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The South Caspian Basin: a review of its evolution from subsidence modelling

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Abstract

The basement surface of the South Caspian depression lies at a depth of 20–25 km, making it one of the deepest basins in the world. It occupies the southern, deep-water, part of the Caspian Sea and two adjacent lowlands: the West Turkmenia in the east and the Lower Kura in the west. The basin can be subdivided into several sub-basins with two main depocentres, one in the northern part of the basin, just on the southern flank of the Apsheron Sill, and one, called the Pre-Alborz trough, located in the south-eastern part of the marine basin.

The sedimentary fill of the South Caspian Basin has been significantly deformed. Part of it is allochthonous and folded, overlying a ductile detachment zone within the Maikop shale (Oligocene–Early Miocene). The folded succession is unconformably overlain by Upper Pliocene–Quaternary neo-autochthonous sediments. An intense shortening event, related to the NNE–SSW convergence of the Arabian plate with Eurasia, affected the region during the Pliocene–Pleistocene. The thickness of Pliocene–Quaternary sediments alone reaches 10 km. They were deposited in a rapidly subsiding basin and were sourced from the surrounding Caucasus, Alborz, and Kopet-Dagh orogens as well as from the nearby Russian Platform.

The thickness of the crust beneath the western central part of the basin is as little as 8 km in the western central part of the basin but exceeds 15 km in the eastern part. Geophysical data and gravimetric modelling provide evidence that the basement of the marine part of the basin comprises a high-velocity, thin complex crust. Subsidence of the basin is in part due to profound thinning of continental crustal or, more likely, to the formation of oceanic crust. This took place in Middle–Late Jurassic times, in the context of back-arc basin development, with possible reactivation during the Cretaceous. However, there remains controversy regarding the timing of oceanic accretion and deep-water deposition in the South Caspian Basin. The present results are based on subsidence analysis complemented by geological data from tectonic units surrounding the South Caspian Basin and on its margins. An additional mechanism, nevertheless, must be invoked to explain the younger, much more rapid Pliocene–Quaternary phase of subsidence that occurred simultaneously with the subsidence of Caucasus-related molasse basins and the uplift and erosion of the Caucasus Orogen. This rapid subsidence phase is probably of compressional origin and a simple elastic model in compression provides comparable amplitudes of

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subsidence. In addition, the South Caspian Basin is surrounded by orogenically loaded crust that adds to basin downwarping. To the north, the basin is bounded by a subduction zone.

Keywords: South Caspian Basin modelling; Caucasus–Alborz–Kopet-Dagh; Back-arc spreading; Syn-compressional subsidence; Plate tectonics

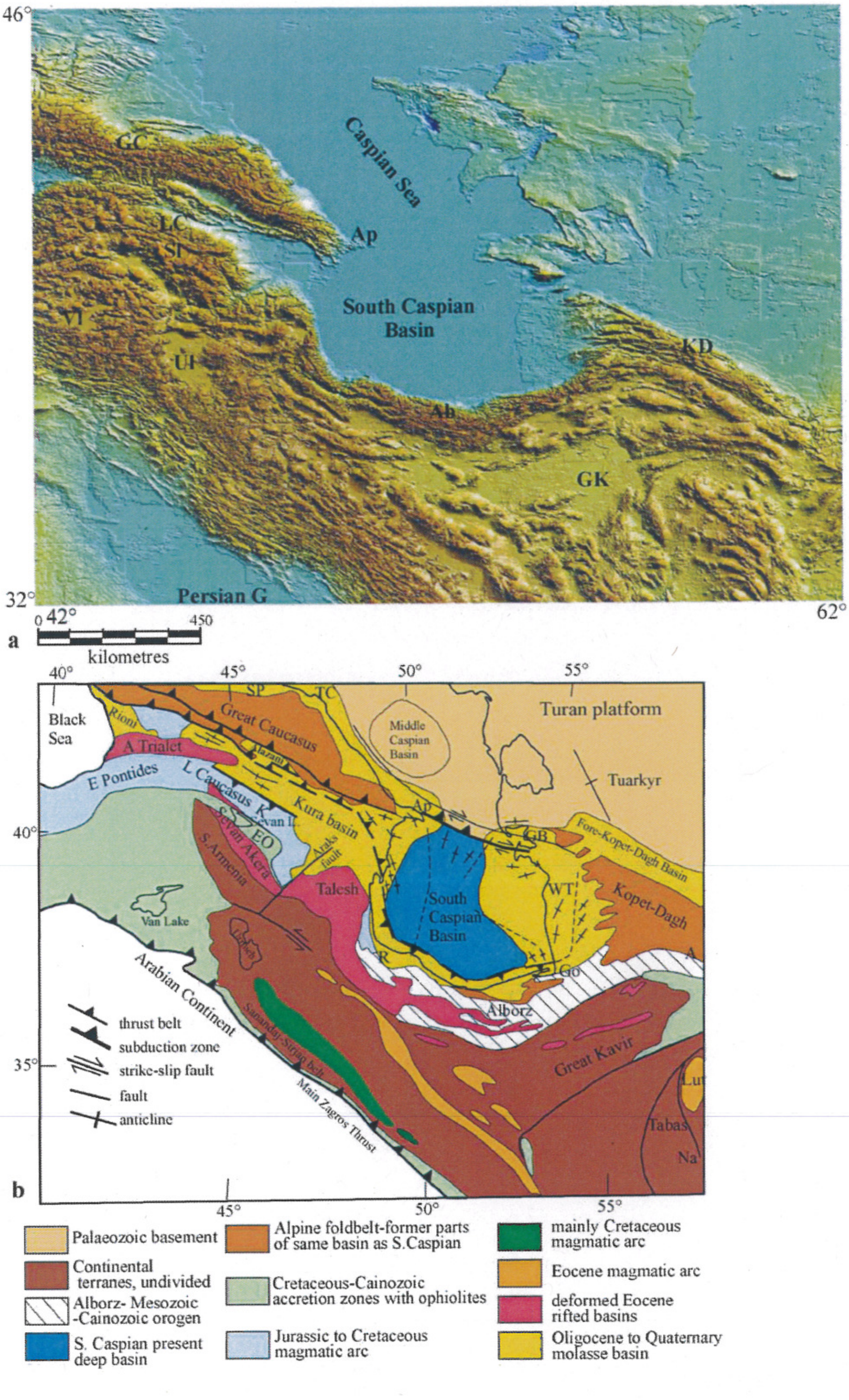
1. Introduction

The sedimentary fill of the South Caspian Basin (SCB) comprises more than 20 km Mesozoic to Cainozoic deposits. About a half of this was deposited in the very short Pliocene–Quaternary interval, within a regional compressional tectonic setting related to the Alpine closure of Neo-Tethys during Arabia–Eurasia collision. Another important compressional event that affected the area of the SCB before its origin was the Eocimmerian or Early Cimmerian orogeny. Cimmerian terranes detached from Gondwana, collided with Eurasia, closing (except in the east), by northwards directed subduction, the Palaeo-Tethys ocean (e.g. Belov, 1981; Sengör et al., 1984; Kazmin and Sborschikov, 1989; Zonenshain et al., 1990; Dercourt et al., 1993, 2000; Ricou, 1996; Stampfli et al., 2001). At the same time, the Neo-Tethys opened to the south of the migrating Cimmerian terranes. Palaeomagnetic data from Iran indicate northwards convergence of the (Cimmerian) blocks in Iran ceased at the end of the Middle Triassic (Saïdi, 1995; Besse et al., 1998). These blocks collided with the Turan plate, which at that time formed the southern margin of the Eurasian continent. Scattered remnants of Palaeo-Tethys indicate that its suture with Eurasia lies in the area of the SCB. These include the Aghdarband suture (Mashad area) in the Binalud mountains north-east of Iran (Stöcklin, 1968; Alavi, 1991; Ruttner, 1993) and Rasht suture (Talesh or Talysh in the Russian literature) in the north-west of Alborz (Davies et al., 1972; Sengör, 1984) (Fig. 1). After accretion, subduction migrated to the south of the accreted Cimmerian blocks and this became the active northern margin of the Neo-Tethys. Long-lived northerly directed subduction of the Neo-Tethys led to opening, during the Mesozoic, of the Great Caucasus–Caspian back-arc basin, the eastern part of which representing the origin of the SCB.

The effects of the middle and late Cimmerian orogenic phases are less evident in the area of the SCB. During the Middle Jurassic, the first of these compressionally affected most of the Black Sea–Pontides area of the Great Caucasus–Caspian back-arc basin. It is recognised by a Bajocian to Bathonian discontinuity, according to location and authors (in any case, pre-Callovian) that was accompanied by subduction related magmatism. The unconformity is mainly documented in Crimea and North Fore–Caucasus (Belov, 1981; Zonenshain et al., 1990). It could be related to collision of blocks or changes in the Neo-Tethys subduction geometry in the Pontides due to plate boundary reorganisations (Stampfli et al., 2001; Ziegler et al., 2001) leading at least in part to closure of the back-arc basin system. The late Cimmerian phase, at the end of Jurassic–Early Cretaceous, is contemporaneous with accretion of several continental blocks, including the Lhasa and Helmand blocks (e.g. Ricou, 1996; Otto, 1997), to Eurasia east of the SCB area.

A Cretaceous to Cainozoic accretionary complex with remnants of oceanic crust has also been recognised in several places. The Sevan–Akera suture is among these remnants; it lies to the south of the Lesser Caucasus Mesozoic volcanic arc (a prolongation of the Ankara–Erzincan suture, south of Eastern Pontides; Okay and Tüysüz, 1999). The Nain–Sabze-

Fig. 1. (a) " Digital Elevation Model " of the area (topography data are issued from the ETOPO 2.5 database of National Geographic Data Center). Ab: Alborz; Ap: Apsheron; GC: Great Caucasus; GK: Great Kavir; KD: Kopet-Dagh; LC: Lesser Caucasus; Sl: Sevan lake; Ul: Urmieh lake; Vl: Van lake. (b) Simplified geologic map of the South-Caspian area showing the main tectonic units. A: Aghdarband; Ap: Apsheron; A Trialet: Achara-Trialet; EO: Erevan-Ordubad; E Pontides: Eastern Pontides; GB: Great Balkhan; Go: Gorgan; K: Karabakh; L Caucasus: Lesser Caucasus; Na: Nayband; R: Rasht; SP: Scythian platform; TC: Terek–Caspian basin; WT: Western Turkmenia.


46°
32°
42°
62°
0
450
kilometres
a
Caspian Sea
GC
LC
Ap
SI
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UI
South Caspian Basin
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Ab
GK
Persian G
40°
45°
50°
55°
35°
b
Black Sea
Rioni
A Trialet
Great Caucasus
SP
TC
Middle Caspian Basin
Turan platform
Tuarkyr
E Pontides
L Caucasus
K
Kura basin
Sevan L.
EO
Sevan Akera
S.Armenia
Araks fault
Talesh
Ap
GB
Fore-Kopet-Dagh Basin
Kopet-Dagh
WT
South Caspian Basin
Van Lake
Urmia
R
Go
A
Alborz
Arabian Continent
Sanandaj-Sirjan belt
Main Zagros Thrust
Great Kavir
Lut
Tabas
Na
thrust belt
subduction zone
strike-slip fault
fault
anticline
Palaeozoic basement
Continental terranes, undivided
Alborz- Mesozoic -Cainozoic orogen
S. Caspian present deep basin
Alpine foldbelt-former parts of same basin as S.Caspian
Cretaceous-Cainozoic accretion zones with ophiolites
Jurassic to Cretaceous magmatic arc
mainly Cretaceous magmatic arc
Eocene magmatic arc
deformed Eocene rifted basins
Oligocene to Quaternary molasse basin

var, Nain–Baft (Stöcklin, 1981), and other such fragments are situated around the central east Iran blocks. They may be the record of the closure of a marginal sea that opened during Jurassic times and/or to the displacement of Cimmerian blocks (hypotheses varying according to the reconstructions, e.g. Dercourt et al., 1993, 2000; Sengör, 1990; Ricou, 1996; Stampfli, 2000; Stampfli et al., 2001). Another possibility is that they are relics of a branch of the Neo-Tethys, assuming that some continental fragments were detached from Gondwana during the Jurassic (e.g. Adamia, 1991; Golonka, 2000b; Golonka et al., 2000).

The subduction of oceanic lithosphere (either a marginal sea or part of the Neo-Tethys) below the Lesser Caucasus volcanic belt was complete at the end of the Cretaceous. After obduction of ophiolites in the Coniacian (Knipper et al., 2001), the Sevan–Akera suture formed in Late Cretaceous or Palaeogene times (Knipper and Sokolov, 1974), as a result of accretion of the South Armenia continental terrane (named also Nakhichevan, Daralagez, Jolfa, Djulfa or Lesser Caucasus block) to the Transcaucasus block (the area situated to the south of the Great Caucasus). On the basis of palaeomagnetic data, the South Armenia block, as is the Sanandaj–Sirjan block (south-west of Iran), is generally thought to have accreted to Eurasia as part of the Eocimmerian accretionary phase, but probably not at their present longitude from the present one (Sengör, 1990; Saïdi, 1995; Stampfli, 1999 personal communication). It is thought that these blocks were subsequently either separated during the Cretaceous prior to renewed collision (Dercourt et al., 1993; Ricou, 1996) or moved laterally along strike–slip faults (Sengör, 1990). Cainozoic rotations possibly accompanied block translation (Saïdi, 1995). Other authors (for example, Bazhenov et al., 1996), have proposed, on the basis also of palaeomagnetic data and Gondwanan Bajocian fauna, that the South Armenia block and perhaps the Sanandaj–Sirjan block were still attached to Gondwana in the Early Jurassic. They then separated from Gondwana, crossing the Neo-Tethys, being accreted to Eurasia in the Late Cretaceous (to Early Palaeogene?) (Knipper and Sokolov, 1974; Zonenshain and Le Pichon, 1986; Adamia, 1991; Golonka, 2000b; Golonka et al., 2000; Knipper et al., 2001). In particular, Zonenshain and Le Pichon (1986) proposed this accretion to have taken place during the Late Senonian with a jump of the subduction zone to the south of the accreted terrane. In this reconstruction, the repositioning of the subduction zone led to the opening of a regionally widespread back-arc basin extending from the Black Sea to the South Caspian Sea during the Eocene.

Little has been published on the general evolution of SCB (Zonenshain and Le Pichon, 1986; Devlin et al., 1999; Golonka, 2000a; Stampfli et al., 2001). Some published work was dedicated to the numerical modelling of subsidence evolution of the SCB (Bagirov et al., 1997; Nadirov et al., 1997; Tagiyev et al., 1997; Korotaev, 1998). The aim of this study is to use new subsidence analysis combined with inferences from other sources about the regional tectonic setting to place better constraints on the timing and possible mechanisms which led to the origin and development of this unique basin.

2. Tectonic surroundings of the South Caspian Basin

The oldest sediments of the SCB are inaccessible because of their great depth. Accordingly, to understand the geodynamic and tectonic context of the formation of this basin and its evolution, it is necessary to consider the geology of the surrounding area. This comprises numerous distinct tectonic units, including platformal regions as well as orogens and their forelands and magmatic belts (Fig. 1), that have been affected by various extensional and compressional phases.

2.1. Platforms

The platforms are the *Turan Platform* to the north-east of the SCB and the *Scythian Platform* to the north-west. They were both affected during the Late Triassic (up to the beginning of the Jurassic according to some authors) by Eocimmerian events (Belov, 1981; Milanovsky, 1991; Sengör et al., 1984; Maksimov, 1992; Nikishin et al., 1998a,b, 2001; Thomas et al., 1999). A weaker compressional event is reported in the Late Jurassic–Early Cretaceous (Thomas et al., 1999; Nikishin et al., 2001). Thereafter, the southern parts of these platforms became the foreland areas, respectively, of the Kopet-Dagh and the Great Caucasus orogens.

2.2. Orogenic belts

The orogens surrounding the SCB include the Great Caucasus, the Alborz, the Kopet-Dagh, and the Great Balkhan. The first of these, the *Great Caucasus* was formed as a result of shortening and closure of a former Jurassic–Eocene back-arc basin, the Great Caucasus Trough, which was underlain by very thin continental (and possibly oceanic locally) crust and opened during the Early Jurassic (e.g. Zonenshain and Le Pichon, 1986; Khain, 1994; Nikishin et al., 2001). An unconformity of pre-Callovian age (Belov, 1981) is interpreted to be the result of partial "closing" of the trough on its northern margin (Zonenshain and Le Pichon, 1986). A new trough opening event, along the southern part of the former deep-water basin, took place during the Callovian–Late Jurassic (Zonenshain and Le Pichon, 1986; Nikishin et al., 2001). Possible tensional events occurred also during the Albian–Cenomanian and the Middle Eocene (Nikishin et al., 1998a,b, 2001). The closure of the Great Caucasus Trough occurred progressively from Late Eocene times, with the main phase of collision being in the Middle Miocene and a major uplift occurring during Late Pliocene–Quaternary times (Belov et al., 1990; Milanovsky, 1991; Nikishin et al., 1998b, 2001; Ershov et al., 1998, 1999, 2003).

