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Meso-Tethyan oceanic sutures and their deformation

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Abstract

On the basis of comparative stratigraphic and paleontologic analysis, supported by some key paleomagnetic data and interpretations, it is shown that during the Mesozoic and the Cenozoic the Eurasian continent grew by accretion of microcontinents. These microcontinents separated basins with oceanic crust from the main ocean. During the Late Cretaceous and the early Cenozoic the collision of the microcontinents with Eurasia resulted in the closure of the basins, and Meso-Tethyan oceanic sutures originated. In the region under consideration, from the Carpathians to Tibet, there are two main Meso-Tethyan sutures: the Carpathian–Lesser Caucasus and the Afghan–Tibet suture. The above-mentioned main structures also had branches, which remained as sutures of small basins: the Kamennopotock, Interpontide, Nain-Baft basins and others.

In the West Carpathians the Carpathian–Lesser Caucasus suture is overlain by a widespread Gemeric–Tatric allochton. From the West Carpathians the suture passes through the Pannonian basin into the Vardar ophiolite zone and farther to the ophiolites of the Izmir–Ankara zone. Being displaced along the North Anatolian right-lateral strike-slip fault, the main suture passes from the Eastern Pontides into the Lesser Caucasus, where it is marked by ophiolites of the Amasia area, the Shirak, Bozum and Zangezur ridges. Sunsequently, it can be observed through the Iranian Qara Dagh mountains to Lake Urumiyeh and the North Anatolian strike-slip fault. Being again shifted along the strike-slip fault, the Carpathian–Lesser Caucasus Meso-Tethyan suture ends in the Western Zagros near the Cenozoic Neo-Tethyan suture.

The Afghan–Tibet Meso-Tethyan suture is situated in the Pamirs in the Rushan–Pshart zone. The east prolongation of the suture has been displaced along the Pamir–Karakorum right-lateral strike-slip fault in Tibet. West of the Pamirs this suture is also displaced along strike-slip faults and continues in the Farahrud zone in Afghanistan, and than passes through the Zabol–Baluch and Daz Murian ophiolite zones and approaches the Neo-Tethyan suture.

The paleomagnetic data allow us to reconstruct the location of those sutures in the Late Cretaceous–early Cenozoic. If one compares the present position of the Carpathian–Lesser Caucasus and Afghan–Tibet Meso-Tethyan sutures with the Late Cretaceous–early Cenozoic reconstructions of these sutures it is possible to trace the inner deformation of the Alpine belt. According to these data, the Carpathian–Lesser Caucasus suture moved northwards over 1200 km ahead of the Arabian–Turkish syntaxis with the displacement direction across the fold belt. At the border of the Dinarides and Hellenides, the displacement decreases to 400–500 km, and its direction is along the fold belt. The amplitude of the suture displacement increases in the Pannonian region and again decreases towards the Eastern Alps. The displacement of the Afghan–Tibet suture was > 2000 km during the Alpine deformation.

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A great ensemble of Cenozoic nappes, folds and structural arcs of the Pamir–Punjab and Arabian–Turkish syntaxes and Carpathian loop is the result of lateral shortening of the Alpine belt through the collision.

1. Introduction

In the Mesozoic and Cenozoic the Eurasian continent was growing due to accretion of microcontinents in the Tethyan ocean; following their amalgamation with Eurasia the Gondwanaland continents were also accreted to Eurasia. During their approach to Eurasia the microcontinents separated marginal basins with oceanic crust from the Tethys (Dercourt et al., 1986). The collision of microcontinents with Eurasia led to closure of the basins and to formation of ophiolitic sutures. In the Jurassic (usually before the Late Jurassic) the Paleo-Tethyan ocean closed (Sengör, 1984). The other important epoch of collision began in the Cretaceous, when other microcontinents were amalgated with Eurasia. The oceanic sutures that formed in the Late Cretaceous and the early Cenozoic are called the sutures of the Meso-Tethyan ocean (Shvolman, 1978). Using this term one should keep in mind that closure of those branches of the Tethys and the existence of the Meso-Tethyan sutures does not mean the disappearance of the Neo-Tethyan ocean which separated the continents of Laurasia and Gondwana Land till late Cenozoic times.

One can establish either this or that territory belongs to the Eurasian or Afro-Arabian domain of Gondwana Land by analyzing Liassic facies and biogeographic communities, for at that time the paleogeographic conditions in the Tethys and on its margins were different from one another. Having studied the Liassic ammonite fauna, Neumayer (1878) distinguished mid-European and Mediterranean faunistic provinces in Europe. The mid-European species of Early Jurassic ammonites are distributed in extra-Alpine Europe. In the Carpathian region, two types of facies of the Liassic deposits contain ammonite fauna. The deposits in a terrigenous carbonate facies belong to the first type. They contain approximately the same quantity of Mediterranean and mid-European ammonite species. The terrigenous material came into these deposits from the northern margin of the Tethys. Deposits of this type formed on the shelf of the Eurasian continent.

The condensed sections of pelagic carbonate deposits in the "ammonitico rosso" facies and its analogues belong to the second type. Mediterranean species of ammonites greatly predominate in these deposits that formed on submarine platforms of the passive margin of the Afro-Arabian sector of Gondwana Land. The deep-water zone of the Tethys hindered the penetration of the European fauna into the Afro-Arabian margin. This "filter" functioned in the Pliensbachian and Toarcian (Geszy, 1984; Rakus, 1988). Nowadays, geologists more often speak of the North Tethys and South Tethys biogeographic provinces, separated by a deep-water "barrier".

The correlation between the biogeographic provinces is well demonstrated by the distribution of Liassic and Bajocian brachiopod communities (Vörös, 1977, 1988; Horvath et al., 1979; Prosorovskaya and Vörös, 1988). Brachiopod locations, where the quantity of species characteristic of one province, is twice or more times the quantity of the species belonging to the other biogeographic province, are shown in Fig. 1. The difference between these biogeographic provinces is also established on the basis of the fauna of the Lias and Dogger benthic foraminifera. Distribution areas of the Pliensbachian genus *Orbitopsella* and of the Bathonian species *Satorina apulensis* are characteristic of the South Tethys province and *Orbitammina elliptica* of the North Tethys province (Baussollet et al., 1985; Sengör et al., 1988).

Let us consider the present location of the Meso-Tethyan sutures in the Alpine belt between the Alps and Tibet.

2. The Afro-Arabian margin of the Meso-Tethys

One can distinguish inner and outer megazones on the African–Arabian margin. Neritic facies (h, Fig. 2) predominate among the Jurassic rocks in the inner megazone. This megazone includes the following tectonic zones: the Dalmatian, Kruja and Karst zones in the Dinarides, the Gavrovo, Tripolitza, Ionian (in Early and Middle Jurassic) and Parnassos zones in the Hellenides, the Ida zone on the island of Crete and the Archangelos zone on the island of Rhodes. The rocks of the megazone formed the Adriatic microcontinent that belonged to the Afro-Arabian domain.

The Jurassic sections are condensed in the outer megazone, pelagic limestones and radiolarites predominating (i, Fig. 2). They formed on the continental slope and in the marginal seas. This

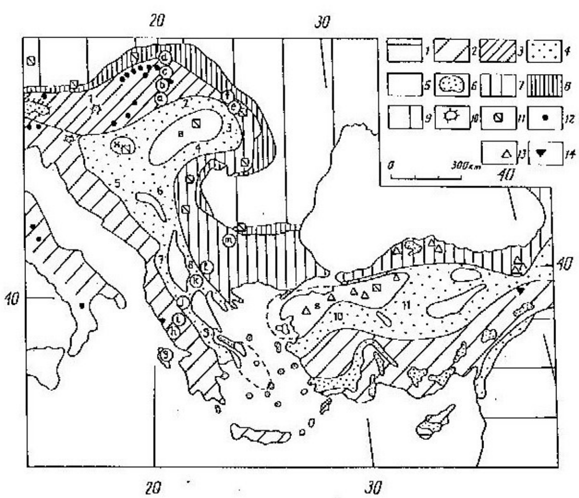

Fig. 1. East Europe and Asia Minor. *1* = Arabian continent; *2–3* = Afro-Arabian margin of the Meso-Tethys; (*3* = Gemeric–Tatric megazone); *4–5* = zone with ophiolites including ophiolite allochthons which are supposed to be connect with the root area: *5* = blocks with old pre-Alpine basement—Bichor (*B*), Mescek (*M*), Sacarya (*S*) and others; main part of them were microcontinents in the Tethys before collision; *6* = ophiolite allochthones, isolated from the root zones; *7–8* = Eurasian margin of the Meso-Tethys: *7* = Rhodope–Iranian megazone, *8* = flysch zones of the Carpathians and Balkans; *9* = Eurasian continent; *10* = ophiolites in tectonic windows; *11–12* = locations of Liassic brachiopods (Vörös, 1977, 1984; Horvath et al., 1979): *11* = North Tethyan communities, *12* = South Tethyan communities; *13–14* = locations of Liassic ammonites in Asia Minor (Bausollet et al., 1975): *13* = North Tethyan communities, *14* = South Tethyan communities.

megazone unites the Budva and Krasta–Cukali zones in the Dinarides, the Pindos zone in the Hellenides, the Ethis and Mangassa zones on the island of Crete, the Adra zone on the island of Carpathos and the Prophit–Illias zone on the island of Rodos. Nappes formed from pelagic Jurassic rocks are also known in the Taurides and in the Inner Zagros. Due to the widespread development of nappes, the rocks of outer and inner megazones do not always occupy a place commensurate with their name in the modern structure.

The Gemeric–Tatric megazone, covering the inner Western Carpathians, the Transdanubian Mountains and a part of the Eastern Carpathians, also belongs to the Afro-Arabian domain. This megazone continues to the west into the Eastern Alps. The Gemeric–Tatric megazone as a whole now occupies an allochthonous position. It overlies the rocks of the Meso-Tethys oceanic crust which crop out in tectonic windows (1, Fig. 1), located near Wechsel and Rechnitz in Austria and near Kosez in Hungary. I am here speaking about the Gemeric–Tatric megazone as a domain that became allochthonous *in toto* in the Cenozoic (Burtman, 1986, 1988). The megazone was also disintegrated into constituent nappes in the Cretaceous. Those nappes were described by Uhlig (1907), Andrusov (Andrusov et al., 1973) and many other geologists.

