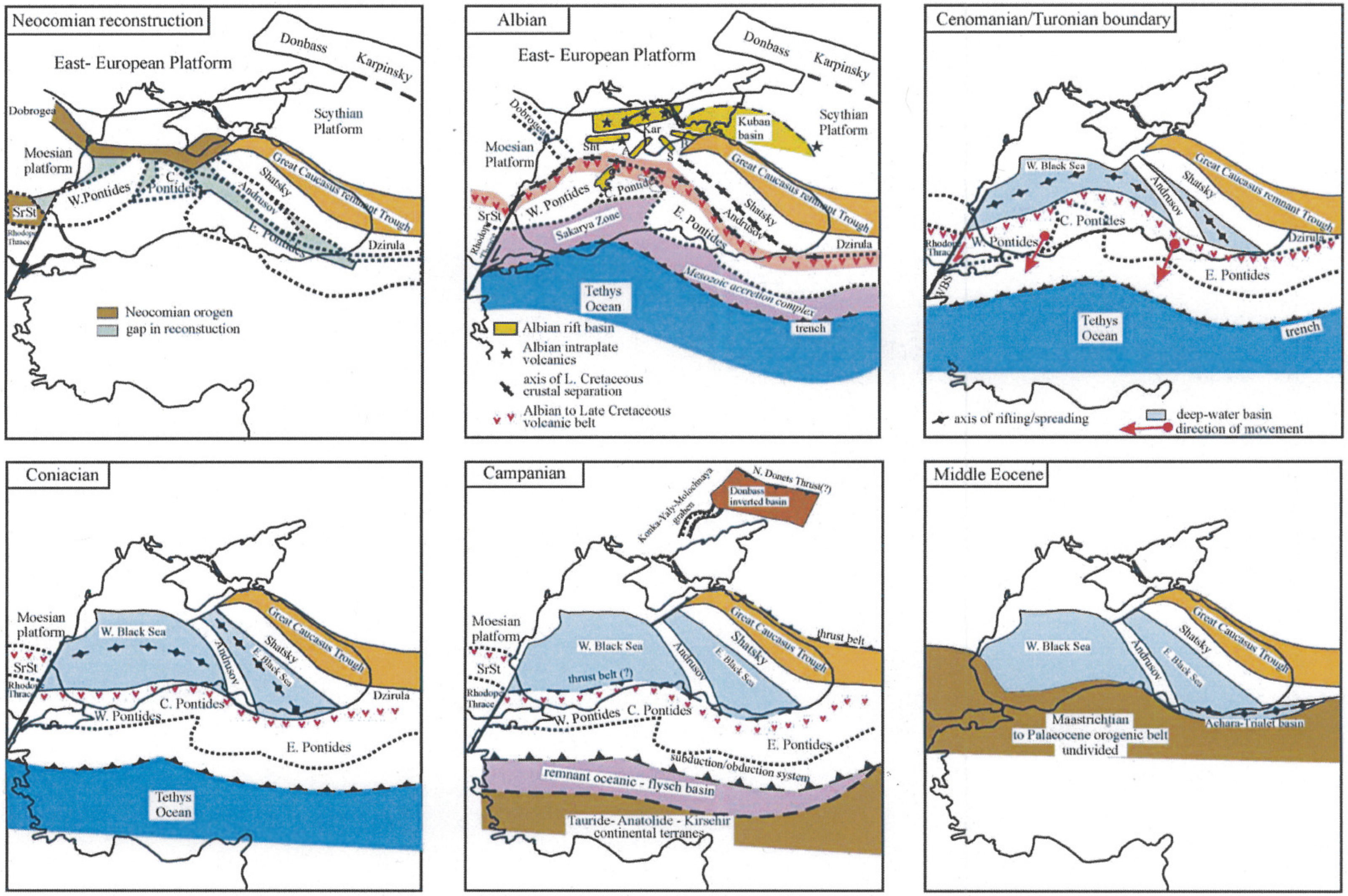

Fig. 3. Palaeotectonic reconstructions for the Black Sea region for: Neocomian (time before start of the opening of the Black Sea basin), Albian, Cenomanian/Turonian, Coniacian, Campanian and Middle Eocene times. SrSt—Srednogorie and Strandzha, Kar—Karkinit trough, Sht—Shtormovaya graben, A—Alma basin, S—Salgir graben, B—Belogorsk basin, WBS—West Black Sea–Saros fault. The proposed former positions of the main continental terranes are shown.

Maastrichtian (Markus and Sharafutdinov, 1989). As proposed in Nikishin et al. (1999, 2001), all of these indications seem to support the idea that Santonian–Campanian to Palaeocene times were not favourable for extension tectonics occurring within the Black Sea.