The *Alborz* is located on the southern border of the SCB. Structural analyses of this area were published some years ago (e.g. Wensink and Varekamp, 1980; Berberian and Berberian, 1981; Berberian, 1983; Haghipour and Aghanabati, 1989; Saïdi, 1995; Alavi, 1996). The orogen has a late Precambrian basement covered by Devonian to Middle Triassic sedimentary layers with a thickness reaching 2–3 km. This cover was only slightly affected by the Eocimmerian orogeny during the Middle–Late Triassic (Davoudzadeh and Schmidt, 1984; Saïdi, 1995; Saïdi et al., 1997). Overlying the unconformity of this age is the Shemshak clastic formation, which contains coal measures. Its age is Norian to Bajocian (e.g. Davoudzadeh and Schmidt, 1984; Stampfli, 1978; Saïdi, 1995). Sediments of the Upper Bajocian to Bathonian age are not present to the north of the Alborz but sedimentation was continuous in the eastern part (near the Kopet-Dagh), where a marine transgression onto the Shemshak formation began in the Late Bajocian (Davoudzadeh and Schmidt, 1984). Callovian and Late Jurassic sediments are represented mainly by carbonates and marls. The Neocomian is not known in the southern part of the Alborz but, in the north-central and eastern Alborz, sedimentation is continuous from the Jurassic to Cretaceous, including marine Neocomian (Davoudzadeh and Schmidt, 1984). Wensink and Varekamp (1980) demonstrated that basaltic volcanism in the central Alborz commenced in the Neocomian (Barremian alkaline olivine basalts) and continued into the Turonian–Lower Senonian (high alumina basalts—highest member of the volcanics). However, the age of these volcanic rocks is not well constrained. Late Cretaceous is represented by different facies; for example, in the western Alborz it consists of volcanics and volcano-sediments, after a lacuna from the Middle Jurassic to the beginning of the Late Cretaceous. A Late Cretaceous granitoid body intruded to the west of central Alborz (Berberian and Berberian, 1981; Haghipour and Aghanabati, 1989). In north-central Alborz, Cretaceous consists of a thick series of carbonates and marls. Possible compressional deformation occurred in pre-Eocene times. The Palaeocene and Lower Eocene have an irregular distribution in the Alborz, from a few metres up to 3000 m, including andesitic and basaltic volcanic rocks in the west, indicating, according to Saïdi (1995), the proximity of an active volcanic margin. Eocene strata are not present in the northern part of the Alborz. An important Eocene trough, filled mainly by the Middle–Upper Eocene Karaj formation, developed along the southern part of the Alborz. It comprises more than 4000 m of volcanics and volcano-sediments in western Alborz, decreasing to 3000 m in the east and passing to tuffs then to shales and sandstones. During the Oligocene and the Miocene, the region was uplifted due to the collision of Arabia with Iranian terranes; at the same time, continental or marine sediments were deposited in internal small basins (Davoudzadeh et al., 1997). Finally, a major uplift of the Alborz occurred from Pliocene to Quaternary times.

The *Kopet-Dagh* is located to the east of the SCB and fringes the north of the easternmost part of the Alborz. The Kopet-Dagh was affected by Eocimmerian deformation and is separated from the Alborz by fragments of the Palaeo-Tethys suture (Agdharband).

Subsequently, a thick sedimentary basin developed within its margins with an almost continuous succession of Jurassic to Tertiary (about 10 km) of marine deep-water to continental sediments (Davoudzadeh and Schmidt, 1984; Milanovsky, 1991; Maksimov, 1992). The Lower–Middle Jurassic, composed of 1.5 km of siltstones and sandstones, is only poorly investigated by wells. It is interpreted as being a strictly marine and even partly deep-water formation (Afshar Harb, 1979; Milanovsky, 1991; Khain, 1994). After a Bathonian lacuna (Afshar Harb, 1979; Lasemi, 1995), a Callovian–Late Jurassic carbonate platform adjacent to a deeper marine basinal environment developed on the northern margin of the basin (Lasemi, 1995). The carbonate succession, over 1200 m in thickness, includes shelf to open marine facies. The Jurassic/Cretaceous boundary is marked, on the basin margin, by an erosional unconformity (Lasemi, 1995) interpreted as indicative of Late Cimmerian deformation (Milanovsky, 1991) but it also corresponds to a global regression. Nevertheless, in the middle of the Kopet-Dagh Basin, sedimentation seems to have been continuous (Khain, 1994). During the Neocomian up to 1.1 km of carbonates, marls, minor shales and evaporites were deposited in a shallow-water environment. In the east, these grade into fluvial sandstones (Moussavi-Harami and Brenner, 1992). In the Barremian–Cenomanian, up to 2.5–3 km of carbonates and clastic deposits enriched by glauconites and phosphorites were deposited. The Turonian–Maastrichtian succession is composed of 1 km of carbonates, marls and clays; the Palaeocene of a few hundred metres of carbonates and marls; and the Eocene consists mainly of 500–1000 m of clays. All data from the beginning of the Cretaceous are derived from the margins of the basin and it is not known if an area with deep-water environment in the main basin depocentre had persisted during these periods. The Kopet-Dagh was folded during the late Alpine Orogeny (Zonenshain et al., 1990), mainly from Middle Miocene, and thrusted onto the southern edge of the Turan Platform, leading to the formation of the Fore Kopet-Dagh Basin to the north of the orogenic belt.

The *Great Balkhan* to the north-east of SCB is considered as the north-western prolongation of the Jurassic Kopet-Dagh basin (Milanovsky, 1991; Maksimov, 1992). Zonenshain and Le Pichon (1986) consider it also as an eastern prolongation of the Great Caucasus Bajocian basin. The Early–Mid Jurassic marine argillaceous and silty arenaceous succession exceeds 4 km in thickness; it is covered by Upper Jurassic shelf carbonates of some hundred metres, themselves overlain by calcareous Neocomian with an important angular discordance (Prozorovskiy, 1985; Khain, 1994). From the Cretaceous onward it formed a separate uplift, which can be regarded as a Late Cimmerian structure and is thrusted towards the north on the Turanian platform edge (Khain, 1994).

2.3. Foreland basins

The foreland basins of the previously mentioned orogenic belts also record information of relevance to the SCB. The *Terek–Caspian molasse basin* is situated to the north-east of the Great Caucasus. During the Middle Jurassic, prior to the development of a pre-Callovian unconformity, this basin is considered to have been the foreland basin of the Middle Jurassic Great Caucasus "orogen" (Koronovsky et al., 1987, 1990; Milanovsky, 1991; Ershov et al., 1998, 1999, 2003; Nikishin et al., 2001). However, the Great Caucasus Basin was not totally closed at this time, although there may have been underthrusting at its northern margin. Subsequently, from Callovian–Oxfordian times, the Terek–Caspian basin started to subside rapidly as a shelf area for the "renewed" Great Caucasus deep-water trough. Subsidence rate was not uniform, sometimes rapid and sometimes slow, from the Cretaceous until the Eocene (Nikishin et al., 1998a,b; Ershov et al., 1998, 1999, 2003). From the latest Eocene–Oligocene, the basin developed as a pre-foreland basin (Ershov et al., 1998, 1999, 2003). Acceleration of subsidence and true foreland basin development began in the Late Miocene, simultaneously with the main collision phase of the Alpine Great Caucasus. The total Oligocene to recent subsidence has been close to 7 km.

The *Fore Kopet-Dagh molasse basin* is located to the north of the Kopet-Dagh orogen with a 40–50-km average width. The Upper Oligocene to Quaternary continental-marine molasse (2–2.5 km) filled it with a discordance between the Miocene and the Pliocene (Khain, 1994). The molasse is underlain by a 6–8 km thick Jurassic–Palaeogene series deposited on the margins of the Kopet-Dagh Basin.

2.4. Magmatic or volcanic belts and associated basins

The *Karabakh magmatic belt* is located in the Transcaucasus area. It comprises Lower Jurassic shales and clastics; a pre-Bajocian gap; a series of Bajocian to Late Cretaceous marine sediments with peaks of volcanic activity in the Bajocian–Bathonian, in the Kimmeridgian and occasionally from the Albian to Maastrichtian (see, for a more detailed compilation, Nikishin et al., 2001).

Towards the west, the *Eastern Pontides* are a prolongation of the Karabakh magmatic belt. They are composed of similar Mesozoic complexes (Yilmaz et al., 1997; Okay and Sahintürk, 1997). The *Achara-Trialet zone* is situated to the north of the Eastern Pontides. It is a (Early?) Middle Jurassic (Levin, 1995) to Cenomanian (Yilmaz et al., 2000) island arc. It is considered by Levin (1995) thereafter to be an intra-arc Cretaceous–Palaeogene rift or, by Lordkipanidze et al. (1984), to be a back-arc or inter-arc basin overlain conformably by Upper Eocene shoshonitic rocks of a mature arc. Yilmaz et al. (2000) proposed a new division of the Eastern Pontides. The southern segment would be a fore-arc and the central part an arc from the Jurassic until the end of the Campanian. The northern segment (the Achara-Trialet zone) is considered to have been an arc until Cenomanian times and then a juvenile back-arc during the Late Cretaceous (pre-Maastrichtian). From Maastrichtian to Early Eocene times only volcanoclastics are present without volcanics. Palaeocene–Early Eocene compression occurred in the Eastern Pontides as a result of the collision of the Pontides with the Taurides (Okay and Sahintürk, 1997; Okay and Tüysüz, 1999). Yilmaz et al. (2000) interpret the Mid Eocene volcanics of the Achara-Trialet zone as resulting from a tensional regime during a post-collisional event subsequent to crustal thickening, similar to an interpretation proposed for comparable rocks in the Eastern Pontides (Yilmaz and Terzioglu, 1994; Yilmaz and Boztug, 1996). This Middle Eocene tensional event in Achara-Trialet represents an important argument for the authors for a Middle Eocene age of opening of the East Black Sea (Kazmin et al., 2000). The Achara-Trialet zone was then affected by a post-Eocene closure followed by thrusting towards the north in the Neogene (Lordkipanidze, 1980; Karyakin, 1989; Banks, 1997; Banks et al., 1997).

The *Erevan-Ordubad Basin* is located in Armenia–Azerbaijan, immediately to the south of the ophiolitic Sevan–Akera suture, north of the South Armenia block. It contains thin Bajocian–Callovian calcareous, thick calcareous Upper Cretaceous (Khain, 1994), and an up to 2–3 km thick sequence of Palaeocene to mainly Eocene flysch-type sediments (Milanovsky, 1991). Eocene rapid subsidence of the basin was accompanied by basaltic, andesitic, and dacitic volcanism, peaking during the Middle Eocene. Moreover, Eocene to Oligocene granites intruded this basin. The rapid subsidence and volcanism imply an extensional origin of the Erevan-Ordubad Basin (Koronovsky et al., 1997). A compressional tectonic setting commenced at the Eocene–Oligocene transition (Milanovsky, 1991) with the main deformation phase being at the end of the Oligocene (Khain, 1994).

The *Talesh Basin* is located close to the south-western margin of the Caspian Sea. Gasanov (1992, 1996) demonstrated that Santonian–Maastrichtian shallow-water carbonates and Danian–Palaeocene clastic sediments containing tuffs underly the Talesh Basin. The Eocene volcaniclastic series attain a thickness of 4–5 km; they were deposited in relatively deep-water and are evidence of rapid basin subsidence. The Middle Eocene series contains olistostroms composed of Cretaceous and Eocene rocks. Volcanic units consist of basalts, andesites, trachy-basalts and trachy-andesites; moreover, there is also a complex consisting of peridodites, gabbro, gabbro-syenites and serpentinites (Ali-Zade et al., 1996; Gasanov, 1992, 1996). The extensional Talesh Basin was identified by Ismail-Zade et al. (1995) as a back-arc basin. It was inverted at the end of the Eocene and during the Neogene (Milanovsky, 1991; Gasanov, 1996). A prolongation of the Talesh Basin lies along the southern margin of the Alborz (Berberian, 1983; Haghipour and Aghanabati, 1989; Saïdi, 1995). The thickness of sediments and the volcanics/tuffs ratio decrease towards the eastern Alborz and no Eocene is present to the north of the Alborz (see above).

Although it is generally assumed by many authors (Zonenshain and Le Pichon, 1986; Milanovsky, 1991; Khain, 1994) that it was connected to the Achara-

Trialet Basin, the Talesh Basin may have also formed the eastern prolongation of the Erevan-Ordubad Basin. Thus, the *Achara-Trialet, Erevan-Ordubad, Talesh and southern Alborz Eocene basins* would have had similar features (Khain, 1994; Nikishin et al., 2001). They began to evolve as a consequence of Cretaceous rifting, were then submitted at the end of the latest Cretaceous–Palaeocene times to a compressional regime during closure in the back-arc domain of the Neo-Tethys subduction zone (Zonenshain and Le Pichon, 1986). Their clearest common feature is a Middle Eocene extensional phase of subsidence that resulted in the development of deep narrow basins (often in marine conditions), accompanied by calc-alkaline volcanism, trending towards alkaline. They were later inverted during Late Eocene, Oligocene and Neogene phases of transpression and collision (Saïdi, 1995; Koronovsky et al., 1997; Okay and Sahintürk, 1997; Yilmaz et al., 1997; Nikishin et al., 2001).

3. The South Caspian Basin

The SCB comprises a deep and thick central basin in the southern part of the Caspian Sea with thinner prolongations to the east and west onshore. The main central basin can be subdivided into several sub-basins (Fig. 2a). One is located in the northern part

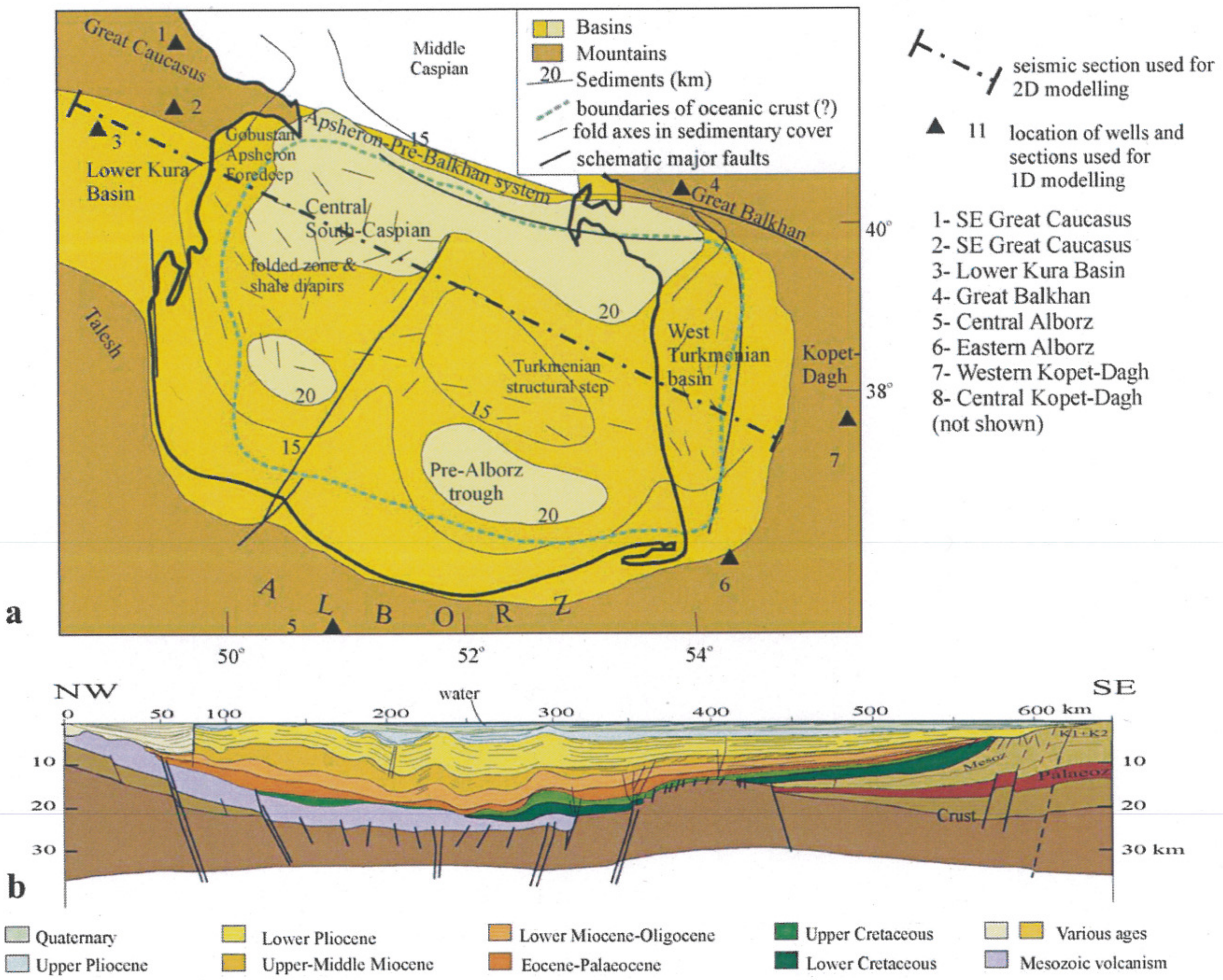

Fig. 2. (a) Map of the basement depth and location of the section and wells or synthetic wells (triangles with numbers) used for the subsidence reconstruction. Isolines of basement depth, contours of "oceanic" crust and axes of folds are from Shikalibeily and Grigoriants (1980) and Berberian (1983). (b) Crustal cross-section of the SCB (location on (a)), after Mamedov (1992) modified.

of the SCB and, attains a depth of perhaps as much as 26 km to the north north-west, near the Apsheron Sill (Knapp et al., 2000). The second major sub-basin is the Pre-Alborz trough, with a thickness of more than 20 km, located in the south-eastern part of the SCB to the north of the Alborz. The thickness of sediments in the south-west corner of the SCB differs substantially in variously published basement depth maps: for example, 20 km according to Malovitsky (1967), Shikalibeily and Grigoriants (1980) and Krylov (1987), 10 km for others (Anonymous, 1994; Volozh, 1995, unpublished map). The western offshore part of the SCB is very thick but the central eastern marine part (between the Pre-Alborz trough and south of Apsheron–Pre-Balkhan depocentre) is occupied by a horst where the basement depth is around 15 km (the "Turkmenian structural step") and where the crust is thicker than elsewhere.

