Evolution of Early Mesozoic back-arc basins in the Black Sea–Caucasus segment of a Tethyan active margin

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Abstract: Six new reconstructions illustrate the evolution of back-arc basins in the Black Sea–Caucasus region from the Mid-Triassic to the end of the Mid-Jurassic. The *c.* 2000 km long Tauric (Küre) basin opened in the Late Permian–Early Triassic as the Pontides–Transcaucasus and Rhodope microcontinents rifted from the Eurasian margin. The oceanic floor of the Tauric basin in the Mid-Triassic was at least 300 km wide. In the east the basin closed near the present-day Caspian Sea and to the west of the West Crimea transform it split into two branches to the south and north of the Moesian platform. The Tauric basin was partly inverted in the Carnian, when several Gondwanian terranes (Iran, South Armenia) collided with the Palaeotethyan subduction zone. Following the initiation of a new subduction zone, the back-arc extension resumed in the Norian–Early Jurassic. Opening of the Izmir–Ankara–Sevan back-arc basin commenced south of the Pontides–Transcaucasus. Simultaneously, rifting began in the Greater Caucasus and continued until the Early Pliensbachian. This was followed by the continental break-up in the Late Pliensbachian–Toarcian. A narrow (100–150 km) strip of oceanic crust had formed by the beginning of the Aalenian. In the Late Aalenian a southward-migrating subduction zone at the southern margin of the Izmir–Ankara–Sevan basin had reached the central part of Neo-Tethys and presumably collided with a mid-oceanic ridge. Subduction was blocked and Africa–Eurasia convergence was compensated by inversion in the Tauric and Greater Caucasus basins. The basins were closed by the end of the Bathonian.

During the last two decades several attempts were made to reconstruct the history of the early Mesozoic back-arc basins in the Black Sea–Caucasus–South Caspian region (Dercourt *et al.* 1985, 1993, 2001; Adamia *et al.* 1990*a*; Kazmin 1990; Ustaömer & Robertson 1993; Stampfli 1996; Banks & Robinson 1997; Kazmin & Natapov 1998; Stampfli *et al.* 1998; Nikishin *et al.* 2001; Stampfli & Borel 2002). Although significant progress has been made, in most of the published works the reconstructions were schematic.

The main problems concern the relationships between the Tauric (Küre) back-arc basin and Greater Caucasus basin, the time and the mode of origin of the latter, and the configuration and the evolution of both basins. In most reconstructions, as listed above, the Greater Caucasus basin is interpreted as an eastward extension of the Tauric basin, although there is reliable evidence that the two basins were separated by the crustal block of the Shatsky rise. The opening of the Greater Caucasus basin in early Jurassic time and its subsequent evolution is usually related to a subduction zone along the southern margin of the Pontides. However, there is convincing evidence that in the Jurassic and Neocomian this margin was passive (Altiner & Koçyiğit 1992; Tüysüz *et al.* 1995; Okay & Şahintürk 1997). Consequently, the interpretation of the early Mesozoic evolution of the Pontides–Caucasus region needs revision.

Restoration of the early Mesozoic history is hampered by a lack of reliable palaeomagnetic data. For reasons still unknown, palaeomagnetic measurements of Jurassic rocks of the Pontides, Transcaucasus and Crimea yield very low inclinations, corresponding to remote southerly positions far from Eurasia (Asanidze & Pechersky 1979; Lauer 1984; Westphal *et al.* 1986; Saribudak 1988; Pechersky & Safronov 1993). The only attempt to reconcile the palaeomagnetic and geological data, by Kazmin & Natapov (1998), was unsuccessful.

In the present paper, controversial palaeomagnetic data on terranes were not used. Movements of terranes relative to the Eurasian margin were instead deduced from geological data; i.e. the time of rifting and collision, the duration of rifting, spreading and subduction periods. Reasonable spreading and subduction rates were assumed. The position of the Eurasian margin was taken from recently published works (Kazmin & Natapov 1998; Daukeev *et al.* 2002), where it was calculated using oceanic magnetic anomalies and plate motion relative to hotspots.

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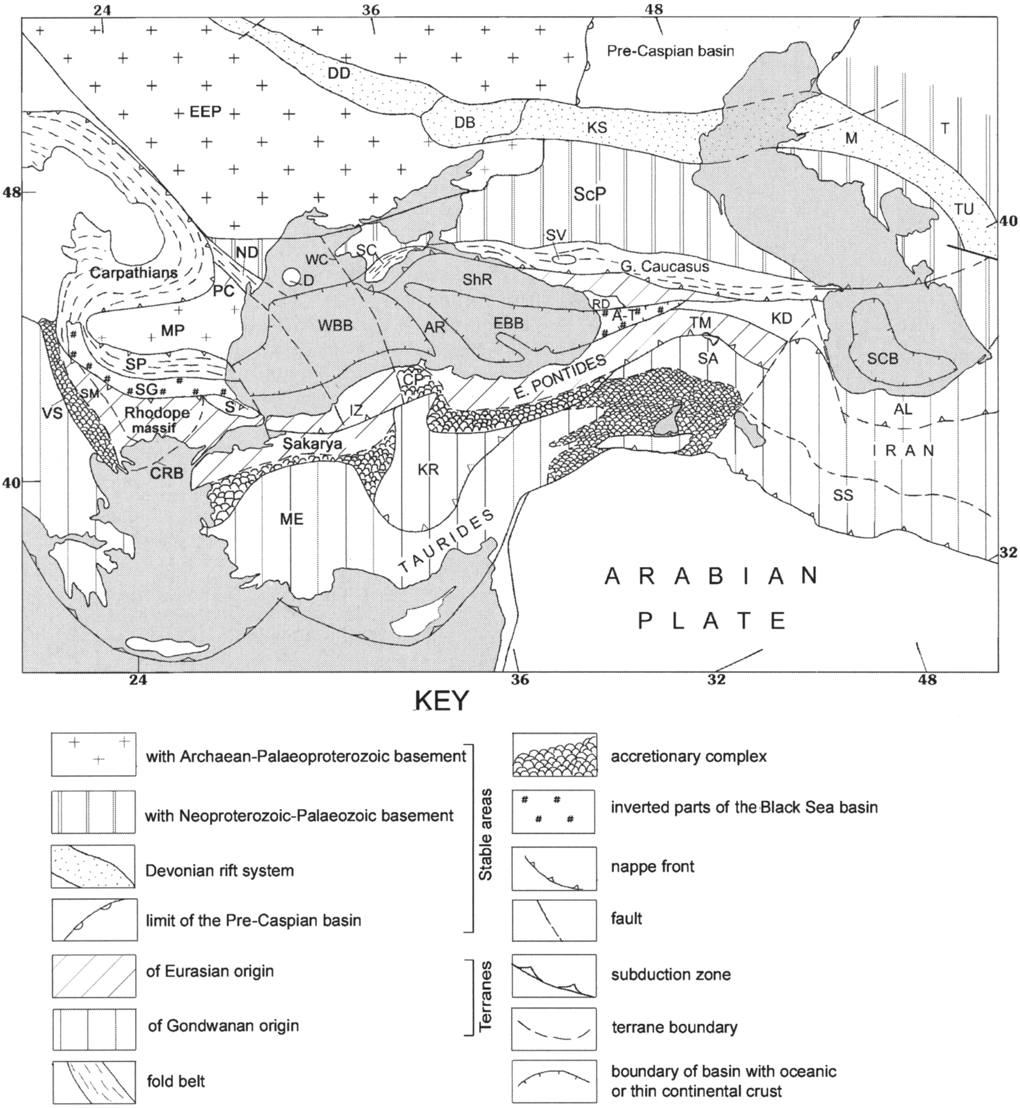

Fig. 1. Main structures of the Alpine Belt in the Black Sea–Caucasus region. AL, Alborz; AR, Andrusov rise; A-T, Adjaro–Trialetia; CI, Central Iran; CRB, Circum Rhodope belt; CP, Central Pontides; DB, Donbass; DD, Dniepr–Donets aulacogen; EBB, Eastern Black Sea basin; EEP, East European platform; EI, East Iran; EP, Eastern Pontides; GC, Greater Caucasus; GCF, Greater Caucasus fold belt; IZ, Istanbul zone; KD, Kura depression; KDB, Kopetdag basin; KM, Kargı massif; KR, Kirşehir massif; KS, Karpinsky swell; M, Mangyshlak; ME, Menderes massif; MG, Manych graben; MP, Moesian platform; ND, North Dobrogea; PB, Pre-Caspian basin; PC, Pechenega–Camena fault; PFB, Palaeozoic fold belt; PR, Paikon Ridge; RD, Rhodope massif; S, Strandja (Istranca) zone; SA, South Armenian terrane; SC, South Crimea; SCB, South Caspian basin; ScP, Scythian platform; SG, Sredna Gora zone; ShR, Shatsky rise; SK, Sakarya (Sakaria) block; SM, Serbomacedonian massif; SP, Stara Planina zone; SS, Sanandaj–Sinjar zone; SV, Svanetia; T, Turanian platform; TB, Tauric basin; TFB, Triassic fold belt; TM, Transcaucasus massif; TU, Tuarkyr; VC, Vardar suture; WBB, Western Black Sea basin; WC, West Crimea fault; WEP, West European Platform.

There are two types of terranes involved in the evolution of the active Eurasian margin and the evolution of the related back-arc basins (see Fig. 1). Terranes (microcontinents) of the first type have a Neoproterozoic basement strongly altered by Hercynian tectonics (Adamia *et al.*

1989; Okay & Şahintürk 1997; Zakariadze *et al.* 1998). The wide development of pre- to syntectonic granitoids (330–280 Ma) and late Palaeozoic molasse with clear Eurasian affinity (Belov 1981) indicates that these blocks were rifted from the late Palaeozoic active margin of Europe. They formed a chain, including the Transcaucasian massif, the Pontides and also blocks of the Andrusov and Shatsky rises, which formed parts of the Pontides–Transcaucasus prior to opening of Mesozoic marginal basins.