(5) Data on the Cretaceous stratigraphy and tectonics of the Pontides are controversial (Görür et al., 1993; Robinson, 1997; Rojay and Altiner, 1998). During the Cenomanian–Turonian, the Pontides underwent regional subsidence and local uplift (Görür et al., 1993; Yilmaz et al., 1997). Okay and Sahintürk (1997) describe the regional uplift of the eastern Pontides during the Cenomanian. Compressional tectonics took place during the Coniacian–Santonian–pre-middle Campanian times with overthrusting of ophiolitic melange onto fore-arc terranes (Yilmaz et al., 1997; Rojay and Altiner, 1998). This compressional deformation was followed by migration of the magmatic arc to the south during the Campanian (Yilmaz et al., 1997). For this event, described as a back-arc tensional phase corresponding to the opening of the western Black Sea basin (Yilmaz et al., 1997), we suggest a southward migration of the magmatic arc during the Campanian, connected with a southward migration of a trench after accretion of ophiolites. Our proposal is supported by new stratigraphic investigations in the central Pontides, which show that the main phase of ophiolitic melange obduction took place before the middle Campanian (Rojay and Altiner, 1998).

(6) Intense magmatism took place in the Senonian along the Pontides volcanic arc and may be connected with the changes of subduction system and compression of the arc rather than with the opening of the Black Sea basin.

(7) Since the Maastrichtian to Early Eocene, the Pontides experienced compressional and collisional tectonics due to accretion of continental terranes (Kirsehir Block, Tauride–Anatolide terranes (Yilmaz et al., 1997)). We do not exclude back-thrusting of the Pontides towards the Black Sea basin at that time, which could account for the rapid subsidence event of a back-arc belt between the active volcanic arc of Pontides and the Black Sea basin during the latest Cretaceous (Yilmaz et al., 1997). In this case, the subsidence should have had a syn-compressional flexural origin (not a back-arc tension).

(8) In Bulgaria, the external Balkanide thrust belt was activated at the end of the Cretaceous (Sinclair et al., 1997); at the same time, some parts of the Moesian Platform were uplifted (Harbury and Cohen, 1997), possibly in response to the build-up of intraplate compressional stresses.

(9) Data on Intra-Pontide suture are very controversial and this is why we ignored this unit in our tectonic restorations. However, it could be a Cretaceous rifted basin that closed at the end of the Cretaceous.

The data discussed above show that the most likely times for the opening of the Black Sea basins is the Cenomanian to Coniacian, that is within approximately 10 million years. Our proposal is that both the western and the eastern Black Sea basins originated during this period. Moreover, during the Senonian, the Black Sea area was in a compressional environment with maximum compression occurring within Maastrichtian–pre Eocene times.

Robinson et al. (1995, 1996) discuss the hypothesis that the eastern Black Sea basin opened during the Palaeocene. This timing is supported by the presence of a regional unconformity between the Maastrichtian and the Eocene strata as observed on seismic data acquired from the eastern Black Sea basin. Robinson proposed that this unconformity was connected with uplift caused by rifting. However, recent data show that this unconformity was more likely to be connected with the latest Cretaceous to pre-Eocene compression tectonics occurring on a larger scale along the southern margin of Eastern Europe (Nikishin et al., 1999).

4. Eocene history of the Black Sea region

Extensional tectonics dominated in the Transcaucasus region and in the Pontides (Yilmaz et al., 1997; Nikishin et al., 2001) during part of the Eocene. The Achara–Trialet basin of Georgia (Fig. 3) subsided during the Late Cretaceous–Palaeocene times and superimposed flysch-type sedimentation on top of the Aptian–Turonian Transcaucasus magmatic arc. During the Eocene, the basin was affected by a new subsidence phase that was accompanied by a large-scale basaltic, alkali-basaltic and andesitic volcanism and the accumulation of volcaniclastic flysch. This volcanic activity peaked during the middle Eocene. The Eocene sequence attains a thickness of 3–5 km (Lordkipanidze, 1980; Karyakin, 1989). A tensional Eocene history of the basin was proposed (Lordkipanidze, 1980; Karyakin, 1989; Koronovsky et al., 1997; Nikishin et al., 2001). This rift basin was connected with the southern part of the eastern Black Sea basin (Fig. 3). During the Late Eocene, compression started in the Achara–Trialet basin. Furthermore, the Achara–Trialet basin was compressionally deformed into a north-verging thrustbelt during the Neogene (Milanovsky, 1968, 1991; Robinson, 1997; Banks and Robinson, 1997). This thrustbelt links up to the west with the external parts of the eastern Pontides.