This work is based on publicly available data from Mamedov (1992) for the cross-section (Fig. 2b) and published papers for the stratigraphic columns. When the present work began, the more recent but not fully published data on the structure of the SCB of Knapp et al. (2000) were not yet available.

3.1. Sedimentary fill

Sediments overlying basement in the centre of the basin are not reached by wells and there are, at present, no reliable, continuous seismic correlation of the deepest horizons towards the basin margins. Consequently, there exist several very different interpretations of the age of the basin. Here, it is supposed that deposition began during the Callovian–Late Jurassic and that the total thickness of Mesozoic sediments is not more than some 5 km. The Mesozoic layer, interpreted as comprising mainly volcanics, is present at the base of the western basin. It extends onshore and, according to Mamedov (1992), may be of Jurassic age. In the eastern part of the SCB (Fig. 2b), Cretaceous sediments are present, which can be extrapolated towards the central part. Upper Cretaceous sediments overly folded units of Jurassic–Cretaceous age.

The bulk of the sedimentary infill of the SCB is, however, Oligocene and younger. The thickest series is represented by the Pliocene–Quaternary with more than 10 km of sediments deposited in a very short time interval (less than 5 Myr). The very rapid deposition prevented normal expulsion of fluids, leading to overpressuring and undercompaction of underlying deposits situated below, mainly the Oligo–Miocene Maikop source rock (Narimanov, 1993). The Maikop is at the source of numerous mud diapirs and volcanoes covering the SCB.

A NNE–SSW directed shortening, related to convergence of the Arabian plate with Eurasia, affected the region during Pliocene–Pleistocene times. Sedimentary cover has been significantly deformed, with large buckle folds developed overlying a ductile detachment zone in the Maikop shale. The surrounding orogens (Caucasus, Alborz, Kopet-Dagh) were uplifted at this time, isolating the SCB and providing by their erosion, a huge source of sediments. These were transported towards the rapidly subsiding SCB through three drainage systems: the palaeo-Amu Darya from the east, the palaeo-Kura from the west, and palaeo-Volga from the north; the last also carried sediments from the Russian Platform. The deltas of these rivers provided favourable locations for reservoirs, with prograding clinoforms, deep-water turbidites and slump systems (Abdullayev, 2000).

The northern boundary of the SCB is the Apsheron–Balkhan fault belt, along the Apsheronian step. It separates the SCB from the middle Caspian province that has a thicker crust, a Late Palaeozoic basement also affected by Eocimmerian orogeny, and sedimentary cover thinner than the SCB. The Apsheron–Balkhan belt connects the Great Caucasus and Kopet-Dagh orogens. Seismic data (Abrams and Narimanov, 1997) show that near the surface the belt is a typical positive transpressional "flower" structure. But extensional focal mechanisms at depth can be interpreted as extension at the top of the SCB basement, subducting northwards since about 5 Ma in the northwestern part of SCB (Granath and Baganz, 1996; Granath et al., 2000). Recent deep seismic reflection data across Apsheron (Knapp et al., 2000) show a gentle deepening of the SCB basement/cover contact, identified by a bright reflection, towards the north to 26–28 km. It is interpreted to image the subduction of the SCB basement below the Apsheron–Balkhan belt.

Fig. 3 presents a schematic comparison of the lithostratigraphic columns and thicknesses observed in the western, central, and eastern parts of the SCB.

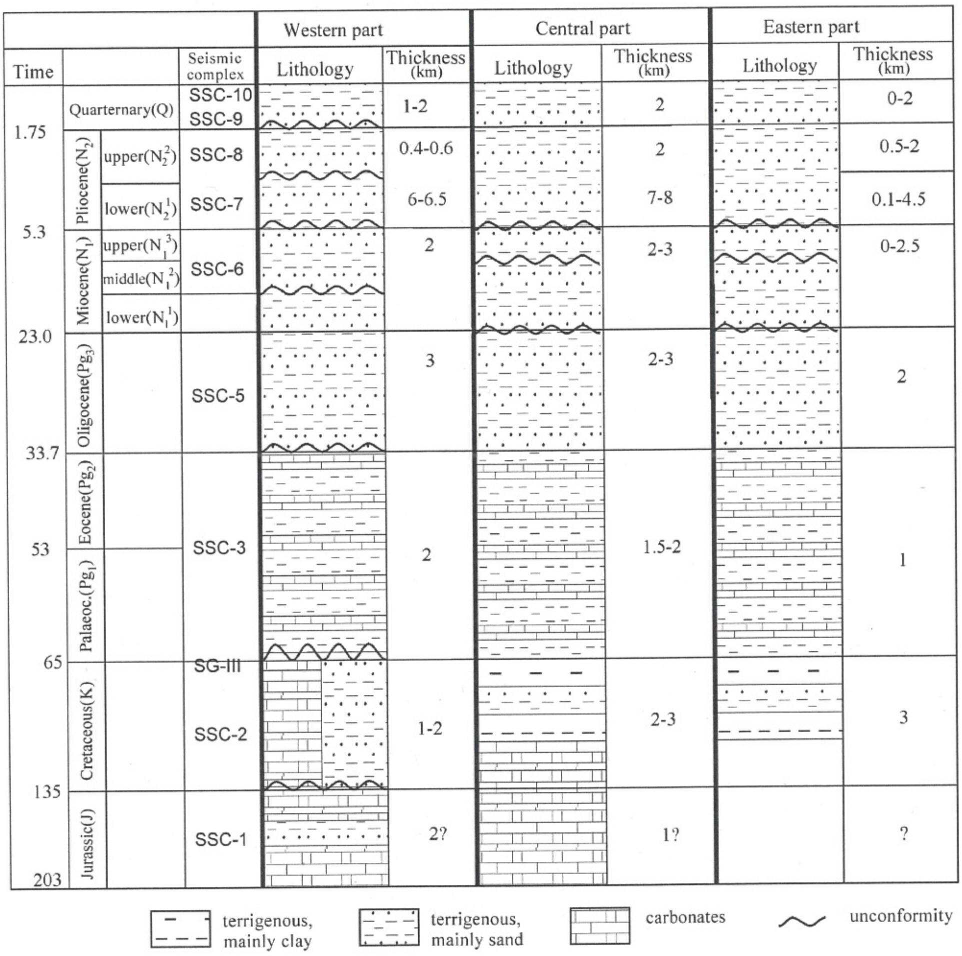

Fig. 3. Comparison of synthetic lithostratigraphic profiles between the western, central and eastern parts of the SCB. SSC are the seismic complexes determined from the western Kura basin. Lithology in the central part is derived from the margins (SSC4 not represented on the column is Eocene, when it exists SSC3 represents the Palaeocene). SG3 corresponds to a reflecting boundary.

Onshore, the western prolongation of the SCB is the Lower Kura Basin, located to the south-east of the Great Caucasus. The Kura basin comprises several sub-basins of different origins, with up to 15 km of sediments, separated by areas with less sediments. The Lower Kura is the easternmost of these sub-basins.

In its recent evolution, after having been a margin of the Great Caucasus Trough, the Kura Basin has become partly a foreland flexural basin, filled with molasse sediments, to the south of the Great Caucasus belt. The thickness of the Oligocene to Quaternary molasse is close to 5–8 km (Milanovsky, 1991; Akhmedov and Guldust, 1981), divided into two main parts. The first of these is Oligocene–Lower Miocene (Maikopian) mainly clay succession and the second consists of Middle Miocene–Quaternary clays and clastics with minor carbonates. Pliocene to Quaternary was the main time of molasse deposition and contains the Productive Series in the Lower Pliocene. Pre-Oligocene deposits are poorly investigated. Carbonates, clays and clastics represent the Palaeocene and Eocene successions, which have an irregular distribution. Eocene sediments are enriched by tuff material.

Volcanic complexes are situated below the Cainozoic strata. These were penetrated by the Saatly superdeep well (8324 m) in Azerbaijan (Nadirov, 1984; Ostroumova and Zeiter, 1987; Shikalibeily et al., 1988; Ali-Zade et al., 1999) and by some other wells (Akhmedov and Guldust, 1981). The Saatly well is situated on a NW–SE zone of uplift (Talesh–Vandam), which correlates with a major positive gravity anomaly. The well contains several kilometres of Mesozoic sediments and documents an important hiatus in sedimentation from Late Cretaceous until Middle Miocene times (Shikalibeily et al., 1988; Ali-Zade et al., 1999). Ali-Zade et al. (1999) recognised several volcanic complexes in this area, the two most important of which are a widespread Lower–Middle Jurassic volcano-sedimentary series of at least 4784 m with basalts, andesites, dolerites, diorites (possibly covering nearly the whole Lower Kura Basin basement) and 390 m of Late Jurassic and 320 m of Early Cretaceous carbonates with basalts sills. The Lower–Middle Jurassic volcanogenic complexes have numerous characteristics in common with a similar series in the north-eastern part of the Lesser Caucasus.

Data on the Mesozoic volcanic complexes of the Kura Basin, together with those from the Transcaucasus volcanic arc, confirm that large scale subduction-related volcanism took place in the Transcaucasus arc in Bajocian to Cretaceous times. In the east, Mesozoic volcanic activity diminished towards central and eastern Alborz (Berberian, 1983). The occurrence of Late Jurassic olivine-bearing basalt along the northern part of the long Middle Jurassic calc-alkaline magmatic belt demonstrates that back-arc (or intra-arc) extension took place during the Late Jurassic and led to the partial splitting of the former Middle Jurassic volcanic belt (Zonenshain and Le Pichon, 1986). From these data, it can be suggested that the onset of oceanic spreading in the SCB may have occurred at this time. The question is where, laterally into the SCB, did the spreading stop. Zonenshain and Le Pichon (1986), for example, interpreted a high velocity anomaly in the crust beneath the Lower Kura Basin (see Fig. 2C of Ershov et al., 2003) as indicating the presence of hidden oceanic crust of the back-arc basin.

The West Turkmenia Basin is located just to the west of the Kopet-Dagh orogen, in the eastern part of the South Caspian Sea and onshore. The basement of this basin is composed of a "block" with a crust thicker than below the central SCB (Fig. 4). The basin contains more than 10–15 km of sediments. Rapid subsidence started no later than the Oligocene, simultaneously with a phase of deformation in the Kopet-Dagh (Maksimov, 1992). A similar long wavelength subsidence pattern exists for the Maikopian in the North Caucasus basins (Ershov et al., 2003) and also from the Oligocene onwards in the Great Kavir Basin (Fig. 1), south of the eastern Alborz (Reyre and Mohafez, 1970). As in the SCB, the West Turkmenia Basin was characterised by rapid subsidence during the Pliocene, leading to the deposition of a red-bed series (3–4 km), analogous to the Pliocene Productive series of Azerbaijan. The SCB, having resulted from these peculiar conditions is thus, besides its astonishing thickness and unusual geodynamic history, mainly of interest for its huge petroleum potential.

3.2. Nature of the crust from geophysical data

A simple gravity model has been constructed to check if crustal scale interpretations taken from Mamedov (1992) were in agreement with the gravimetric data along the section studied. These interpretations involve the total sedimentary thickness and its average density, the thickness of the crust and its velocity structure, and the topography of the basement and Moho along the section studied (western part of the section is modified from the data of Baranova et al., 1991). The results (Fig. 4) show that the interpretations are compatible with the gravity data. It was not possible to make a more detailed model with a better fit, especially near the eastern half of the section where crust is thicker (Fig. 2b). The interpreted geological section is in fact a compilation of several projected seismic lines. The exact position of the seismic lines was not available for publication; the gravity profile was drawn simply as a straight line from a map Bouguer anomalies between the two ends of the geological section.

The crust in the central part of the SCB is thin (about 8 km) and lacks an upper low velocity layer (Shikalibeily and Grigoriants, 1980; Mangino and Priestley, 1998) (Fig. 2b). In the Kura depression, Moho depth varies from 45 to 60 km with an abrupt step to the central SCB depression. To the east, the crustal thickness increases more gradually towards Turkmenia.

The South Caspian behaves as a relatively rigid aseismic block (Priestley et al., 1994). Very few

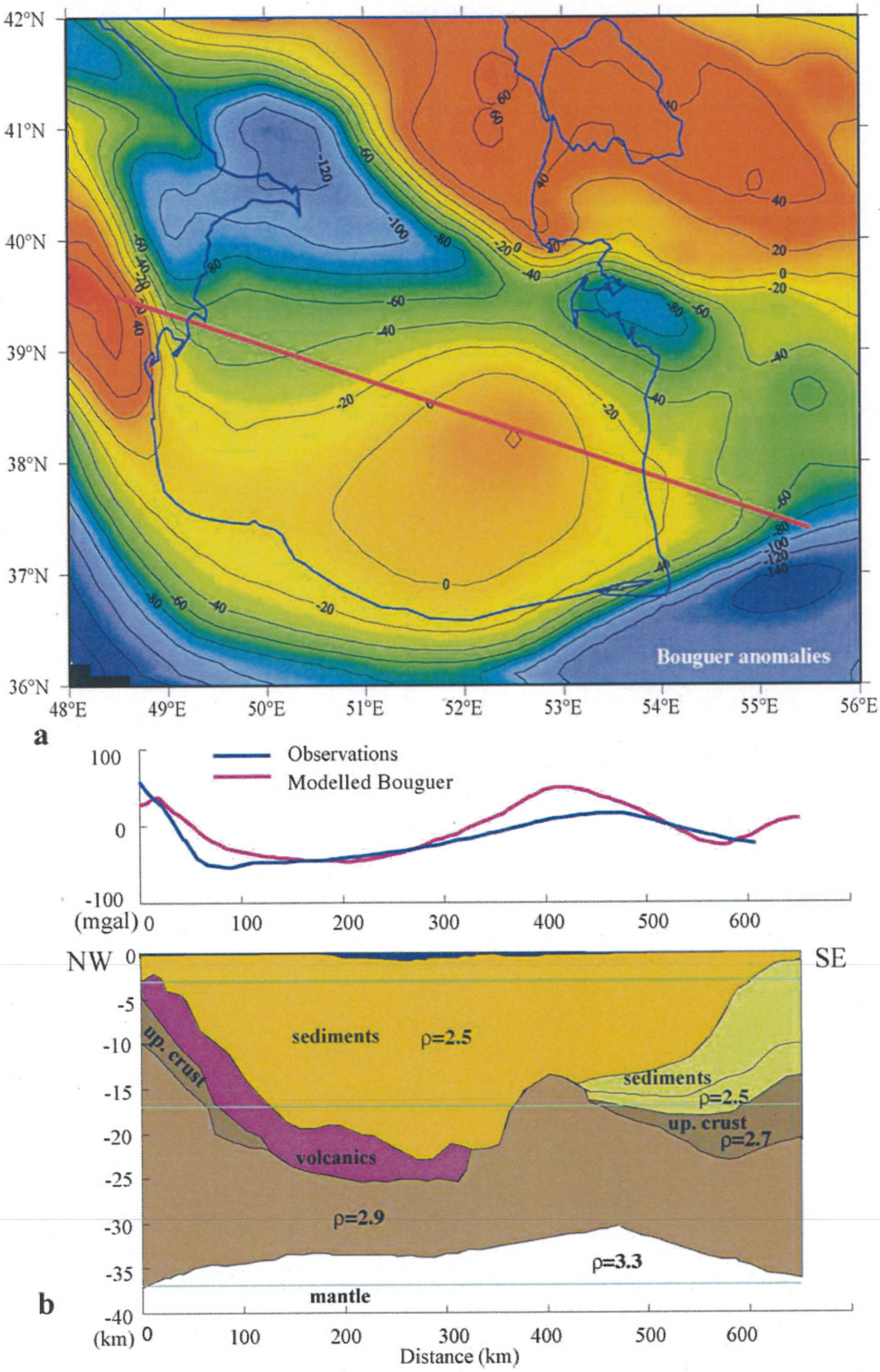

Fig. 4. Check of crustal structure of the SCB along the NW–SE section (cf. Fig. 2b) by a simple gravimetric model. (a) Simplified map of Bouguer gravity anomalies (drawn from gridded data of the Geological Survey of Russia); (b) gravimetric model (crustal section from Mamedov, 1992, western part of the section is modified with the data of Baranova et al., 1991).

earthquakes occur in the SCB given that deformed and deforming, seismically very active areas surround it. Most authors accordingly consider that the crust underlying the SCB is of oceanic origin. Mangino and Priestley (1998) compiled the existing geophysical data and argued that most of the available information supported an oceanic affinity but could not conclude such definitively. In effect, the SCB crust is thicker than a normal oceanic crust and, from its seismic velocities, could also be lower continental crust or, alternatively, highly thinned and intruded continental crust. In this case, Mangino and Priestley (1998) attributed the resistance of the SCB crust to deformation as resulting from the shallow depth to the Moho and not necessarily to its oceanic composition.

