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DEEP BASINS OF THE BLACK SEA AND CASPIAN SEA AS REMNANTS OF MESOZOIC BACK-ARC BASINS

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ABSTRACT

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We review the geological and geophysical structural framework of the deep Black Sea and Caspian Sea basins. Based on seismic evidence and subsidence history, we conclude that the deep basins have an oceanic crust formed in a marginal sea environment. We propose that the present deep basins are remnants of a much greater marginal sea formed during three separate episodes during the Mesozoic: in the Middle Jurassic, Upper Jurassic and Late Cretaceous. A tentative sketch of the geologic evolution of the area is presented. The marginal sea reached its greatest extent in the Early Tertiary when it was about 900 km wide and 3000 km long. The central part of the marginal sea has since disappeared during the collision between the Arabian promontory and the Eurasian margin.

INTRODUCTION

The deep basins of the Black Sea and South Caspian Sea have no low-velocity crustal ("granitic") layer. An oceanic-type crust with a compressive velocity v_p of 6.8 to 7 km s^{-1} underlies a sedimentary layer which is 15 km thick in the Black Sea and about 20 km thick in the South Caspian Sea (Gegelyantz et al., 1958; Neprochnov, 1966). For a long period, it was assumed that both basins were very young (Muratov, 1972): Late Miocene or even Pliocene–Quaternary. Lately, such an idea was accepted by Yanshin et al. (1980) and Shlezinger (1981) who proposed basification of former continental basement by eclogitization of granite rocks. A completely opposite hypothesis was proposed by Sorokhtin (1979) and Vardapetyan (1981) who consider the Black Sea and Caspian Sea basin as remnants of Early Mesozoic Tethys floor. The main evidence for their hypothesis is the great thickness of sediments that manifests the magnitude of the subsidence and suggests a long life for these basins. Their Jurassic estimate for the age of the basement is obtained by assuming that it is

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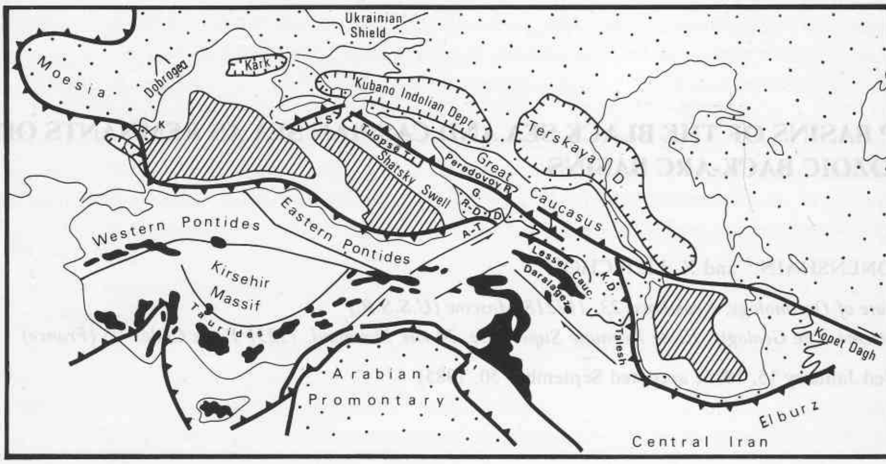

Fig. 1. Positions of deep basins of the Black Sea and Caspian Sea within the Mediterranean belt. Deep basins are striped. Black–ophiolite; dots—undeformed Cenozoic sequences; toothed line—front of the main alpine thrusts; line with arrow—strik—slip fault. Barbed line—contour of young depressions. *A–T* —Adjaro–Trialetian, *B*—Burgass synclinorium, *D*—Dzirula massif, *G*—Gagro–Chjawa zone, *K*—Lower Kamchian trough, *K.D.*—Kura Depression, *Kark*—Karkinitsky trough, *R-D*—Rionian depression, *S*—Sorokhtin deep.

oceanic and follows the "normal" oceanic square root of age subsidence in the absence of sediments and that it is in local isostatic equilibrium. Similar considerations made on the basis of heat flow by A. Golmshtok (pers. commun., 1983) also led to a Jurassic age.

However, the Early Mesozoic age proposed by these authors is not supported by geologic data which suggest that these basins did not exist before the Mid Late Jurassic as Upper Jurassic conglomerates of Crimea received clastics from the south, that is from the area now occupied by the Black Sea, whereas the Rhaetian–Liassic Shemshak formation in the Elburz Mts. was fed by detritics derived from the north, from the area now occupied by the South Caspian basin. Apolskii (1974) interpreted the basins of the Black Sea and South Caspian Sea as formed along en-échelon extensional faults by sinistral strike–slip from the Kopet Dag to the Carpathians. Finally Adamia et al. (1974) considered the Black Sea and Caspian Sea basins as remnants of back-arc basins resulting from spreading behind a Late Cretaceous–Paleogene volcanic island arc. The same hypothesis was made by Letouzey et al. (1977). Recent detailed study of deep sea basins, especially of the Black Sea basin, using in particular multichannel seismic reflection, indicates that these basins were indeed formed by extension during the Middle to Late Mesozoic.

The deep basins of the Black Sea and Caspian Sea are respectively located to the west and east of the Arabian promontory, which faces the Caucasus (Fig. 1). For the last 80 m.y., this promontory has moved northward with Eurasia (Savostin et al., 1986). As a result, the Turkey and Iran micro-plates are now moving apart

(McKenzie, 1978) while the crust of the deep-sea basins is thrust below the Eurasian margin at a rate of about 1 cm yr^{-1} in the region of Crimea and the Caucasus (Vardapetyan, 1979). Fold chains surround the deep-sea basins, especially along their southern boundary. Locally, their northern boundary cuts the folded structures as along the western coast of the Black Sea or the eastern margin of the Caspian basin. However, the general shape of the chain of Crimea, the Caucasus, Balkans and Kopet Dag follows the outlines of the deep basins.

STRUCTURES AROUND THE BLACK SEA

The basins of the Black Sea seem to act as a structural boundary for the Late Cretaceous and Paleogene structures. These structures were formed after the first collision of continental fragments of Gondwanaland with the Eurasian active margin in the Middle–Late Cretaceous (Dercourt et al., 1986). The Late Cretaceous–Paleogene volcanic belt is continuously traced south of the basin, from the Balkans through the Pontides in Turkey, the Adjaro–Trialetian zone and Talesh in the Caucasus to the Elburz and Central Iran (see Fig. 1). North of the basins, no volcanism of that time is detected, but flysch is widespread (for instance, the flysch zone of the southern slope of the Great Caucasus). This fact alone is significant evidence in favour of the hypothesis of deep basins as remnants of marginal seas formed behind island arcs.

In the Caucasus, north of the Sevan–Akkera zone, which corresponds to the suture of the Tethys Ocean, there are three large structures formed before the Late Cretaceous–Paleogene volcanic belt: (1) the Jurassic and Cretaceous volcanic belt of the Lesser Caucasus, or Somkheto–Agdam, (2) the Dzirula massif and (3) the anticlinorium of the Great Caucasus.

The Lesser Caucasian (Somkheto–Agdam) volcanic belt consists of calc-alkaline series that resulted from the activity of a Jurassic–Cretaceous island arc (Adamia et al., 1977; Lordkipanidze, 1980). The belt continues to the west with contemporaneous volcanics in the Eastern Pontides although the connection is covered by Paleogene volcanics and sediments of the Adjaro–Trialetian zone. The Lesser Caucasian belt seems to be mainly underlain by the metamorphic series of the Transcaucasian massif though locally it rests upon ophiolites corresponding to a former oceanic floor.