Less clear is the situation of the Rhodope massif. Traditionally its crust was described as Precambrian, strongly affected by Hercynian and Alpine tectonometamorphic events (Kronberg *et al.* 1970; Jones *et al.* 1992; Kozhoukharova 1996). According to others, the massif is an Alpine metamorphic complex formed by Cenozoic subduction–accretion processes (Barr *et al.* 1999; see also Himmerkus *et al.* 2006). Perhaps a compromise solution is acceptable: the Rhodope massif was perhaps a part of the Palaeozoic margin of Eurasia, to which magmatic material was added during the Alpine cycle. In the following reconstruction we envisage that a Triassic back-arc basin opened between the Rhodope massif and the Moesian platform and that the Rhodope massif was a part of a 'Rhodope–Pontide fragment' (Şengör 1984).

Terranes of the second type can be seen as fragments of Gondwana that collided with the Eurasian margin during the Mesozoic and Early Cenozoic. The largest of these fragments, Iran, belonged to the ribbon-like Cimmerian continent (Şengör 1979) and had its western extension as a chain of blocks including the South Armenian terrane (Dercourt *et al.* 1986), probably the Kirşehir massif and some smaller fragments.

As there are few data on Alpine accretion, the present-day size of post-late Triassic terranes is assumed with some corrections (e.g. straightening of Alpine bends, approximate enlargement of partly underthrust terranes) in the following reconstructions.

Early–Mid-Triassic reconstruction (Fig. 2)

Many workers suggested that a large basin existed in the Triassic and Jurassic between the Scythian platform and the Pontides (Şengör & Yılmaz 1981; Şengör 1984; Adamia *et al.* 1990*a*; Kazmin 1990; Ustaömer & Robertson 1993; Stampfli 1996; Stampfli *et al.* 1998). Şengör viewed this basin as a relict of Palaeotethys. However, later studies demonstrated convincingly that this large basin was formed behind a north-dipping subduction zone in which the Palaeotethyan crust was consumed. The Karakaya accretionary complex, including rocks that originated in abyssal, carbonate platform and trench settings, formed related to this subduction. In the Central Pontides a Triassic magmatic arc (the Çangaldag arc) and a back-arc basin were reconstructed (Pickett & Robertson 1996; Ustaömer & Robertson 1997; Robertson 2002). The basin has been given different names: the Küre (Ustaömer & Robertson 1993, 1997; Nikishin *et al.* 2001; Stampfli & Borel 2002) or Tauric basin (Kazmin 1990). Fragments of its oceanic crust and sediments crop out in the fold belts of North Dobrogea and South Crimea, in the Strandja zone and in the Central Pontides. They were also penetrated by drill-holes in the northwestern shelf edge of the Black Sea.

In the Tulchea zone of the North Dobrogea fold belt a continuous succession of sediments of early–mid-Triassic to mid-Jurassic age marks the northern passive margin of the basin (Gradinaru 1988, 1995). The facies become progressively deeper towards the axial zone of the belt, where late Triassic–early Jurassic flysch-type units are known. These sediments and a unit of mid-ocean ridge basalts (MORB) (Stampfli *et al.* 1998) intercalated with the deep-sea carbonates form north-vergent tectonic slices within the Niculitel nappe pile. The age of the basalts ranges from the late Early Triassic (Scythian) to Carnian (Sandulescu 1995).

Very similar to the flysch-type units of the North Dobrogea is the Tauric Series of South Crimea. This comprises proximal and distal turbidites formed on the south-facing slope and rise (Mazarovich & Mileev 1989). The oldest sediments belong to the Ladinian, and the youngest to the Mid-Jurassic. In the Norian and Early Jurassic parts of the succession there are intercalated lavas and tuffs ranging from basalts and andesite–basalts to a acidic varieties. Drilling shows that the sediments of the Tauric Series extend along the Black Sea shelf edge towards North Dobrogea (Ulanovskaya & Shevchenko 1992), thus marking the northern margin of the Triassic–Jurassic basin.

The ophiolites and associated rocks of the Küre area in the Central Pontides were first described as slices of the Palaeotethyan crust (Yılmaz & Şengör 1985). Detailed structural and geochemical studies later demonstrated that two types of ophiolites are present (Ustaömer & Robertson 1997; Robertson 2002). The first type is represented by dismembered ophiolites in the Karakaya accretionary complex. Ophiolites of the second type are interpreted as tectonic slices of oceanic crust formed in a back-arc (Küre) basin. They are covered by phyllites and

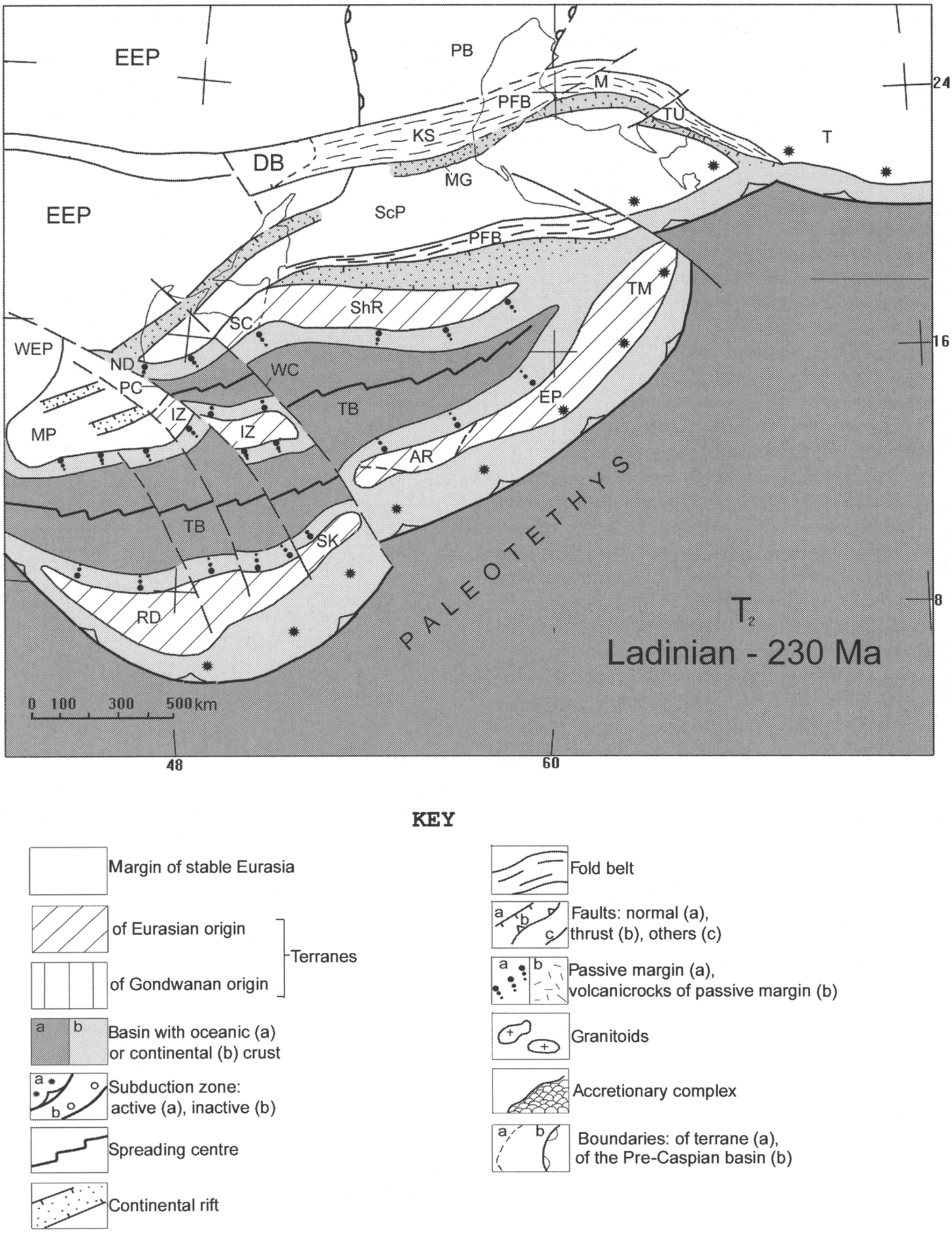

Fig. 2. Mid-Triassic reconstruction. Time of the maximum opening of the Tauric basin (Ladinian). Abbreviations as in Figure 1.

flysch-type sediments (the Akgöl Formation), of mid-Triassic to mid-Jurassic age accumulated. (Ustaömer & Robertson 1997; Robertson 2002). A close resemblance of the Akgöl Formation to the Tauric Series of Crimea has been emphasized (Bocaletti & Manetti 1988).

Data on the western extension of the Tauric basin come from the Strandja zone, on the southernmost periphery of the Balkanides. Here, low-grade metamorphic rocks of Early Triassic–Mid-Jurassic age unconformably overlie a metagranitic basement, intruded by 300 Ma granites (Okay *et al.* 2001). The cover and the basement form a series of north-vergent nappes. Ustaömer & Robertson (1993) were the first to suggest that early Mesozoic sediments were deposited on a south-facing passive margin of a back-arc basin that opened between the Moesian platform and the Rhodope massif. This interpretation was later confirmed by restoration of the pre-deformation structure, but the margin was referred to as 'Palaeotethyan' (Banks 1997; Okay *et al.* 2001). However, geological data show that in the Triassic the northern margin of Palaeotethys from Kunlun to the Pontides was active (Kazmin & Natapov *et al.* 1998). Most probably the Palaeotethyan subduction zone extended westward to south of the Rhodope massif (Golonka 2000; Dercourt *et al.* 2001). We support, therefore, the earlier suggestion that the western branch of the Tauric basin opened between Rhodope and Moesia.

The above data confirm that in Triassic time a large Tauric basin with oceanic-type crust existed between the Rhodope–Pontide fragment and the Scythian platform. The eastern part of the Tauric basin opened between the Eastern Pontides–Transcaucasus microcontinent and the Shatsky rise and its eastern extension, the Dzirula massif. To the east the Tauric basin narrowed and closed at the longitude of the present-day western coast of the Caspian Sea. Further east, in the south Turan, the Triassic active margin was of Andean type. Back-arc extension there (if any), resulted only in opening of small epicontinental basins (Boulin 1990). Accordingly, the Pontides–Transcaucasus block occupied a diagonal position relative to the Eurasian margin; this places constraints on the width of the Tauric basin at the longitude of the Crimea–Central Pontides.