This rifting cycle is also reflected in the middle Eocene extensional collapse of the eastern Pontides during which a marine volcaniclastic sequence was deposited (Okay and Sahintürk, 1997; Yilmaz et al., 1997). This tensional event was followed by compression during the Oligocene (Yilmaz et al., 1997).

In Bulgaria, thrusting of the Balkanides onto the western Black Sea basin occurred in the middle and Late Eocene times. This thrusting led to the formation of the Kamchia foreland basin (Sinclair et al., 1997).

5. Oligocene to Quaternary tectonic history of the Black Sea region

The tectonics of the Black Sea region, from Oligocene to recent, has been described in numerous papers (Milanovsky, 1968, 1991; Tugolesov et al., 1985; Beloussov and Volvovsky, 1989; Robinson, 1997; Ershov et al., 1998, 1999; Nikishin et al., 1998, 2001). The period from the Oligocene to Quaternary is characterised by multiple phases of compression connected with collision with the Arabian continent that affected the Caucasus–Pontides–Black Sea region (Dercourt et al., 1993). The Great Caucasus trough began to close since the Late Eocene as a result of the underthrusting of its crust below the Scythian Platform. The closure of the deep-water basin occurred during Late Eocene to Oligocene and Early Miocene times and was followed by an asynchronous collision and general thrusting and folding since approximately middle Miocene to recent times. Large-scale uplift of the mountains started during the Late Sarmatian (nearly 10 Ma) (Milanovsky, 1968, 1991). The Tuapse foredeep basin, situated between the Great Caucasus orogen and the Black Sea basin, also began to subside during the Oligocene with its development being controlled by the thrustbelt (Tugolesov et al., 1985; Robinson, 1997). In the Transcaucasus and Turkish regions, uplift started during the Oligocene, but an acceleration of the deformation started during the Late Sarmatian. Large-scale synorogenic basalt–andesite to rhyolite volcanism took place during Late Sarmatian times along the orogenic belt in the Caucasus–Turkey area (Milanovsky, 1968, 1991; Yilmaz et al., 1997). Our proposal is that the main time span of crustal thickening and heating in the Caucasus–Turkey region took place since the Late Sarmatian and lasted for the past 10 million years.

Seismic profiles show that the thrust–fold belt occurs between the Pontides and the deep-water part of the Black Sea basin (Tugolesov et al., 1985; Robinson, 1997). Although the timing of deformation is poorly known, there is evidence of compressional tectonics within Pliocene sediments (Robinson et al., 1995). Moreover, the onset of folding may have began in the Oligocene, though latest Cretaceous to Early Eocene deformational events cannot be excluded.

Some inversion structures have been observed on the Odessa Shelf and in northern Crimea (Muratov, 1969; Tugolesov et al., 1985; Robinson and Kerusov, 1997). The timing of this deformation is mainly pre-middle Miocene, with multiple phases of compressional deformation starting possibly in the Late Eocene (Robinson and Kerusov, 1997).

In Bulgaria, an extension phase took place during the end of Eocene to Oligocene in the Rhodope Massif. This was accompanied by exhumation of metamorphic rocks (37–35 Ma), and calc-alkaline and shoshonitic volcanism (37–25.5 Ma) (Yanev and Bardintzeff, 1997; Krohe et al., 1999). Another short volcanic phase took place between 22 and 19 Ma and produced typical alkaline rocks (basanites) in the Rhodope–Moesian region (Yanev and Bardintzeff, 1997). The data show that the Bulgarian region was in a tensional regime during latest Eocene to Early Miocene.

Data on the recent stress field show that the Black Sea region is situated within a compressional environment (Nikitina, 1997; Sassi and Faure, 1997). Collisional tectonics lasted along the Izmir–Ankara suture until Late Miocene (Bozkurt and Satir, 1999). The dextral strike-slip movement of the North Anatolian Fault began since around the end of the Miocene (Robertson and Grasso, 1995; Yilmaz et al., 1997), with a total displacement of almost 50 km (Yilmaz et al., 1997).

In conclusion, regional compression, folding and thrusting started during the Late Eocene–Oligocene and was contemporaneous with an extension phase that took place in the Rhodopean area. General uplift of the recent mountains (and crustal thickening) that surround the Black Sea began nearly 10 million years ago and was accompanied by widespread volcanism and crustal heating. This event is related to a final slab detachment that occurred at the end of subduction of the Tethyan oceanic lithosphere.

6. Black Sea basin burial history modelling

A standard backstripping procedure has been used for the modelling of burial history (Steckler and Watts, 1978). The 1D/2D models obtained comprise restoration of palaeo-profiles for different time slices and include decompaction of sedimentary units and palaeo-waterdepth profiles (Ershov, 1997; Ershov et al., 1998; Korotaev, 1998). In order to produce 2D reconstructions, each section was subdivided into a set of pseudo-wells for 1D reconstruction that included decompaction. Initially, a palaeo-waterdepth correction was applied to each well, followed by a reconstitution through time made for each section.