Pelagic limestones and radiolarites are widely distributed together with crinoidal limestones among the Jurassic deposits of the Gemeric–Tatric megazone (a–c, Fig. 3). Pelagic rocks developed in sections of Lias and Dogger in the Tatric zone, Dogger and Malm in the Fatric zone, Lias in the Veporic zone and Dogger in the Hronic zone.

The South Tethys community of Liassic brachiopods is located in Austroalpine nappes, in the Southern Alps, in the Apennines and in the Hellenides (Fig. 1). This community is characteristic of the inner zones of the Western Carpathians and Transdanubian Mountains. The South Tethys ammonites of the Lias are also distributed in the Transdanubian mountains (Geszy, 1973). In the outer zones of the Dinarides, South Tethys benthic foraminifera of Early and Middle Jurassic

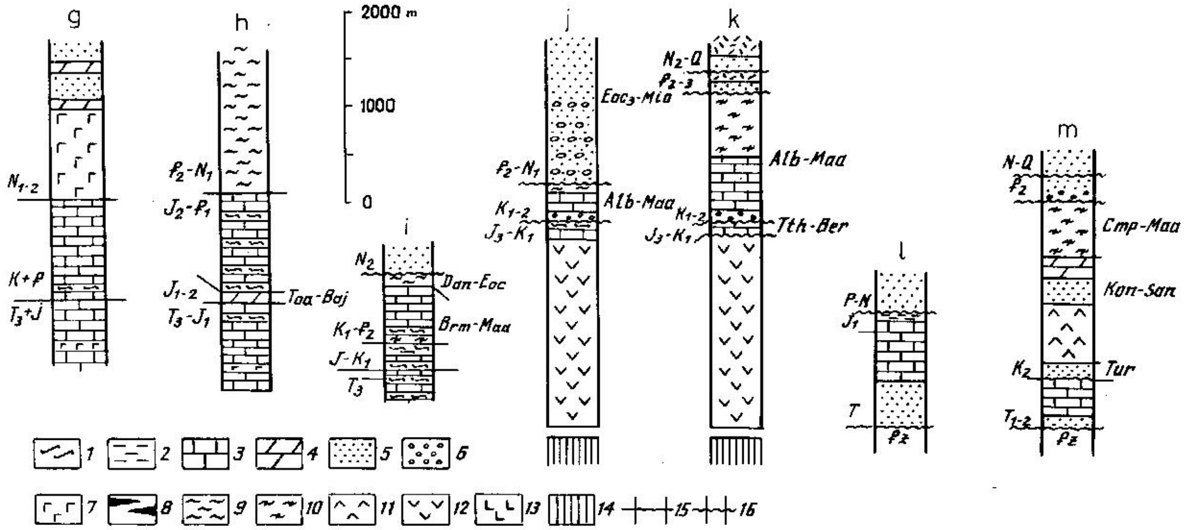

Fig. 2. Stratigraphic sections of tectonic units of the Hellenides and Balcanides: *g* = Paksos (Foreapulian); *h* = Ionian; *i* = Pindos; *j* = Othrys–Pelagonian; *k* = Vardar; *l* = Serbian–Macedonian; *m* = Srednegorie. The locations of the sections are shown in Fig. 1. Data sources: Hristov, 1960; Aubouin, 1965; Gocev et al., 1974; Aiello et al., 1977; Rassios et al., 1983. *1* = cherts; *2* = clays; *3* = marls; *4* = limestones; *5* = sandstones; *6* = conglomerates; *7* = gypsum; *8* = coals; *9* = terrigenous flysch; *10* = carbonate flysch; *11* = andesite and mixed volcanics; *12* = basic volcanics; *13* = gabbro; *14* = ultramafics; *15* = stratigraphic gap; *16* = angular unconformity.

age occur (Baussollet et al., 1985). In Asia Minor, in the Taurids, the South Tethys biogeographic province is established from the benthic foraminifera and ammonites (Baussollet et al., 1975; Sengör et al., 1988).

3. The Eurasian margin of the Meso-Tethys

The rocks of pre-Alpine continental crust, still covered by Mesozoic deposits, are well developed on the Eurasian margin. Jurassic deposits of the megazone usually unconformably overlie pre-Paleozoic, Paleozoic and Triassic rocks. The Early Jurassic deposits are represented by coastal and continental coal-bearing and terrigenous deposits. The marine Liassic beds contain a fauna of the North Tethys type. This marginal Rhodope–Iranian megazone includes the Marmaros nappes of the Eastern Carpathians, the Getic nappes of the Southern Carpathians, the Serbo-Macedonian and Rhodope massifs, the Srednegorie and Stara Planina zones of the Balkanides, the Sakarya continent and the Eastern Pontides in Asia Minor, and the Samheto–Kafan, Artvin–Bolnisi, Adjara–Trialetian, Georgian (Dzirula), Gagra–Djava and Talesh zones of the Caucasus, the Alborz and northwestern Iran.

In the Getic nappe there is a layer (500 m) of conglomerates, sandstones and shales with coaly interlayers. Ammonites in the shales and fossil plants in the coal date them as Early Jurassic (Nastaseanu et al., 1981). The Jurassic rocks of the Eastern Pontides have a similar structure where the Lias is represented by a rhythmic stratum of terrigenous rocks with fossil plants and coal horizons (the Kelkit formation). The fauna of the Liassic ammonites is of the European type both in the Western and Eastern Pontides, having been described from many stations (Fig. 1). Late Jurassic ammonites of the Pontides also belong to the North Tethys biogeographic province (Enay, 1974; Baussolet et al., 1975).

Liassic deposits are also present in the Dzirula crystalline massif. Their sequence is shown in locality n (Figs. 4, 5). In the lower part of the

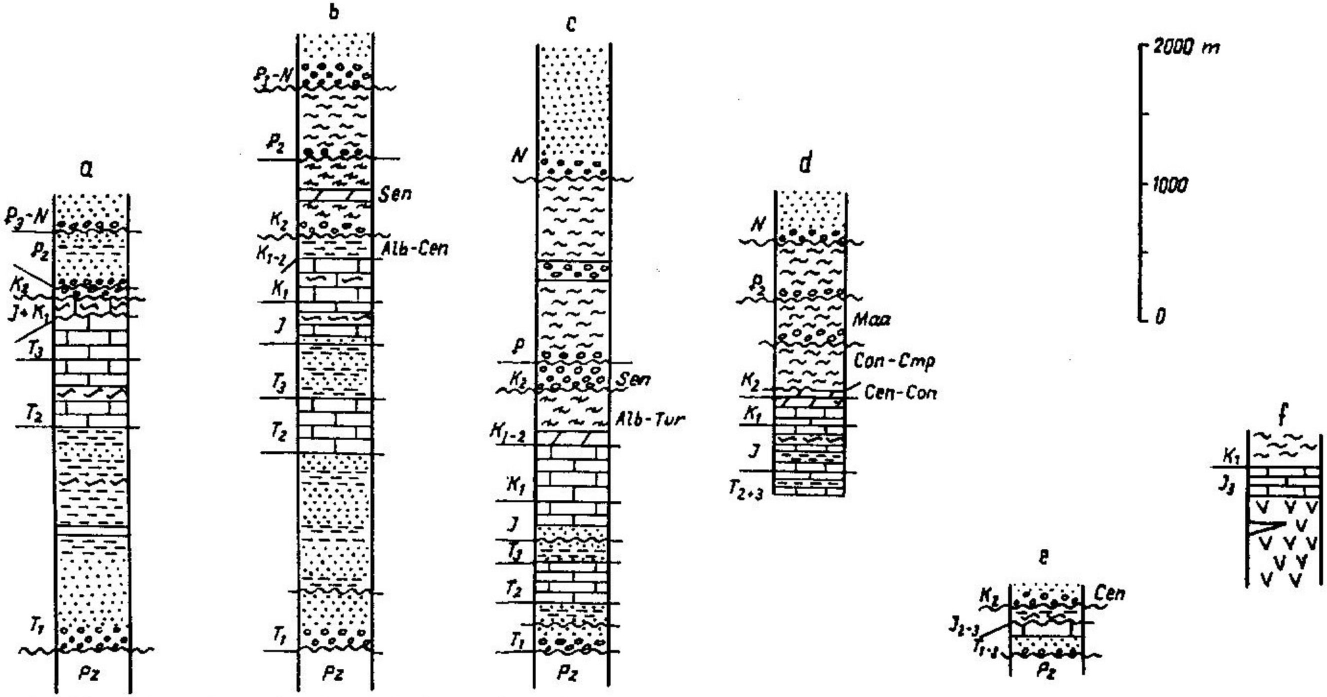

Fig. 3. Stratigraphic sections of tectonic units of the Carpathian region: *a* = Gemeric; *b* = Veporic; *c* = Tatric, *d* = Pieniny; *e* = Marmaros block (Delovets unit); *f* = Kamennopotock. Location of the sections is shown in Fig. 1. Data sources: Lomize, 1968, Bysova et al., 1971; Andrusov and Samuel, 1973; Birkenmajer, 1977. Symbols are given in Fig. 2.

section a layer of tuffites, sandstones and conglomerates occurs, with interlayers of coaly shales with a *Ginkgo* and *Pteropsida* flora. Quartzitic and arkosic sandstones with coal occurrences lie over the tuffites and directly over metamorphic rocks. The upper part of the section is formed from limestones with Middle and Late Liassic ammonites and brachiopods. The ammonite fauna is a mixture of Mediterranean and mid-European species, which is common for the North Tethys biogeographic province (Rostovtsev and Azarjan, 1971; Rostovtsev, 1978).