More recently, Granath et al. (2000) concluded from gravity data that no continental crust is present below any part of the South Caspian Sea. They interpreted the negative anomaly (reaching − 150 mGals), observed in the north-western part of Apsheron, as a South Caspian crustal root under the Apsheron peninsular area, displacing the normal lithospheric mantle caused by the underthrusting or northwards subduction of the SCB crust. Their model (an EW section) shows a thickened sedimentary layer in the east caused by compressive deformation within the Mesozoic sediments (as seen in Fig. 2b), on which is superimposed the Amu Darya delta. The anomalous crust continues below the Turkmen shelf and the Tertiary section is the same offshore and onshore Turkmenistan. Recent seismic reflection data in the vicinity of Apsheron show a highly reflective layer interpreted as an 8 km thick basement, consistent with an oceanic crustal affinity in this area (Knapp et al., 2000).

The general consensus, therefore, is that the crust underlying the SCB, at least in its marine part, is likely oceanic. Related problematic issues include the timing of spreading, in one phase or during several episodes, and the possible lateral continuation of the oceanic basin below the Lower Kura Basin in the west and Turkmenia or even the Kopet-Dagh in the east.

3.3. Burial history

Having checked the crustal profile, subsidence analysis of the SCB was carried out. The age of the oldest deposits, in the deep central basin, is not known. Moreover, assuming that these are underlain by oceanic crust, the first stages of the evolution of the basin, prior to sea-floor spreading (i.e. the rifting phase) are not recorded and not possible to analyse in the central basin. Therefore, in order to characterise the rifting phase and to infer the timing of oceanic opening, attention was focused on subsidence curves recorded in the different units surrounding the SCB that were formerly margins or prolongations of the SCB.

On the margins of the SCB, there are only a few places where Jurassic sediments are observable (sometimes only Upper Jurassic), the Great Caucasus, Great Balkhan, Alborz or on its onshore prolongations, the Lower Kura Basin and the Kopet-Dagh. Subsidence curves constructed from synthetic wells (Fig. 5) in these surrounding areas are compared to those from the lateral E–W extensions of the SCB. The synthetic wells were compiled from field data or, sometimes, unpublished seismic data. Odin's (1994) geological time scale was chosen to allow direct comparison to basin modelling carried out during the Peri-Tethys Programme. Clay–sand ratios in Cainozoic rocks and correlations with depth–porosity are taken from Bredehoeft et al. (1988). The backstripping method is similar to the one used by Steckler and Watts (1978). The 1D subsidence curves were constructed with air-loaded local isostasy (Airy type); no corrections were made for sea-level and water-depth variations.

Increases in tectonic subsidence rate punctuate tectonic phases during basin evolution. After a possible foreland subsidence stage in the Alborz, beginning with deposition of the Shemshak in the Late Triassic, there seems to be a transition to a rifting phase in Early(?) Jurassic to Bathonian times, from western Alborz (Jolfa) to eastern Alborz and Kopet-Dagh (Saïdi et al., 1997). The first Jurassic sediments are observable only in the curves of the Alborz (Saïdi et al., 1997; Fig. 5). Rifting could have begun in the Alborz during the Sinemurian as it did in the Great Caucasus (see Ershov et al., 2003). However, the lack of chrono-stratigraphic resolution for the Shemshak formation allows only an average subsidence rate during the long time interval from the Norian to the Bajocian–Bathonian. The synthetic wells from the eastern Great Caucasus and the Great Balkhan indicate that subsidence began with the first

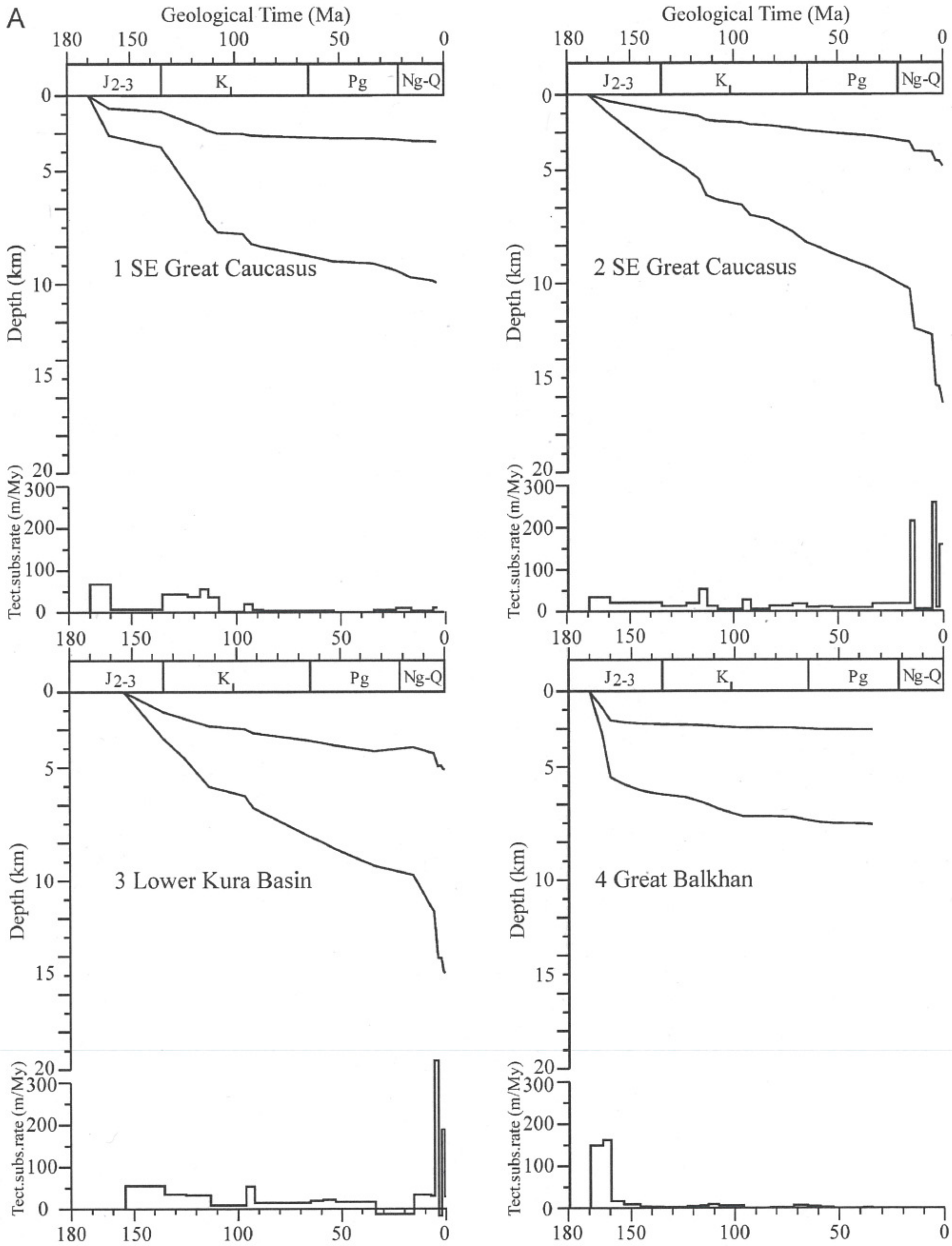

Fig. 5. Subsidence curves of the synthetic wells studied (location on Fig. 2a) with air-loaded tectonic subsidence, time scale is from Odin (1994). Stratigraphy for summary section of Great Balkhan, Kopet-Dagh (synthetic columns without precise locations) are after Luppov, 1972, for Kura depression—after Akhmedov and Guldust, 1981 and Azerbaijan Atlas, Anonymous, 1994. Alborz sections are after Saïdi (1995), Caucasus from Ershov et al. (2003). Correlation between seismic reflectors and stratigraphical units is after Mamedov (1992).

recorded sediments in the Bajocian–Bathonian (as there is no records here of the Sinemurian rifting). However, the subsidence rate increases in time in association with active extensional faulting and magmatic intrusive activity.

These various areas were the margins of the eastern part of the Great Caucasus Trough, the future SCB. Alborz was situated on the southern margin of the SCB. The northern margin went from the Great Caucasus to the Great Balkhan and north of the

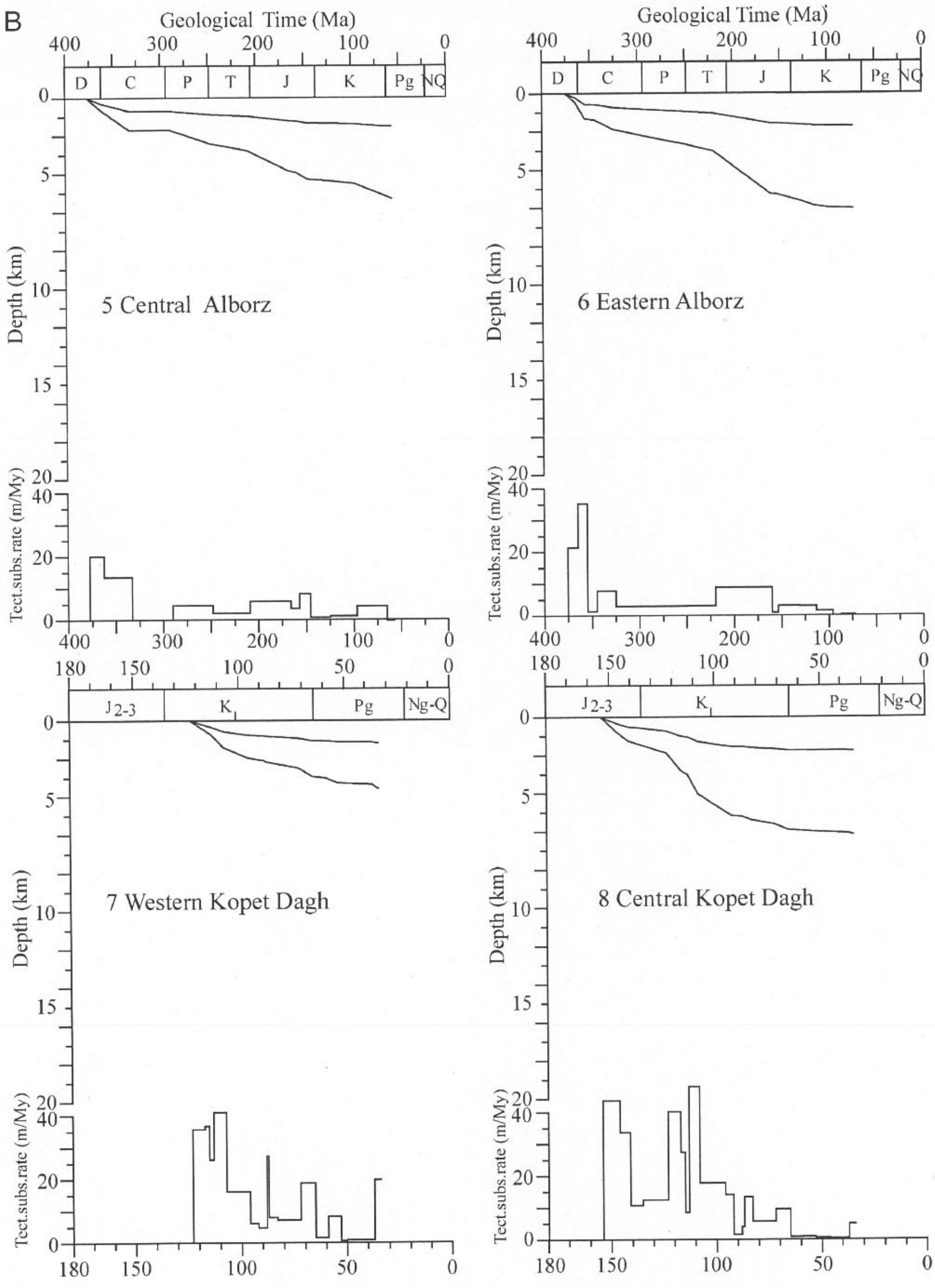

Fig. 5 (*continued*).

Kopet-Dagh. Sengör et al. (1984) identified an extensional event beginning during the Middle Jurassic in the Kopet-Dagh. It can thus be concluded that the Early(?)–Middle Jurassic period of subsidence correlates with rifting of the SCB. The Bajocian and/or Bathonian successions are absent in some parts of the margins (north Alborz, north Kopet-Dagh) but sedimentation is continuous in the Kopet-Dagh and Gorgan (south-east corner of the Caspian Sea). Either this is evidence of the Mid-Cimmerian orogeny observed in the west North Caucasus, with compression affecting the course of the rifting, or the unconformity is a "break-up unconformity" related to rift shoulder uplift at the transition from rifting to sea-floor spreading. In fact, either can be linked to the reorganisation of blocks and plates that provided the space for the SCB opening.

In the Late Jurassic, subsidence slows down on the margins of the SCB, especially the northern ones (Great Caucasus, Great Balkhan). For the Alborz, the decrease of the tectonic subsidence rate after deposition of the Shemshak formation (Mid–Late Jurassic), is less important than on the northern margins. But it is probably due to the absence of stratigraphic precision in the Shemshak, which is considered as an average deposit during Late Triassic–early Mid Jurassic period.

In contrast, subsidence is greatly accelerated in the basinal area (Lower Kura, Kopet-Dagh, inner central zone of the Great Caucasus). This event may be a sign of cessation of rifting on the margins when spreading begins in the basin. So from the Callovian to the Tithonian–Berriasian subsidence is fast in the basins, slow on the margins. We conclude, therefore, that oceanic spreading began at least in some parts of the SCB when extension diminishes on the margins.

Another acceleration of subsidence, associated with basalts, occurred in the Early Cretaceous (Neocomian–Barremian–Albian) in the Kopet-Dagh, in Alborz and Lower Kura. This corresponds to a period of rapid subsidence also recorded in the Great Caucasus Trough and in the Eastern Pontides. It is widely distributed either on the margins or in basinal prolongations of the SCB. It could represent a general reactivation of extension and opening. However, as its age varies slightly according to the points of observation, a more accurate stratigraphy is necessary.

In the Eocene, a phase of subsidence exists in the area of Jolfa (near the South Armenia suture), in central and eastern Alborz (Saïdi et al., 1997). It corresponds to the extensional opening of basins and is characterised by volcanism. Northward, in basinal areas, it is difficult to know whether an extensional phase occurred at this time because it depends on knowing how deep the water was, but this is poorly constrained. Nevertheless, we think that extension was mainly concentrated in the southern margins (see discussion below).

The peak of tectonic subsidence of the SCB is observed in the Pliocene–Quaternary. The total magnitude of tectonic subsidence during this time reaches 2 km, with very high rates of subsidence (Brunet et al., 1997; Fig. 5).

To image partly the Cainozoic evolution of the basin, we have made a 2D reconstruction (Fig. 6) along the NW–SE section of the SCB (Fig. 2b). It is flexurally backstripped, taking into account decompaction of sediments and some assumptions of water-depth through time. We assumed a depth of 2500 m in the Oligocene in the SCB, similar to a basin with oceanic crust partly filled by sediments, but this value is not constrained. From the Oligocene to the present, we suppose that the depositional water-depth decreased regularly to its modern depth of 900 m. We could also assume a much greater water-depth at the beginning of the Cainozoic and a progressive filling of a very deep basin before a renewal of tectonic subsidence during the Pliocene–Quaternary.

Our data are not sufficiently precise to proceed further in the interpretation. Supplementary studies are necessary, especially in Iran. It is indeed in the south-east corner of SCB where elements of the lateral prolongation of the marine basin are more likely to be found. Additional studies in this area might allow observations of the opening of the basinal part of the SCB and not only its margins.

Further south, before an important anti-clockwise rotation (Davoudzadeh et al., 1981; Saïdi, 1995) and concomitant westward displacement of the Tabas and Lut blocks, the Nayband area may have been a south-east prolongation of one of the basins, either the SCB or a basin further south. Indeed, this region displays active subsidence in the Jurassic with 5500 m of sediments deposited, including 2700 m of Bathono–Callovian only (Saïdi, 1995; Saïdi et al., 1997). The Nayband area became a passive margin in the Late Jurassic (Saïdi et al., 1997). It is now isolated in the middle of Iran after the rotation of the adjoining tectonic units. This area also deserves further study to clarify the geodynamics of Iran.