Pseudo-well data, obtained from seismogeological sections (profile nos. 181, 281, 581) collected by the research vessel "Professor Shtockman" (Beloussov and Volvovsky, 1989) were also used to complete 1D and 2D modelling of the Black Sea burial history. Model results are shown in Figs. 4 and 5.

The main problem that arose during the construction of the models was the undefined value of the palaeo-water depth during basin evolution which has reached up to several kilometres. There are no reliable data available to make a precise estimation of palaeo-water depths of the Black Sea basin. Deep wells penetrate only the upper part of the sedimentary cover, and most of the information come from the interpretation of seismic profiles. Therefore, all palaeo-bathymetries proposed for the basin are hypothetical. A recent proposal put forward by Spadini et al. (1997) is based on the hypothesis that during Mid-Cretaceous times, at the time of oceanic crust formation, water depth in the western Black Sea basin reached up to 5 km. However, a relatively large sea level fall took place during Early Sarmatian and the water depth had decreased up to 1 km. We made other assumptions, based on the following facts:

- A deep-water basin existed in pre-Oligocene times, a fact supported by data obtained from seismic profiles and sustained by the presence of Palaeocene–Eocene–Late Cretaceous(?) passive margins around this basin (Tugolesov et al., 1985; Beloussov and Volvovsky, 1989).
- The proposed timing for the rifting in the Black Sea basin and oceanic crust formation or large-scale crustal thinning are Aptian–Albian, and Cenomanian to Coniacian, respectively.
- Recent active back-arc basins with very young oceanic crust, such as the Tyrrhenian Sea, have present water depths of approximately 3–3.5 km. A similar Cenomanian-to-Coniacian water depth for the Black Sea is proposed here.
- The Senonian-to-Eocene Black Sea basin would have been similar to the Algero–Provençal basin with water depth of approximately 2.5–3 km.

Despite the lack of accurate data on water depth changes during Oligocene to Quaternary times, fragments of pelagic Oligocene clays transported to the sea bottom by recent mud volcanoes within the Black Sea have been collected (Ivanov, 1999). Data on palaeo-canyons on the passive margins and on the palaeogeographic history of the shelf regions provide no information on large-scale (more than a few hundred metres) water depth falls or rises during post-Eocene times. As a consequence, the water depth we propose for Oligocene–Neogene times is similar to the recent one, but with a main sea level fall (not more than a few hundred metres) taking place during Early Pliocene (nearly 5.2–3.4 Ma) (Milanovsky, 1968; Jones and Simmons, 1997; Ershov et al., 1998).

The main differences between our reconstruction and the one proposed by Spadini et al. (1997) lies in the maximum basin depth and the absence in our model of a water depth fall during Sarmatian time.