The Dzirula massif has a Hercynian core, thus showing European affinities rather than Gondwanian ones. The Jurassic–Cretaceous sequence lies unconformably above it and forms the sedimentary cover of the massif. Its lower parts consist of volcanics especially abundant during the Bajocian. Volcanics belong to the shoshonitic series and were formed in an islands-arc environment (Lordkipanidze, 1980). Volcanic activity almost ceased after the Bajocian. The above deposits compose a sedimentary sequence which is continuous up to and including the Eocene and which was only deformed during and after the Miocene. Upper Jurassic evaporites and redbeds are covered by Cretaceous flysch-like deposits indicating considerable subsidence of the massif. The whole rock complex, Bajocian lavas included, continues from the Dzirula massif onto the southern slope of the Great Caucasus and then westward in the so-called Gagro-Chjawa zone which runs into the Black Sea near Gagra. The Dzirula massif corresponds to a microcontinent which, since the Jurassic, has experienced considerable subsidence behind the Lesser Caucasian arc.

The Great Caucasus is an anticlinorium with a folded Paleozoic core at its center (Peredovoi Ridge). Jurassic and Cretaceous thick shales with deep-water cherts (i.e., Cenomanian–Turonian Ananurat horizon) are most typical of it. Lower–Middle Liassic (Sinemurian) calc-alkaline volcanics of island arc (andesites, dacites) are present in the lower part of the shale sequence. Locally, they rest upon the Hercynian folded basement. Lordkipanidze (1980) supposes the occurrence during the Liassic of an independent Great Caucasian island arc probably inherited from the Triassic time. From the Toarcian to the Bajocian, oceanic type tholeiites and also rhyolites are widely spread. These lavas are evidence for distension and formation of a large marginal basin with oceanic crust or strongly thinned continental crust at the location of the Great Caucasus (Adamia et al., 1977, 1982; Gamkrelidze, 1982). Part of this basin was probably inherited from the Late Paleozoic and Triassic because the deposits of that age, composing the Dizs series of Svanetia, are exposed in a conformable section under Jurassic shales. The northern flank of the Great Caucasian basin had already been deformed by the Late Jurassic, during the Cimmerian orogeny; it then became part of the Eurasian continental margin. The continuous shelf carbonate sequence of Jurassic, Cretaceous and Paleogene age was deposited on this margin. The orogeny of the northern flank indicates partial consumption of the crust of the Great Caucasian basin. Its central and southern parts continued their subsidence until the Oligocene (up to the Maikopian) when the orogeny of the entire distended series of the Great Caucasus first started, as uplift began in its central portion.

The main suture along which the complexes of the Dzirula massif were converging with the complexes of the Great Caucasus coincides with the Main Thrust of the southern slope of the Great Caucasus to the east, and, to the west, with the fault zone that follows the southern boundary of the Hercynian core of Peredovoi Ridge. Displacements and deformations occurred till Recent and continue at present as indicated by the seismicity belt which follows the southern slope of the Great Caucasus and is traced along the Black Sea shore. Focal mechanism solutions indicate compression across the Great Caucasus (Vardapetyan, 1979). Some of the Caucasian structures continue into the Black Sea: such as the Adjaro–Trialetian zone, the Dzirula massif and the Gagro-Chjawa zone.

The Crimea highland greatly differs from the Great Caucasus in the widespread occurrence of Triassic–Liassic flysch (Tauric series) and olistostromes (Eskiordian suite) and also in the general manifestation of the Cimmerian orogeny in mid-Jurassic.

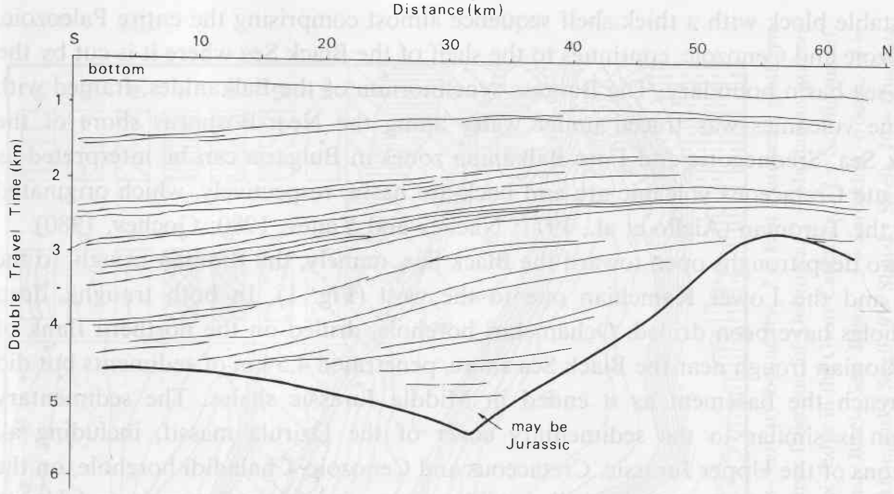

Fig. 2. Tracing of seismic reflection cross-section through the Karkinitsky trough, NW Black Sea shelf (after Tugolessov et al., 1983). The structure suggests the existence of a block tilted northward along southern facing fault.

However, the Lower–Middle Jurassic volcanics there consist of basalts and basaltic andesites similar to the Great Caucasian ones. Bajocian basalts of the shoshonitic series are an obvious continuation of the volcanics of the Gagro-Chjawa zone of the Great Caucasus and the missing link is presumably buried under the Black Sea basin.

North of the Crimea mountains, in Stepnoi Crimea, and on the northwestern shelf of the Black Sea, a 5-km deep Karkinitsky or North Crimea trough was revealed by drilling and seismic profiling (Figs. 1 and 3, Tugolessov et al., 1983a). It is composed of Cretaceous (probably including Upper Jurassic) and Cenozoic sequences sharply unconform on the Hercynian fold basement of the Scythyan platform. Cretaceous deposits also rest unconformably upon the southern flank of the trough on the Triassic–Jurassic fold basement of the Crimea highland. In the Albian and Cenomanian–Turonian, calc-alkaline volcanics appear within the sedimentary cover. In borehole "Ilichevskaya-2", on the saddle between the Karkinitsky trough and the Black Sea basin, their thickness reaches 500 m; they are composed of andesites, diabases and tuffs (Tugolessov et al., 1983a). Multi-channel seismic reflection data show that the basement of the Karkinitsky trough is split by faults into blocks dipping northward, thus presenting a typical pattern of extension and thinning of the continental crust (Fig. 2).

The structures surrounding the Black sea to the west, namely, Dobrogea, the Moesian platform and the northern Balkans are cut by the Black Sea basin. The Cimmerian structures of Dobrogea can be traced through the western shelf of the Black Sea to their continuation in a Crimea highland. The Moesian platform which is a stable block with a thick shelf sequence almost comprising the entire Paleozoic, Mesozoic and Cenozoic, continues to the shelf of the Black Sea where it is cut by the deep-sea basin boundary. The Burgass synclinorium of the Balkanides, framed with Eocene volcanics was traced under water along the Near-Bosporus shore of the Black Sea. Srednegorie and Fore-Balkanian zones in Bulgaria can be interpreted as the Late Cretaceous volcanic arc and back-arc basin, respectively, which originated after the Turonian (Aiello et al., 1977; Nachev and Yanev, 1980; Gochev, 1980).

Two deep troughs open toward the Black Sea, namely, the Rionian trough, to the east, and the Lower Kamchian one to the west (Fig. 1). In both troughs, deep boreholes have been drilled. Ochamchiri borehole, drilled on the northern flank of the Rionian trough near the Black Sea shore, penetrated 4.5 km of sediments but did not reach the basement as it ended in Middle Jurassic shales. The sedimentary section is similar to the sedimentary cover of the Dzirula massif, including all horizons of the Upper Jurassic, Cretaceous and Cenozoic. Chaladidi borehole, on the southern flank of the trough, drilled Albian volcanics below Cenozoic and Upper Cretaceous sediments; further down, Lower Cretaceous carbonate sediments rest on Upper Jurassic basalts at a depth of 4 km. M.B. Lordkipanidze (pers. commun., 1983) considers Jurassic lava as part of an alkaline-basaltic series marking extensional conditions. Choloki borehole, located on the continuation of the Adjaro–Trialetian zone, penetrated Eocene effusives at a depth of 3.5 km.