When sea-floor spreading began (if spreading occurred in the SCB as we have assumed) is difficult to document precisely from the subsidence curves only. Obviously, it must be inferred and discussed in

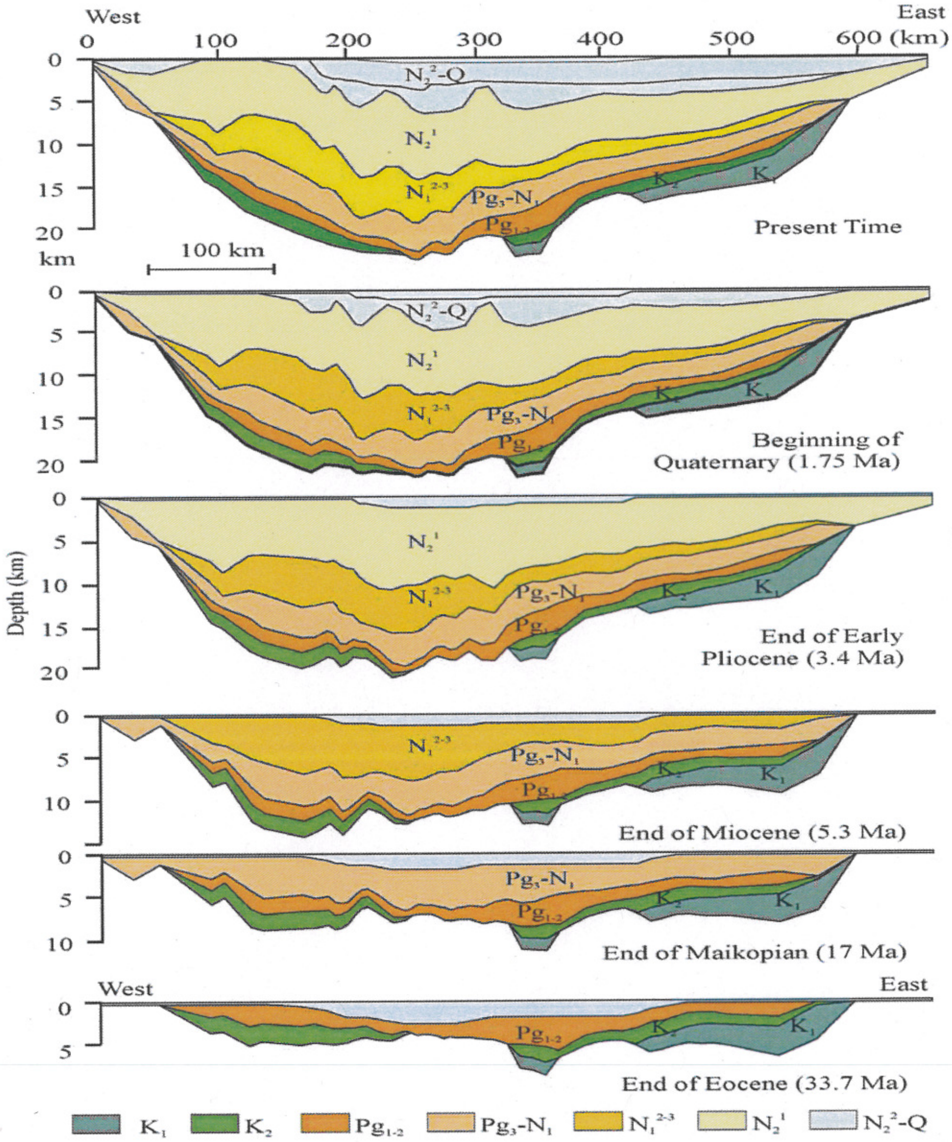

Fig. 6. Backstripping reconstruction, from the end of Eocene to Present, along the NW–SE section (location on Fig. 2a), taking into account decompaction and estimation of palaeobathymetries (cf. Fig. 3).

combination with the regional geodynamic evolution of the whole of the study area.

4. Timing of the opening of the South Caspian Basin

4.1. Argumentation on the age of opening

Various ages and mechanisms have been suggested to explain the opening of the SCB. Most authors have proposed that the SCB is a Mesozoic to Palaeogene back-arc basin. Authors favouring the Mesozoic suggest either a Late Jurassic or a Cretaceous age of opening (e.g. Zonenshain and Le Pichon, 1986; Dercourt et al., 1986, 1993, 2000; Mamedov, 1992; Granath and Baganz, 1996; Granath et al., 2000; Otto, 1997; Nikishin et al., 1998a,b, 2001; Stampfli et al., 2001; Ziegler et al., 2001). Authors favouring the Palaeogene include Berberian and Berberian (1981), Abrams and Narimanov (1997), and Boulin (1991).

Berberian (1983), proposed a Late Mesozoic–Early Tertiary age for the oceanic crust of the SCB, Kazmin (1991) an Eocene one. Golonka (2000a,b) postulates an opening in several steps, beginning in the Jurassic that continued going on during the Cretaceous and Eocene. Finally, the last phase of opening of the Pre-Alborz trough would have occurred in the middle Miocene by northward movement of the South Caspian Microcontinent, a block situated in an area of uplift and thick crust in the east of the SCB marine basin (Golonka, 2000a,b). However, geophysical data indicate that this area likely also has crust of oceanic affinity, though thicker than normal (see above).

There are also models suggesting an earlier opening or a pull-apart mechanism. For example, Berberian (1983) hypothesised that the South Caspian Sea is a remnant of the Palaeo-Tethys Ocean trapped during the Eocimmerian convergence, with the age of its crust therefore being Palaeozoic Early Mesozoic. Another variation of a model with a pull-apart basin opening in the context of large translational movement of blocks during Cretaceous–Palaeogene times is described by Apol'skiy (1974). Similarly, Sengör (1990) proposed a pull-apart opening of the SCB along a major Cretaceous dextral strike–slip fault system of the Kopet-Dagh/Alborz/Apsheron Sill/Caucasus. Movement of the middle eastern Cimmeride collage of Iranian blocks would have occurred from a position in front of the Turan platform to the present longitude. Nevertheless, this last model is not incompatible with back-arc opening, as lateral displacements of blocks have accompanied the northwards subduction of the Neo-Tethys. "Anonymous" (1994) and Nadirov et al. (1997) proposed, in supplement to Sengör, a westward subduction of the SCB from Middle Jurassic to Late Cretaceous times.

We agree with a Mesozoic age of basin opening, as proposed by many authors, but we need to constrain better the age of opening. In order to determine a more precise tectonic history using the existing data, we can discuss the following items.

(1) Partial closing of the Palaeo-Tethys with the accretion of Gondwanan blocks to Eurasia in the Middle–Late Triassic produced a widespread Eocimmerian orogeny in the Caspian–Caucasus–Iranian region (Belov, 1981; Kazmin and Sborschikov, 1989; Dercourt et al., 1993, 2000; Stampfli et al., 2001); the SCB was closed at this time. Orogeny created elevated areas that were eroded and provided sediments found in the Shemshak deposits of North–East Iran following collision. Otto (1997) speaks of the extensional collapse of the Eocimmerian orogen affecting a large area from Iran to the Mangyshlak (to the north-east of Middle Caspian Sea). It led to the localised deposition of sediments (last member of Shemshak in Iran) in small, extensional post-collisional basins from the Early Jurassic to the beginning of Middle Jurassic.

According to Stöcklin (1974), Stampfli (1978), Zonenshain and Le Pichon (1986), and Alavi (1996), the Shemshak depocentres located in the Alborz are foreland basins situated to the south of the belt, since the origin of sediments was identified as from the north. In this case, a continent, part of the Eocimmerian belt, occupied the position of the present SCB, meaning that the basin could only opened after the Bajocian, after deposition of the Shemshak.

In another interpretation, Rad (1986) suggests that, to the east of the Alborz, sediments originated from the south and that a delta depocentre of Shemshak is located to the north of the present Alborz. Sediment thickness varies from 1000 m in the south to 3000 m in the north, towards the opening basin of the South Caspian. Sediments were transported by turbidite currents to the deep sea (in Kopet-Dagh). There was rapid subsidence but, nevertheless, a transgression (passing laterally to the Kashafrud formation in the Kopet-Dagh). The Middle Jurassic transgression, at the end of the Shemshak stage, corresponds to the establishment of a carbonate platform (Rad, 1986) on the margins of the SCB. In this model, the SCB was already deep by the cessation of deposition of the Shemshak, with rifting well developed and close to oceanic opening.

More generally, extension occurred in the Great Caucasus and in the Kopet-Dagh (Zonenshain and Le Pichon, 1986) in a context of a back arc basin developed through the Transcaucasus and Alborz. This is understood to be the rifting stage of the SCB. It follows that the oceanic opening of SCB must be later than the very beginning of Middle Jurassic.

(2) During the Early Eocene, from the Lesser Caucasus to Iran, subduction appears to have taken place south of its previous Mesozoic position. This was due to the accretion of the South-Armenia block,

and perhaps the Sanandaj–Sirjan block, during the Late Cretaceous or Early Palaeogene. This was accompanied by a shift of the zone of volcanic activity towards the south and also of the area of extension into a back-arc situation. The large distance between the subduction zone and the area of back-arc extension area is frequently a difficulty for this theory of Jurassic basin opening. But this is considering the present configuration of the area. The Lesser Caucasus–South Armenia block (covered by marine Coniacian) had not yet been accreted (Bazhenov et al., 1996; Golonka, 2000a,b) and the Iranian blocks were situated further eastward (Saïdi, 1995) and probably in relative positions different from the present ones (in part Sengör, 1990; Golonka, 2000a,b; Stampfli, 1999 personal communication; Stampfli et al., 2001). Thus, in the Jurassic, the distance between the subduction and the "back-arc" zone of extension was probably considerably less (Fig. 7) than it appears from the present block disposition. After accretion and/or lateral displacement of blocks in the Late Cretaceous–Palaeogene, Neo-Tethys subduction was active more to the south than before (Fig. 7). Extension and volcanism occurred in the (Middle–Late) Eocene in small, elongated basins: the Achara-Trialet, Erevan-Ordubad, Talesh and south of Alborz (Fig. 7). Normal faulting, accompanying an important phase of extension, is observed especially in the Talesh Basin, southwest of Alborz (NIOC and GSA, 1978). This is also the case in the Great Kavir Basin (south-east of Alborz) where extension and deposition of 2000 m of tuffs are observed (Reyre and Mohafez, 1970). Deep-water flysch and andesites passing laterally to tuffs often filled these basins. The basins appear to be situated primarily near the locus of the preceding end of Cretaceous–Palaeocene collision. Volcanics seem to be absent in the Eocene deposits of the Lower Kura (Anonymous, 1994) and Kopet-Dagh in the east (thus in the prolongations of the SCB).

Nevertheless, Eocene rift-like basins surround the southern margin of the SCB. It means that the SCB existed during Eocene times as a strong lithospheric block (Fig. 7), not deformed by Eocene extension and around which the Eocene basins opened and later were shaped during Cainozoic collision. The general arcuate pattern of the Talesh and Alborz follows the shape of the more resistant oceanic basement of the SCB (Berberian, 1983).

The cause of the Eocene extension needs to be clarified: it occurs either in a back-arc setting, after closing and reopening of the area of suture (perhaps the gravitational collapse of an area of thicker crust), or in a strike–slip setting with lateral displacement of accreted blocks. So the earliest and latest limits for the time of opening of the SCB are fixed from Middle Jurassic to Palaeocene.

(3) The Senonian to Palaeocene was predominantly a time of compressional tectonics in this entire region, from Turkey to the Caucasus area and East European Craton (Nikishin et al., 1999, 2001); this means that a pre-Senonian opening of the South Caspian is more realistic.

(4) Subduction related to Transcaucasus magmatic arc was active from the Bajocian until the Late Cretaceous; it follows that the SCB developed in a back-arc environment. The Middle Jurassic basalts of the Kura Basin and Neocomian basalts of the Alborz could be related to back-arc extensional events. Renewed subsidence began along the southern margin of the Great Caucasus Trough and Kopet-Dagh from the Callovian–Oxfordian and cooling-related subsidence lasted until the end of the Eocene. From the Callovian to the Eocene, the Great Caucasus Trough, the SCB, and the Kopet-Dagh Basin were united as a single deep-water basin (Zonenshain and Le Pichon, 1986). This was supposed to have been underlain by thinned continental (in the main part of the Great Caucasus Trough as no remnant of oceanic crust has been identified there) to oceanic crust locally in the SCB.

All the factors discussed above lead to the conclusion that after an Early(?) to Middle Jurassic rifting event, the SCB opened sometime between Callovian and Middle Cretaceous times. An important change in tectonic subsidence pattern occurred between the Bajocian–Bathonian and the Callovo–Oxfordian, with an increase of subsidence rates in the basin and a reduction on the margins. This event is well documented for the Pre-Caucasus area (Bolotov, 1996; Nikishin et al., 1998a; Ershov et al., 1998), for the SCB margins, and the Kopet-Dagh (cf. Fig. 5). It is probably an indication of the onset of spreading in the SCB part of the large back-arc deep-water basin that included the Great Caucasus Trough, the SCB, and the Kopet-Dagh Basin. During the Early Cretaceous, after a compressive phase (end of Jurassic/beginning of Cretaceous), opening of the SCB may have been reactivated during a second

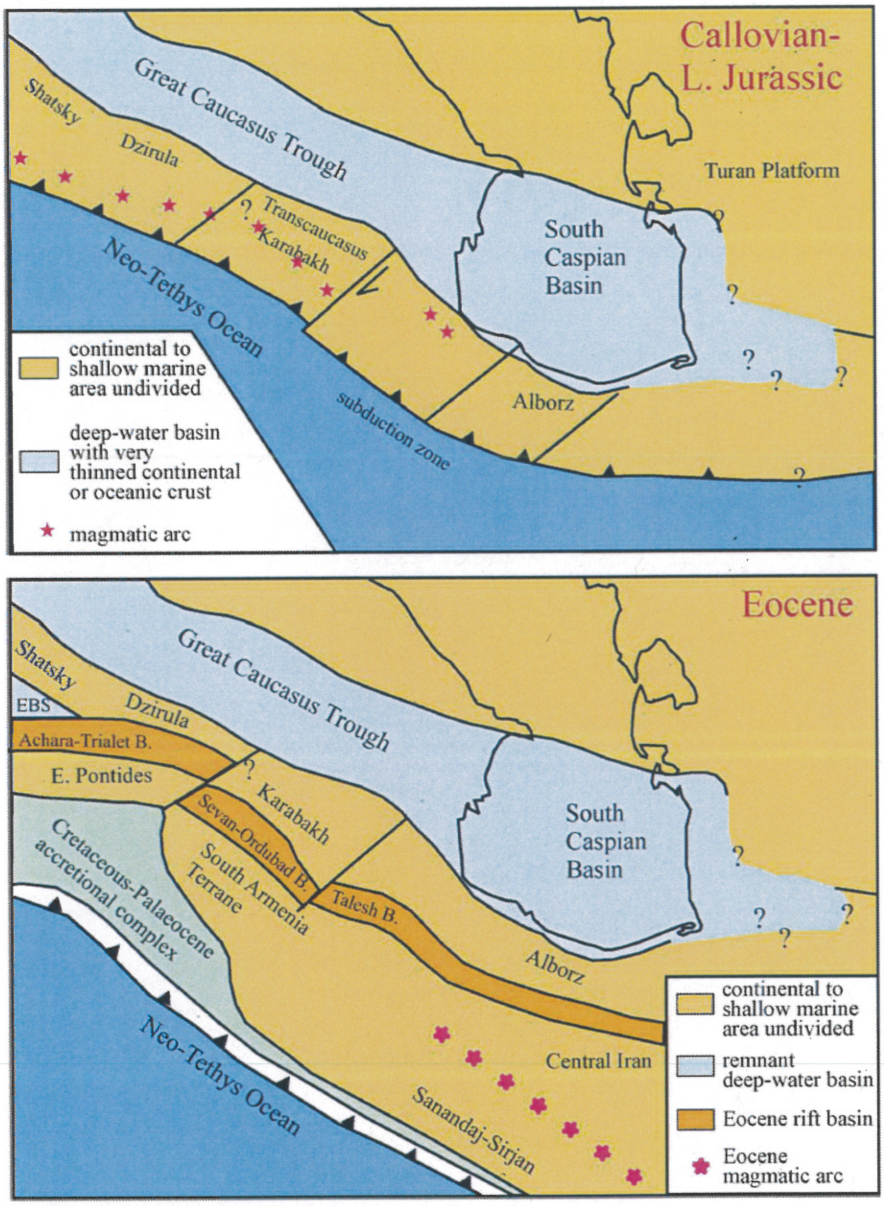

Fig. 7. Tentative reconstruction of the geodynamic setting of the South Caspian area during Callovian–Late Jurassic and Eocene stages; EBS: Eastern Black Sea.

extensional phase, perhaps in a strike–slip setting although more precise studies are necessary.