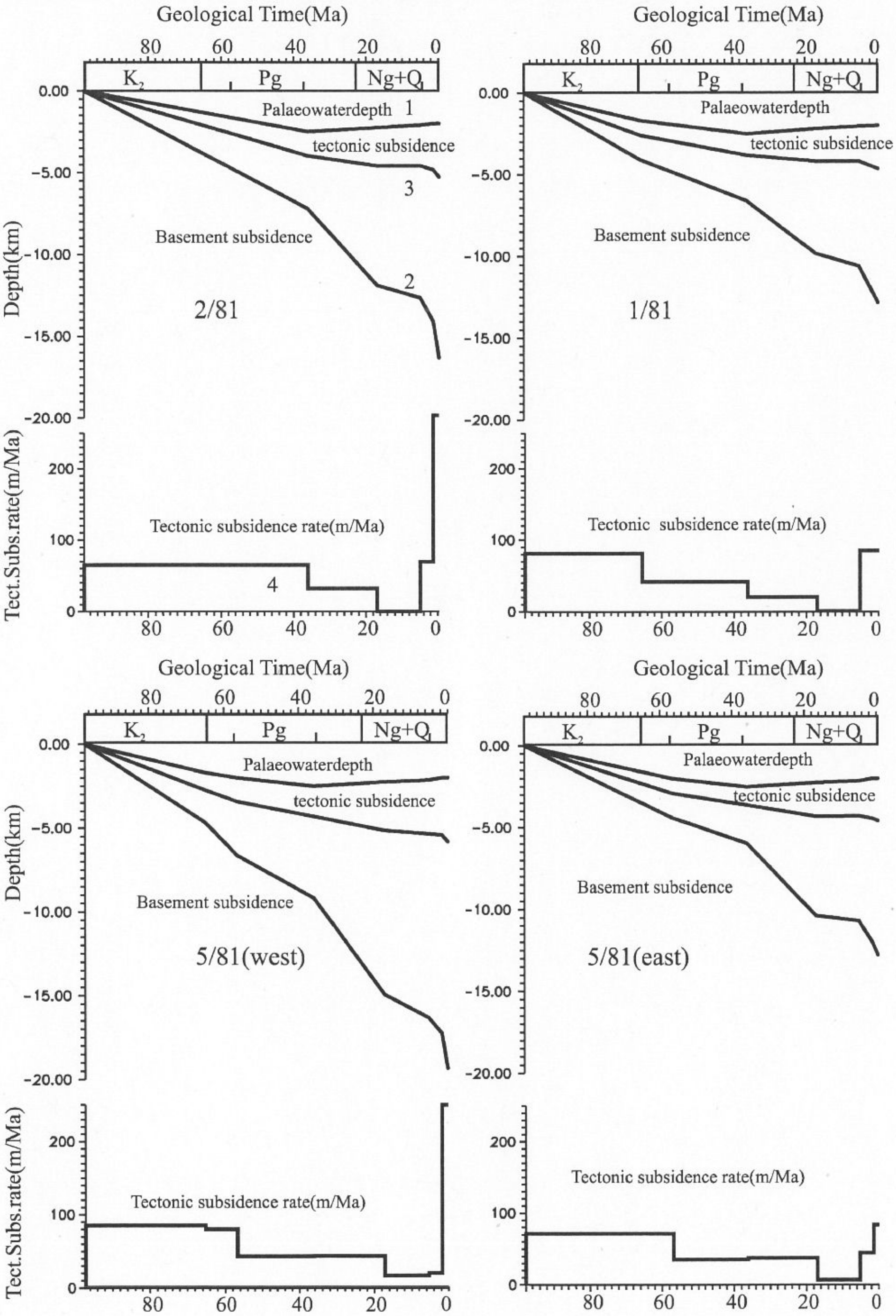

Fig. 4. 1D modelling of the burial history of Black Sea basin. 1—Palaeo-water depth, 2—basement subsidence, 3—tectonic subsidence, 4—rate of tectonic subsidence. Time scale is from Odin (1994).