It has been suggested that the Tauric basin extended directly eastwards into the Greater Caucasus (Stampfli *et al.* 1998; Nikishin *et al.* 2001), where Permo-Triassic sediments are usually included in the upper part of the Dizi Series of Svanetia (Somin & Belov 1967; Adamia 1968). Later studies have demonstrated that these sediments form an individual complex separated from the Palaeozoic sediments of the Dizi Series by a period of intensive folding and metamorphism, and that they accumulated in a back-arc basin north of the Transcaucasian massif (Kazmin & Sborshchikov 1989). However, this basin is not seen as a direct extension of the Tauric basin. In Svanetia, Triassic sediments are shallow-water quartzitic and arkosic clastic rocks, lacking volcanics rocks, and have nothing in common with the flysch and ophiolites of the Tauric basin. In the Late Triassic the Svanetia basin was inverted and unconformably overlapped by early Jurassic sediments, whereas the Tauric basin existed until the Mid-Jurassic. Finally, in the Mid-Jurassic, northward subduction of the Tauric basin was accompanied by formation of a volcanic arc on the Dzirula massif and Shatsky rise. This means that the Tauric basin was located south of the Shatsky rise, whereas the Permo-Triassic basin of the Greater Caucasus was to the north of it (Fig. 2).

The western part of the Tauric Basin consisted of two branches. The North Dobrogea branch opened between the Scythian platform and a continental fragment that was rifted from it and located within the Pontides as the Istanbul zone. In the Istanbul zone Neoproterozoic basement is covered by platform-type Ordovician and younger Phanerozoic sediments, representing part of the south-facing Palaeozoic passive margin of eastern Europe (Şengör 1984; Ustaömer & Robertson 1993; Okay *et al.* 1994; Yılmaz *et al.* 1997). The passive margin can be traced through the north Crimea to the Bechasyn zone of Fore-Caucasus, which is geologically identical to the Istanbul zone.

Data on the southern branch of the Tauric basin between the Moesian platform and the Rhodope massif are very limited. According to Banks (1997) and Okay *et al.* (2001), the northern passive margin of this branch originated in the 'earliest Triassic', whereas the final closure of the basin began in the late Mid-Jurassic. A brief period of inversion in the Carnian was followed by the accumulation of the Late Triassic Lipachka flysch. The significance of this event has been interpreted either as a transition to a compressional regime, or as a resumption of extension. The second interpretation is preferable in our opinion (see the next section for details).

There is no evidence for the development of oceanic crust in this basin, but its width could be considerable, taking into account its long, Early Triassic–Early Jurassic, period of existence.

A sharp change in the Tauric basin structure coincides with one of the major transverse features of the region, the West Crimea fault. Another major fault, the Pechenega–Camena strike-slip fault, was also active in the Early Mesozoic, constituting the southwestern transform boundary of the North Dobrogea basin (Gradinaru 1988). Both faults belong to a southeastern extension of the Tornquist lineament and were instrumental in the subsequent opening and evolution of the Western Black Sea basin, acting as NW-trending transforms (Okay *et al.* 1994; Kazmin *et al.* 2000). As these faults were convex to the east, the Tauric basin narrowed and probably closed westwards, i.e. towards the pole of opening. The zone of the Pechenega–Camena and West Crimea faults corresponds, accordingly, to the Eurelian Equator.

To estimate the probable width of the Tauric basin, the following considerations can be used. The period of spreading in the North Dobrogea branch lasted from the late Early Triassic (Scythian) to the Early Carnian, i.e. for about 15–16 Ma. At a spreading rate of 1 or 2 cm a^{-1}, the newly formed crust was about 150–300 km wide. If the southern branch was opening at the same rate, the total width of the oceanic-type basement could have reached 300–600 km. Of the previously published reconstructions, the closest to that presented in Figure 3, although more schematic, is that by Ustaömer & Robertson (1993).

Late Triassic stage (Fig. 3)

The Tauric basin was partly inverted in the Carnian. At that time a number of the Gondwana-derived microcontinents collided with the Eurasian margin. This event is well dated. In northern Iran the Lower–Middle Triassic carbonates of 'Tethyan' type changed abruptly in the Late Carnian–Norian to continental coal-bearing clastic deposits of the Shemshak Formation, typical of the adjacent regions of Eurasia (Dercourt *et al.* 1986). North of the Alborz, in the area of the future south Caspian basin, the Cimmerian fold belt was formed. The frontal nappes of this belt, containing ophiolites, are known in western Alborz and in the Aladag–Binalud (easternmost extension of Alborz) (Alavi 1996). To the west the fold belt extended into the Greater Caucasus, where the Permian-Triassic (Svanetian) rift basin was inverted.

West of Iran several smaller continental fragments docked with the Transcaucasus and Eastern Pontides. In one, the South Armenian microcontinent, the Tethyan Palaeozoic–Triassic succession and a Late Triassic transition to the Shemshak facies is well documented (Dercourt *et al.* 1986). South Armenia was either an extension of Iran or constituted an independent block. Less certain is the position of the Kirşehir massif. According to Tüysüz *et al.* (1995), this block was

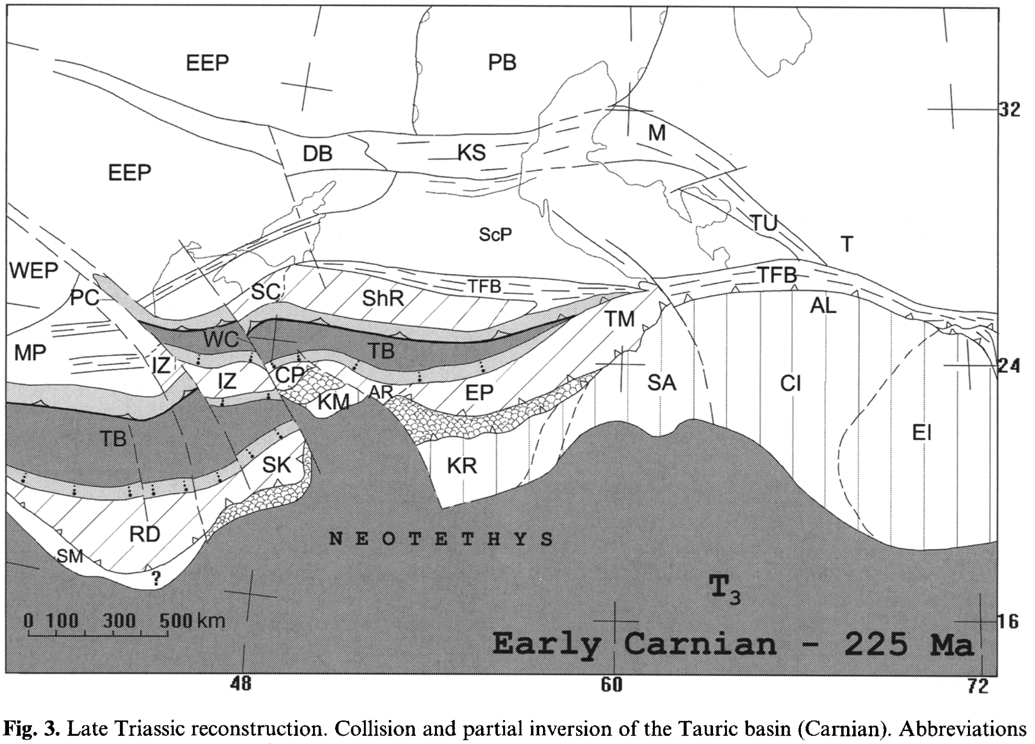

Fig. 3. Late Triassic reconstruction. Collision and partial inversion of the Tauric basin (Carnian). Abbreviations as in Figure 1; legend as in Figure 2.

rifted from the Sakarya massif in the Early Jurassic to open the Izmir–Ankara Ocean. On the other hand, in most reconstructions (e.g. Dercourt *et al.* 1985, 1993, 2001; Golonka 2000) the Kirşehir massif is regarded as a fragment of Gondwana, rifted from its margin in the Permian or Early Mesozoic. Because of very strong Alpine magmatic history and structural remobilization (Whitney *et al.* 2001), the history of Kirşehir is still poorly understood, so its inclusion in Cimmeria, as suggested here, is hypothetical. In the same category as South Armenia possibly belongs the Kargi massif, a carbonate platform within the Triassic accretionary complex of the Central Pontides (Ustaömer & Robertson 1997).

The effect of collision in the Pontides was mild: the eastern branch of the Tauric basin remained open but probably reduced. In the flysch sequences of the Crimea and the Central Pontides (Tauric Series; Akgöl Formation.) sedimentation continued from the Mid-Triassic (Ladinian?) to Carnian and Norian without a visible break or deformation that could be attributed to closure of the basin. As mentioned above, the western branch between the Moesian platform and the Rhodope massif was also not closed until the end of the Mid-Jurassic, although a short period of compression perhaps led to some shortening of the basin (Banks 1997; Okay *et al.* 2001). The compression resulted either in underthrusting (subduction?) at the northern margin of the basin (Fig. 3) and/or in the overthrusting of the Kirklareli nappe at the southern margin (Şengör 1984). In the North Dobrogea branch of the Tauric basin the onset of accumulation of late Triassic flysch is usually regarded as marking a transition from extension to compression (E. Gradinaru, pers. comm.).

The late Triassic compression was not restricted to the collision zone or back-arc basin but affected the adjacent portions of the Moesian and Scythian platforms. In the continental rift system extending from the Moesian platform to Mangyshlak and Tuarkyr (see above) the marine sediments were folded and faulted, and in some cases low-angle detachments developed (Tari *et al.* 1997; Volozh *et al.* 1999; Orel 2001). The inversion terminated in emergence and cessation of marine sedimentation. The Triassic rifts, and also the adjacent late Palaeozoic fold belts of the Karpinsky swell and Mangyshlak, were deformed and uplifted (Volozh *et al.* 1999).

Late Triassic–early Mid-Jurassic stage (Figs 4 & 5)

As a result of the late Triassic collision, the Palaeotethyan subduction zone was blocked and a new subduction zone developed south of the accreted microcontinents. This event was followed by extension, which led to rifting and then opening of the new back-arc basins in a wide back-arc region: in Iran, south of the Pontides–Transcaucasus and in the Greater Caucasus.

In Iran, the period of extension began in the Late Carnian. A system of east–west-trending continental rifts transected this territory, extending into the East Iran block. At present, the early Mesozoic rifts in this block strike NE. This implies rotation of east Iran by 90–130° anticlockwise in post-Triassic time, as confirmed by palaeomagnetic data (Soffel & Förster 1984). A spectacular discovery was made of a thick early to mid-Triassic marine clastic sequence with ammonites in central Iran, in the Anarek area (Aistov *et al.* 1984; Ruttner 1984) (Fig. 6). When rotated clockwise with the rest of east Iran, this area return to its initial position along the active Eurasian margin, where sediments of this type are known in the Aghdaraband area (Ruttner 1984).