4.2. Water-depth in the South Caspian Basin

For the present study, we have made the choice of a deep-water depositional environment within the SCB from the Callovian. Having adopted the hypothesis that oceanic crust began to be formed in Callovo-Oxfordian times, we constructed a hypothetical subsidence curve (Fig. 8) in the centre of the SCB. The sedimentary succession taken for this area is from that observed in the cross-section (Fig. 2b). The requisite palaeobathymetries for determining tectonic subsidence were derived from the law of subsidence of oceanic basins proposed by Parson and Sclater (1977),

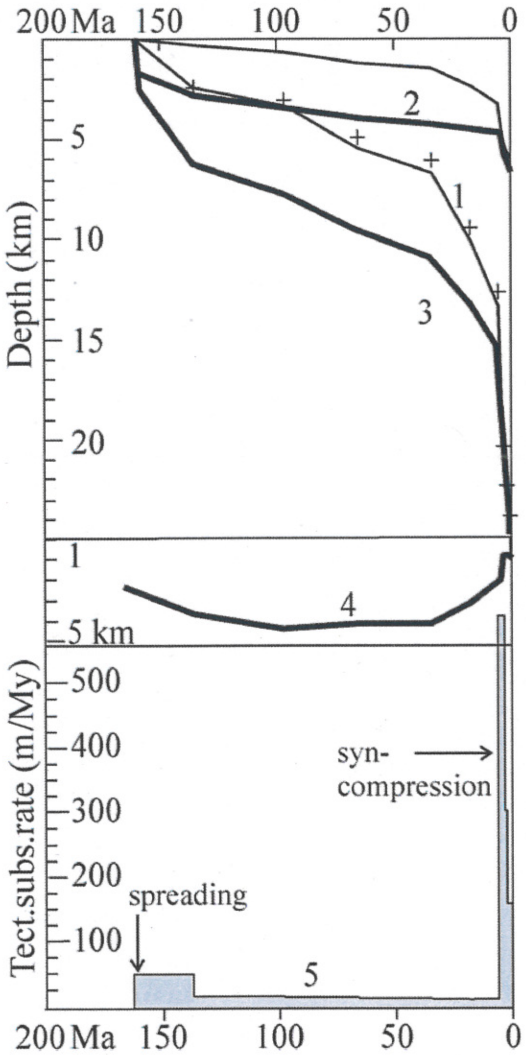

Fig. 8. Construction of a subsidence curve in the centre of the basin, taking palaeowater-depths of an oceanic crust of Callovian age (water-depths assumed with a law of oceanic basin evolution after Parson and Sclater, 1977): Depth (m) = 2500 + 350 (Age Ma)$^{1/2}$. (1) Basement deepening without corrections of water-depth only decompaction (crosses without decompaction); (2) tectonic subsidence with corrections of sea-level and bathymetry; (3) basement deepening with corrections; (4) bathymetry; (5) tectonic subsidence rate m/My (corrected).

viz. Depth (m) = 2500 + 350 (Age Ma)$^{1/2}$ until the Pliocene. The sea-level curve used for corrections (see Saïdi et al., 1997) is the first-order curve of Haq et al. (1987).

This has led us to assume a water-depth of more than 4000 m during Cretaceous–Palaeogene times that subsequently decreases as a result of sedimentary infill of the basin. During the Pliocene–Quaternary, the tectonic subsidence curve deviates from the oceanic crust cooling curve with a rapid increase of subsidence of almost 2 km. From seismic data on the slopes of the SCB, there is a possibility that the Pliocene environment was not deep, which would imply an even greater increase of tectonic subsidence during the last phase of Pliocene–Quaternary basin development.

Uncertainties related to the adopted palaeobathymetry evolution are great and represent the most important contribution to the errors associated with determining tectonic subsidence. When the synthetic wells are situated in a deep marine environment, subsidence curves and rates in the basin are strongly dependent on assumptions regarding absolute water-depths and how these change. Indeed, several problems are unsolved. First, during the Mesozoic, we need to determine when the basin actually began to be "deep". Second, the actual depth of water in a "deep" basin is extremely uncertain, depending, for example, on whether it is underlain by oceanic crust or not and on the age of this crust. Finally, it is important to know how the "deep" basin evolved in time, whether, for example, great depth was long-lived or whether it was punctuated by periods of significant shallowing.

According to facies and slope changes, water depth seems to increase during the Bathono–Callovian, in the first stage of basin evolution, from north-east Iran towards a deep SCB in Late Jurassic (Stampfli, 1978, 1999 personal communication). If it is assumed that oceanic crust formation began at this time, the bathymetry should remain deep until partial filling of the basin in the Cainozoic. The data available from areas surrounding the SCB seem to indicate a decrease of palaeobathymetry in the Cretaceous. But, in fact, it is possible that all our observations concern only the margins (even for points studied in the Kopet-Dagh) and never the deep oceanic basin itself or its lateral continuation.

To improve our knowledge of the evolution of the SCB, further studies are necessary of the Shemshak formation, the Cretaceous and the Eocene, especially in Iran. Better chronostratigraphic data are required for the Shemshak formation to provide more precise determination of the times of changes in depositional setting and the occurrence of important tectonic events (faulting, intrusion, etc.).

As for the Cretaceous, a basinal area should be looked for near the south-east corner of the SCB (in Gorgan area, near the Kopet-Dagh) and the basin extension towards the east–south-east must be clarified. The location, tectonic framework, and timing of the Early Cretaceous extensional event should also be better constrained.

The geometry and the character of the Eocene extensional volcanic basins must be further specified, from Achara-Trialet to south of the Alborz without forgetting a similar basin bordering the Sanandaj Sirjan block to the north-east of the Zagros. All these basins may have comprised a single complex before lateral and rotational displacement of different tectonic units.

5. Regional dynamics of the Oligocene to Quaternary South Caspian Basin subsidence

The subsidence evolution of the SCB clearly comprises two main stages. The first largely corresponds with extension and cooling from the Jurassic to the Pliocene. The second appears as a short, rapid phase of subsidence during Pliocene–Quaternary times in a generally compressive setting. What follows is a discussion of how the progressive shortening of the area can be linked to the movements of especially the Arabian plate towards the north and how this led to the second subsidence phase.

5.1. Oligocene to Quaternary regional context

The latest evolutionary stage of the SCB, including its main period of subsidence, took place in an environment of general compression which lasted from Oligocene until Recent times (e.g. Milanovsky, 1991; Kopp, 1997, 2000). It is related to the final collision of Arabia, and of the Indian block in the east, with the Eurasian margin, beginning in the Late Eocene–Oligocene and lasting until the present. The Kopet-Dagh foldbelt began to evolve from Oligocene times, together with other block movements towards the SCB (Milanovsky, 1991). This coincided with the development of the Apsheron–Balkhan transpression belt along the northern margin of the SCB. A general north-eastward oriented compression, combined with dextral strike–slip motion occurring in the Kopet-Dagh range (Jackson and McKenzie, 1984), drove the overthrusting of the Kopet-Dagh range onto the Turan–Kazakh platform. This was accompanied by a large-scale lithospheric forebulge to the north of the Kopet-Dagh (Nikishin et al., 1997; Thomas et al., 1999) and the development of a foredeep (Milanovsky, 1991; Otto, 1997).

The Alborz appears to have been partly overthrusted onto the southern part of the SCB as a result of the partitioning of oblique slip into strike–slip and thrust motions (Priestley et al., 1994). The SCB began to be subducted northwards below the Apsheron–Balkhan structure in Recent times (Granath and Baganz, 1996; Granath et al., 2000). Priestley et al. (1994) first hypothesised northwards subduction on the basis of earthquake focal mechanisms.

The Talesh trough has been inverted since the latest Eocene–Oligocene; mountain uplift has occurred during the Neogene along with clockwise rotation towards the SCB. The closing of the Great Caucasus Trough was accompanied by the development of foreland basins (Ershov et al., 2003). Folding and general inversion and the onset of uplift migrated from the central part of the Great Caucasus to the east during the Neogene (Milanovsky, 1968, 1991). Within the general shortening context, the triangular blocks of Kura, situated to the west, as well as Turkmenia to the east of the SCB, had a tendency to escape towards the centre of the SCB (Kopp, 1997, 2000), and to thrust over the SCB as they converged. The SCB was accordingly loaded on almost all its margins during this NS to NE–SW shortening. The final rapid and important subsidence phase in the SCB occurred in Pliocene–Quaternary, at the climax of shortening and compression, when the surrounding belts were being uplifted with their intense erosion providing the source material for sedimentary infill of the SCB.

In this general context, factors that could be driving mechanisms for the rapid subsidence of the SCB during Late Cainozoic–Quaternary times are as follows, without any specific order of importance or chronology:

(1) crustal/lithospheric thrusting towards the SCB, of the Caucasus, Alborz and Kopet-Dagh;
(2) gravitational loading of the surrounding orogens by the development of lithospheric roots (Caucasus, Kopet-Dagh and Alborz);

(3) syn-compressional downward "buckling" of the basin lithosphere, development of lithospheric syncline;
(4) onset of subduction of the SCB towards the north below Apsheron;
(5) possible reorganization of mantle movements over a wide area due to upper mantle flow and upper mantle cooling related Zagros subduction.

The quantification of the effects of these different mechanisms would require extremely complex models and detailed further discussion is far beyond the scope of this paper. However, consideration of a simple model, allows an order of magnitude estimate to be made of the Pliocene–Quaternary subsidence to be explained in great part by a syn-compressional flexure; in reality, the other mechanisms are very probably involved as well.

5.2. Flexural model of the Pliocene–Quaternary subsidence

Given that the oceanic/sub-oceanic crust of the SCB was formed in the Jurassic at latest Early Cretaceous times, by the Oligocene it was about 100 million years old; the lithosphere was therefore cold and strong during the whole compressional epoch.

The rapid phase of Plio-Quaternary basin subsidence was contemporaneous with an increase of compression and orogenesis on the borders of the basin. A recent episode of oceanic spreading cannot be advanced to explain this subsidence because related thermal cooling could not produce such an effect so quickly. We therefore examine the hypothesis that a flexural response of the basin lithosphere to an increase of imposed compressional force can largely explain the rapid syn-compressional subsidence (Berberian, 1983; Zonenshain and Le Pichon, 1986; Korotaev, 1998; Korotaev et al., 1999). This kind of mechanism has been invoked to explain a rapid Pliocene–Quaternary subsidence stage in the North Sea (Kooi et al., 1991) and in the Barents Sea (Korotaev et al., 1998). The predicted behaviour is also similar for Late Neogene subsidence of the Black Sea (Cloetingh et al., 2003; Nikishin et al., 2003). The key feature of the proposed mechanism is the presence of a lithospheric inhomogeneity of an appropriate wavelength, which results in a pre-existing equivalent elastic plate flexure (i.e. that of an "idealised" elastic plate whose response to loading is similar to the response of the lithosphere) (see Nikishin et al., 2003). The mechanical properties of the equivalent elastic plate are determined by the thickness (effective elastic thickness, EET) and configuration of the effective middle surface (EMS) of the loaded lithosphere. Both EET and EMS depend on thermal regime and crustal thickness (Burov and Diament, 1995; Ershov, 1999). Therefore, the requisite pre-existing deflection of the plate can be created not only by a pre-existing flexural deformation, but also by changes of thermal regime and crustal thickness (Ershov, 1999). An increase of compressional force induces an amplification of pre-existing deflections. The amplitude of the induced subsidence is proportional to the amplitudes of the force and the pre-existing deflection and depends on the wavelength of the latter (Cloetingh et al., 1985; Cloetingh and Kooi, 1992).

The lithosphere of the investigated SCB displays both structural and thermal heterogeneity. The lithosphere in the centre of the basin is surmised to be of oceanic (sub-oceanic) nature of Middle–Late Jurassic age. It is relatively cold and strong. On the other hand, recent orogens are developing on the borders of the basin (Caucasus, Alborz). These are areas of thick continental crust, high heat flow and active volcanism; it follows that the lithosphere of the orogenic areas is mechanically weak. We have used an algorithm developed by Ershov (1999) to calculate the effective elastic plate thickness (EET) along a north–south cross-section through the central part of the basin (Fig. 9). The crustal section is synthesised from the Moho map of Volvovsky and Starostenko (1996), the basement map and map of the oceanic crust location (Shikalibeily and Grigoriants, 1980; Berberian, 1983) and modified according to the data along the east–west crustal sections presented by Chernov (1990), Mamedov (1992), and the north–south section of Baranova et al. (1991). Heat flow data were extracted from the map of Belousov et al. (1991).

The EMS of an elastic plate mechanically equivalent to the South Caspian lithosphere is flexed downwards at the centre of the basin with an amplitude of about 25 km (Fig. 9). EET is about 10 km at the edges and about 40 km in the centre (Nadirov et al., 1997 investigated a similar, flexural model for the evolution of the SCB that has slightly lower parameters). Compressional forces

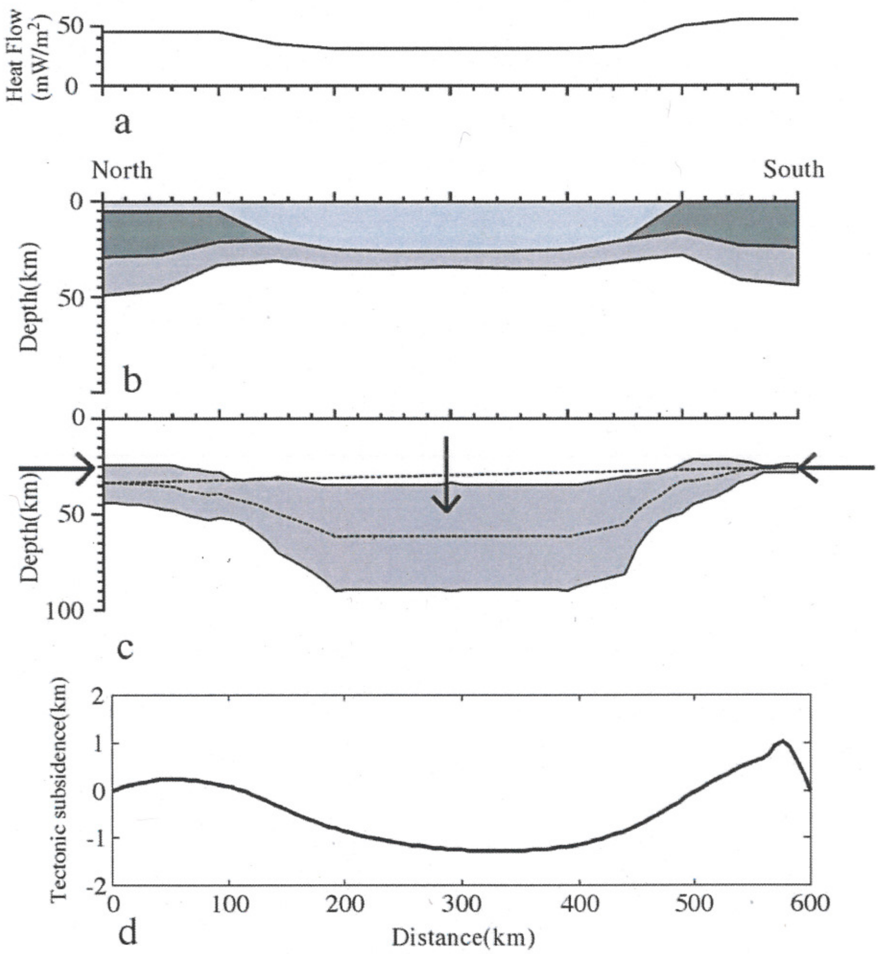

Fig. 9. Results of flexural modelling of the South Caspian Basin for the Pliocene–Quaternary stage with a force 10^{13} N/m. (a) Heat flow; (b) crustal section of a north–south synthetic profile (values taken from general maps of sediments and Moho depths); (c) configuration of effective elastic plate and effective middle surface (dashed line); (d) profile of tectonic subsidence obtained by the compression of the basin.

of 5×10^{13} N/m induce a tectonic subsidence with a maximum of about 2 km in the centre of the basin. This is in agreement with the "observed" tectonic subsidence from the burial history restoration (Fig. 8). The amplitude of the force is high in comparison with "typical" plate-tectonic forces ($2–10 \times 10^{12}$ N/m). This is partly a consequence of the choice of the elastic rheology. Cloetingh et al. (1989), for example, discussed the effects of adopting more realistic brittle–ductile rheologies on inferred stress levels. It can also be explained by the effects of stress concentrations on rigid inclusions, such as demonstrated by Nikitina (1997) and Brunet et al. (1997). Thus, the proposed, simple, 2D model can explain the order of magnitude of rapid syn-compressional subsidence in the SCB; nevertheless, the necessary tectonic force is quite high. An additional factor worth noting is that the problem is typically 3D, since the SCB is being thrusted under almost all its margins and subducted towards the north at least in its north-west corner. A much more complex model is needed, taking into account the loading by overthrusting on two or three of the SCB's four margins as well as bending associated with the subduction of the SCB towards the north-west below the Apsheron ridge.

6. Conclusions

We have deduced the history of the tectonic evolution of the SCB from an analysis of tectonic subsidence on its margins. Data from the Alborz indicate that a rifting phase on the southern margin of SCB took place in Early(?) to Middle Jurassic times, perhaps at the same time as the opening of the Great Caucasus Trough, which began in the Sinemurian. The north margin of the latter was prolongated into the SCB northern margin, towards the Great Balkhan. This opening occurred in the framework of the back-arc extension of a very long basin, as demonstrated by the presence of significant Middle Jurassic volcanism from the Pontides to Alborz through the Lesser Caucasus.

The Callovian–Late Jurassic phase of subsidence, mainly seen in the basinal area and in the lateral extensions of the SCB, is interpreted to be evidence of the beginning of oceanic crust formation. Depending on how wide the eventual oceanic basin was, the question remains of how deep the basin was at the end of Jurassic and whether it remained deep or not during the Cretaceous.

An Aptian–Albian phase of subsidence acceleration occurred on the margins of the SCB and it remains to be determined whether spreading possibly occurred in several discrete episodes or whether it occurred at different times in different parts of the SCB. This question is intimately linked with estimations of palaeobathymetry of the basin.

During Cretaceous to Eocene times, the basin generally displays gentle thermal subsidence affected by changes in ambient tectonic stress (compression events at the Jurassic/Cretaceous boundary and during the Late Cretaceous to Palaeocene) and was filled by deep-water sediments (mainly carbonates and turbidites). In the Eocene, according to some authors the age of opening of the SCB, we infer that active extension affected areas to the south of the SCB. This formed a series of extensional basins, in succession the Achara-Trialet to Sevan-Ordubad, Talesh, and to the south of the Alborz. These basins were filled by several kilometres of volcanics and tuffs and seem to have opened near the location of the Late Cretaceous–Early Palaeogene closure by terrane accretion of an earlier oceanic basin.