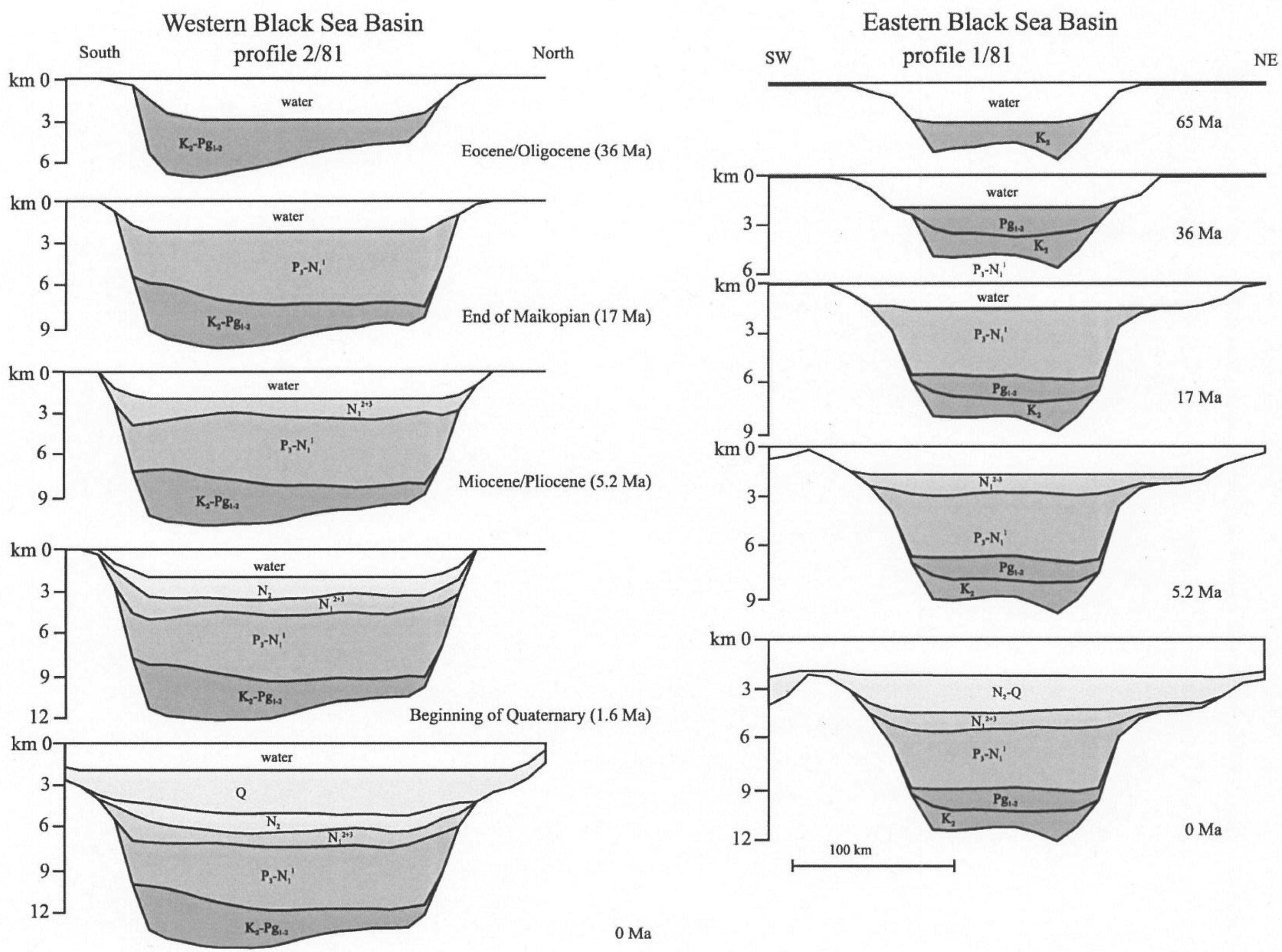

Fig. 5. Results of 2D modelling (with compaction correction and palaeo-waterdepth correction) for eastern Black Sea and western Black Sea basins.

For each of our 1D burial history reconstructions, we can recognize some stages of subsidence:

- Late Cretaceous–Eocene (97–36 Ma): the water depth at the end of stage is 2.5 km. Tectonic subsidence is up to 4.3 km in the western Black Sea basin and 3.8 km in eastern Black Sea basin. The rate of tectonic subsidence reaches 70–80 m/Ma.
- Oligocene–Miocene (36–5.2 Ma): water depth at the beginning of this stage is 2.5 km, and 2.25 km at the end. Tectonic subsidence is 0.6 and 0.4 km in the western and eastern Black Sea basins, respectively. The rate of tectonic subsidence is 20–30 m/Ma.
- Pliocene–Quaternary stage (5.2–0 Ma): the water depth is close to the modern depth at 2.0–2.2 km. The tectonic subsidence is 0.45–0.5 km in both basins with a rate reaching 85 m/Ma.

Our model, as well as those of Spadini et al. (1997) and Robinson et al. (1996), demonstrates an acceleration of basement subsidence in Pliocene–Quaternary time. The differences between these models lie mainly in the estimation of palaeo-water depth. Robinson et al. (1996) did not take into account palaeo-water depth and this results in an Oligocene acceleration of subsidence. In our model, the great thickness of Oligocene sediments is the result of filling of a deep-water basin. This conclusion is supported by seismostratigraphic analysis (Beloussov and Volvovsky, 1989). In addition, the model presented here differs from that of Spadini et al. (1997), in terms of the initial palaeo-water depth that has been assumed and in the occurrence or not of a large sea level fall in Early Sarmatian time.

7. Model of the Pliocene–Quaternary rapid subsidence of the Black Sea basin

This section focuses on the stage of rapid subsidence during Pliocene–Quaternary. Although some models have been proposed to explain this subsidence, such as the model including phase transitions in the crust (Artyushkov, 1993) or another including convection (Smolyaninova et al., 1996), they are hypothetical and have no strong data to support them. Taking that statement aside, we do not exclude the possibilities of phase transitions or convection-related movements. The model presented in this paper is based on the simple assumption that the Pliocene–Quaternary rapid subsidence took place in a regional compressional environment. This type of model was introduced by Cloetingh et al. (1989) for the North Sea region, and developed in terms of lithospheric folding for the Late Cainozoic deformation in the northern Peri-Tethyan region (Nikishin et al., 1997; Korotaev, 1998). Hence, a model of syn-compressional downward bending of the Black Sea lithosphere is proposed here as an explanation for the rapid Pliocene–Quaternary subsidence. An identical proposal is made for the Pliocene–Quaternary subsidence event in the South Caspian basin (Brunet et al., 2003).

The lithosphere may be modelled in terms of its rheological properties as a thin elastic plate, referred to here as the equivalent elastic plate. This physical property of the equivalent elastic plate is determined by two parameters: the effective elastic thickness of the lithosphere (EET) (Burov and Diament, 1995) and the location of the effective middle surface of the lithosphere (EMS) (Ershov, 1999). The lithosphere underlying the Black Sea contains great heterogeneities. For example, the oceanic or severely thinned continental crust in the central part of the basin has an EET of almost 50–100 km, whereas in the marginal parts of basin, where there is thick and heated continental crust, the EET is close to 25–30 km. This means that the EMS is flexed downwards in the centre of the basin with an amplitude of about 25 km. Such pre-deformational heterogeneities of the lithosphere can induce downward flexural movements under regional compression (Fig. 6). The algorithms used here to calculate EET and EMS were presented by Ershov (1999).