In the Artvin–Bolnisi zone the Liassic deposits lie on metamorphic rocks of the Loki and Khrami massifs. The lower part of the Loki section is formed from sandstones with coal lenses and plant remains and also ammonites of Sinemurean–Pliensbachian age, overlain by a stratum with a rhythmic intercalation of sandstones, siltstones and argillites with Pliensbachian–Aalenian ammonites (p, Fig. 5). Liassic sections (q, Fig. 5) have an analogous structure in the Somhet–Kafan tectonic zone (Gasanov, 1967; Panov, 1978).

In the Iranian Talesh, the Alborz and in northwestern Iran, Rhaetian, Lias and partly Dogger are represented by a thick terrigenous coal-bearing stratum known as the Shemshak formation. The formation is formed from large-

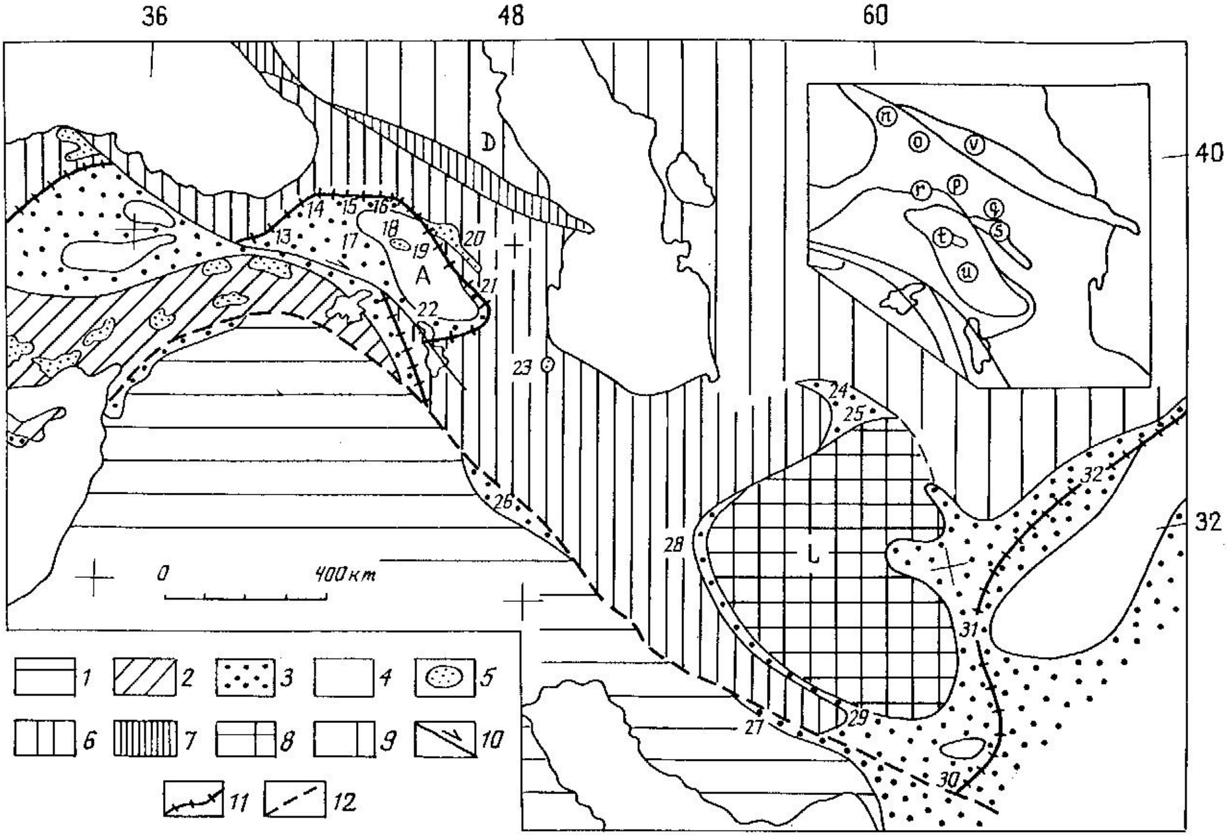

Fig. 4. The Caucasus and Iran. *1* = Arabian continent; *2* = Afro-Arabian margin of the Meso-Tethys; *3–4* = zone with ophiolites, it includes ophiolite allochthones which are supposed to be connect with the root area: *4* = blocks with pre-Alpine basement; *5* = ophiolite allochthones, isolated from the root zones; *6–7* = Eurasian margin of the Mesozoic Tethys: *6* = Rhodope–Iranian megazone, *7* = flysch zone of the Caucasus; *8* = Central–Iranian massif; *9* = Eurasian continent; *10* = North Anatolian fault; *11–12* = oceanic sutures: *11* = Meso-Tethyan, *12* = Neo-Tethyan. The inset shows location of stratigraphic sections in the Caucasus.

rhythmic, alluvial, lake-swamp and coastal-marine deposits alternating in the section. In the Iranian Talesh the section is made up of argillites, siltstones and sandstones with conglomerate and limestone interlayers. These deposits contain an early Liassic flora and ammonites and pelecypodas ranging from Sinemurian to Bajocian age (Davies et al., 1972). The flora of the Shemshak formation belongs to the Eurasian biogeographic province.

The Lias section in the Alborz is similar to the Talesh section (Stampfli, 1978). In the Central Alborz are coal-bearing horizons among these deposits, containing productive coal beds. The lower part of the formation contains an early Liassic flora, the middle section Toarcian–Aalenian ammonites, and the upper part a Middle Jurassic flora. The Shemshak formation is overlain by limestones with a late Bajocian marine fauna (Poliansky, 1980; Davoudzadeh and Schmidt, 1981). The environment in which the Shemshak formation formed was continental. The sediments were brought by rivers flowing from north to south. The South Alborz territory was the site of formation of alluvial-deltaic and lagunal-deltaic deposits.

West of the Alborz, the Shemshak formation extends as far as Lake Urumiyeh. South of the Alborz the deposits under consideration are distributed as far as the Main thrust of the Zagros (Poliansky, 1980; Davoudzadeh and Schmidt, 1981).

In northeastern Iran, in the Binalud mountains, clastic continental-coastal Rhaeto-Liassic deposits occur. In the Iranian Kopeh Dagh, these deposits (the Kashafrud formation) overlie with

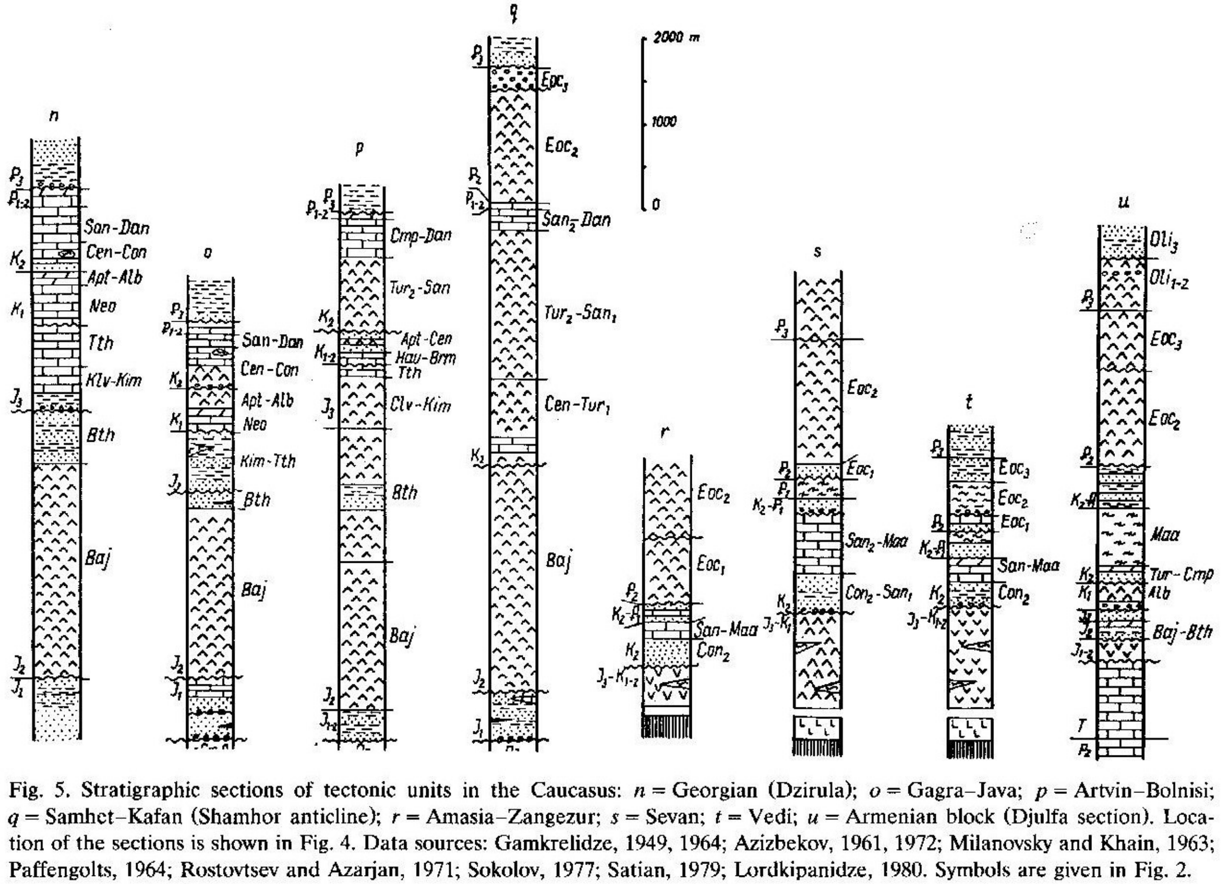

Fig. 5. Stratigraphic sections of tectonic units in the Caucasus: n = Georgian (Dzirula); o = Gagra–Java; p = Artvin–Bolnisi; q = Samhet–Kafan (Shamhor anticline); r = Amasia–Zangezur; s = Sevan; t = Vedi; u = Armenian block (Djulfa section). Location of the sections is shown in Fig. 4. Data sources: Gamkrelidze, 1949, 1964; Azizbekov, 1961, 1972; Milanovsky and Khain, 1963; Paffengolts, 1964; Rostovtsev and Azarjan, 1971; Sokolov, 1977; Satian, 1979; Lordkipanidze, 1980. Symbols are given in Fig. 2.

an angular unconformity deposits including a Late Triassic fauna (Stöcklin, 1974). In the eastern Kopeh Dagh (in the mountains of Mozduran) the thickness of the coal-bearing deposits of the Rhaetian, Lias and partly Dogger reaches 2000 m (Valbe, 1967). Farther east, the coal-bearing Lias is developed in the northern piedmont of the Hindu Kush, in the Afghan–Tadjik depression, and in the Central and Northern Pamirs (Androsov et al., 1977). North of these regions the coal-bearing Jurassic is distributed on the Turan platform and in the Tien-Shan.