During Oligocene–Miocene times, rapid subsidence of the SCB took place, with deposition of clay being dominant within the basin. The start of the rapid subsidence phase was coincident with the onset of Alpine compression in the region. The former Kopet-Dagh basin was affected by crustal shortening and right-lateral transpression tectonics. The Great Caucasus Trough was also actively closing at this time; underthrusting of basinal crust to the north occurred during the Oligocene, becoming a more irregular collision of the former basin margins during the Miocene.

All the compressional units surrounding the SCB progressively settled, leading to the basin's last phase of evolution. During the Pliocene–Quaternary, an unusually rapid subsidence of the SCB took place; this was coincident with the uplift of the Great Caucasus, Kopet-Dagh and Alborz mountains. This general context of convergence, involving large horizontal compressive forces, may explain the short phase of rapid subsidence as a combination of loading of the edges of the basin and of syn-compressional downflexing of the entire basin. At the same time, the orogenic uplifts were intensely eroded, providing a large influx of sediments. These were transported towards the SCB by several major rivers and filled the SCB during its extraordinarily rapid phase of subsidence, allowing the formation of a prolific hydrocarbon system.

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References

Abdullayev, N.R., 2000. Seismic stratigraphy of the Upper Pliocene and Quaternary deposits in the South Caspian Basin. Journal of Petroleum Science and Engineering 28, 207–226.

Abrams, M.A., Narimanov, A.A., 1997. Geochemical evaluation of hydrocarbons and their potential sources in the western South Caspian depression, Republic of Azerbaijan. Marine and Petroleum Geology 14, 451–468.

Adamia, S.A., 1991. The Caucasus Oil and Gas Province. Occasional Publications. ESRI, New Ser. 7 (I-II), Part I, 53–74.

Afshar Harb, A., 1979. The stratigraphy, tectonics and petroleum geology of the Kopet-Dagh region, northern Iran. PhD Thesis, Imperial College, London, 316 pp.

Akhmedov, A.G., Guldust, M.A., 1981. Tectonic criteria of perspectives of oil–gas productivity of Kura depression. Soviet Geology 2, 34–45 (in Russian).

Alavi, M., 1991. Sedimentary and structural characteristics of the Paleo-Tethys remnants in northeastern Iran. Geological Society of America Bulletin 103, 983–992.

Alavi, M., 1996. Tectonostratigraphic synthesis and structural style of the Alborz Mountain system in Northern Iran. Journal of Geodynamics 21 (1), 1–33.

Ali-Zade, A.A., Ali-Zade, K.A., Aliyulla, Kh.A., 1996. Region XI. Azerbaijan. In: Krasheninnikov, V.A., Akhmetiev, M.A. (Eds.), Late Eocene–Early Oligocene geological and biotic events on the territory of the Former Soviet Union: Part I. The regional geology of the Upper Eocene and Lower Oligocene. GEOS, Moscow, pp. 112–130 (in Russian).

Ali-Zade, A.A., Khain, V.E., Ismail-Zade, A.D. (Eds.), 1999. The Saatly Ultra-Deep Well. The study of deep structure of the Kura intermontane depression according to the data of the drilling of the Saatly ultra-deep well SG-1. Nafta-Press, Baku. 288 pp. In Russian.

Anonymous, A.D., 1994. Petroleum Potential of Azerbaijan. Atlas BEICIP-FRANLAB, Rueil Malmaison, France.

Apol'skiy, O.P., 1974. Origin of the Black and South Caspian Sea troughs. Geotectonics, 310–311 (fasc. 5).

Bagirov, E., Nadirov, R., Lerche, I., 1997. Hydrocarbon evolution for a north–south section of the South Caspian Basin. Marine and Petroleum Geology 14, 773–854.

Banks, C.J., 1997. Mesozoic strike–slip back-arc basins of the Western Black Sea region. In: Robinson, A.G. (Ed.), Regional and Petroleum Geology of the Black Sea and Surrounding Region. Memoir vol. 68. American Association of Petroleum Geologists, Tulsa, OK, pp. 53–62.

Banks, C.J., Robinson, A.G., Williams, M.P., 1997. Structure and regional tectonics of the Achara-Trialet fold belt and the adjacent Rioni and Kartli Foreland Basins, Republic of Georgia. In: Robinson, A.G. (Ed.), Regional and Petroleum Geology of the Black Sea and Surrounding Region. Memoir vol. 68. American Association of Petroleum Geologists, Tulsa, OK, pp. 331–346.

Baranova, E.P., Kosminskaya, I.P., Pavlenkova, N.I., 1991. A reinterpretation of south Caspian DSS data. Geophysical Journal 10, 666–677.

Bazhenov, M.L., Burtman, V.S., Levashova, N.L., 1996. Lower and middle Jurassic paleomagnetic results from the south Lesser Caucasus and the evolution of the Mesozoic Tethys ocean. Earth and Planetary Science Letters 14, 79–89.

Belousov, V.V., Pavlenkova, N.I., Kvyatkovskaya, G.N. (Eds.), 1991. Deep Structure of USSR Territory. Nauka, Moscow. 224 pp. In Russian.

Belov, A.A., 1981. Tectonic History of the Alpine Fold Belt in the Paleozoic Nauka, Moscow. In Russian.

Belov, A.A., Burtman, V.S., Zinkevich, V.P., Knipper, A.L., Lobkovsky, L.I., Lukianov, A.V., Mazarovich, A.O., Makarov, V.I., Markov, M.S., Perfiliev, A.S., Puscharovsky, Yu.M., Rikhter, A.V., Rozen, O.M., Ruzhentsev, S.V., Savelieva, G.N., Samygin, S.G., Sokolov, S.D., Trifonov, V.G., Scherba, I.G., Baranov, G.I., Dobrzhinetskaya, L.F., Dotduev, S.I., Zlobin, V.L., Kurenkov, S.A., Rachkov, V.S., 1990. Tectonic layering of Lithosphere and Regional Geological Investigations Nauka, Moscow. In Russian.

Berberian, M., 1983. The southern Caspian: a compressional depression floored by a trapped, modified oceanic crust. Canadian Journal of Earth Sciences 20, 163–183.

Berberian, F., Berberian, M., 1981. Tectono-plutonic episodes in Iran. In: Gupta, H.K., Delany, F.M. (Eds.), Zagros. Hindu Kush, Himalaya Geodynamic Evolution. Geodyn. Ser., vol. 3. American Geophysical Union, Washington, DC, pp. 5–32.

Besse, J., Torcq, F., Gallet, Y., Ricou, L.-E., Krystyn, L., Saïdi, A., 1998. Late Permian to Late Triassic paleomagnetic data from Iran: constraints on the migration of the Iranian block through the Tethyan Ocean and initial destruction of Pangaea. Geophysical Journal International 135, 77–92.

Bolotov, S.N., 1996. Mesozoic–Cainozoic geological history of the Scythian Platform, and quantitative characteristics of the main stages of the evolution according to computer modelling. PhD Thesis. Moscow State Univ., Moscow, Russia (in Russian).

Boulin, J., 1991. Structures in Southwest Asia and evolution of the eastern Tethys. Tectonophysics 196, 211–268.

Bredehoeft, J.D., Djevanshir, R.D., Belitz, K.R., 1988. Lateral fluid flow in a compacting sand-shale sequence: South Caspian Basin. American Association of Petroleum Geologists Bulletin 72, 416–424.

Brunet, M.-F., Rebetsky, Y.L., Nikitina, Y.S., 1997. First steps of modelling in the South Caspian Sea. Peri-Tethys Programme Annual Meeting, June 10–12, 1997, Rabat, Morocco, 25–26.

Burov, E.B., Diament, M., 1995. The effective elastic thickness (Te) of continental lithosphere: what does it really mean? Journal of Geophysical Research 100, 3905–3927.

Chernov, A.A., 1990. Structure of Pre-Pliocene sediments of South Caspian. In: Lebedev, N.I. (Ed.), Caspian Sea: Questions of Geology and Geomorphology. Nauka, Moscow, pp. 105–110 (in Russian).

Cloetingh, S., Kooi, H., 1992. Intraplate stresses and dynamical aspects of rifted basins. Tectonophysics 208, 167–185.

Cloetingh, S., McQueen, H., Lambeck, K., 1985. On a tectonic mechanism for regional sealevel variations. Earth and Planetary Science Letters 75, 157–166.

Cloetingh, S., Kooi, H., Groenewoud, W., 1989. Intraplate stresses and sedimentary basin evolution. In: Pride, R.A. (Ed.), Origin and Evolution of Sedimentary Basins and Their Energy and Mineral Resources. Geophys. Monogr., vol. 48. American Geophysical Union, Washington, DC, pp. 1–16.

Cloetingh, S., Spadini, G., Van Wees, J.D., Beekman, F., 2003. Thermo-mechanical modeling of Black Sea Basin (de)formation. Sedimentary Geology, this issue.

Davies, R., Jones, C.R., Hamzepour, B., Clark, G.C., 1972. Geology of Masuleh Sheet, D3. Geological Survey of Iran, Tehran. 198 pp.
Davoudzadeh, M., Schmidt, K., 1984. A review of the mesozoic paleogeography and paleotectonic evolution of Iran. Neues Jahrbuch fuer Geologie und Paläontologie, Abhandlungen 168, 182–207.
Davoudzadeh, M., Soffel, H., Schmidt, K., 1981. On the rotation of the Central–East Iran microplate. Neues Jahrbuch fuer Geologie und Paläontologie, Monatshefte 3, 180–192.
Davoudzadeh, M., Lammerer, B., Weber-Diefenbach, K., 1997. Paleogeography, stratigraphy, and tectonics of the tertiary of Iran. Neues Jahrbuch fuer Geologie und Paläontologie, Abhandlungen 205, 33–67.
Dercourt, J., Zonenshain, L.P., Ricou, L.E., Kazmin, V.G., Le Pichon, X., Knipper, A.L., Grandjacquet, C., Sbortshikov, I.M., Geyssant, J., Lepvrier, C., Pechersky, D.H., Boulin, J., Sibuet, J.C., Savostin, L.A., Sorokhtin, O., Westphal, M., Bazhenov, M.L., Lauer, J.P., Biju-Duval, B., 1986. Geological evolution of the Tethys belt from the Atlantic to the Pamirs since the Lias. Tectonophysics 123, 241–315.
Dercourt, J., Ricou, L.E., Vrielynck, B. (Eds.), 1993. Atlas Tethys, Paleoenvironmental Maps. Gauthier-Villars, Paris. 307 pp. 14 maps, 1 plate.
Dercourt, J., Gaetani, M., Vrielynck, B., Barrier, E., Biju-Duval, B., Brunet, M.-F., Cadet, J.-P., Crasquin, S., Sandulescu, M. (Eds.), 2000. Atlas Peri-Tethys. Palaeogeographical Maps. CCGM/CGMW, Paris (24 maps and explanatory notes: I–XX; 269 pp).
Devlin, W.J., Cogswell, J.M., Gaskins, G.M., Isaksen, G.H., Pitcher, D.M., Puls, D.P., Stanley, K.O., Wall, G.R.T., 1999. South Caspian basin: young cool, and full of promise. GSA Today 9 (7), 1–9.
Ershov, A.V., 1999. Effective middle surface of lithosphere. Earth and Planetary Science Letters 173, 129–141.
Ershov, A.V., Brunet, M.-F., Nikishin, A.M., Bolotov, S.N., Korotaev, M.V., Kosova, S.S., 1998. Evolution of the eastern Fore–Caucasus basin during the Cenozoic collision: burial history and dynamic modelling. In: Crasquin-Soleau, S., Barrier, E. (Eds.), Peri-Tethys Memoir 4: Epicratonic Basins of Peri-Tethyan Platforms. Mém. Mus. natn. Hist. nat., Paris, vol. 179, pp. 111–130.
Ershov, A.V., Brunet, M.-F., Korotaev, M.V., Nikishin, A.M., Bolotov, S.N., 1999. Late Cenozoic burial history and dynamics of the Northern Caucasus molasse basin: implications for foreland basin modelling. In: Stephenson, R.A., Wilson, M., Starostenko, V.I. (Eds.), EUROPROBE Georift. Vol. 2: Intraplate Tectonics and Basin Dynamics of the Eastern European Craton and Its Margins. Tectonophysics, vol. 313, pp. 219–241.
Ershov, A.V., Brunet, M.-F., Nikishin, A.M., Bolotov, S.N., Nazarevich, B.P., Korotaev, M.V., 2003. Northern Caucasus basin: thermal history and synthesis of subsidence models. Sedimentary Geology, this issue.
Gasanov, T.Ab., 1992. The post-Eocene Nappes of the Talysh zone in the Lesser Caucasus. Geotectonics 26, 324–331.
Gasanov, T.Ab., 1996. Geodynamics of Ophiolites in the Structure of the Lesser Caucasus and Iran. Elm, Baku. In Russian.
Golonka, J., 2000a. Geodynamic evolution of the South Caspian Basin. Istanbul 2000, AAPG's Inaugural Regional International Conference, July 9–12, 2000. Istanbul, Turkey, pp. 40–45 (Abstract volume).
Golonka, J., 2000b. Cambrian–Neogene Plate Tectonic Maps Wydawnictwo Uniwersytetu Jagiellonskiego, Krakow. 125 pp. 37 plates.
Golonka, J., Osczypko, N., Slaczka, A., 2000. Late Carboniferous Neogene geodynamic evolution and paleogeography of the circum-Carpathian region and adjacent area. Annales Societatis Geologorum Poloniae 70, 107–136.
Granath, J.W., Baganz, O.W., 1996. A review of Neogene subsidence mechanisms for the south Caspian basin. 3rd Annual Meeting and Fieldtrip of IGCP Project n°369 "Comparative evolution of Peri-Tethyan rift basins", Centennial Celebration of EGSMA, November 19–24, Cairo, Egypt, pp. 24–25.
Granath, J.W., Soofi, K.A., Baganz, O.W., Bagirov, E., 2000. Gravity modelling and its implications to the tectonics of the South Caspian Basin. Istanbul 2000, AAPG's Inaugural Regional International Conference, July 9–12, 2000. Istanbul, Turkey, pp. 46–49 (Abstract volume).
Haghipour, A., Aghanabati, A., 1989. Geological map of Iran, scale 1:2,500,000. Ministry of mines and metals. Geological survey of Iran (second edition).
Haq, B.U., Hardenbol, J., Vail, P.R., 1987. Chronology of fluctuating sea levels since the triassic. Science 235, 1156–1167.
Ismail-Zade, A.D., Veliev, M.M., Mamedov, M.M., Bagirov, A.E., 1995. Tectonic position of the Talysh mafic–ultramafic complex. Geotectonics 29, 184–186.
Jackson, J., McKenzie, D., 1984. Active tectonics of the Alpine Himalayan belt between western Turkey and Pakistan. Geophysical Journal of the Royal Astronomical Society 77, 185–264.
Karyakin, Yu.V., 1989. Geodynamics of Volcanic Complexes Formation in the Lesser Caucasus Nauka, Moscow. 150 pp. In Russian.
Kazmin, V.G., 1991. Collision and rifting in the Tethys Ocean: geodynamic implications. Tectonophysics 196, 371–384.
Kazmin, V.G., Sborschikov, I.M., 1989. Paleozoic and early Mesozoic deformations on the Caucasus, and their place in the Tethys history. In: Belov, A.A., Satian, M.A. (Eds.), Geodynamics of the Caucasus. Nauka, Moscow, pp. 46–54 (in Russian).
Kazmin, V.G., Schreider, A.A., Bulychev, A.A., 2000. Early stages of evolution of the Black Sea. In: Bozkurt, B., Winchester, J.A., Piper, J.D.A. (Eds.), Tectonics and Magmatism in Turkey and the Surrounding Area. Special Publications, vol. 173. Geological Society, London, pp. 235–249.
Khain, V.E., 1994. Geology of the Northern Eurasia (ex-USSR). Second Part of the Geology of the USSR. Phanerozoic Fold Belts and Young Platforms. Gebrüder Borntraeger, Berlin. 390 pp.
Knapp, J.H., Diaconescu, C.C., Connor, J.A., McBride, J.H., Simmons, M.D., 2000. Deep seismic exploration of the South Caspian Basin: lithospheric-scale imaging of the world's deepest basin. Istanbul 2000, AAPG's Inaugural Regional International Conference, July 9–12, 2000, Istanbul, Turkey, pp. 35–37 (Abstract volume).
Knipper, A.L., Sokolov, S.D., 1974. Pre-upper Senonian tectonic sheets in the Lesser Caucasus. Geotectonics 1, 357–361.
Knipper, A.L., Sharaskin, A.Ya., Voznesenskii, A.I., 2001. Ophio-

lite-clastic breccias of the Lesser Caucasus: structural features and origin. Geotectonics 35, 199–206.

Kooi, H., Hettema, M., Cloetingh, S., 1991. Lithospheric dynamics and the rapid Pliocene–Quaternary subsidence phase in the southern North-Sea basin. Tectonophysics 192, 145–159.

Kopp, M.L., 1997. Lateral escape structures in the Alpine–Himalayan collisional belt. Scientific World. Transactions of the Institution of Geology RAS, Moscow 506. 314 pp.

Kopp, M.L., 2000. The recent deformations of the Scythian and Southern East European platforms as a result of pressure form the Arabian plate. Geotectonics 34, 106–120.