The shape and amplitude of the subsidence induced by syn-compressional downward bending of the lithosphere depend on several parameters: the width of the area of thinned continental or oceanic crust, the thickness of the sedimentary cover, the thermal regime and the applied force. The working ranges of these parameters (i.e., ranges where the modelled amplitude and geometry of subsidence are similar to that observed in the Black Sea basin) were estimated by Korotaev (1998). A moderate intraplate force (5×10^{12}–10^{13} N/m) can induce 0.3–2 km of tectonic subsidence in a

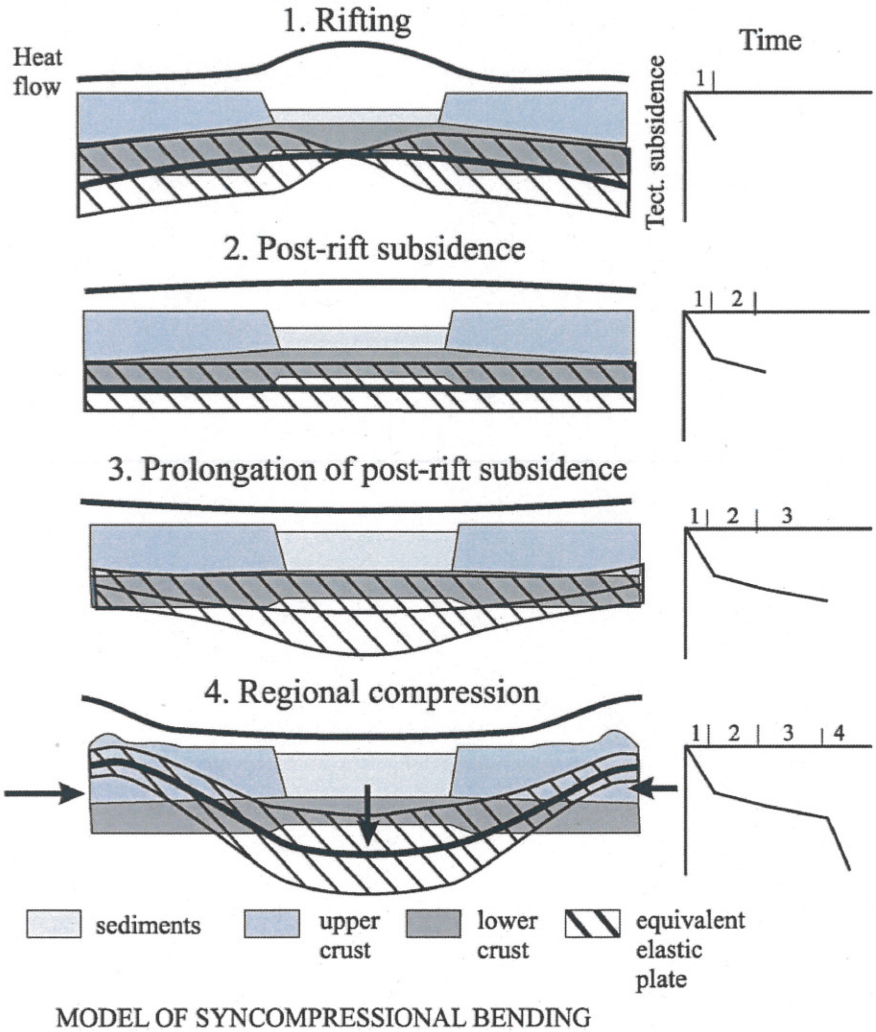

Fig. 6. Schematic representation of the model of syn-compressional bending of the lithosphere.

basin of 200–300 km width with oceanic or very thinned continental crust covered by 15–25 km of sediments. These values correspond to those observed in the Black Sea basin and the proposed model can be applied to the investigated area.

For the rheological modelling of the Black Sea lithosphere, we used two crustal cross-sections through the western and eastern Black Sea (Fig. 7). In the western Black Sea, sedimentary thickness varies from 3 to 19 km, the upper ("granitic") crust is absent in the central part of the basin, and the Moho depth is 25 km in the centre and 40 km at the margins. Values of heat flow are 30 mW/m^2 in the centre of the profile and 50 mW/m^2 on the edges. In the eastern Black Sea, the sedimentary thickness varies from 2 to 15 km, the upper crust is 5–7 km thick and the lower crust 6–8 km thick in the central part. Moho depth is 25 km in the centre and 40 km at the margins. Heat flow is 30 mW/m^2 in the centre of the profile and 50 mW/m^2 on the edges.