In the Carpathians and in the Balkans, the Rodope–Iranian megazone was separated from the Eurasian continent by marginal sea basins during the Jurassic and Early Cretaceous. Flysch deposits were formed in these basins. Now they constitute the Rahov–Belotisza zone, the Severin nappe and the Fore-Balkan zone. The suture of the basin which was closed in the Middle Cretaceous lies in the Eastern and Southern Carpathians. In the Southern Carpathians, ophiolites belonging to the Severin nappe occur. The ophiolite complex consists of ultramafics, gabbro and Late Jurassic oceanic basalts. The Getic nappe of crystalline rocks lies on the Severin allochthon. Neocomian molasse contain pebbles of rocks from the Getic and Severin nappes (Burchfiel, 1976; Nastaseanu et al., 1981).

In the Eastern Carpathians, the Kamennopotock (Black-shale) nappe has a similar structural position (under the Marmoros crystalline nappes). This allochthon contains Late Jurassic basalts (f, Fig. 3), but gabbro and ultramafics are not seen. Nappes are covered by molasse of Cenomanian age (Byzova et al., 1983; Bazhenov and Burtman, 1990). Farther to the south this suture seems to extend into the Porec zone of eastern Serbia.

In the Caucasus, the deposits of the marginal sea are distributed in the flysch zone of the Greater Caucasus (Fig. 4). Jurassic flysch deposits formed in the Toarcian–Aalenian and Callovian–Kimmeridgian (Fig. 6). Volcanogenic rocks are widely distributed in Jurassic deposits. Lavas, tuffs and tuffites of a calc-alkaline composition are observed in the sections of Early and middle Liassic age. In the Pliensbachian–Toarcian and Bajocian, tholeiitic pillow lavas and hyaloclastics

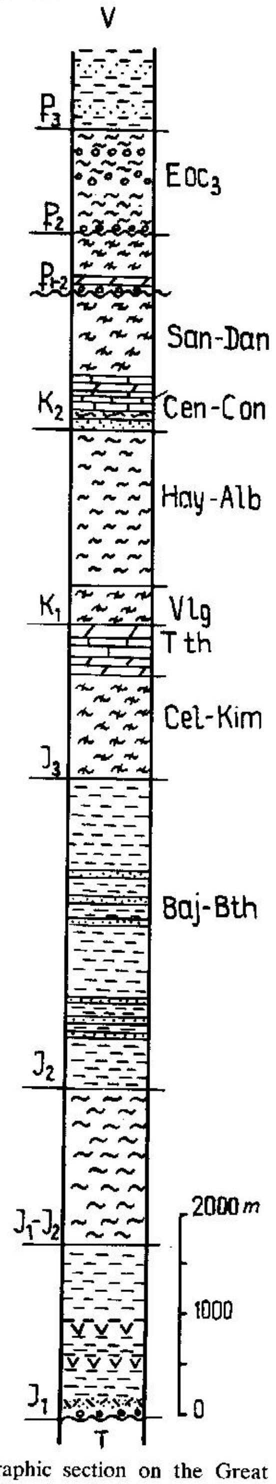

Fig. 6. Stratigraphic section on the Great Caucasus flysch zone (Mestia–Tianet unit). Data sources: Gamkrelidze, 1964; Gamkrelidze and Gamkrelidze, 1977. Symbols are given in Fig. 2.

formed in the Chatlin–Laily tectonic zone of the Greater Caucasus. The volcanics do not form continuous strata and everywhere they alternate with sedimentary deposits predominating in the section. The Liassic–Bajocian deposits are also penetrated by numerous sills and diabase dikes (Lordkipanidze, 1980). Liassic and Dogger ammonite fauna of the Greater Caucasus proves that this region belonged to the North Tethys biogeographic province (Rostovtsev and Azarjan, 1971; Rostovtsev, 1978).

East of the Caspian Sea no deep-sea basin separated the marginal Rhodope–Iranian megazone from the rest of Eurasia, because here neither ophiolites nor flysch along the northern margin of the Alpine zone are present, which would separate this zone from the Turan and Tien-Shan zones.

4. Ophiolite zones and Meso-Tethyan sutures

4.1. The Carpathian–Lesser Caucasus suture

In the greater part of the studied area the Eurasian and Afro-Arabian domains are separated by a megazone with developed oceanic volcanics and ophiolites. In the majority of cases, oceanic volcanics are of Jurassic, Early Cretaceous, and, in some places, Triassic age.

The East European segment

The megazone under consideration includes the Zlatibor (5, Fig. 1), Serbian (7) and Kapaonic (6) zones in the Dinarides, and the Vardar (8) and Orthys–Pelagonian (9) zones in the Hellenides (j, k, Fig. 2). Paleozoic and more ancient rock massifs lie among ophiolites. Parts of these massifs were likely microcontinents in the Tethys.

In the southern part of the Pannonian region, a Jurassic ophiolite association was developed in the Mures zone (4, Fig. 1) of the Apuseni Mountains. The ophiolitic zone is traced to the northeast under the younger deposits of the Transylvanian basin. Transylvanian ophiolitic nappes are the remnants of that zone (3, Fig. 1). The ophiolitic rocks of the Transylvanian nappes are of different ages, extending from Middle Triassic to Neocomian (Russo-Sandulescu et al., 1984). Coarse-grained molasses of the uppermost Albian (Vraconian) and the Cenomanian unconformably covered ophiolitic nappes and other units in the Apuseni Mountains and in the Inner zone of the Eastern Carpathians (Burchfiel, 1976; Bleahu et al., 1981; Sandulescu et al., 1981).

Toward the west, Mures ophiolites are traced under younger deposits up to the area of Belgrade, i.e., to the Vardar–Kapaonic ophiolite zone. The other branch of the Vardar–Kapaonic ophiolite zone continues to the northwest along the Danube valley.

In the northern part of the Pannonian region, the rocks of the ophiolite association are distributed in the Bükk mountains (2, Fig. 1), where mafic pillow lavas, gabbro and ultramafics crop out. The lavas are likely of Liassic age. Between the Bükk and Matra mountains, near the Darno line, gabbro-diabases, diabase dikes, pillow lavas, radiolarites and tuffs crop out. Petrochemical and geochemical studies of volcanics suggest that they are abyssal tholeiites (Onuoha, 1977; Embey-Isztin, 1980). South of the Balaton–Darno line, the rocks of the ophiolite association were penetrated by boreholes in the basement of the Great Hungarian depression in areas between the rivers Drava and Danube, Danube and Tisza and east of the river Tisza. Gabbro, spilites and diabases covered by radiolarites are discovered in the drill cores (Horvath et al., 1979). To the northeast similar sections were penetrated by boreholes in the basement of the Carpathian foredeep (Danilovich, 1981). Judging by petrologic characteristics, volcanics of this ophiolite association are oceanic tholeiites.

The zone under consideration, where rocks of the Mesozoic ophiolite association were developed, has an E–W strike. It approaches the Dinarides ophiolite zone in the area of Zagreb. One can consider that the belt of ophiolite development extends from the Subpelagonian zone of the Hellenides through the Zlatibor and Serbian zones of the Dinarides into the northern part of the Pannonian Basin. The Balaton–Darno line serves as a northwestern border of this belt, though north of the line Mesozoic ophiolite remains (Meliata formation) may lie as an allochthon on the Gemeric (Horvath et al., 1979) or may crop out in tectonic windows (Kozur, 1984). The zones with oceanic volcanics and ophiolites (Zlatibor, Igal–Bükk, Mures–Transilvanian) outline the borders of the megazone in the Pannonian basin. Such zones likely separate the Bichor and Mecsek massifs with pre-Alpine continental basement.

At the Bihor (B, Fig. 1) and Mecsek (M, Fig. 1) massifs, Early Liassic rocks are represented by coal-bearing paralic deposits common to the marginal sections of the Eurasian continent. The Liassic ammonite and brachiopod communities, studied in these middle massifs, belong to the North Tethys biogeographic province (Horvath et al., 1979; Geszy, 1984; Prosorovskaya and Vörös, 1988).

North of the Pannonian basin the Pieniny Klippen Belt serves as a border between the African and Eurasian domains (d, Fig. 3). In this zone the Jurassic is represented by siliceous-carbonate deposits in neritic and pelagic facies. In a cross section, the neritic Jurassic facies are distributed symmetrically in relation to the pelagic rocks, and this observation allows the recognition of the relicts of the deep-water basin and the delineation of submarine highs. The structure of the zone is extremely compressed. Now the zone is hundreds of meters to several kilometers wide and its length is over 700 km.

Ophiolite outcrops are absent in Pieniny Klippen Belt. The clasts of the ophiolite rock association are known only among the clastic material in the Cretaceous and Paleocene sandstones and grits. The Pieniny Klippen Belt is likely a suture of a Tethyan marginal basin. This zone acquired the role of a border between the domains of African and Eurasian origin only in the Neogene when the Gemeric–Tatric megazone was moved to it modern position (Burtman, 1986, 1988). The original suture of the Meso-Tethyan oceanic basin is now covered by the allochthonous Gemeric–Tatric zone.

The location of the Meso-Tethyan suture is clearer south of the Pannonian basin, where ophiolites form nappes overthrusted towards southwest. The root zone of these nappes is in the Vardar zone (8, Fig. 1), along which the Meso-Tethyan suture is located. One can trace the described megazone on the territory of the Aegean sea by ophiolite outcrops on the islands of the Cyclades archipelago. The allochthonous remnants of ophiolite nappes also lie on the islands of Crete and Rhodos.