In the Lower Kamchian trough to the west, with its northeastern Varna depression prolongation, drilling in boreholes Igneada, Kamchia, Tulenevo, Kamsakra, etc. penetrated sediments of various thicknesses extending down into the Neogene, Paleogene and locally into the Cretaceous. It should be noted that an unconformity is traced almost everywhere at the bottom of Cretaceous or Upper Jurassic–Lower Cretaceous deposits (Bomarskaya, 1979). The Varna depression only began its rapid subsidence in the Eocene, the subsidence being controlled by faults (Foose and Mannheim, 1975; Golovinsky and Glumov, 1976).

DEEP STRUCTURE OF BLACK SEA BASINS

Better knowledge of the deep structure of the Black Sea was obtained through a 5-yr research project using multichannel seismic reflection profiling (MSR) with common depth point technique (CDP) on a net of profiles at 25 km spacing. The work was performed by E.M. Khakhalev. Results of the research are presented by Tugolessov et al. (1983a, b), Pustilnikov et al. (1980), Terekhov (1979) and Gorshkov (1983). Earlier research was performed in the Black Sea in connection with DSDP Leg 42 in 1975 (Letouzey and Montadert, 1978; Neprochnov, 1980). Geophysical and geological studies of the Black Sea have been carried out by many researchers (Neprochnov, 1966; Goncharov et al., 1972; Neprochnov et al., 1974; Ross et al., 1974; Letouzey et al., 1977; Shimkus, 1977; etc.). The structural map compiled on

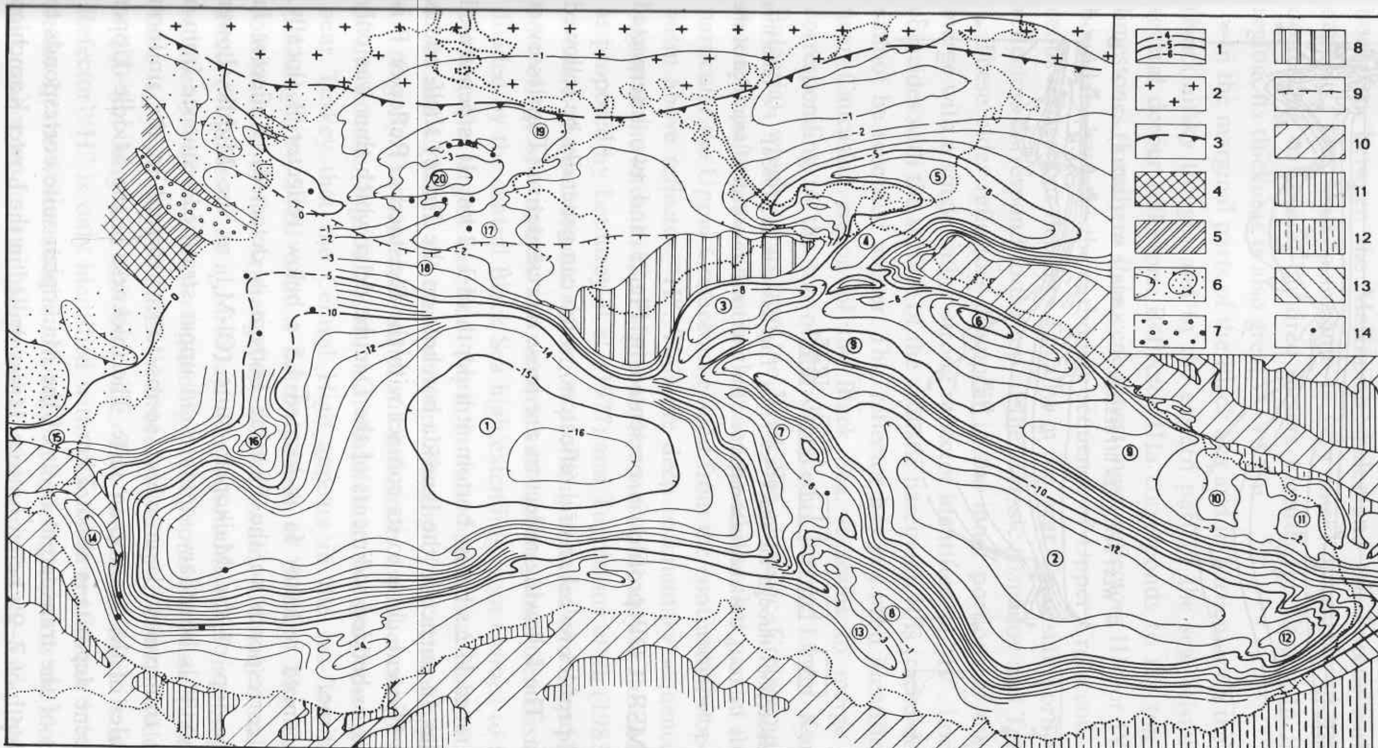

Fig. 3. Structure of the Black Sea (after Tugolessov et al., 1983a). *1* = isobaths (in km) of Cretaceous–Tertiary boundary, *2* = East European platform, *3* = limit of the East European platform, *4* = Riphean of Dobrogea, *5* = Paleozoic of Dobrogea, *6* = Upper Paleozoic of Dobrogea, *7* = Maidanian graben, *8* = Crimea, *9* = limit of the Crimea–Dobrogea Triassic trough, *10* = Alpides, *11* = Paleozoic of Peredovoi Ridge in the Great Caucasus, *12* = Szirula massif, *13* = Adjaro–Trialetian zone, *14* = boreholes. Figures in circles: *1* = West Black Sea basin, *2* = East Black Sea Basin, *3* = Sorotkhin trough, *4* = Kerch–Tamanian trough, *5* = Indolo-Kubanian trough, *6* = Tuapse trough, *7* = Andrussov Swell, *8* = Arkhangelsky Swell, *9* = Shatsky Swell, *10* = Gudauta Swell, *11* = Ochamchiri Swell, *12* = Gurian trough, *13* = Sinopan trough, *14* = Burgass trough, *15* = Lower Kemchian trough, *16* = Polshkov Rise, *17* = Kalamitsky Swell, *18* = South-Kalamitsky scarp, *19* = Karkinitsky trough, *20* = Mikhailovsky depression.

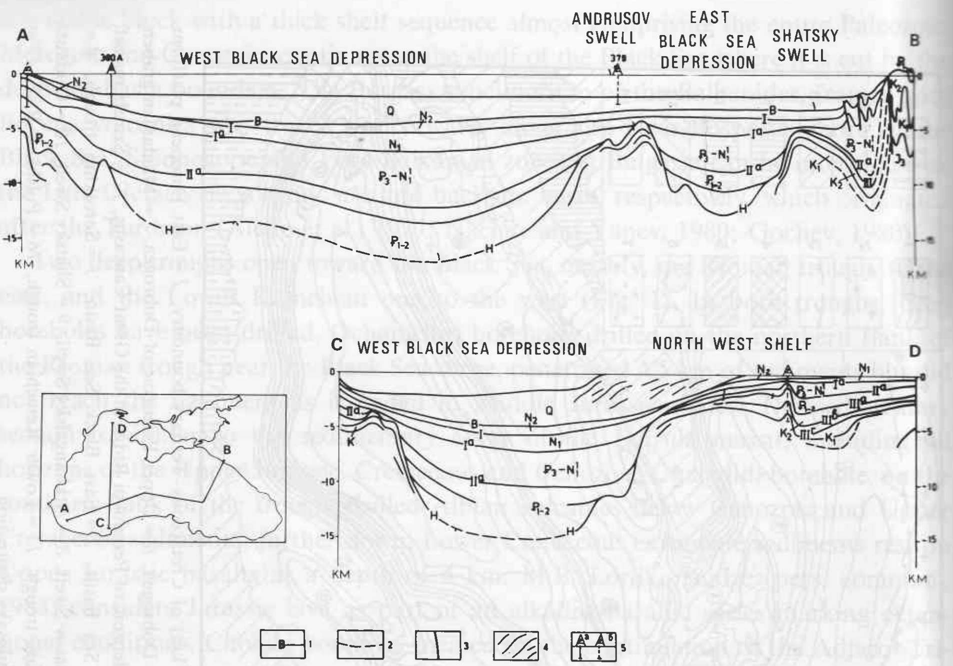

Fig. 4. Geological cross-sections through the Black Sea (after Tugolessov et al., 1983a). *1* = reflectors (identification is given in the text), *2* = uncertain reflectors, *3* = fault, *4* = cross-bedded sequence of the Paleo-Danube delta, *5* = boreholes.

the basis of the MSR CDP profiles is presented in Fig. 3 and two interpreted sections in Fig. 4.