The back-arc rifting in Iran was a reaction to the onset of subduction at its southwestern margin. The evidence for the newly formed active margin comes from the northwestern part of the Sanandaj–Sinjar zone, where upper Triassic–lower Jurassic turbidites and 'schistes lustres', intercalated with andesitic–basaltic pillow lavas, are known in the Mahabad and Esfahan area (National Iranian Oil Company 1975–1979; Cherven 1986) (Fig. 6). Following previous reconstructions (e.g. Dercourt *et al.* 1993), we believe that the Sanandaj–Sinjar block originally constituted the southwestern margin of Iran.

In Norian–early Jurassic time the Alborz was a rapidly subsiding coastal plain, on which 3000–4000 m of the coal-bearing Shemshak clastic deposits accumulated in a paralic setting. The source of the terrigenous material was to the north, where the Cimmerian fold belt was eroded (Berberian & King 1981; Davoudzadeh & Schmidt 1981, 1984; Lensch *et al.* 1984). South of the Alborz a marine basin probably occupied, an east–west rift, separating the Alborz from the rest of Iran.

West of Iran, at the southern margin of the Rhodope–Pontide fragment, extension behind a newly formed subduction zone led first to rifting (Fig. 4) and then to opening of the Izmir–Ankara–Sevan basin (Fig. 5). The oldest continental sediments, related to the rift stage, are known at the margin of the Sakarya block, where they were dated as Hettangian (Altiner & Koçiyğit 1992; Koçiyğit 1998). Upwards, they pass into the marine sequence of the passive margin, which existed through the Jurassic and

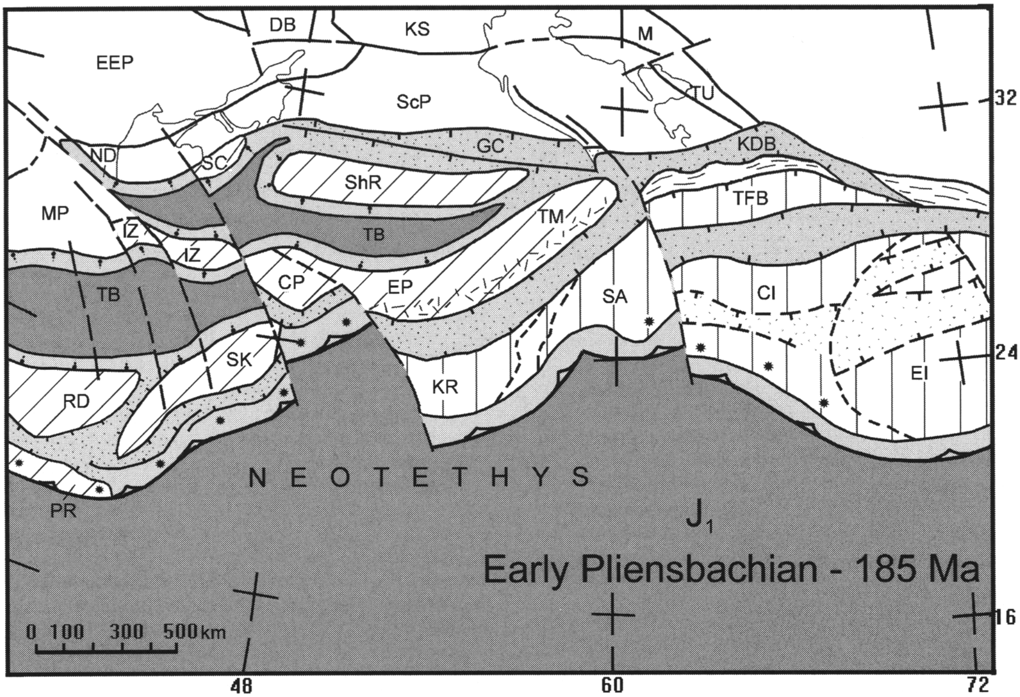

Fig. 4. Early Jurassic reconstruction. The end of the first stage of extension (Norian–Early Pliensbachian). Abbreviations as in Figure 1; legend as in Figure 2.

Early Cretaceous. According to Okay & Şahintürk (1997), the marine transgression started in the Eastern Pontides in the Early Pliensbachian and spread from the south. A thick series of volcaniclastic sediments, intercalated with beds of 'ammonitico rosso' limestones and rare flows of the andesitic–basaltic lavas (the Kelkit Formation) was deposited on the subsiding passive margin. Geochemical parameters indicate an intraplate setting of volcanism. To the east, in the Transcaucasus, the earliest continental rift sediments are also dated as Hettangian (Panov 2000). The marine volcanic–sedimentary complex, extending from the Eastern Pontides, has been penetrated by drill-holes. Thin units of rhyolites and dacites are intercalated there with transgressive sediments of Pliensbachian–Toarcian age (Lordkipanidze 1986). The Pliensbachian transgression probably coincided with the transition from rifting to spreading.

It has been suggested that the passive margin was formed as a result of rifting of an unknown microcontinent from the Pontides (Okay & Şahintürk 1997). In our reconstruction the rifted microcontinent included South Armenia, the Kirşehir massif(?) and, perhaps, some other blocks (Fig. 4). Behind the southward-migrating trench–arc system and continental fragments, the Izmir–Ankara–Sevan back-arc basin began to open. Evidence of an island arc formed on a rifted continental fragment comes from the northwestern part of the South Armenian block. According to Agamalyan (1987), on the western slope of the Tsachkunyak ridge in this area a thick (up to 6000 m) pile of lavas and volcaniclastic rocks, the Aparan Series, rests with a normal contact on the Precambrian basement. The basal unit, containing intercalations of shales and sandstones, was dated to the Toarcian–Aalenian. The overlying volcanic succession is only tentatively dated as Mid-Jurassic, although K/Ar determinations from lavas in the uppermost unit have yielded latest Jurassic to early Cretaceous ages. In the lower part of the succession basalts or andesitic–basalts are the main rock types; the upper part is built essentially of tuffs, tuffites, lava-breccia and olistostromes. Pre-late Cretaceous intrusions of tonalites, quartz porphyry and granites cut the volcanic succession. Limited petrological and geochemical studies point to an island arc setting of volcanism. No data on Jurassic volcanic activity are known from the Kirşehir massif.

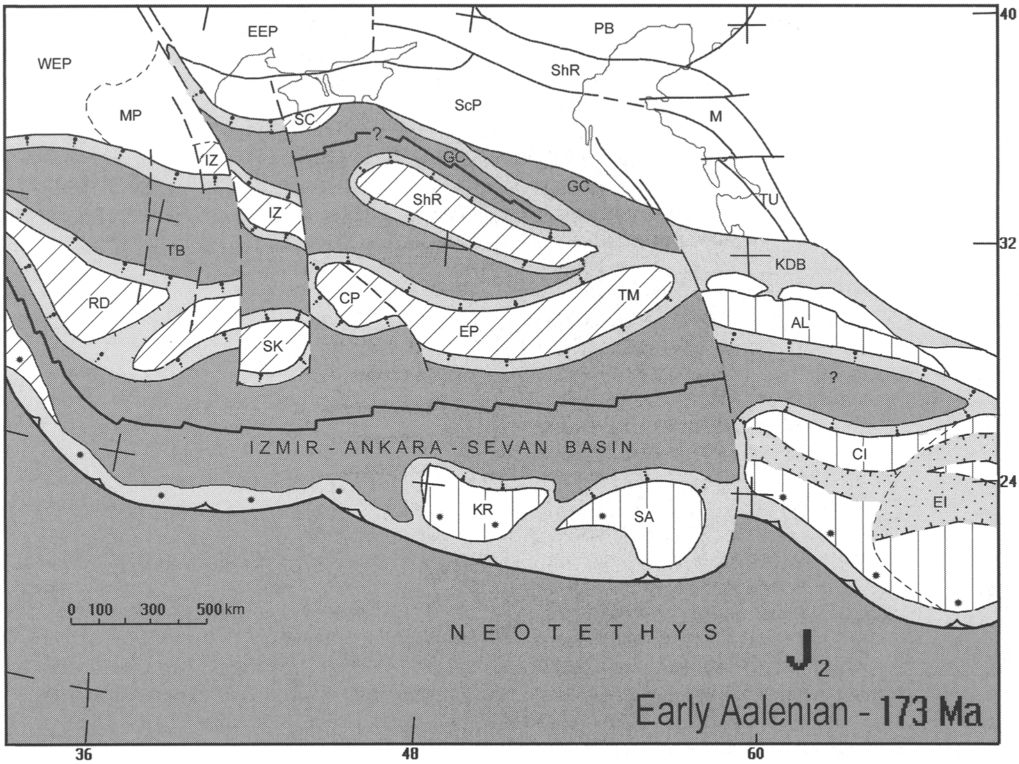

Fig. 5. Mid-Jurassic reconstruction. The end of the second stage of extension (Late Pliensbachian–Early Aalenian). Abbreviations as in Figure 1; legend as in Figure 2.

The Izmir–Ankara–Sevan 'back-arc basin' extended to the southern periphery of the Rhodope–Serbomacedonian massif. A terminal western part of Neo-Tethys between the Serbomacedonian massif and the Pelagonian block (the Vardar or Axios basin) was studied recently in detail (Brown & Robertson 2004). It was demonstrated that in the Mid-Jurassic (or earlier) a continental fragment, the Paikon ridge, was rifted from the Serbomacedonian margin, following the onset of the eastward subduction of the Vadar oceanic crust. A volcanic arc formed on top of the Paikon ridge, while spreading and opening of the Guevgueli back-arc basin was in progress during the Mid–Late Jurassic. The time and style of evolution in this part of Greece are surprisingly similar to those deduced for the Izmir–Ankara–Sevan basin. The Guevgueli basin, or a branch of it, extended eastward to Thrace (NE Greece) to form the Jurassic–Early Cretaceous Circum-Rhodope belt (Magganas 2002).