The Asia Minor segment

In Asia Minor the main Meso-Tethyan suture occurs in the Izmir–Ankara ophiolite zone (10–11, Figs. 1, 4). This zone is a complex body containing melanges of different types, ultramafic massifs and massifs with the pre-Alpine continental crust. The melange with the serpentine matrix, which is developed in the Izmir–Ankara zone, includes blocks of the ophiolite association rocks and of sedimentary rocks. Blocks of basalts contain interlayers of jaspers and limestones with the fossils of Jurassic and Early Cretaceous age. Obduction of the ophiolites took place in the Late Cretacious and the late Maastrichtian deposits unconformably overlie ophiolites and other units (Kocyigit, 1990; Tüysuz, 1990, 1993; Yilmaz, 1990). Continental collision began here in the Late Cretaceous (Tüysuz, 1993) and likely continued to the early Cenozoic (Görür et al., 1985).

The northern branch of the main ophiolite megazone separates the Sakarya Massif from the Rhodope–Iranian megazone. Cretaceous ophiolites and ophiolitic melange are covered by the Maastrichtian olistostroma formation in that Interpontidian suture zone. This branch of the Tethys was also consumed through the Maastrichtian and the early Cenozoic (Sengör and Yilmaz, 1981; Yilmaz et al., 1982; Yilmaz, 1990; Okay, 1991). Biogeographic investigations of the Jurassic fauna shows, that the deep-water "barrier" between Eurasia and Gondwana land was south of the Sakarya microcontinent in the Izmir–Ankara branch of the Tethyan ocean (Fig. 1).

In the Eastern Pontides the ophiolite megazone is represented in the Kelkit river valley (12, Fig. 4), where a Cretaceous ophiolite association is distributed (Sengör et al., 1980).

The Lesser Caucasus segment

The Izmir–Ankara ophiolite zone bifurcates to the east. The northern (Lesser Caucasus) branch of the ophiolite zone extends from the Kelkit river valley to the northeast into the river Coruh basin. In the upper reaches of the rivers Kelkit and Coruh and in the Karasu (Euphrates) river valley serpentine melange is developed with blocks of basalts, radiolarites and sedimentary rocks. Senonian rudist limestones unconformably overlie this melange in the Coruh–Karasu interfluve (13, Fig. 4). Two main sutures—Paleo-Tethyan and Meso-Tethyan—are joined in that area (Sengör et al., 1980; Tüyzüs, 1990).

Farther to the east the described zone extends to Armenia. In the Amasia area (15, Fig. 4) serpentine melange is distributed. It contains large bodies of peridotites, dunites, pyroxenites, gabbro, gabbro-norites and troctolites. Other blocks in the melange are built by mafic lavas, radiolarites, metamorphic rocks and also by Cretaceous and Paleocene sedimentary rocks. The melange is tectonically overlain by Late Cretaceous limestones. Within this Amasia melange zone it was observed that the melange is unconformably overlain by Senonian basal conglomerates (Sokolov, 1974). The Amasia ophiolite zone has an imbricated structure. The tectonic slices dip steeply. Both slices and knockers in the serpentinite melange are oriented parallel with the strike of the ophiolite zone. Mylonites are widely developed. The inner structure of the Amasia zone corresponds with its location in the place of the suture.

The eastern extension of the described ophiolite zone is in the Shirak and Basum ridges (16, Fig. 4). This territory is known to contain many ultramafic outcrops, which are mainly located in Cretaceous rocks and have tectonic contacts against them. Peridotites and gabbro crop out on the south slope of the Shirak ridge. To the east serpentinized pyroxenites, olivine gabbro, basalts and radiolarites are distributed in the Bazum ridge. Farther east ophiolites are penetrated by holes drilled in the northwestern end of Lake Sevan and on its south coast (19, Fig. 4), where the hole went through pyroxenites, gabbro and diabases (Satian, 1984).

The Lesser Caucasus ophiolite zone continues from Lake Sevan under the cover of young deposits into the Zangezur ridge, where the ophiolite association forms a tectonic melange developed in the Zangezur fault zone. Here peridotites, serpentinites, olivine gabbro, troctolites, spilites, andesites and radiolarites occur. The upper age limit of this association is determined from ophiolite pebbles in Senonian conglomerates (Satian, 1984). Southeast of the Zangezur ridge, in the Qara Dagh mountains in Iran (21, Fig. 4) ophiolites are overlain by Late Cretaceous pelagic limestones. The ophiolites build nappes, the obduction is dated as Late Cretaceous (Berberian, 1983).

Northeast of the Lesser Caucasus ophiolitic zone ophiolites of the Sevan zone (20, Fig. 4) occur. The Sevan ophiolites (s, Fig. 5) form a system of nappes torn off from their roots. These nappes overlie the rocks of the Rhodope–Iranian megazone. The volcanic-siliceous part of the ophiolite association contains a fauna of Middle and Late Jurassic, Early Cretaceous and early Late Cretaceous age. The ophiolite overthrusting was in the Cenomanian–Coniacian (Sokolov, 1974; Knipper, 1975; Satian, 1984). In the Sevan zone, melange gabbro-diabases and andesite porphyrites were also discovered in the ophiolite. Their isotopic age (K-Ar) was determined at 291 ± 3 Ma (Gasanov, 1985).

Southwest of the Lesser Caucasus ophiolite zone there is the Vedi zone of allochthonous ophiolites (18, Fig. 4; t, Fig. 5). The ophiolite nappes lie here on the Armenian block. It is highly probable that the Lesser Caucasus (Zangezur) ophiolite zone is the root zone of Sevan and Vedi ophiolite allochthons (Knipper and Khain, 1980).

In northeastern Iran a very interesting tectonic junction occurs, but unfortunately the data on its geology are scarce. Northwest of this region, the Lesser Caucasus ophiolite zone contains suture of two oceans (Paleo-Tethys and Meso-Tethys), and to the east and south sutures of that oceans diverge. In any case, in the Western Alborz the Eurasian coal-bearing Liassic formation stratigraphically lies over the Late and Early Triassic carbonate rocks, which are typical for Gondwana Land (Davies et al., 1972).

At the northwestern termination of the Alborz in the Iranian Talesh (Bogrovdag), the rocks of the ophiolite association are known, cropping out west of the city of Rasht (23, Fig. 4). Data are scanty on these ophiolites. Peridotites and gabbros are mentioned, and serpentinite protrusions in Liassic rocks are described. There are pebbles of ultrabasites in Jurassic conglomerates. The stratigraphic section of the Iranian Talesh contains Silurian and Devonian basalts, andesites and tuffs (Davies et al., 1972; Stocklin, 1974; Berberian and King, 1981). North of the Iranian Talesh and the Alborz there are no Gondwana facies of late Paleozoic and Triassic age. Thus, there is a good reason to draw the suture of the Paleo-Tethys through the northern part of the Iranian Talesh and the Caspian sea into Khorasan in northeastern Iran. Some investigators consider that the Rasht ophiolites represent this suture (Berberian and King, 1981).

The trace of the Meso-Tethyan suture is difficult to follow south of Zangezur and the Iranian Qara Gagh mountains owing to scarce data on this region. We proved above that this suture is the southern border of the distribution of Liassic terrigenous and paralic coal-bearing deposits. Having taken this circumstance into account, we can locate the suture in Iran. We saw that coal-bearing Lias is developed north and northeast of the Lesser Caucasus ophiolite zone. These deposits are also distributed southeast of the Armenian highlands in the Iranian Talesh. South of the Armenian highlands, the coal-bearing Lias is developed in Iran up to Lake Urumiyeh in the west. Hence, the development area of continental and paralic Liassic facies surrounds the Armenian highlands in the north, east and south.

The Main Zagros Thrust serves as the southwestern border of development of the facies under consideration. Beyond this border, the Lias and Dogger of the High Zagros are represented by marine, mainly carbonate deposits. One can thus conclude that the Meso-Tethyan suture passes south of the Armenian median massif from the ophiolites of the Zangezur ridge and the Iranian Qara Dagh mountains to Lake Urumiyeh and farther to the Sanandai–Sirjan zone. Most likely, this suture passes through the Khoi ophiolites, which are located at the northern end of Lake Urumiyeh.

The Khoi ophiolites (22, Fig. 4) constitute a melange formed from ultramafics, radiolarites, diabases, tuffs, pelagic limestones and shales. These rocks contain a Late Cretaceous fauna. The melange is overlain by Eocene flysch (Stöcklin, 1974). The Khoi ophiolites likely serve as a link between the northern (Lesser Caucasus) and southern branches of the ophiolite megazone. The southern branch extends along the western border of the Armenian median massif to the Karakese ophiolites (17, Fig. 4) and extends farther southeastwards between the lakes Van and Urumiyeh to the Zagros. The ophiolite zones, it seems, surround the Armenian highlands, which thus formed a microcontinent in the early Mesozoic.

The pre-Alpine basement of the Armenian block is made up of platform deposits of the Gondwana-land type (Paleozoic and Triassic) and by supposed Precambrian metamorphic rocks. The Lias is represented by basalts and tuffs (u, Fig. 5); volcanics have a weak alkaline tholeiitic composition with a high titanium content. They are similar to intercontinental rift basalts (Knipper, 1975). The volcanics are overlain by clays, sandstones and limestones containing Bajocian and Batonian ammonites. As the ammonite fauna shows, the deposits belong to the South Tethyan biogeographic province (Rostovtsev and Azarjan, 1971; Rostovtsev, 1978). Hence, the Armenian microcontinent belongs to the Afro-Arabian row and in the Jurassic it was located in the South Tethys.

Paleomagnetic investigation of the Early Jurassic rocks from the Armenian block have made (Bazhenov et al., 1991). According those data the distance between the Armenian microcontinent and the southern border of Eurasia in the Early Jurassic did not exceed 500–600 km in case when the microcontinent was in juxtaposition with the Great Caucasus area. The reconstruction with a more western position of the Armenian microcontinent in the Jurassic (Kazmin and Knipper, 1989) can solve the contradiction between paleomagnetic and paleobiological data.