The MSR CDP profiles reveal several reflectors, which can generally be followed over the whole sea. The following reflectors are most representative (Tugolessov et al., 1983b).

Reflector B is traced 1–2 s below bottom at a depth of 1–2 km. It is penetrated by several boreholes, in particular the Igneada borehole on the Turkey shelf, and is located within the Pliocene close to its contact with the Pleistocene. Reflector B is traced below the cross-bedded sediments of the Danube delta which thus was only formed in Recent time.

The next pronounced reflector Ia is located 1.5 s below reflector B (locally, reflector I which corresponds to the Pliocene bottom is detected). Reflector Ia corresponds to the top of the Maikopian suite (Ol-M_{c_1} at the Middle/Lower Miocene boundary). This boundary separates an upper stratified sequence from a lower acoustically transparent layer which seems to correspond to the uniform non-calcareous shales of the Maikopian suite. The thickness of the Middle–Upper Miocene and Pliocene deposits is about 2 km.

The lower limit of the transparent layer of the Maikopian suite corresponds to reflector IIa at a depth of 7–9 s below bottom. It was drilled in the Lower Kamchian trough in Bulgaria on the northwestern sea shelf (Ilichevskaya borehole) and in the Rionian trough in the Transcaucasian region. This reflector corresponds to the boundary between the Maikopian shales and the Eocene carbonate terrigenous deposits. The Maikopian suite is very thick; it reaches 3–5 km in depressions decreasing to several hundred meters towards the flanks. In the Fore-Caucasian region, its thickness is also great, 3–3.5 km.

In the marginal parts of the Black Sea and in particular on the northwestern shelf (Karkinitsky trough) and in the eastern part of the sea, seismic profiles revealed several deeper reflectors. Reflector IIIa corresponds to the top of the Cretaceous limestones (locally to Paleocene limestones); reflectors IIIb or IV correspond to the boundary between the carbonate sequence of Upper Cretaceous and Lower Cretaceous terrigenous deposits. Locally, in the near Caucasian portion of the Sea, some reflectors correspond to the top of the Jurassic (Ismailov and Terekhov, 1983).

These older reflectors disappear in the inner portions of the sea as they often merge with a single rather strong reflector identified as "H". Locally, reflector "H" coincides with the surface of the acoustic basement. It is traced down to 12–14 s but cannot be recorded deeper. This reflector has not been identified precisely. In the near Caucasian region of the Black Sea, it seems to merge with reflector III corresponding to the top of the Cretaceous. In the Lower Kamchian trough (in Bulgaria), which opens into the Black Sea, Jurassic–Cretaceous sediments widely occur; therefore, we cannot exclude the possibility that part of the Mesozoic and in particular the Upper Cretaceous extend from the continental slope into the deep-sea basin above reflector "H". Thus the deep sedimentary sequence in the Black Sea begins at the latest in the Paleogene, but most probably at the end of the Cretaceous as proposed by Letouzey et al. (1977) and Tugolessov et al. (1983a, b).

Multi-channel seismic reflection profiling detected two deep basins in the central part of the sea, namely the western and eastern Black Sea basins. The basins are divided by the central Black Sea high extending from Crimea to the Ordu region in Turkey. The largest basement depth in these basins is at least 16 km, in the western basin and 13 km in the eastern basin. The top of the central high is at a depth of 5 km (Fig. 3). The basement relief still influences the relief of reflector IIb at the base of the Eocene (Fig. 5) but it is not manifest in the sea-floor topography. It is only near Turkey that the Central High appears in the sea-floor topography as the Arkhangelsky Swell with a bottom depth of 500 m. The buried northwestern continuation of the Arkhangelsky Swell (towards Crimea) is called the Andrussov Swell (Tugolessov et al., 1983a). The basement relief is completely covered by sediments, reflector Ia corresponding to the top of the Maikopian suite. The western basin is 550 km long in an E–W direction and up to 250 km wide; the eastern basin is 500 km long and 100 km wide. The Andrussov Swell is about 80 km wide. Refector "H" is only identified with reasonable confidence in the shallow eastern basin. In the western basin, it is only well recorded below the slopes down to 15 km and cannot be detected deeper within the center of the basin. Neogene–Quaternary

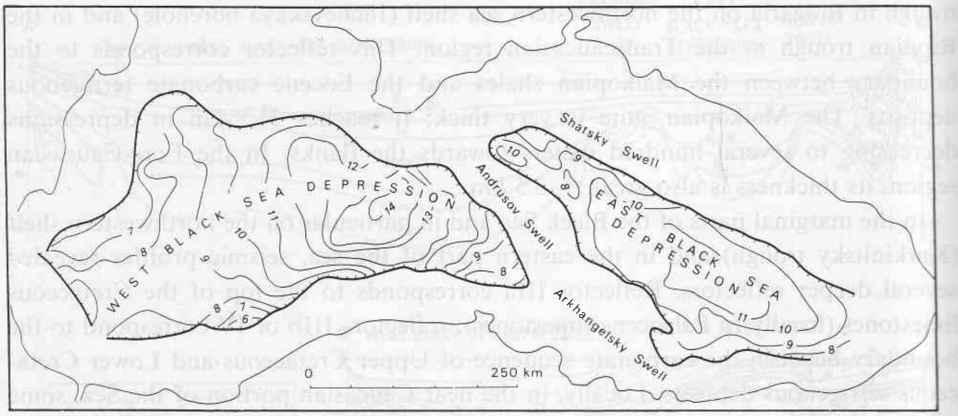

Fig. 5. Isopleth map showing depth (in km) to the IIb reflector near the base of the Eocene in the Black Sea. After E.M. Khakhalev a.o. Toothed line shows boundaries of the Black Sea deep basins.

sediments are 4–6 km thick; the Maikopian suite is 3 to 3.5 km thick in the eastern basin and 6 km thick in the western basin. Eocene and, probably, Paleocene–Upper Cretaceous deposits are 4 to 5 km thick.

In the basins, reflectors lie almost horizontally except for reflector "H" which may dip at an angle of 10–20°. The seismic stratigraphy gives convincing evidence for a gradual subsidence of the basins during the Cenozoic. Some profiles crossing the Andrussov Swell show that reflector "H" has a block faulted structure, the tilted blocks being down faulted towards the basins. Thus, subsidence was most probably caused by extension and thinning of the crust.

Deep seismic sounding (Neprochnov, 1966; Neprochnov et al., 1974) revealed the absence of a low velocity crustal ("granitic") layer under the deep basins. The total crustal thickness down to the 8.1–8.2 km s^{-1} horizon is about 20 km, 14–15 km of them being taken up by sediments. Therefore, the igneous crust is 5 to 6 km thick, which corresponds to the thickness of an oceanic crust. The thickness of the crust increases towards the margin to 40 km (Fig. 6). Refraction data (Neprochnov and Neprochnova, 1980) indicate that down to 1–2 km under the sea-floor, the P-wave velocity in sediments is 1.6–1.8 km s^{-1}; below, in the 3 to 5 km thick Neogene and Maikopian suite, it is 3 km s^{-1} and 4 km s^{-1} in deeper layers.