Koronovsky, N.V., Kozhevnikov, A.V., Panov, D.I., Vishnevsky, L.E., Guschin, A.I., Nikitin, M.Yu., Voronina, A.A., Stafeev, A.N., Panina, L.V., Kopaevich, L.F., Star, M.A., Belyakov, G.A., 1987. The history of geologic development and structure formation of the central part of Terek–Caspian trough. In: Milanovsky, E.E., Koronovsky, N.V. (Eds.), Geology and Mineral Resources of the Great Caucasus. Nauka, Moscow, pp. 147–174 (in Russian).

Koronovsky, N.V., Guschun, A.I., Nikitin, M.Yu., Panina, L.V., Stafeev, A.N., 1990. Gelogical development and origin of the recent structure of the Terek–Caspian foreland basin. In: Leonov, Yu.G., Khain, V.E. (Eds.), Tectonics of Orogenic Structures of the Caucasus and Middle Asia. Nauka, Moscow, pp. 4–35 (in Russian).

Koronovsky, N.V., Lomize, M.G., Guschin, A.I., Zarschikov, A.A., Panina, L.V., Romanovskaya, M.A., 1997. Main events in tectonic evolution of the Caucasus segment of the Mediterranean fold belt. Vestnik Moskovskogo Universiteta. Seria Geologia 4, 5–12 (in Russian).

Korotaev, M.V., 1998. Sedimentary basins in compressional environments: modelling of rapid subsidence phases. PhD Thesis, Moscow State Univ., Moscow, Russia (in Russian).

Korotaev, M.V., Nikishin, A.M., Ershov, A.V., 1998. Flexural modelling of the Barents Sea neotectonics. Abstracts of 6th Zonenshain Conference on plate Tectonics, Moscow, pp. 51–52.

Korotaev, M., Ershov, A., Nikishin, A., Brunet, M.-F., 1999. Sedimentary basins in compressional environment—modelling of the rapid subsidence. EAGE Meeting, June, 1999, Helsinki, Finland. Extended abstract.

Krylov, N.A. (Ed.), 1987. Structure and Oil-and-Gas Content of Basins of Internal Seas. Nauka, Moscow. 248 pp. In Russian.

Lasemi, Y., 1995. Platform carbonates of the Upper Jurassic Mozduran Formation in the Kopet Dagh basin, NE Iran-facies, palaeoenvironments and sequences. Sedimentary Geology 99, 151–164.

Levin, L.E., 1995. Volcanogenic and volcaniclastic reservoir rocks in Mesozoic–Cenozoic Island Arcs: examples from the Caucasus and the NW Pacific. Journal of Petroleum Geology 18, 267–288.

Lordkipanidze, M.B., 1980. Alpine Volcanism and Geodynamics of the Central Segment of the Mediterranean Fold Belt. Metsniereba, Tbilisi. 162 pp. In Russian.

Lordkipanidze, M., Gugushilli, V.I., Mikadze, G.A., Batiashvili, I., 1984. Volcanism and postvolcanic processes on the rift and island arc stages of evolution of the Adjara-Trialetia. Geologicke Prace Spravy 80, 91–100.

Luppov, N.P., 1972. Geology of USSR. Vol. 22 Turkmenian SSR. Geology Nedra, Moscow. 768 pp. In Russian.

Maksimov, S.P. (Ed.), 1992. Geological structure and economic minerals of the USSR. Vol. 6. Kazakhstan and Middle Asia. Book 3. Platform mantle of the epi-Paleozoic plates and depressions of Middle Asia and South Kazakhstan. Nedra, Moscow. 148 pp. In Russian.

Malovitsky, Ya.P., 1967. Perspectives for Oil and Gas in the Caspian Sea. Izd. Oil Ministry USSR, Ser. Oil and Gas Geology and Geophysics, Moscow. 78 pp. In Russian.

Mamedov, P.Z., 1992. Seismostratigraphical investigations of geological structure of sedimentary cover of South caspian superdepression and perspectives of oil–gas productivity. PhD Thesis, Baku, 40 pp. (in Russian).

Mangino, S., Priestley, K., 1998. The crustal structure of the southern Caspian region. Geophysical Journal International 133, 630–648.

Milanovsky, E.E., 1968. Neotectonics of the Caucasus. Nedra, Moscow. 683 pp. In Russian.

Milanovsky, E.E., 1991. Geology of the USSR. Part 3 Moscow Univ. Press, Moscow. 272 pp. In Russian.

Moussavi-Harami, R., Brenner, R.L., 1992. Geohistory analysis and petroleum reservoir characteristics of lower cretaceous (neocomian) sandstones, eastern Kopet-Dagh basin, northeastern Iran. American Association of Petroleum Geologists Bulletin 76, 1200–1208.

Nadirov, S.G. (Ed.), 1984. Geology and exploration of oil and gas fields of Azerbaijan and investigation of Saatly SG-1. Trudy AzNIPIneft, Baku (in Russian).

Nadirov, R.S., Bagirov, E., Tagiyev, M., Lerche, I., 1997. Flexural plate subsidence, sedimentation rates, and structural development of the super-deep South Caspian basin. Marine and Petroleum Geology 14, 383–400.

Narimanov, A.A., 1993. The petroleum systems of the South Caspian basin. In: Doré, A.G., Augustson, J.H., Hermanrud, C., Stewart, D.J., Sylta, O. (Eds.), Basin Modelling Advances and Applications. Norw. Petrol. Forening, Sp. Publ., vol. 3, Elsevier, Amsterdam, pp. 599–608.

Nikishin, A.M., Brunet, M.-F., Cloetingh, S., Ershov, A.V., 1997. Northern Peri-Tethyan Cenozoic intraplate deformations: influence of the Tethyan collision belt on the Eurasian continent from Paris to Tian-Shan. C. R. Acad. Sci. Paris 324 (IIa), 49–57.

Nikishin, A.M., Cloetingh, S., Bolotov, S.N., Baraboshkin, E.Y., Kopaevich, L.F., Nazarevich, B.P., Panov, D.I., Brunet, M.-F., Ershov, A.V., Il'ina, V.V., Kosova, S.S., Stephenson, R.A., 1998a. Scythian platform: chronostratigraphy and polyphase stages of tectonic history. In: Crasquin-Soleau, S., Barrier, E. (Eds.), Peri-Tethys Memoir 3: Stratigraphy and Evolution of Peri-Tethyan Platforms. Mém. Mus. natn. Hist. nat., Paris, vol. 177, pp. 151–162.

Nikishin, A.M., Cloetingh, S., Brunet, M.-F., Stephenson, R., Bolotov, S.N., Ershov, A.V., 1998b. Scythian Platform and Black Sea region: Mesozoic–Cenozoic tectonic and dynamics. In: Crasquin-Soleau, S., Barrier, E. (Eds.), Peri-Tethys Memoir 3: Stratigraphy and Evolution of Peri-Tethyan Platforms. Mém. Mus. natn. Hist. nat., Paris, vol. 177, 163–176.

Nikishin, A.M., Ziegler, P.A., Stephenson, R.A., Ustinova, M.A.,

1999. Santonian to Paleocene tectonics of the East-European Craton and adjacent areas. Bulletin de l'Institut Royal des Sciences Naturelles de Belgique. Science de la Terre 69 (Suppl. A), 147–159.

Nikishin, A., Ziegler, P.A., Panov, D.I., Nazarevich, B.P., Brunet, M.-F., Stephenson, R.A., Bolotov, S.N., Korotaev, M.V., Tikhomirov, P.L., 2001. Mesozoic and Cenozoic evolution of the Scythian Platform–Black-Sea–Caucasus domain. In: Ziegler, P.A., Cavazza, W., Robertson, A.H.F., Crasquin-Soleau, S. (Eds.), Peri-Tethys Memoir 6: Peri-Tethyan Rift/Wrench Basins and Passive Margins. Mém. Mus. natn. Hist. nat., Paris, vol. 186, pp. 295–346.

Nikishin, A.M., Korotaev, M.V., Ershov, A.V., Brunet, M.-F., 2003. The Black Sea Basin: tectonic history and Neogene–Quaternary rapid subsidence modelling. Sedimentary Geology, this issue.

Nikitina, Y., 1997. Modélisation analytique et numérique appliquée à la clarification du mécanisme de la déformation dans la région Caucase-Iran. PhD Thesis, Univ. Pierre et Marie Curie, Paris, France, Mém. Sc. Terre no. 97-05, 227 pp. (in French, with English Abstr.).

National Iranian Oil Company (NIOC) and Geological Survery of Iran (GSA), 1978. Geological map of Iran sheet no. 1 North–West Iran 1: 1,000,000. Compiled by Huber, H., Eftekhar-Nezhad, J.

Odin, G.S., 1994. Geological time scale (1994). C. R. Acad. Sci. Paris 318 (ser II), 59–71.

Okay, A.I., Sahintürk, Ö., 1997. Geology of the Eastern Pontides. In: Robinson, A.G. (Ed.), Regional and Petroleum Geology of the Black Sea and Surrounding Region. Memoir vol. 68 American Association of Petroleum Geologists, Tulsa, OK, pp. 291–311.

Okay, A.I., Tüysüz, O., 1999. Tethyan sutures of northern Turkey. In: Durand, B., Jolivet, L., Horváth, F., Séranne, M. (Eds.), The Mediterranean Basins: Tertiary Extension within the Alpine Orogen. Special Publications, vol. 156, Geological Society, London, pp. 475–515.

Ostroumova, A.S., Zeiter, I.Ya., 1987. Volcanic Series in Saatly superdeep well. Soviet Geology 7, 79–89 (in Russian).

Otto, S.C., 1997. Mesozoic–Cenozoic history of deformation and petroleum systems in sedimentary basins of central Asia: implications of collisions on the Eurasian margin. Petroleum Geoscience 3, 327–341.

Parson, B., Sclater, J.G., 1977. An analysis of the variation of ocean floor bathymetry and heat flow with age. Journal of Geophysical Research 82, 803–827.

Priestley, K., Baker, C., Jackson, J., 1994. Implications of earthquake focal mechanism data for the active tectonics of the south Caspian Basin and surrounding regions. Geophysical Journal International 118, 111–141.

Prozorovskiy, V.A., 1985. Role of Kimmerian movements in structural development of West-Central Asia. Geotectonics 19, 491–496.

Rad, F.K., 1986. A Jurassic Delta in the Eastern Alborz, NE Iran. Journal of Petroleum Geology 9 (3), 281–294.

Reyre, D., Mohafez, S., 1970. Une première contribution des accords NIOC-ERAP à la connaissance géologique de l'Iran. Première partie. Revue de l'Institut Français du Pétrole XXV, 687–713 (in French, with English Abstr.).

Ricou, L.E., 1996. The plate tectonic history of the past Tethys Ocean. In: Nairn, A.E.M., Ricou, L.E., Vrielynck, B., Dercourt, J. (Eds.), The Ocean Basins and Margins. The Tethys Ocean, vol. 8, Plenum, New York, pp. 3–70.

Ruttner, A.W., 1993. Southern borderland of Triassic Laurasia in NE Iran. Geologische Rundschau 82, 110–120.

Saïdi, A., 1995. Calendrier de la migration permo-triasique et morcellement mésozoïque des éléments continentaux de l'Iran. PhD Thesis, Univ. Pierre et Marie Curie, Paris, France, 298 pp. (in French).

Saïdi, A., Brunet, M.-F., Ricou, L.E., 1997. Continental accretion of the Iran block to Eurasia as seen from late Paleozoic to early Cretaceous subsidence curves. Geodinamica Acta 10, 89–208.

Sengör, A.M.C., 1984. The Cimmeride orogenic system and the tectonics of Eurasia. Geological Society of America Special Paper 195, 181–241.

Sengör, A.M.C., 1990. A new model for the late Paleozoic–Mesozoic tectonic evolution of Iran and implications for Oman. The Geology and Tectonics of the Oman Region. Geological Society Special Publication, vol. 49, 797–831.

Sengör, A.M.C., Yilmaz, Y., Sungurlu, O., 1984. Tectonics of the Mediterranean Cimmerides: nature and evolution of the western termination of Paleo-Tethys. In: Dixon, T.H., Robertson, A.H.F. (Eds.), The Geological Evolution of the Eastern Mediterranean. Geological Society of London Special Publication, vol. 17, pp. 77–112.

Shikalibeily, E.S., Grigoriants, B.V., 1980. Principal features of the crustal structure of the South-Caspian Basin and the conditions of its formation. Tectonophysics 69, 113–121.

Shikalibeily, E.S., Abdullaev, R.N., Ali-Zade, A., 1988. Geological results of the super-deep well of Saatly. Sovetskaya Geologiya 11, 61–66 (in Russian).

Stampfli, G., 1978. Etude géologique générale de l'Alborz oriental au Sud de Gonbad-e-Qabus (Iran NE). Doctorate Dissertation, Univ. Genève, Switzerland, 328 pp. (in French).

Stampfli, G., 2000. Tethyan oceans. In: Bozkurt, E., Winchester, J.A., Pieper, J.D.A. (Eds.), Tectonics and Magmatism in Turkey and the Surrounding Area. Special Publications, vol. 173, Geological Society, London, pp. 1–23.

Stampfli, G.M., Mosar, J., Favre, P., Pillevuit, A., Vannay, J.-C., 2001. Late Palaeozoic to Mesozoic evolution of the Western Tethyan realm: the Neothethys–East Mediterranean basin connection. In: Ziegler, P.A., Cavazza, W., Robertson, A.H.F., Crasquin-Soleau, S. (Eds.), Peri-Tethys Memoir 6: Peri-Tethyan Rift/Wrench Basins and Passive Margins. Mém. Mus. natn. Hist. nat., Paris, vol. 186, pp. 51–108.

Steckler, M.S., Watts, A.B., 1978. Subsidence of the Atlantic type continental margin off New York. Earth and Planetary Science Letters 41, 1–13.

Stöcklin, J., 1968. Structural history and tectonics of Iran: a review. American Association of Petroleum Geologists Bulletin 52, 1229–1258.

Stöcklin, J., 1974. Possible ancient continental margin in Iran. In: Burk, C.A., Drake, C.L. (Eds.), The Geology of Continental Margins. Springer Verlag, pp. 873–887.

Stöcklin, J., 1981. A brief report on geodynamics in Iran. In: Gupta, H.K., Delany, F.M. (Eds.), Zagros. Hindu Kush, Himalaya Geodynamic Evolution. American Geophysical Union, Geodyn. Ser., vol. 3, pp. 70–74.

Tagiyev, M.F., Nadirov, R.S., Bagirov, E.B., 1997. Geohistory, thermal history and hydrocarbon generation history of the northwest South Caspian Basin. Marine and Petroleum Geology 14, 363–382.

Thomas, J.C., Grasso, J.R., Bossu, R., Martinod, J., Nurtaev, B., 1999. Recent deformation in the Turan and South Kazakh platforms, western central Asia, and its relation to Arabia–Asia and India–Asia collisions. Tectonics 18, 201–214.

Volvovsky, B.S., Starostenko, V.I. (Eds.), 1996. Geophysical Parameters of the South Sector of Alpine Orogen. Naukova Dumka, Kiev. In Russian.

Wensink, H., Varekamp, J.C., 1980. Paleomagnetism of basalts from Alborz: Iran part of Asia in the Cretaceous. Tectonophysics 68, 113–129.

Yilmaz, A., Terzioglu, N., 1994. The geotectonic setting of Upper Cretaceous–Tertiary volcanism along the Eastern Pontian zone. International Volcanological Congress, p. 39 (Abstracts).

Yilmaz, S., Boztug, D., 1996. Space and time relations of three plutonic phases in the Eastern Pontides, Turquey. International Geology Review 38, 935–956.

Yilmaz, Y., Tüysüz, O., Yigitbas, E., Can Genç, S., Sengör, A.M.C., 1997. Geology and tectonic evolution of the Pontides. In: Robinson, A.G. (Ed.), Regional and Petroleum Geology of the Black Sea and Surrounding Region. Memoir vol. 68 American Association of Petroleum Geologists, Tulsa, OK, pp. 183–226.

Yilmaz, A., Adamia, S., Chabukiani, A., Chkhotua, T., Erdogan, K., Tuzcu, S., Karabiyikoglu, M.F., 2000. Structural correlation of the southern Transcaucasus (Georgia)–eastern Pontides (Turkey). In: Bozkurt, B., Winchester, J.A., Piper, J.D.A. (Eds.), Tectonics and Magmatism in Turkey and the Surrounding Area. Special Publications, vol. 173, Geological Society, London, pp. 171–182.

Ziegler, P.A., Cloetingh, S., Guiraud, R., Stampfli, G.M., 2001. Peri-Tethyan platforms: constraints on dynamics of rifting and basin inversion. In: Ziegler, P.A., Cavazza, W., Robertson, A.H.F., Crasquin-Soleau, S. (Eds.), Peri-Tethys Memoir 6: Peri-Tethyan Rift/Wrench Basins and Passive Margins. Mém. Mus. natn. Hist. nat., Paris, vol. 186, pp. 9–49.

Zonenshain, L.P., Le Pichon, X., 1986. Deep basins of the Black Sea and Caspian Sea as remnants of Mesozoic back-arc basins. Tectonophysics 123, 181–211.

Zonenshain, L.P., Kuzmin, M.I., Natapov, L.M., 1990. Geology of the USSR: a plate-tectonic synthesis. In: Page, B.M. (Ed.), American Geophysical Union, Geodyn. Ser., Washington, DC, vol. 21, 242 pp.