Thus, the Meso-Tethyan suture likely surrounds the Armenian block from the north, east and south. South of Lake Urumiyeh it reaches the Neo-Tethyan suture. The Neo-Tethyan suture formed in the late Cenozoic owing to the collision of the Arabian continent with Eurasia. It outlines the northern part of the Arabian platform and then continues to the southeast along the border of the Sanandai–Sirjan zone and the High Zagros, then through Makran and Baluchestan into the Himalayas.

4.2. Mesozoic ophiolites of southern and eastern Iran

Mesozoic ophiolite nappes developed near the Neo-Tethyan suture. They were obducted onto the Arabian platform at the end of the Cretaceous. These ophiolites are the remains of the peri-Arabian oceanic basin which formed in the Permo-Triassic as a result of splitting of the Arabian platform margin. The fragments of the Mesozoic oceanic crust are distributed throughout the whole periphery of the Arabian platform from Oman in the southeast to Syria and Cyprus. In the Zagros the Mesozoic ophiolites are preserved near the Neo-Tethyan suture in the city of Kermanshah and near Lake Neyriz (26, 27, Fig. 5; Ricou, 1976; Braud, 1987).

Mesozoic ophiolites are widely developed at the margins of the Central Iranian block (L, Fig. 4). The eastern part of this block is formed by grabens and horsts. The grabens are filled by thick Jurassic and Cretaceous deposits. The Lias is represented by the coal-bearing Nayband formation. The lower part of its section contains Late Triassic and Liassic floras (Poliansky, 1980). In the northern part of the Central Iranian block among terrigenous rocks interlayers of limestones with a Sinemurian marine fauna occur. The coal-bearing formation is overlain by limestones which contain goniotites of late Toarcian–Bajocian age. This fauna discovered in the southern part of the Central Iranian block (near the city of Kerman) proves that the region belongs to the North Tethyan biogeographic province (Hallam, 1975).

The following ophiolite zones are located along the borders of the Central Iranian block: the Nain–Baft zone (28–29, Fig. 4) along the western border, the Doruneh–Joghatai zone (24–25) along the northern border, the Daz Murian zone (30) along the southern border and the Zabol–Baluch zone (31, Fig. 4) on the eastern border. Melange is developed in these zones. It includes ultramafics, mafic lavas and pelagic sedimentary rocks. The age of the sedimentary rocks in clasts ranges from late Turonian to Maastrichtian. The determination of K-Ar ages on basic volcanics and gabbros also date these rocks as Late Cretaceous (Berberian and King, 1981; Desmons and Beccaluva, 1983).

A large area of ophiolitic melange is developed at the northern border of the Central Iranian block (25, Fig. 4). To the north, in the vicinity of the city of Sabzevar the ophiolites form several bodies on Joghatai ridge (24, Fig. 4). Besides the ophiolitic melange, a complete section of the ophiolite association is seen here. Its lower part is occupied by harzburgites, dunites, lherzolites and peridotites, followed by cumulate gabbro-norites and troctolites. An association of diabasic sheeted dikes lies higher, overlain by a volcanogenic-sedimentary horizon of mafic pillow lavas (20%), pyroclastic rocks (70%), radiolarites and pelagic limestones with Campanian–Maastrichtian *Globotruncana*. Volcanics have a low titanium content. Amphibolites, glaucophane and green shales and coarse clastic deposits with clasts of the ophiolite association also developed in the Sabzevar region. Paleocene limestones overlie these deposits (Knipper, 1975; Lensch et al., 1977; Desmonds and Beccaluva, 1983).

There is a tendency to connect by an oceanic suture the ophiolite zone of the Joghatai ridge with ophiolites of Karadag and Zangezur, drawing this suture along the Alborz (Berberian and King, 1981; Berberian, 1983). In this connection the development of Cretaceous volcanics is mentioned in the Alborz where they have a tholeiitic and alkaline composition. However, the Late Alpine fold structure of the region serves as the main ground for these assumptions. On the whole, there are no adequate reasons for drawing of a Mesozoic oceanic suture along the southern flank of the Alborz.

The ultramafic rocks (harzburgites, dunites, lherzolites) are represented in the Zabol–Baluch zone located on the eastern border of the Central Iranian massif. They form an ophiolitic melange in which gabbro is also present. The melange is overlain by pillow basalts with horizons of hyaloclastites, radiolarites and pelagic limestones with Tithonian–Maastrichtian microfauna. The basalt petrochemistry reveals their closeness to tholeites of midoceanic ridges. The ophiolites developed at the southern border of the Central Iranian block were forming in an oceanic environment (Desmonds and Beccaluva, 1983; Tirrul et al., 1983). In the Zabol–Baluch zone the ophiolites are associated with a thick horizon of the Late Cretaceous flysch containing radiolarites and other deep-water deposits. These deposits greatly differ from carbonate deposits of the same age developed in the inner part of the Central Iranian block. Zones of Mesozoic oceanic crust development were connected with each other, almost completely surrounding the Central Iranian block. Now the sutures of this oceanic are represented by faults with ophiolites.

4.3. *The Afghan–Tibet suture*

East of the Central Iranian block, a Meso-Tethyan suture is located in southern Afghanistan. Continental and paralic coal-bearing Lias facies developed widely in northern Afghanistan, north of the Harirud–Hindu Kush fault. They are also known directly south of this fault, in the Rude–Kafgan area. The Meso-Tethyan suture should be looked for to the south, in the Farahrud zone (32, Fig. 4). The rocks of the ophiolite association crop out on the northern and southeastern borders of the Farahrud zone and its central part. The largest ultramafic bodies are located near the Hashrud and Helmand faults along the southeastern border of the zone. The ultramafics (serpentinites, peridotites, dunites) are located among thick Late Jurassic–Early Cretaceous deposits, which are represented by mafic and intermediate lavas with strata of siltstones, phyllites, silicons and limestones (Dronov, 1980).

The Meso-Tethyan suture passes along the Farahrud zone in a southwesterly direction to the Central Iranian block. Farther, the suture is traced via the Zabol–Baluch and Daz Murian ophiolite zones up to the Neo-Tethyan suture.

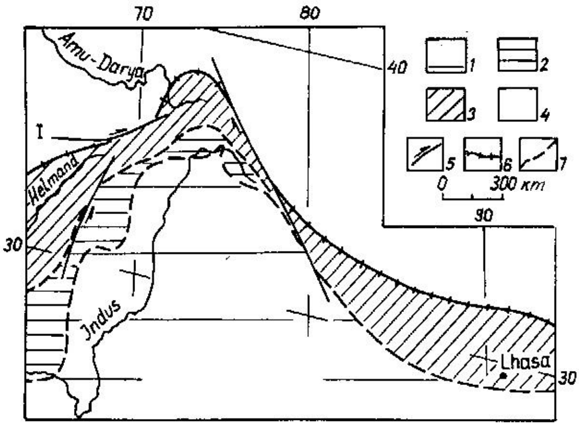

Fig. 7. The Pamirs and Tibet. *1* = Indian continent; *2* = rocks of Neo-Tethyan island arcs and marginal seas; *3* = Hilmand–Lhasa microcontinent; *4* = Eurasian continent; *5* = strike-slip faults; *6–7* = oceanic sutures: *6* = Meso-Tethyan, *7* = Neo-Tethyan.

The Farahrud zone is truncated in northeastern direction being cut by the Hindu Kush faults. The Harirud–Hindu Kush fault and its western extension—the Zebak–Minjan fault—are right-lateral strike-slip faults. The extension of the Meso-Tethyan suture is shifted along the strike slip and is situated in Afghan Badakhshan (Fig. 7). Farther on, the suture passes into the Rushan–Pshart fault zone separating the Central and Southern Pamirs (Pashkov and Shvolman, 1979).

Continental coal-bearing deposits (5000 m) with Late Triassic, Liassic and Dogger flora developed in Afghan Badakhshan north of the Meso-Tethyan suture (Dronov, 1980). These deposits extend into the Central Pamirs. Deposits of continental slopes of both borders of the Meso-Tethys are preserved in the Rushan–Pshart zone and near it. Remnants of Triassic and Jurassic ophiolites are known in the Rushan–Pshart zone and south of it (Shvolman, 1978, 1980; Pashkov and Shvolman, 1979; Dronov, 1986, 1988).

The eastern extension of the Meso-Tethyan suture is shifted by the Pamir–Karakorum strike-slip fault into Tibet (Fig. 7), from where it passes over to Southeast Asia (Peive et al., 1964; Burtman, 1982; Sengör, 1984; Shvolman and Pashkov, 1986; Pearce and Deng, 1988; Girardeau et al., 1989; Burtman and Molnar, 1993). In Tibet, the Meso-Tethyan suture separates the Lhasa block from northern Tibet. In the Jurassic, the Lhasa block was a part of a large Helmand–Lhasa microcontinent, which included also the Southern Pamirs and southern Afghanistan (Sengör, 1984). The fauna in the Jurassic limestones of the southeastern Pamirs, which were deposited on the southern margin of the Afghan–Tibet Mesozoic oceanic basin, belongs to the North Tethyan biogeographic province (Dronov and Andreeva, 1962). Hence, the Helmand–Lhasa microcontinent must have been near Eurasia in Jurassic times.

5. Meso-Tethyan sutures and Alpine deformations

In the late Cenozoic, a greater part of the Tethys ocean was closed due to joining of Indian and Arabian continents to Eurasia. As a result, sutures of the Neo-Tethys came into being; the main ones are shown in Figs. 5 and 8. The collision of continents resulted in large tectonic deformations and the present structure of the Alpine belt was formed. While the process was going on, the Meso-Tethyan sutures were also deformed. Since the fold structure of the Alpine belt was formed by shortening, let us try to evaluate the range of magnitude of this shortening. Our task is to study lateral deformations, which reflect the movements of the masses in the earth crust.

The scale of such movements can be estimated by comparing the present position of the Meso-Tethyan sutures with the reconstructed locations of these sutures in the geological past. Such a comparison offers the possibility to trace back the inner deformation of the Alpine belt in the Cenozoic. The displacements due to shortening of oceanic basins are not considered, for the Meso-Tethyan sutures are contained wholly within the area of continental crust formed in early Cenozoic and earlier.