Rheological modelling of the Black Sea lithosphere shows that EET in the central part of the basin is 60–70 km and EET on the margin of the basin is 30 km. EMS is downflexured with amplitude of about 15–25 km. Under compression with a 5×10^{12} N/m force, the central part of the modelled profile subsides of 0.4 km, a figure similar to the tectonic subsidence obtained from modelling of the burial history. Our

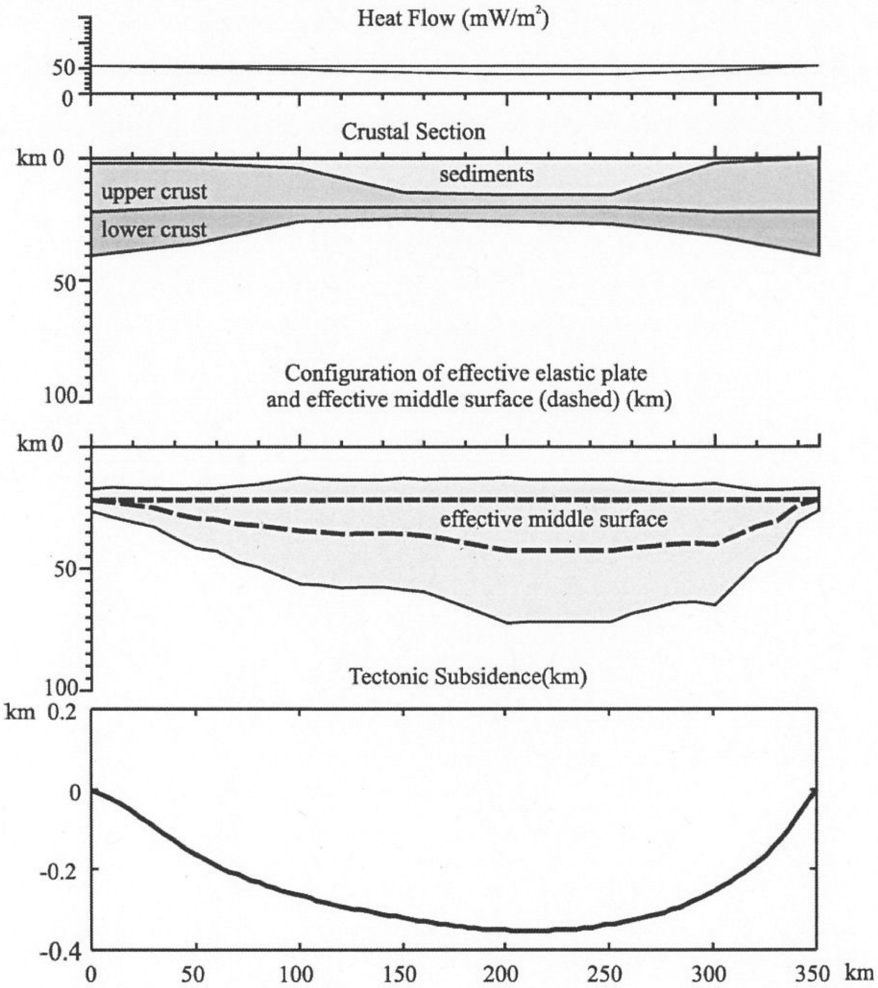

Fig. 7. Rheological modelling of the Black Sea basins. 1—Heat flow, 2—crustal section, 3—profile of effective elastic plate and location of effective middle surface, 4—profile of the modelled tectonic subsidence of the basin.

results show that the rapid Pliocene–Quaternary subsidence of the Black Sea basin can be explained by down-bending of the lithosphere due to compression.

8. Conclusions

(1) The western and eastern Black Sea basins originated mainly along a former Albian volcanic arc. The opening of the basins has taken place since Cenomanian to Coniacian times in a back-arc geodynamic regime due to trench retreat of the Pontides subduction system. The West Black Sea–Saros Fault could be the main strike-slip fault that controlled the opening of the basin.

(2) The Black Sea region was in a compressional environment from Senonian to Palaeocene times. This was followed by dominant Eocene tension connected with orogenic collapse in Turkey.

(3) Compression affected the Black Sea region since the Late Eocene with the exception of the Rhodopean–Moesian area.

(4) The rapid Pliocene–Quaternary subsidence of the Black Sea basin took place in a compressional environment.

(5) The kinematic and dynamic modelling of Black Sea basin subsidence shows that a compression-induced downward bending of the lithosphere could be the reason for the rapid (non-thermal) Pliocene–Quaternary subsidence phase.

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