In the western branch of the Tauric basin shortening stopped and extension and opening(?) was renewed. The renewed extension was marked by rifting on the northern (Moesian) margin of the basin in the Carnian–Norian (Dabovski & Georgiev 1996; Georgiev & Byrne 1995; Sinclair *et al.* 1997). At present, the Upper Triassic–Middle Jurassic sediments of this margin crop out in one of the nappes of Stara Planina, known as the Kotel zone. As demonstrated by geological and geophysical data, the sediments (clay–carbonate shales, flysch), accumulated at a south-facing rift margin, dominated by the Golitza master fault. The Golitza and associated normal faults dissected the Early–Mid-Triassic carbonate platform (Sinclair *et al.* 1997, p. 96, fig. 5); that is, the new continental slope was formed further north then the initial Early Triassic slope. Where the southern margin of the Tauric basin was located at the time is unknown. In any case, the Late Triassic (Lipachka) flysch spread as far south as the Strandja zone (Şengör 1984; Banks 1997).

There is no direct evidence of the situation in the eastern branch of the Tauric basin. However, termination of shortening and even reopening

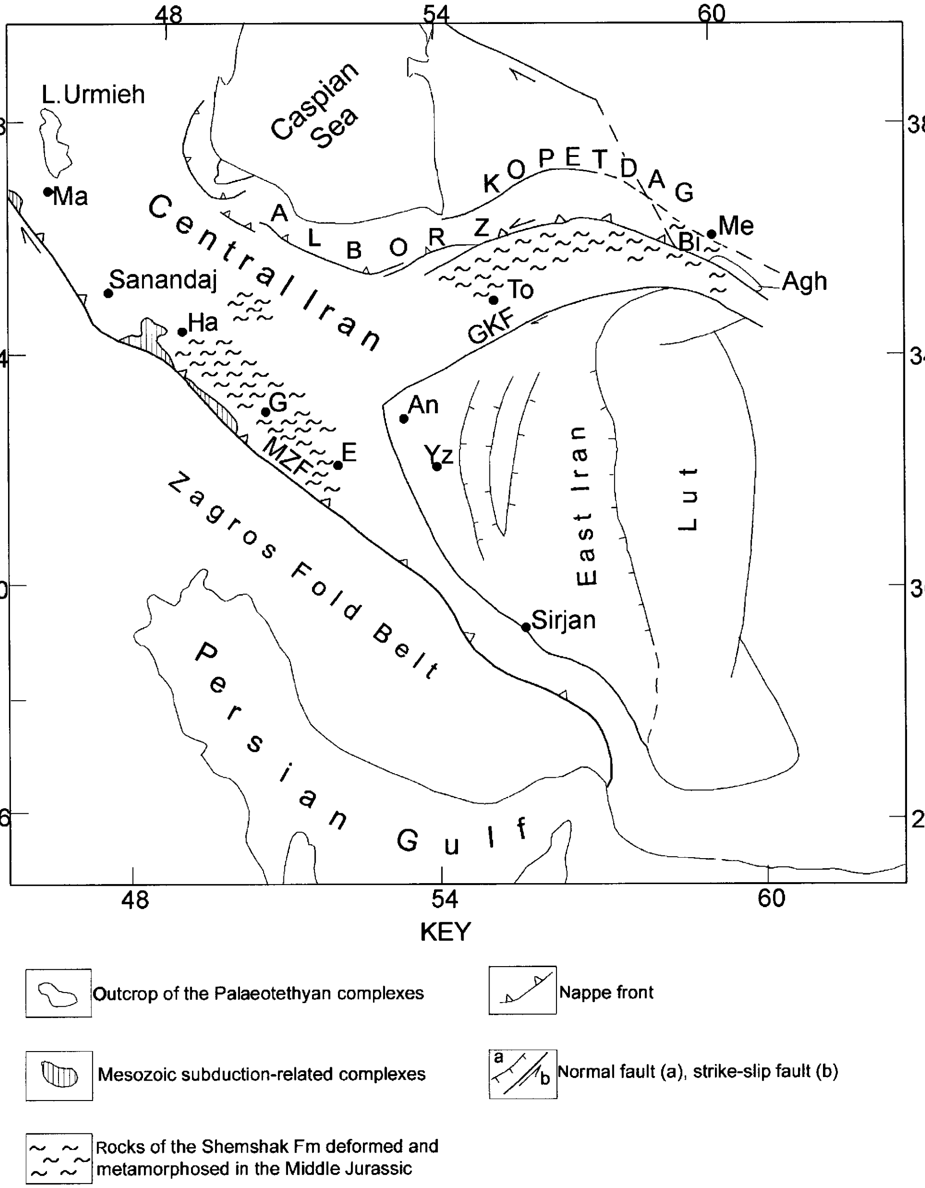

Fig. 6. Index map of Iran (after Davoudzadeh & Schmidt 1984, with some additions). An, Anarek; Agh, Agharaband; Bi, Binalud; E, Esfahan; G, Golpaigan; GKF, Great Kevir Fault; Ha, Hamadan; Ma, Mahabad; MZF, Main Zagros Fault; To, Torbat; Yz, Yezd.

are likely, because at the same time rifting started in the Greater Caucasus to the north and at the Pontides–Transcaucasus margin to the south of the Tauric basin. Thus, the whole region, it seems, was affected by extension. In this geodynamic setting, late Triassic (Norian) volcanism on the Scythian platform is particularly important. Following Carnian compressional deformation, sedimentation resumed there in the Norian in several widely dispersed subsiding basins, locally of irregular shape (e.g. the Nogaisk basin in the eastern Fore-Caucasus) and sometimes in reactivated early–mid-Triassic rifts (e.g. the East Manych grabens). Associated with shallow to moderately deep-water sediments, there are andesites, rhyolites, ignimbrites and volcaniclastic rocks. According to Nikishin *et al.* (2001), volcanic rocks have a calc-alkaline affinity, although Nikishin *et al.* noted that the data available are not sufficient for a reliable determination. Following V. E. Khain (1979), they viewed the Norian volcanic rocks as a subduction-related volcanic belt. As an alternative interpretation it can be suggested that volcanism and subsidence reflected regional extension and rifting in a wide area, far from the newly formed subduction zone, a situation also characteristic of the early Triassic extension.

The Greater Caucasus basin is usually described as a long and narrow continental rift (Nikishin *et al.* 1998, 2001). However, several workers advocated limited spreading at the later stages of the basin's evolution (Adamia *et al.* 1987; Prutsky & Lavrishchev 1989; Dotduev 1989). The late Precambrian–Palaeozoic basement of the basin is exposed in the NW of the Greater Caucasus, forming a 'crystalline core' (Fig. 7a). On the southern flank this 'core' overthrusts a thick Mesozoic succession of the southern slope along the Main Caucasian thrust. To the SE and NE the basement complexes plunge below the overlying Jurassic sediments. Thus, the Main Caucasian thrust divides the basin into two sub-basins: southwestern and northeastern.

The age of the basal Jurassic beds, transgressing the basement, is Sinemurian or younger, and this is usually accepted as the age of the Greater Caucasus basin (Nikishin *et al.* 2001; Panov 2000). However, in the deeper part of the northeastern sub-basin (e.g. along the northern tributaries of the Alazani river) Sinemurian microfossils occur within a monotonous shale sequences far upward stratigraphically from the unexposed basement. The lowermost Jurassic (Hettangian) succession is likely to be present there (Panov 2000). Hettangian and even Rhaetian sediments were described in the lower part of a continuous Jurassic succession in the southwestern sub-basin in Svanetia. A Triassic age for the lowermost part of the section was first established by the discovery of foraminifers (Saidova *et al.* 1988). Later a continuous succession, from the Rhaetian to Hettangian and Sinemurian, was proved by studies of palynomorphs (Adamia *et al.* 1990*b*). Shallow-water Upper Triassic sediments were also described in the westernmost part of the southwestern sub-basin (Krasnaya Polyana area) by Slavin (1958). Although contacts with the adjacent Jurassic rocks are tectonic, Triassic sediments may belong to the basal part of the Jurassic succession, as in Svanetia. There is enough evidence, in our opinion, to date the onset of rifting in the Greater Caucasus basin as earliest Jurassic or even latest Triassic.

The rift basin was bounded in the north by a master fault, which evolved along the Palaeozoic Tyrnyauz–Pshekish suture. Another major south-dipping fault (the future Main Caucasus thrust) transected the basin obliquely, dividing it into two sub-basins (Fig. 7b and c). The associated monoclinal block had a maximum altitude in the NW, its surface gradually subsiding to the SE and NE. In general, the structure resembled that of the Baikal rift, where the diagonal monoclinal block of the Olkhon Island and Academician ridge separates the Northern and Central basins.

In the southwestern sub-basin rifting propagated from the west, where the Greater Caucasus basin somehow connected with the Tauric basin (Fig. 5). The period of rifting lasted for about 22 Ma (Rhaetian–Early Pliensbachian). The onset of spreading, in the Late Pliensbachian–Toarcian, was marked by eruption of MORB in the axial zone of the southwestern sub-basin (Lordkipanidze 1980, 1986; Adamia *et al.* 1987), rapid subsidence and deposition of bathyal clays of the Tsiklauri horizon (Panov 2000), and transgression of the adjacent Scythian passive margin (Nikishin *et al.* 1998, 2001), which we interpret as a break-up unconformity. Spreading continued (perhaps sporadically) until the Early Aalenian, i.e. for about 14–15 Ma. The strip of newly formed crust was hardly wider than 100–150 km, because the subsequent closure of the Greater Caucasus basin was not accompanied by supra-subduction volcanism. Accordingly, the spreading rate was about 1 cm a^{-1}.

Mid-Jurassic stage (Figs 8 and 9)

Major changes in the evolution of the marginal seas occurred in the Late Aalenian, when a period of compression began. As a result, almost all of the marginal basins were closed. The onset of compression in different basins was diachronous, from the Late Aalenian to Bathonian, perhaps as a result of the great complexity of the regional geological structure.