5.1. The Carpathian–Lesser Caucasus suture

The reconstruction of the Carpathian–Lesser Caucasus suture (Fig. 8) is based on data on paleomagnetic declinations in Late Cretaceous rocks (Table 1) and results of investigations on the origin of structural arcs in orogenic belts (Burtman, 1986, 1989). Paleomagnetic data of the paleolatitude, obtained in the Lesser Caucasus (Bazhenov and Burtman, 1989, 1990) for the Late Cretaceous rocks in the Sevan ridge close to the Meso-Tethyan suture, were also used. According to these data, at the end of the Cretaceous, the Lesser Caucasian part of the Meso-Tethyan suture was located 600–1200 km south of its present position. While coordinating the data from the Caucasus with the reconstruction of the same suture in the Carpathian–Balkan region, the maximal displacement value of the suture in the Caucasus is more preferable, because this minimizes the sinuosity of the trend line and thus eliminates unnecessary along-strike shortening of it.

So, the displacement of the Carpathian–Lesser Caucasus suture reaches its maximum in the apex of the Arabian–Turkish syntaxis, where the magnitude is about 1200 km and the displacement direction is across the Alpine belt. To the west the magnitude diminishes to 400–500 km at the border between the Hellenides and Dinarides. Here the displacement direction is along the belt (Fig. 8). Farther on the magnitude increases in the Carpathian loop area and again decreases towards the Eastern Alps. Migration of the suture is the result of shortening of the earth crust in the North Alpine belt, while its fold and nappe structure was forming.

5.2. Formation of the Arabian–Turkish syntaxis

As the Meso-Tethyan sutures are situated inside the Alpine belt, the magnitude of their displacement is less than the shortening value across the belt. Such shortening can be determined in the Arabian–Turkish syntaxis by comparing paleomagnetic latitudes defined for the Late Cretaceous on the Arabian platform (Ron et al., 1984) and in Dagestan (D, Fig. 4) on the northern slope of the Greater Caucasus ridge at the northern border of the Alpine belt (Bazhenov and Burtman, 1989, 1990). According to these data, the distance between the northern edge of the Ara-

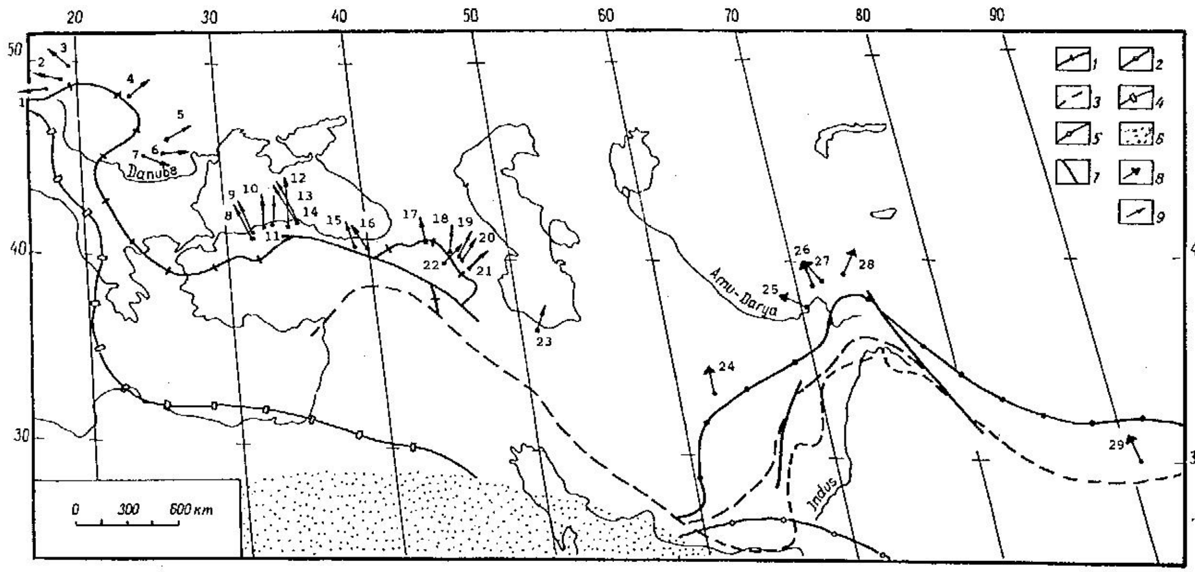

Fig. 8. Deformation of Meso-Tethyan sutures in the Cenozoic. *1–2* = present position of Meso-Tethyan sutures: *1* = Carpathian–Lesser Caucasus suture, *2* = Afghan–Tibet suture; *3* = Neo-Tethyan sutures; *4–6* = reconstruction for the Late Cretaceous–early Cenozoic: *4* = Carpathian–Lesser Caucasus suture, *5* = Afghan–Tibet suture, *6* = ocean Tethys; *7* = strike-slip faults; *8–9* = paleomagnetic declinations in the rocks: *8* = Paleocene, *9* = Late Cretaceous.

bian platform and the territory of Dagestan was 22 ± 4° in the Late Cretaceous and at present the geographic latitude of those areas differ only by 6°. Hence, the Alpine belt got narrower here by 1800 ± 450 km. One can also determine the value of the lateral shortening of the Alpine belt by using the data on movement of the Afro-Arabian plate, provided by the magnetic anomalies in the present oceans. According to these data the distance between Arabia and Eurasia decreased by approximately 2000 km in the Cenozoic (Aubouin et al., 1986).

Folding and nappe emplacement occurred due to shortening of the Alpine belt and lateral mass flows emerged as a result of tectonic escape. Let us consider the structural plan of the Arabian–Turkish syntaxes, where the axis of the Alpine folds delineate a few major structural arcs (Fig. 9).

The East Taurus arc is outlined by a fold axis of the Eastern Taurus and southern Kurdistan. The following structural elements run parallel within this arc: folds in the cover of the Arabian platform; folds in the Alpine belt; and the border of the Arabian plate. Such relationships indicate development of folds of the structural arc as a result of the direct pressure of the edge of the Arabian plate.

East and north of the East Taurus arc we find a system of conjugated arcs, including the Khorasan, Elburz, Lesser Caucasus and Trabzon arcs (Fig. 9). This system of conjugate arcs (the KELT system of arcs) is disharmonic relative to the East Taurus arc and borders it along the strike slip. The KELT system of arcs is also disharmonic relative to the structures developed north of this system. Paleomagnetic study shows that the Lesser Caucasus arc has secondary origin: it was oroclinally bend after development of the folds that outline the arc (Bazhenov and Burtman, 1989, 1990). The Elburz and Trabson arcs, conjugate with the Lesser Caucasus arc, probably also have a secondary origin. Paleomagnetic study in the Eastern Pontides (Van der Voo, 1968) speaks in favor of this origin for the Trabzon structural arc.

The West Kopet Dag and South Caspian structural arcs bound the KELT system of arcs in the northeast. The West Kopet Dag arc is disharmonic relative to the Khorasan and Elburz arcs, the South Caspian arc is disharmonic relative to the Lesser Caucasus and Elburz arcs. Paleomagnetic study (Bazhenov, 1987) of the West Kopet Dag arc revealed that the folds have an unchanged strike here. Consequently, the transverse position of the South Caspian arc relative to the Khorasan arc is primary. Such relations are likely between the South Caspian and Lesser Caucasus arcs.

Thus, the folds of the East Taurus arc formed under the direct influence of the edge of the Arabian plate, but the disharmonic secondary arcs (oroclines) developed as a result of mass flow along the Alpine belt. Relations between the Lesser Caucasus and East Taurus arcs indicates mass flow from the crest of the syntaxis in an easterly directions. Rock masses were also extruded in a westerly direction as indicated by recent structure of Asia Minor (Sengor and Yilmaz, 1981). The West Kopet Dag and South Caspian arcs are derivatives with respect to the arcs of the KELT system. The folds within these arcs developed as a result of extrusion of mass from crests of the Lesser Caucasus and the Khorasan structural arcs (Kopp, 1979; Burtman, 1989).

The onset of the orogenesis in the region of the future syntaxis is Late Eocene–Early Oligocene in age. By this times the Arabian plate had approached Eurasia and than moved northwesterly along its boundary in the interval 35–20 Ma (Aubouin et al., 1986). The Arabian plate squeezed Minor Asian earth crust masses to the northwest and rotated them counterclockwise, which initiated the formation of the Carpathian loop (Burtman, 1986, 1988). But in the Caucasian transaction, interaction of the plates was limited and deformations were relatively weak.

In the Early Miocene, the direction of the displacement of the Arabian plate relative to

Table 1
Paleomagnetic data from Late Cretaceous rocks based on a Late Cretaceous reconstruction of the Carpathyan–Lesser Caucasus suture

No.	References	D	α_{95}	λ_c	β
1	Bazhenov et al., 1980	266	6	13	+107
2	Marschalko and Pagac, 1980	286	17	13	+87
3	Krs et al., 1979	313	3	13	+60
4	Bazhenov et al., 1980	52	6	15	−37
5	Bazhenov and Burtman, 1990	65	10	14	−51
6	Bazhenov and Burtman, 1990	89	8	14	−75
7	Bazhenov and Burtman, 1990	111	10	14	−97
8	Saribudak, 1989	339	8	17	+35
9	Saribudak, 1989	343	18	17	+31
10	Saribudak, 1989	3	8	17	+14
11	Saribudak, 1989	8	8	17	+9
12	Saribudak, 1989	2	4	17	+15
13	Saribudak, 1989	335	5	17	+42
14	Saribudak, 1989	341	10	17	+36
15	Orbay and Bayburdi, 1979	347	10	18	+31
16	Van der Voo, 1968	334	10	18	+44
17	Bazhenov and Burtman, 1989, 1990	354	3	20	+26
18	Bazhenov and Burtman, 1989, 1990	13	3	20	+7
19	Bazhenov and Burtman, 1989, 1990	35	12	20	−15
20	Bazhenov and Burtman, 1989, 1990	37	5	20	−17
21	Bazhenov and Burtman, 1989, 1990	56	12	20	−36
22	Pecherskii and Nguen, 1978	50	13	20	−30
23	Wensink and Varecamp, 1980	25	6	20	−5

D = declination of primary magnitization; α_{95} = accuracy of NRM determination with a 95% probability; λ_c = direction of the Cretaceous paleomeridian in degrees relative to the present-day meridian; β = angle of rotation of the structural zone after the Late Cretaceous (+ = counterclockwise rotation, − = clockwise rotation).
Numbers refer to location given in Fig. 8.