The areas surrounding the two deep basins are characterized by generalized subsidence during the Neogene and occurrence of more ancient Mesozoic rocks than in the deep basins. They include the shelf areas which may cover deep troughs, as the Burgass and Lower Kamchian troughs to the west (Fig. 3) and two deep water areas which correspond to structural highs. These are the Central Black Sea high, with the Andrussov Swell to the north and the Arkhangelsky Swell, to the south, and the East Black Sea high with two NW–SE structural units, the Shatsky Swell and the Tuapse

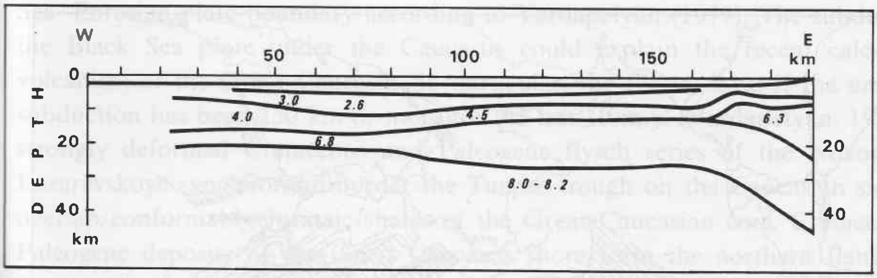

Fig. 6. Deep seismic sounding cross-section through the eastern part of the Black Sea (after Neprochnov et al., 1974). Figures correspond to V_p in m s^{-1}.

trough along the Caucasus (see Fig. 3). These structural highs are now buried below sediment although they had a clear morphological identity in the Maikopian.

The Central Black Sea high structure is well outlined by reflector "H" (see Fig. 4). Two and locally three additional reflectors are identified below it. The overlapping basins sediments are unconformable (Tugolessov et al., 1983a). These horizons probably lie within the Cretaceous and Jurassic. Cretaceous sediments were drilled by Shimkus (1977) from the Arkhangelsky swell foothills. The Arkhangelsky swell is probably non-volcanic as it is characterized by small negative anomalies (Ross et al., 1974; Neprochnov et al., 1974, see Fig. 7).

The Shatsky swell includes the Gudanta and Ochamchiri domes (Fig. 3) which are in the prolongation of the Dzirula massif structures in Georgia, composed of Cretaceous and Jurassic sequences. Figure 7 shows that there is a large (300 T) positive magnetic anomaly which could be explained by the presence of volcanic rocks. This would suggest that Middle Jurassic, mainly Bajocian volcanics of the Dzirula massif and the Gagro-Chjawa zone of the southern slope of the Great Caucasus extend over the Shatsky swell. We consider the Shatsky swell as a link between the domains of Jurassic volcanics of the Caucasus and Crimea highland.

The Tuapse and Sorotkhin troughs extending between the Shatsky swell and the mountains of the Great Caucasus and Crimea highland are narrow (25 to 30 km) with up to 10 km deep depressions (Terekhov, 1979; Pustilnikov et al., 1980; Tugolessov et al., 1983a). They are filled by Mesozoic and Cenozoic sediments. The Sochi–Adler depression opens into the Tuapse trough and exposes Miocene sediments at the surface. The Sorotkhin trough is in the prolongation of the Kerch–Tamanian trough and of the Mesozoic–Cenozoic Indolo-Kubanian trough of the Fore-Caucasian region which separates the Great Caucasus from Crimea. The continental margins of the first two troughs are zones of intensive deformation (Fig. 8) and have been interpreted as a young accretionary prism associated with subduction of the Black Sea floor under Crimea and the Caucasus (Ushakov et al., 1977). A negative −80 mGal free-air anomaly is compatible with this interpretation. The present seismicity is confined to the slopes of the troughs which lie along the Black

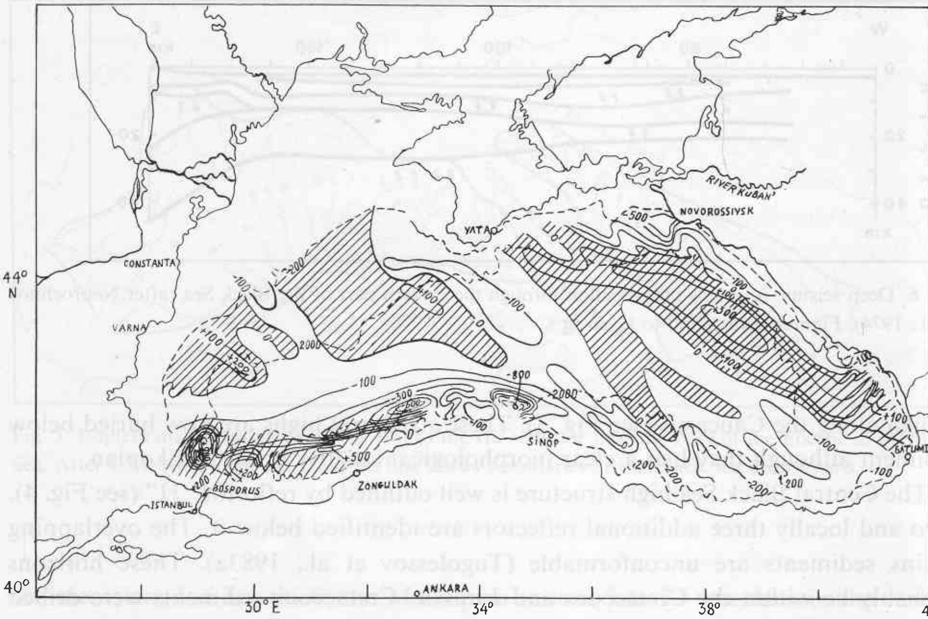

Fig. 7. Chart of magnetic anomalies over the Black Sea (after Ross et al., 1974). Contour interval is 100 nT. Positive anomaly areas are striped.

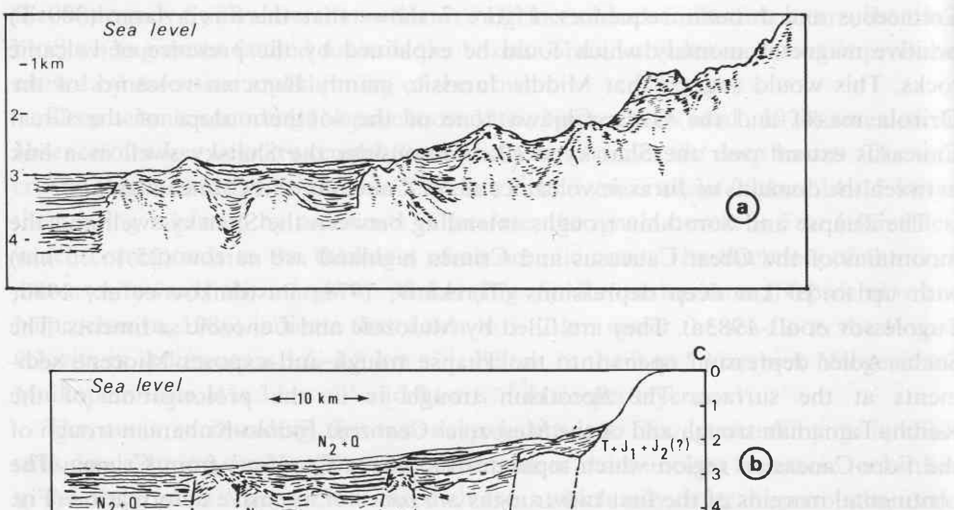

Fig. 8. Seismic reflection sections through the Great Caucasus–Crimea margin. A. Cross-section near Gelendjik (after Pustilnikov et al., 1980). B. Cross-section near Yalta (after Terekhov, 1979).