Mid-Jurassic deformation was very important in Iran. In the Sanandaj–Sirjan zone (the Esfahan–Golpaygan–Hamadan area; see Fig. 6) the sediments of the Shemshak Formation were folded, slightly metamorphosed and intruded by diorite–granodiorite plutons with ages of 165–175 Ma (Davoudzadeh & Schmidt 1984). Similar deformation, magmatism and metamorphism affected Late Triassic–Early Jurassic sediments in a wide belt between the East Alborz and Great Kevir fault and the Binalud ridge (National

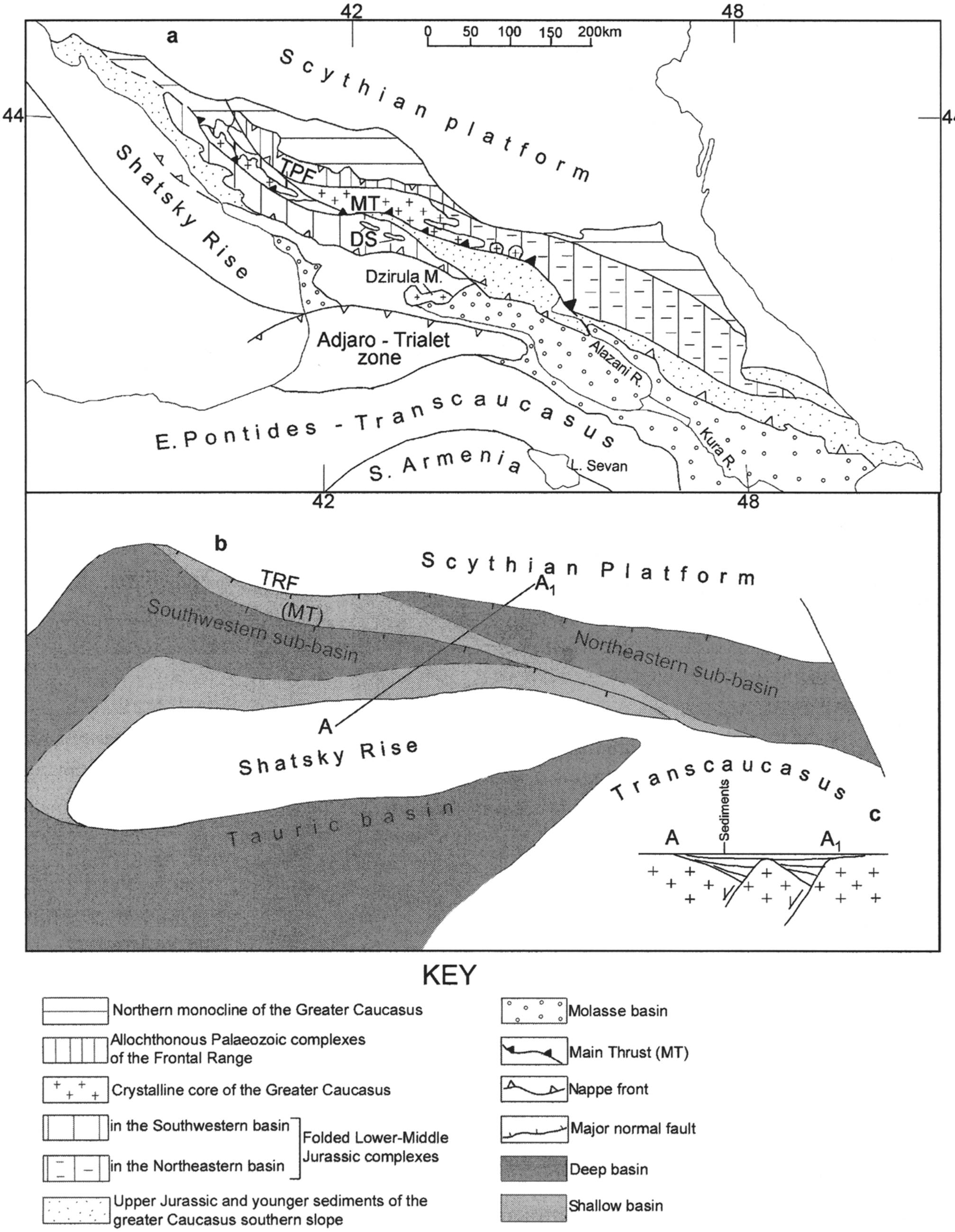

Fig. 7. (a) Main geological features of the Caucasus. **(b)** Reconstruction of the Greater Caucasus basin for early Aalenian time (without scale). **(c)** Tentative cross-section. DS, Dizi Series; MT, Main Thrust; TPF, Tyrnyauz–Pshekish Fault.

Iranian Oil Company 1975–1979; Lammerer *et al.* 1984; Lensch *et al.* 1984; Alavi 1996). It appears that Central Iran was involved in Mid-Jurassic deformation and magmatism, whereas in the Central and Western Alborz this event resulted only in uplift and emergence (Delaloye *et al.* 1981; Alavi 1996). The whole of Iran was peneplaned in post-Mid-Jurassic time and then

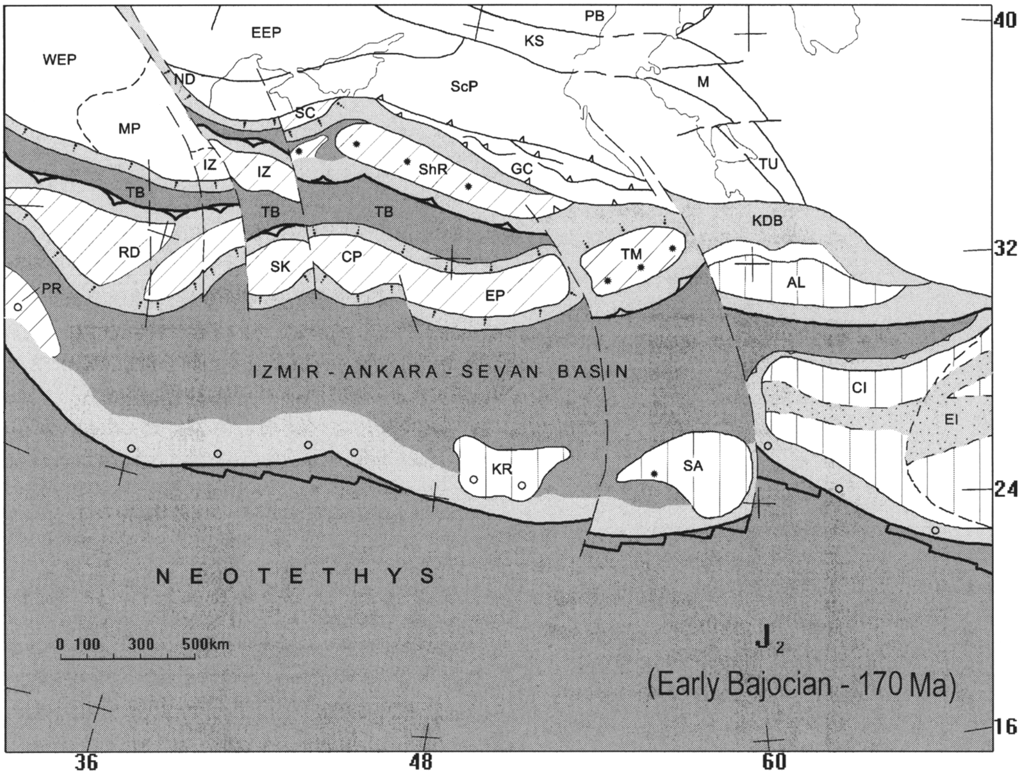

Fig. 8. Mid-Jurassic reconstruction. The end of the first stage of compression (Late Aalenian–Bajocian). Abbreviations as in Figure 1; legend as in Figure 2.

covered by a diachronous transgression during Late Jurassic–Early Cretaceous time.

The cause of the Mid-Jurassic deformation in Iran, as well as in the whole region, will be discussed later. Here, we wish to emphasize that deformation in Central–Eastern Iran may be of 'internal' origin and did not depend directly on events at its margins. It was noted that the Mid-Jurassic tectonomagmatic belt ran parallel to the northern margin of Iran, i.e. to the Alborz, and its origin was attributed to 'cratonization of the magmatic arc' (Davoudzadeh & Schmidt 1984). Sharp differences between the Alborz ('passive margin') and Central Iran ('arc') suggest that an important role in the 'cratonization' was played by the closure of the Intra-Iranian basin, as tentatively demonstrated in Figs 8 and 9.

A good record of the closure of the eastern branch of the Tauric basin is preserved in the Central Pontides and the South Crimea (Yılmaz & Şengör 1985; Bocaletti & Manetti 1988; Ustaömer & Robertson 1997; Nikishin *et al.* 1998, 2001). The oceanic crust of the Tauric basin was subducted below the Shatsky rise, on which the Bajocian volcanic arc was formed. Southward subduction of the Greater Caucasus basin should be excluded for two reasons: (1) in a narrow Greater Caucasus basin there was either no or very little oceanic crust present to generate arc magmatism lasting for about 4.0 Ma; (2) Jurassic deformation on the southern slope of the Greater Caucasus was strongly south-vergent, which is inconsistent with south-directed subduction below the Shatsky rise. In western Georgia, the Bajocian calc-alkaline arc volcanites and associated intrusions have been studied in detail on the Dzirula basement uplift (Lordkipanidze 1980, 1986; Adamia *et al.* 1990*a*) and also traced by onshore and offshore drilling to the adjacent part of the Shatsky rise. Further NW, Jurassic volcanic complexes are marked by characteristic magnetic anomalies on the Shatsky rise (Kazmin & Lobkovsky 2003). Finally, fragmemts of arc-related complexes (lavas, tuffs, volcaniclastic rocks and small dioritic plutons) crop out along the Black Sea Coast in the South Crimea, most probably in an allochthonous unit. Intrusive

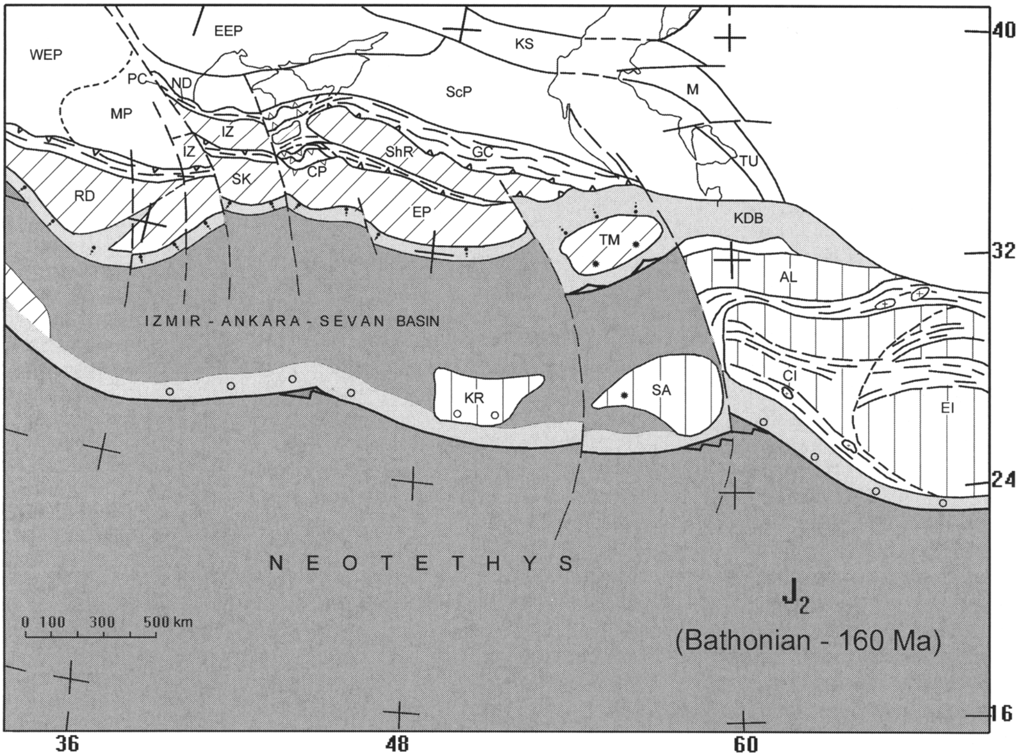

Fig. 9. Mid-Jurassic reconstruction. The end of the second stage of compression (Bathonian). Abbreviations as in Figure 1; legend as in Figure 2.

rocks of diorite–granodiorite composition and large volcanic centres (seamounts?) developed synchronously in the back-arc region of the southern slope of the Greater Caucasus. The abrupt termination of volcanic activity at the end of the Bajocian marks the collision of the arc with the Pontides.

In the Tauric Series of Crimea two stages of deformation are usually distinguished (Mazarorich & Mileev 1989; Nikishin *et al.* 1998, 2001). The early pre-Bajocian stage correlates with the onset of north-directed subduction in the Tauric basin. Following collision of the Pontides with the Shatsky rise volcanic arc and its western extension (volcanic complexes of the Crimea–Black Sea shore), a small remaining basin was compressed and finally deformed in the Bathonian. During this stage south-vergent thrusting of the Tauric Series took place. Ustaömer & Robertson (1997) demonstrated that in the Central Pontides north-vergent thrusts dominate the Küre complex and can be attributed to accretion during closure of the Tauric basin. On the other hand, opposite-verging structures in the Pontides and Crimea may have originated during a final stage of collision when convergence was directed to both sides of the relict basin towards the Shatsky rise and its western extention.

As a result of collision, Crimea was welded to the Central Pontides and an orogenic belt was formed and then eroded. Products of erosion are known as the Demerji conglomerate in the South Crimea and the Muzun conglomerate in the Pontides. It was demonstrated long ago that the source of exotic blocks in the Demerji conglomerate was to the south (Chernov 1971), i.e. in the Pontides.

In the Greater Caucasus basin the same two main stages of deformation are documented. During the pre-Bajocian stage the northeastern sub-basin was closed (Fig. 8). A system of south-vergent thrusts formed within Jurassic sediments, resembling the structure of an accretionary prism (Panov 2000). The southwestern sub-basin (south of the MCT (Main Caucasus Thrust) remained undeformed and sedimentation there continued until the Bathonian. In pre-Callovian time the sedimentary pile was thrust southward (Panov &

Prutsky 1983; Panov 2000) (Fig. 9). In front of the newly formed orogenic belt a narrow foredeep originated as a result of elastic bending of the lithosphere of the Shatsky rise. Late Jurassic–Early Cretaceous carbonates and siliclastic turbidites began to accumulate in an asymmetric trough.

A very different evolutionary trend characterized the Transcaucasus massif and the adjacent (southeastern) portion of the Greater Caucasus basin. The Bajocian volcanic arc formed on the Transcaucasian massif; however, subducting lithosphere belonged there not to the Tauric but instead to the Izmir–Ankara–Sevan basin (Figs. 8 and 9). The arc was not affected by Mid-Jurassic deformation: magmatic activity continued uninterrupted through the Late Jurassic and part of the Neocomian (Lordkipanidze 1980, 1986; Kazmin *et al.* 1986). Sedimentation on the northern margin of the massif was also continuous, indicating that the southeastern part of the Greater Caucasus basin was not closed during Mid-Jurassic inversion. Further east, this part of the basin extended through the South Caspian to the Kopetdag basin, where no Mid-Jurassic deformation is reported and sedimentation continued uninterrupted from the Mid- to Late Jurassic (Lensch *et al.* 1984).

Closure and deformation of the western branch of the Tauric basin are dated as post-Mid-Jurassic and pre-Cenomanian (Banks 1997; Okay *et al.* 2001). The youngest rocks in the north-vergent nappes of the Strandja zone have a mid-Jurassic age. A 155 Ma Rb–Sr age (biotite whole-rock) from the metagranitic basement of the Zwezdets nappe in Strandja dates regional metamorphism as Oxfordian–Kimmeridgian (Okay *et al.* 2001). Two events may provide additional information on the time of deformation: (1) At the northern margin of the basin (the Kotel zone) a transition from basinal to shallow-water facies took place in the Callovian (Georgiev & Byrne 1995); (2) in front of the Strandja nappes the Nish–Trojan foredeep evolved in the Late Jurassic and Neocomian (Okay *et al.* 2001). Its position and age are similar to the foredeep at the southern slope of the Greater Caucasus (see above). In both areas compressional deformation occurred penecontemporaneously at the end of the Mid-Jurassic to the beginning of the Late Jurassic.

No precise data are available on mid-Jurassic deformation in the North Dobrogea. Indirect evidence comes from studies by Gradinaru (1988, 1995), who documented opening of the rift basin along the Pechenega–Camena fault in a transtensional setting and simultaneous transgression on the adjacent part of the Moesian platform in the Late Bathonian. These events apparently postdate the closure of the North Dobrogea branch of the Tauric basin. Accordingly, the time of its closure is pre-Late Bathonian, i.e. probably simultaneous with the final deformation in South Crimea.

Discussion

Four major epochs can thus be distinguished in the early Mesozoic history of the northwestern margin of Tethys. The first epoch lasted for about 20–22 Ma, from the Scythian to the Early Carnian. This was a time of spreading and opening of the Tauric basin and associated basins of the North Dobrogea and the Greater Caucasus. Spreading in the Tauric basin was preceded by rifting, but evidence of this event is very limited. In the Istanbul zone of the Western Pontides there is the north–south-trending Kocaeli basin, which may represent a failed rift associated with opening of the Küre (Tauric) basin (Ustaömer & Robertson 1993, 1997). The Kocaeli basin is filled by red clastic deposits with alkaline lavas at the base (Late Permian?) and the marine succession is dated from the Early Scythian to Carnian. Continental rifting of the Scythian and Turonian platforms also commenced in the Late Permian and evolved in the Early–Mid-Triassic (Orel 2001; Glumov *et al.* 2004). The Late Permian is provisionally accepted as the time of initial rifting of the Tauric basin.

The Tauric basin opened behind the north-dipping Palaeotethyan subduction zone (Pickett & Robertson 1996; Ustaömer & Robertson 1993, 1997; Robertson 2002). Subduction commenced at the southern margin of the Pontides–Transcaucasus microcontinent after its collision with the Scythian margin in the Viséan. The time of collision is constrained by the synchronous development of Serpukhovian–Bashkirian molasse on the Scythian margin and in the Transcaucasus (E. V. Khain 1979; Belov 1981). Intrusions of granodiorites and granites, together with subaerial volcanism in the Transcaucasus (*c.* 320–250 Ma) were related to late Palaeozoic northward subduction below this massif (Adamia *et al.* 1982, 1989). A question is why the back-arc extension only began in the Late Permian? The Late Palaeozoic evolution of the active margin was interrupted in the Early Permian by a strong compressional event. At that time north-vergent thrusting affected Palaeozoic sediments of the Fore-Caucasus and the Karpinsky swell (Volozh *et al.* 1999; Glumov *et al.* 2004). Early Permian compression is known in other parts of the active margin of southern