Sea–Eurasian plate boundary according to Vardapetyan (1979). The subduction of the Black Sea plate under the Caucasus could explain the recent calc-alkaline volcanism of the Great Caucasus, in particular, the Elburz lava, if the amount of subduction has been 150 km or more for the last 10 m.y. (Vardapetyan, 1979). The strongly deformed Cretaceous and Paleogene flysch series of the Novorossiisk–Lazarevskoye synclinorium border the Tuapse trough on the Caucasian side; they overlap conformably Jurassic shales of the Great Caucasian core. Cretaceous and Paleogene deposits of the Great Caucasus shore form the northern flank of the Tuapse trough. Thus, we suppose that the Tuapse trough forms the prolongation of the Great Caucasian structures into the sea.

TENTATIVE GEOLOGICAL HISTORY OF BLACK SEA BASINS

The data discussed above lead us to distinguish three episodes in the formation of the Black Sea basins. The first episode embraced the first half of the Jurassic and coincided with the opening of a basin at the location of the present Great Caucasus. In the Black Sea, sediments of this basin are only preserved in the Tuapse trough. The Ochamchiri borehole drilled through the sequence of this basin cover from Jurassic to Neogene. Most of this old basin floor was probably subducted under the Eurasian margin, predominantly in post-Maikopian time.

A second distensive episode occurred at the end of the Jurassic and beginning of the Cretaceous. It is marked by the appearance of alkaline basalts associated with distension in the Rionian lowland, by unconformities at the base of the Cretaceous and Upper Jurassic in Bulgaria, by the unconformity at the bottom of the sedimentary cover and by the extensional structures in the Karkinitsky trough, and finally by the unconformity at the base of the Upper Jurassic in the Crimea Highland. The Karkinitsky and Lower Kamchian troughs, the Andrussov swell and most of the Shatsky swell, which probably have a continental crust, were affected by this episode of rifting and subsidence.

The third episode of distension occurred at the end of Cretaceous or the beginning of Paleogene and was the time of formation of two small oceanic basins. This episode is marked by the appearance of rift structures in the Adjaro–Trialetian zone of the Lesser Caucasus (Adamia et al., 1974; Lordkipanidze, 1980) and in the Burgass synclinorium and Varna depression in Bulgaria (Nachev and Yanev, 1980; Gochev, 1980).

Partial closure of the Black Sea basins may have begun 30 to 35 m.y. ago in Maikopian time as the Maikopian shales accumulated. The Andrussov and Shatsky swells which were above sea-level prior to that time, were immerged at that time and became buried below sediment at the end of the Maikopian, 15 m.y. ago. Rising mountain chains had cut the links with open ocean and the basins had become euxinic.

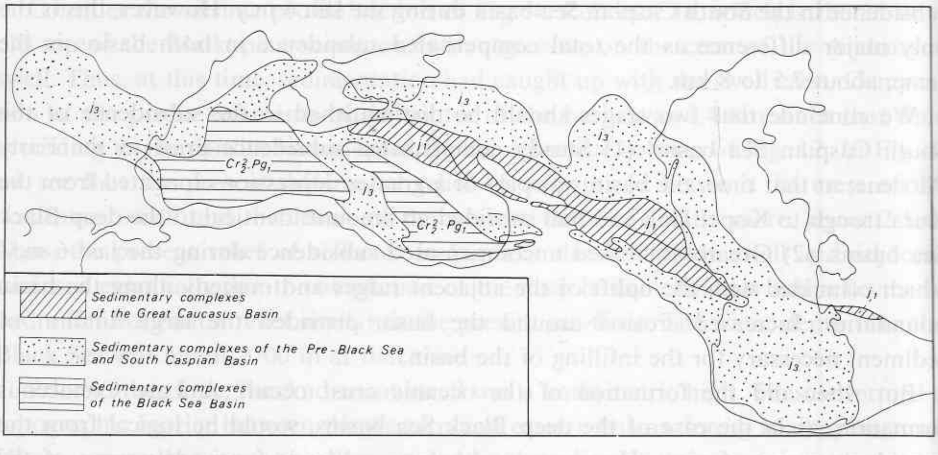

Fig. 13. Proposed distribution of sedimentary sequences belonging to the basins of different ages.

Formation of this basin as the result of distension of some more ancient basement is marked by the presence of the Great Caucasus Gojkht basalts which belong to the low-potassium tholeiitic series. Abundance of diabasic sills and dykes within the Liassic supports this interpretation. The overlying Great Caucasus Jurassic and Cretaceous sequence is composed of thick mostly terrigenous deposits which we interpret as marginal sea deposits.

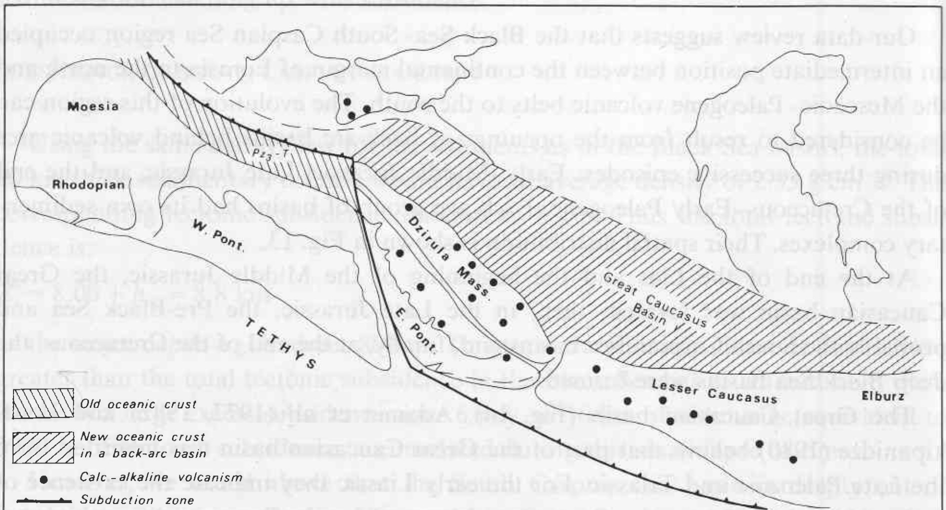

Fig. 14. Reconstruction for the Bajocian (180 m.y.). The Great Caucasus basin exists behind the East Pontides–Lesser Caucasus volcanic arc. A part of the basin was newly created in the Toarcian–Aalenian due to back-arc spreading.

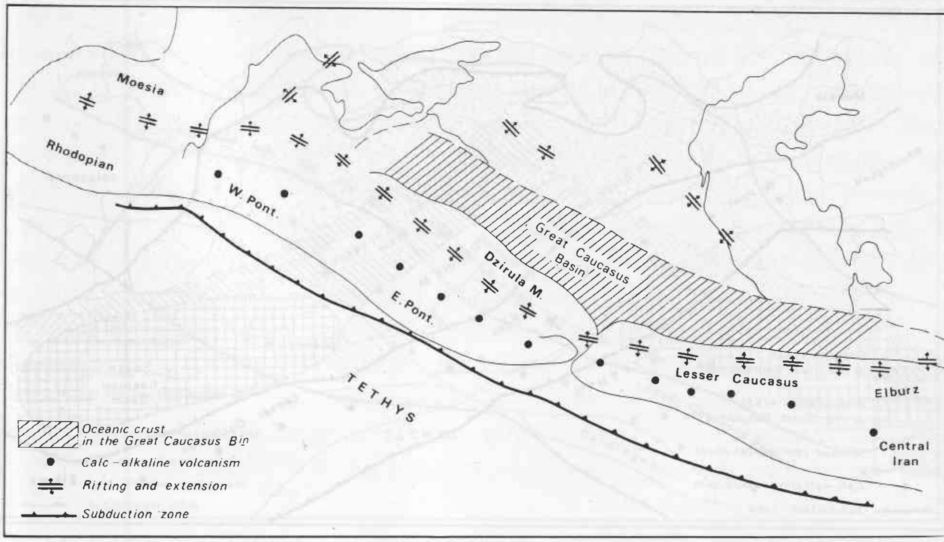

Fig. 15. Reconstruction for the Late Jurassic (145 m.y.). A new back-arc basin is opening behind the Pontides–Lesser Caucasus volcanic arc.

A volcanic arc is situated to the south of the Great Caucasian marginal basin during the Bajocian (Fig. 14). The presence of the arc is revealed by a Bajocian volcanic sequence in the Lesser Caucasus, East Pont, Dzirula massif, Gagro-Chjawa zone and in the Crimea Highland. As previously mentioned the Caucasus volcanism was connected with that of Crimea by a probable volcanism on the Shatsky Swell in the Black Sea. The Tuapse trough, to the west, was also part of the Great Caucasian basin. To the east, this basin can be traced north of the Great Balkan on the eastern coast of the Caspian Sea. The approximate size of the Great Caucasian basin was 1700 to 1800 km in length and 300 km in width. Thus, the central part of the basin may have had an oceanic basement.

Pre-Black Sea and South Caspian basins (Figs. 15 and 16). The Great Caucasian basin seemed to have been slightly reduced in size between the Middle Jurassic and the beginning of the Late Jurassic as the Cimmerian orogeny is manifest along the entire northern flank of the basin. In the westernmost part of the basin, which was adjacent to Crimea, it was caused by the collision of the Moesian plate with Crimea, whereas deformation in Caucasus was caused by underthrusting of part of the Great Caucasian basin under the Eurasian margin (see Fig. 14). As volcanism has not been observed on this margin, the underthrusting probably did not exceed 150 km.

Almost immediately after the Cimmerian orogeny, a new period of extension started to the north of the Lesser Caucasian island arc (Fig. 15). It was characterized by wide occurrence of alkaline basalts in the Rionian lowland depression, by the end of volcanic activity in the Bajocian arc of the Dzirula massif and the Gagro-Chjawa

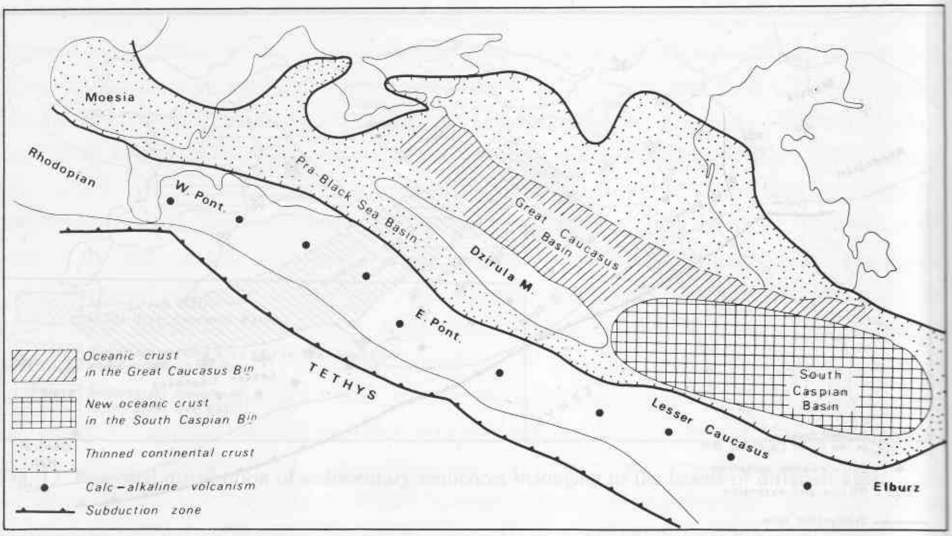

Fig. 16. Reconstruction for the Neocomian (130 m.y.). The Pre-Black Sea Basin to the west and the South Caspian Sea basin to the east were formed at this time. Oceanic crust probably only appeared within the South Caspian Sea basin. Most of the Pre-Black Sea basin has a highly thinned continental basement.

zone, by the occurrence of evaporites and formation of tilted blocks in the basement of the Karkinitsky trough, by subsidence of the Cimmerian basement of Crimea and by the beginning of subsidence of the Lower Kamchian trough, and by the beginning of deposition of the Kopet Dag sedimentary cover. As a result, a northern band was split off the Lesser Caucasian arc and later became a remnant arc. Between the southern active arc of Lesser Caucasus and Pontides, on one side, and the northern remnant arc, on the other, inter-arc basins were formed (Fig. 16). To the west, within the Pre-Black Sea basin, the extension did not lead to complete break-up of the basement, and no new oceanic floor seemed to have formed. The amount of extension is probably of the order of 120 to 150 km within the 300–350 km wide band. To the east, the extension is probably considerably greater as we assume that it resulted in the formation of the South Caspian oceanic basin. We propose that it included the present deep South Caspian Sea basin as well as a westward continuation which was later consumed during the convergence of the Lesser and Greater Caucasus. The oceanic basin was about 300 km wide, based on the size of the present South Caspian Sea basin. To the east, in the Kopet Dag, the oceanic basin wedged out. The total length of the Neocomian back-arc basins was at least 3000 km and the width was about 600 km, including the regions of thinned continental crust.

The Black Sea basins (Figs. 17 and 18). An important event in the history of this segment of the Mediterranean belt occurred in the early Senonian when the southern Gondwanian continental blocks first collided with the Lesser Caucasian island arc

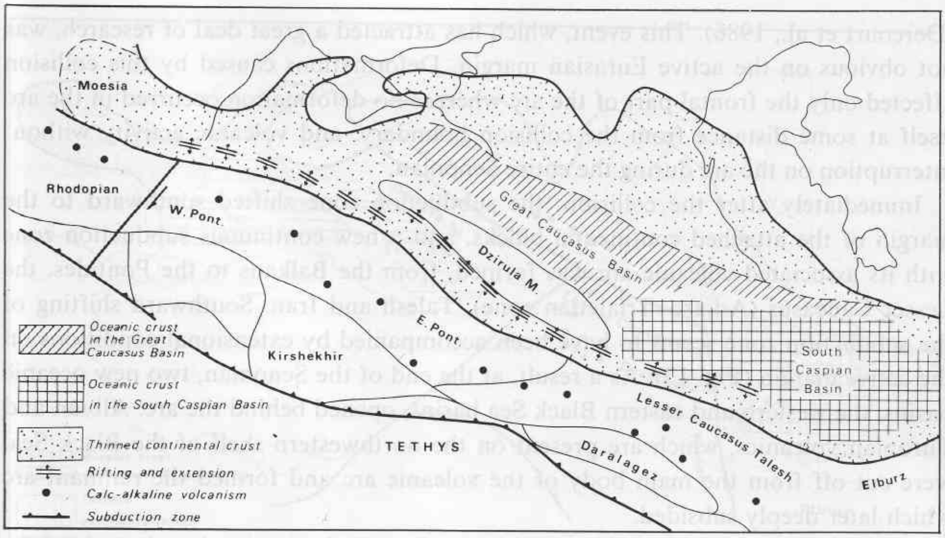

Fig. 17. Reconstruction for the Late Cretaceous (80 m.y.). The subduction zone flipped southward after collision of the Kirshehir and Daralagez blocks with the former subduction zone, accompanied by ophiolitic obduction. As a result, extensional conditions became possible behind the volcanic arc. The Adjaro–Trialetian and Talesh rift zones were manifestations of that extension.