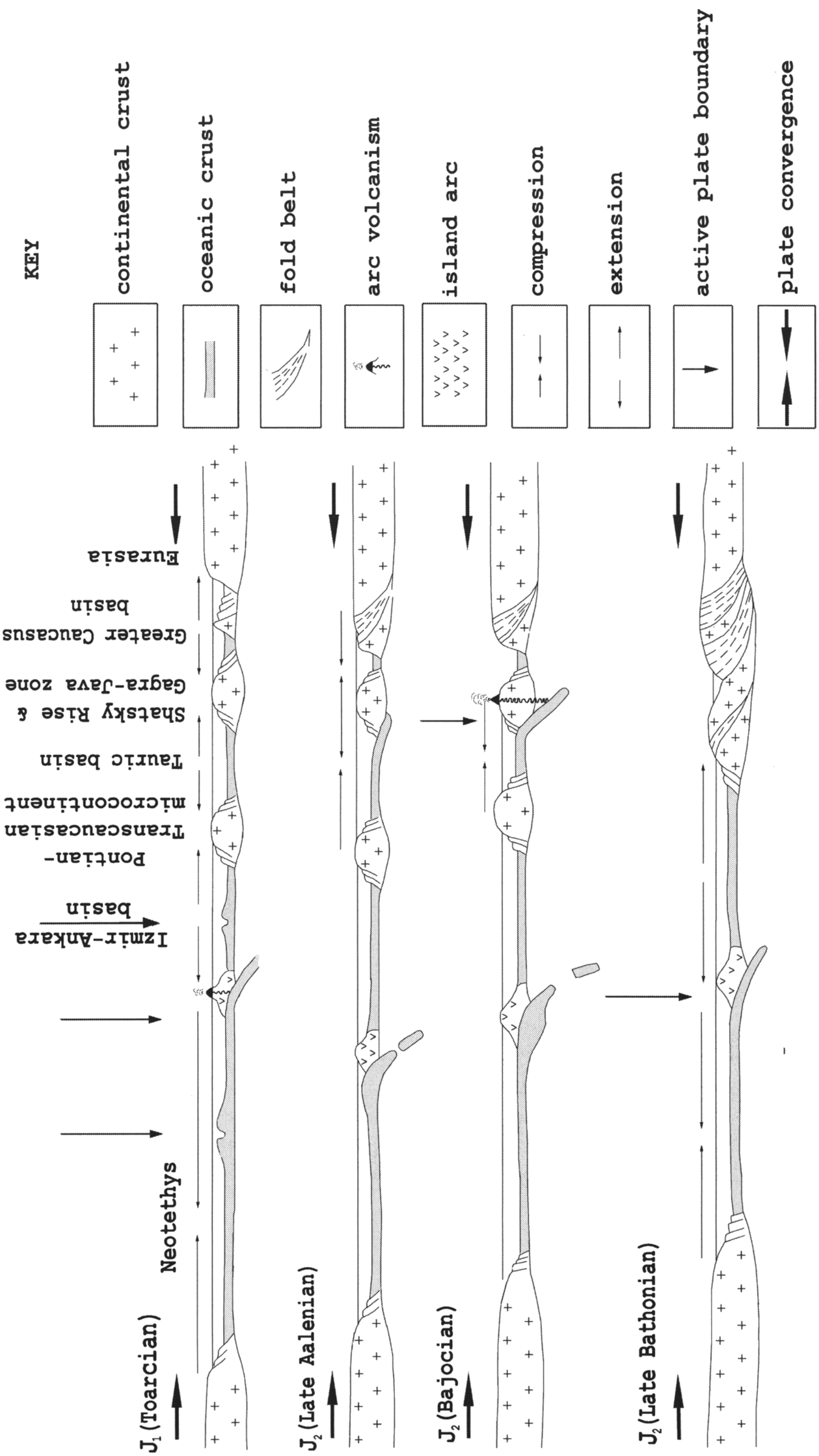

Fig. 10. Scheme illustrating evolution of the Black Sea–Caucasus sector of the Tethyan active margin in the Mid-Jurassic.

Eurasia. Kazmin (2002) noted that this event correlates with rifting of the Gondwana passive margin and formation of a new spreading centre behind a chain of separated microcontinents. It was speculated that compression at the active margin was caused by a trench–mid-ocean ridge collision and that rifting resulted from the slab-pull transmitted to the passive margin at a time, when there was no spreading centre in Tethys. According to this idea, subduction resumed at the margin of the Pontides–Transcaucasus massif in the Late Permian and was immediately followed by back-arc extension, perhaps as a result of a strong intensive roll-back effect.

The Tauric basin was partly inverted in the Carnian, when fragments of Cimmeria collided with the active margin. The main fragment of Cimmeria, Iran, had a western extension as a chain of blocks, including South Armenia, the Kirşehir massif and the Kargi platform. Perhaps because of the small size of these blocks the effect of collision in the Tauric basin was relatively mild; the basin was shortened but not closed. Small-scale shortening explains the lack of Carnian arc magmatism.

In Figure 3, underthrusting or subduction is shown at the northern margin of the Tauric basin. However, this interpretation is arbitrary. South-directed underthrusting of ophiolites and sediments of the Küre (Tauric) basin was described by Ustaömer & Robertson (1997) in their reconstruction of Central Pontides. More information is needed to determine if this structure formed in the Carnian or much later.

Following accretion of Cimmerian fragments, a subduction zone originated south of the accreted microcontinents, and a new phase of extension in a back-arc area began. The main manifestations of this extension include rifting in Iran in Carnian–Norian time (Davoudzadeh & Schmidt 1984); rifting and formation of the Golitza passive margin (Kotel zone) in Carnian–Norian time on the southern periphery of the Moesian platform (Dabovski & Georgiev 1996; Sinclair *et al.* 1997), and opening of the Greater Caucasus basin in the Latest Triassic(?)–Early Jurassic. However, these events were of secondary importance compared with rifting and opening of the Izmir–Ankara–Sevan basin between the Pontides–Transcaucasus and the fragments of Cimmeria, which started in Hettangian time (Altiner & Koçyğit 1992; Koçyğit 1998; Panov 2004). Continental rifting was followed by transgression and deposition of neritic then pelagic carbonates. In the Eastern Pontides and Transcaucasus subsidence and an extensive north-directed transgression started in the Early Pliensbachian (Lordkipanidze 1986; Okay & Şahintürk 1997), resulting, in our opinion, with the break-up of the continental lithosphere and the onset of spreading in the Izmir–Ankara–Sevan basin. The width of the newly formed back-arc basin is unknown. However, if the opening continued until the mid-Cretaceous, i.e. to the onset of subduction at the Pontide margin, its width could be very considerable. It cannot be excluded that in the narrow western part of Neotethys (the Vardar, or Axios basin) migrating island arcs collided with the northwestern Neotethyan passive margin, as suggested by Dercourt *et al.* (1986). New data do not contradict this suggestion (Brown & Robertson 2004).

In our reconstructon the Tauric basin evolved continuously from the Late Carnian to the Early Aalenian, i.e. for about 45 Ma. According to Nikishin *et al.* (1998, 2001), this uninterrupted evolution was punctured by an episode of compression and inversion in the Rhaetian–Hettangian. No convincing evidence of this event can be found in the western or eastern branches of the Tauric basin. In the former, sedimentation was continuous, at least from the Carnian–Norian to the Mid-Jurassic (Dabovsky & Georgiev 1996; Banks 1997; Okay *et al.* 2001). In the latter, no major deformation is known inside the Tauric series and its counterparts in the Central Pontides. A suspected stratigraphic lacuna in the Tauric series, corresponding to the Rhaetian–Hettangian interval (Nikishin *et al.* 1998, 2001); if present, this by no means proves the closure and inversion of the basin, but may reflect erosion or non-deposition on the continental slopes. As shown above, the Greater Caucasus basin originated in the latest Triassic–earliest Jurassic, i.e. at the time of the problematic inversion. We conclude that no Rhaetian–Hettangian inversion affected the Tauric basin.

A period of compression and closure of back-arc basins began in the Late Aalenian and continued for about 8.5 Ma until the end of the Bathonian. As a result, the Tauric and Greater Caucasus basins closed and fold belts formed in their place. The process was accompanied by pre- to post-collisional magmatic activity. South-vergent structures dominated the eastern Tauric and Greater Caucasus basins, whereas in the western Tauric basin the structure was north-vergent. The change of polarity coincides with the West Crimea fault.

The evolution of the part of the Tauric basin between the Sakarya and Istanbul blocks is still a matter of discussion. A controversy exists concerning the timing of suturing of these two blocks along the severely deformed Armutlu–Almacik zone. According to Okay *et al.* (1994), the blocks collided in the Early Eocene, thus putting an end to the opening of the Western Black Sea basin behind the Istanbul zone. Yılmaz *et al.* (1997) associated the formation of the Armutlu–Almacik zone with closure of a branch of Neo-Tethys, the Intra-Pontide Ocean, and dated this event and the emplacement of ophiolites as 'post-Turonian and pre-Late Campanian'. Recent detailed studies (Robertson & Ustaomer 2004) confirmed the existence of a discrete Intra-Pontide oceanic basin that opened in the Triassic and closed in the Turonian. However, Elmas & Yiğitbaş (2001) argued that the Sakarya and Istanbul blocks were welded together in pre-late Jurassic time and that the ophiolites were emplaced along younger strike-slip faults. It is possible that the Intra-Pontide Ocean was part of the Tauric (Küre) basin, situated between the Istanbul and Sakarya blocks (fig. 2; see also Ustaömer & Robertson 1993, p. 234, fig. 10). It was possibly closed in the Mid-Jurassic together with the whole Tauric basin. However, one cannot exclude that it reopened in the Late Jurassic in connection with dextral motion on the Pechenega–Camena fault in a transtensional setting (Gradinaru 1995).

Comparison of the reconstructions in Figures 5 and 9 shows that the minimum Mid-Jurassic shortening along the Pontides–Greater Caucasus transect was about 300–400 km at a convergence rate of 3.5–4.5 cm a^{-1}. The motion of Africa–Arabia relative to Eurasia at this time was essentially left-lateral (Savostin *et al.* 1986), and the convergence between the two plates totals only a few hundred kilometres (Dercourt *et al.* 1985, 1993). Most, or all, of this convergence, was probably compensated by the closure of the back-arc basins.

What caused the compression at the northern margin of Tethys in the Mid-Jurassic? As no collision with continental blocks occurred at that time, one must look for a tentative explanation at the remote plate boundaries. At the beginning of the Mid-Jurassic, spreading in the Izmir–Ankara–Sevan basin had already been active for about 16–17 Ma (from the Pliensbachian to Early Aalenian). At a rate of 5–6 cm a^{-1} the width of the basin reached 800–1000 km (Fig. 10; also see Figs. 8 and 9). According to global reconstructions, the width of Tethys in its westernmost part was about 2000–2200 km (Golonka *et al.* 1996; Golonka 2000; Dercourt *et al.* 2001). As a result, the southward-migrating arc system at the southern front of the Izmir–Ankara–Sevan basin was able to collide with a Tethyan mid-ocean ridge. When subduction was temporarily blocked, convergence between the main plates was compensated by shortening and closure of the back-arc basins. Compression at the northern Tethyan margin terminated at the end of the Bathonian, when subduction at the southern front of the Izmir–Ankara–Sevan basin was renewed.

Conclusions

Evolution of the early Mesozoic back-arc basins in the Black Sea–Caucasus region was governed by several factors, as follows.

(1) Two major periods of extension and opening of the Tauric and associated basins (Permian(?)–Early Triassic and Late Triassic–Early Jurassic) immediately followed formation of new subduction zones. This implies that the initiation of subduction was succeeded by the rapid sinking of a dense slab composed of the old oceanic lithosphere at the margin of the Palaeozoic or Permian–Triassic ocean. Extension created by resulting roll-back affected a wide (up to 1000 km) area of the back-arc region.

(2) Partial inversion of the Tauric and associated basins in the Carnian was related to closure of Palaeo-Tethys and collision of the Cimmerian fragments with the active margin of Eurasia.

(3) The major compressional event in the Mid-Jurassic resulted in deformation and closure of the Tauric and Greater Caucasus back-arc basins. This event probably coincided with ridge–trench collision at the southern margin of the opening Izmir–Ankara–Sevan basin. For a period when the intra-oceanic subduction zone was blocked, Africa–Eurasia convergence was compensated by shortening and closure of the back-arc basins.

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