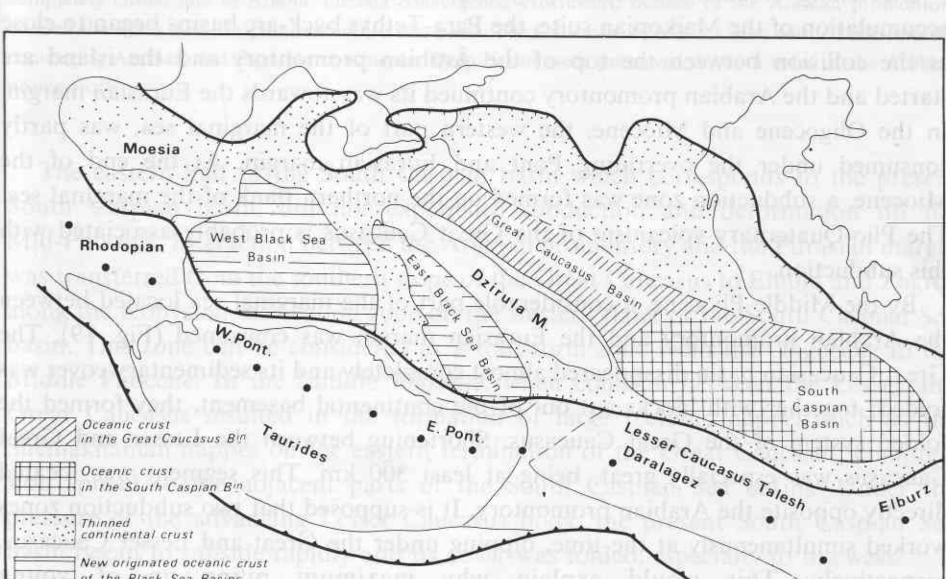

Fig. 18. Reconstruction for the Eocene (50 m.y.). The western and eastern Black Sea oceanic basins had been formed by that time. The Black Sea–Caspian oceanic realm had its maximal extent. It began its shortening during the Oligocene.

(Dercourt et al., 1986). This event, which has attracted a great deal of research, was not obvious on the active Eurasian margin. Deformations caused by this collision affected only the frontal part of the arc whereas no deformation occurred in the arc itself at some distance from the collision boundary, and volcanic activity without interruption on the arc during the entire Senonian.

Immediately after the collision, the subduction zone shifted southward to the margin of the attached continental blocks, and a new continuous subduction zone with its associated volcanic arc was formed, from the Balkans to the Pontides, the Lesser Caucasus (Adjaro–Trialetian zone), Talesh and Iran. Southward shifting of the subduction zone seems to have been accompanied by extensional conditions on the active margin (Fig. 17). As a result, at the end of the Senonian, two new oceanic basins, the western and eastern Black Sea basins, opened behind the arc. Albian and Turonian volcanics, which are present on the northwestern shelf of the Black Sea, were cut off from the main body of the volcanic arc and formed the remnant arc which later deeply subsided.

During the Eocene, a large marginal basin (Para-Tethys) existed to the rear of the volcanic arc (Fig. 18). It was 3000 km long and about 900 km wide. Four deep basins with oceanic crust were present. They are: the Great Caucasian, South Caspian Sea, western and eastern Black Sea basins.

Shortening of back-arc basins (Fig. 19). After the Oligocene, that is after the accumulation of the Maikopian suite, the Para-Tethys back-arc basins began to close as the collision between the top of the Arabian promontory and the island arc started and the Arabian promontory continued its way towards the Eurasian margin. In the Oligocene and Miocene, the western part of the marginal sea, was partly consumed under the overriding Pont and Eurasian margin. At the end of the Miocene, a subduction zone was formed on the northern flank of the marginal sea. The Plio-Quaternary volcanism of the Great Caucasus is probably associated with this subduction.

By the Middle Pliocene, a considerable part of the marginal sea located between the Arabian promontory and the Eurasian margin was consumed (Fig. 19). The Great Caucasian basin disappeared almost completely and its sedimentary cover was folded; together with blocks cut out of the continental basement, they formed the folded system of the Great Caucasus. Shortening between the Lesser and Great Caucasus was especially great, being at least 300 km. This segment was located directly opposite the Arabian promontory. It is supposed that two subduction zones worked simultaneously at the time, dipping under the Great and Lesser Caucasus, respectively. This would explain why maximum piling up of young Pliocene–Quaternary calc-alkaline lava of the Armenian Highland is confined to this segment. The western half of the South Caspian basin was consumed on the tip of the Arabian promontory. An anomalously dense slab under the Kura lowland can be considered to be a remnant, now subsided, oceanic basement of the western South Caspian basin.

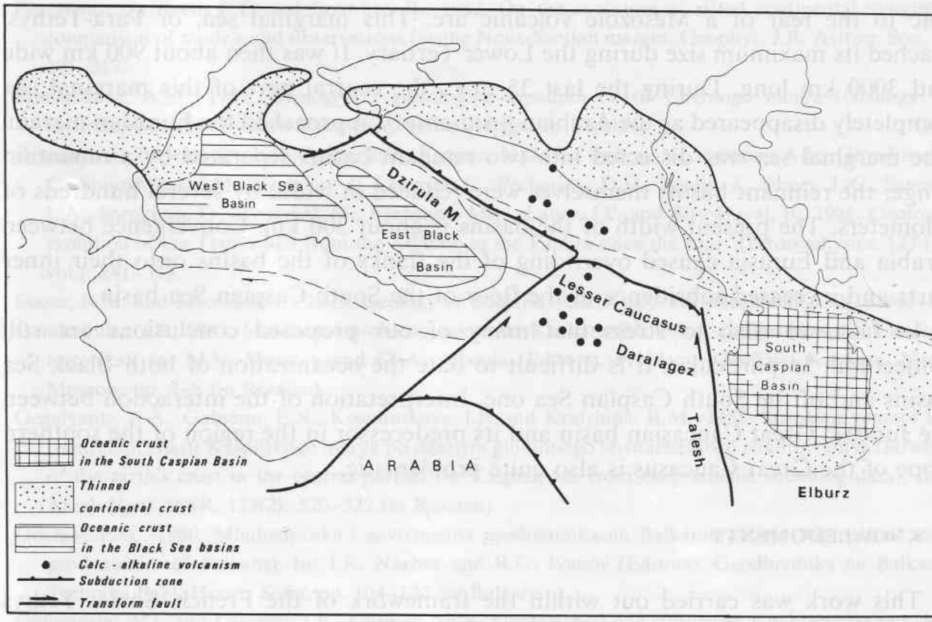

Fig. 19. Reconstruction for the Middle Pliocene (3.5 m.y.). The Great Caucasus basin was almost completely closed due to Arabia–Eurasia convergence. Northward motion of the Arabian promontory divided the Black Sea–Caspian realm into two parts. The Caucasus mountains were formed between the converging Arabian and Eurasian continents. Deep basins were preserved on both sides of the Arabian promontory.

The eastern half of the South Caspian basin which corresponds to the present South Caspian basin, did not experience subduction and deformation till the Mid-Pliocene. Interaction between the Arabian promontory and the Eurasian margin was transferred from the southern slope of the Great Caucasus to Elburz and Zagros along the Kobystan dislocation zone on the western flank of the South Caspian Sea basin. This zone can be considered as a transform zone from the Oligocene to the Middle Pliocene. In the Middle Pliocene, when collision between the Lesser and Great Caucasus resulted in the formation of large tectonic nappes such as the Shemakhanian nappes on the eastern termination of the Great Caucasus, compression spread to the adjacent parts of the South Caspian Sea basins. Under the pressure of the advancing Lesser Caucasus body, the present South Caspian Sea basin began to subside rapidly and its cover was folded, especially to the west.

CONCLUSION

The present basins of the Black Sea and South Caspian Sea are remnants of a much greater marginal sea formed during three separate episodes during the Mesozoic to the rear of a Mesozoic volcanic arc. This marginal sea, or Para-Tethys, reached its maximum size during the Lower Tertiary. It was then about 900 km wide and 3000 km long. During the last 35 m.y., the central part of this marginal sea completely disappeared as the Arabian promontory approached the Eurasian margin. The marginal sea was dissected into two remnant basins separated by a mountain range; the remnant basins themselves were reduced in width by several hundreds of kilometers. The present width of the basins is about 300 km. Convergence between Arabia and Eurasia caused overriding of the flanks of the basins onto their inner parts and increased subsidence of the floor of the South Caspian Sea basin.

In fact, we wish to stress that many of our proposed conclusions are still conjectural. In particular, it is difficult to date the oceanization of both Black Sea basins and of the South Caspian Sea one. Interpretation of the interaction between the Jurassic Great Caucasian basin and its predecessor in the region of the southern slope of the Great Caucasus is also quite problematic.

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Special Issue

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