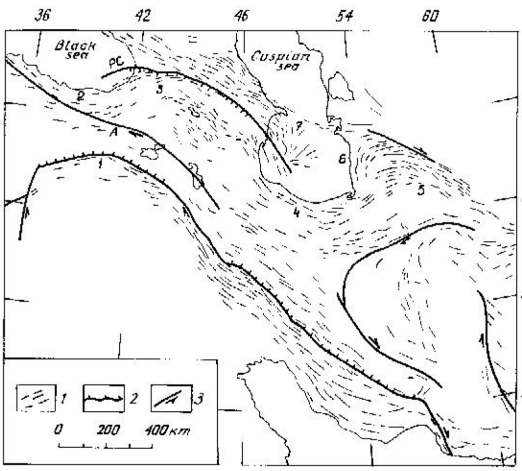

Fig. 9. Structural outline of the Arabian–Turkish syntaxis. *1* = fold axes; *2* = main thrusts; *3* = main strike slips; *PC* = Pontus–Caspian fault; *A* = North Anatolian fault. Structural arcs: *1* = East Taurus; *2* = Trabson; *3* = Lesser Caucasus; *4* = Elburz; *5* = Khorasan, *6* = West Kopet Dag; *7* = South Caspian.

Eurasia changed to the north-northeast and collision of these plates began. It is natural that the most intense Early Miocene deformation occurred on the southern boundary of the Alpine belt where Taurus and Zagros nappes and folds of the East Taurus structural arc formed. Development of folds of this arc continued later and Pliocene deposits are folded in the marginal part of the Arabian plate.

The folds of the Lesser Caucasus and northwestern Iran developed in the Miocene and were then deformed in the structural arcs of the KELT system. Northward overthrusting of the Lesser Caucasus along the Pontus–Caspian thrust (Fig. 9) is associated with development of the Lesser Caucasus arc (Burtman, 1989). The youngest deposits covered by the overthrust have a Late Miocene age. This enables us to assign the onset of development of the arcs of the KELT system to the Miocene, although the main stage of their development is later. A sharp angular disconformity between Early and Late Pliocene deposits (before the Akchagylian) is seen in the outer part of the Lesser Caucasus arc (in the Adjar–Trialet zone).

The folds of the South Caspian and West Kopet Dag arcs formed on the Quaternary in the late stage of development of the KELT system of arcs.

5.3. The Afghan–Tibet suture

Near the Afghan–Tibet suture there are no paleomagnetic data on Late Cretaceous rocks. The results of the paleomagnetic study of Early Cretaceous rocks in the Northern Pamirs (25–28, Fig. 8) testify that this zone moved 600–700 km to the north (in relation to Eurasia) through the Alpine deformations (Bazhenov and Burtman, 1981, 1982, 1986, 1990). These data concern the outer zone of the Alpine fold belt which is located north of the Afghan–Tibet suture. They show that the Meso-Tethyan suture moved over 600 km to the north, but this value is not the full amplitude of its displacement, and no paleomagnetic data are available from the immediate northern margin of the suture.

Paleomagnetic data were obtained from Aptian–Albian rocks of the Takena formation and on Paleogene andesites (60–48 Ma) which are

Table 2
Paleomagnetic data from Paleogene rocks based on the reconstruction of the Afghan–Tibet suture

No.	References	Age	D	α_{95}	λ_p	β
24	Krumsiek, 1976	Eoc	355	–	20	+25
25	Bazhenov and Burtman, 1986, 1990	Eoc–Mio	305	7	19	+74
26	Bazhenov and Burtman, 1986, 1990	Oli–Mio	352	7	19	+27
27	Bazhenov and Burtman, 1986, 1990	Oli	329	7	19	+50
28	Bazhenov and Burtman, 1986, 1990	Eoc–Mio	37	5	19	−18
29	Achache et al., 1984	Pal–Eoc	351	10	18	+27

λ_p = direction of the Paleogene paleomeridian; other symbols as in Table 1.
Numbers refer to location given in Fig. 8.

distributed in the Lhasa block (29, Fig. 8; Table 2), south of the Afghan–Tibet suture (Achache et al., 1984). The Takena formation deposits formed after closing of the Meso-Tethys and construction of the suture. The paleomagnetic studies showed that during the Late Cretaceous, Paleocene and Early Eocene the paleolatitude of the Lhasa block had not changed, but for the last 50 Ma this block moved 2000 + 850 km to the north in relation to stable Eurasia. The data on the magnetic anomalies of the Indian ocean bottom also show that the Indian plate approached Eurasia by 2000–2500 km after the collision had started (Norton and Sclater, 1979; Westphal and Pozzi, 1983). A great ensemble of Cenozoic nappes, folds and structural arcs of the Pamir–Punjab syntaxis (Burtman, 1982; Bazhenov and Burtman, 1986) is the result of lateral shortening of the Alpine belt through the collision.

6. Conclusions

In the region under consideration there were two northern branches of the Tethys, Carpathian–Lesser Caucasus and Afghan–Tibet, which were closed in the Cretaceous–early Cenozoic. The above-mentioned main branches also had branches, which remained as sutures of small basins: the Kamennopotock, Interpontide, Nain-Baft basins and others.

In the Western Carpathians, the Carpathian–Lesser Caucasus suture is overlain by a widespread allochton—the Gemeric-Tatric massif. From the Western Carpathians the suture passes through the Pannonian basin in the Vardar ophiolite zone and farther to the ophiolites of the Izmir–Ankara zone. Being displaced along the North Anatolian right-lateral strike-slip fault, the main suture passes from the Eastern Pontides into the Lesser Caucasus, where it is marked by ophiolites of the Amasia area, the Shirak, Bozum and Zangezur ridges. Then it can be observed through the Iranian Qara Dagh mountains to Lake Urumiyeh and the North Anatolian strike-slip fault. Being again shifted along the strike-slip fault, the Carpathian–Lesser Caucasus Meso-Tethyan suture ends in the Western Zagros near the Cenozoic Neo-Tethyan suture.

The Afghan–Tibet Meso-Tethyan suture is situated in the Pamirs in the Rushan–Pshart zone. The east prolongation of the suture being displaced along the Pamir–Karakorum right-lateral strike-slip fault in Tibet. West of the Pamirs this suture also is displaced along strike-slip faults and continues in the Farahrud zone in Afghanistan, and than passes through the Zabol–Baluch and Daz Murian ophiolite zones and approaches the Neo-Tethyan suture.

Comparing the present position of the Carpathian–Lesser Caucasus and Afghan–Tibet Meso-Tethyan sutures with the Late Cretaceous–early Cenozoic reconstructions of these sutures made it possible to trace the inner deformation of the Alpine belt. During the Alpine deformation the displacement of the Carpathian–Lesser Caucasus suture reaches 1200 km in the top of the Arabian–Turkish syntaxis, the displacement is directed across the fold belt. At the border of the Hellenides and Dinarides the displacement amplitude decreases to 400–500 km and the movement is left-laterally directed along the fold belt. The displacement amplitude increases in the Pannonian area and again decreases towards the Eastern Alps. The displacement of the Afghan–Tibet suture exceeds 2000 km.

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References

Adamia, Sh.A., Asanidze, B.Z., Gambashidze, R.A., Nadareishwili, G.Sh., Nguen, T.K.T. and Petherskii, D.M.,

1979. Paleomagnetism of the Upper Cretaceous rocks of South Georgia and its geological interpretations. Izv. Akad. Nauk SSSR, Ser. Geol., 5: 46–57 (in Russian).

Aiello, E., Bartolini, C., Boccaletti, M., Gocev, P., Karaguleva, J., Kostadinov, V. and Manetty, P., 1977. Sedimentary features of the Srednegorie zone (Bulgarie): an Upper Cretaceous intra-arc basin. Sediment. Geol., 19: 36–68.

Androsov, B.N., Kolchanov, V.P. and Kulakov, B.B., 1977. Coal in Northern Foothills of the Hindu Cuss. Nauka, Moscow, 135 pp. (in Russian).

Andrusov, D. and Samuel O., 1973. Cretaceous–Paleocene of the West Carpathians. 10th Congr. Carpathian–Balcan Geol. Assoc. Guideb. Geol. Inst., Bratislava, 62 pp.

Andrusov, D., Bystricky, J. and Fusan, O., 1973. Outline of structure of the West Carpathians. 10th Congr. Carpathian–Balcan Geol. Assoc. Guideb. Geol. Inst., Bratislava, 44 pp.

Archache, J., Courtillot, V. and Xiu, Z.Y., 1984. Paleogeographic and tectonic evolution of southern Tibet since middle Cretaceous time: new paleomagnetic data and synthesis. J. Geophys. Res., B89: 10,311–10,339.

Aubouin, J., 1965. Geosynclines. Elsevier, Amsterdam, 335 pp.

Aubouin, J., Le Pichon, X. and Monin, A.S. (Editors), 1986. Evolution of the Tethys. Tectonophysics, 123: 1–316.

Asizbekov, S.A., 1961. Geology of the Nahichevan ASSR. Gosgeolizdat, Moscow, 502 pp. (in Russian).

Azisbekov, S.A. (Editor), 1972. Azerbaidjan SSR. Geology of the USSR, vol. 47. Nedra, Moscow, 520 pp. (in Russian).

Baussollet, J.P., Bergougnan, H. and Enay, R., 1975. Repartitions des faunes et facies Liasiques dans l'East de la Turquie. C.R. Acad. Sci., Ser. D, 280: 583–586.

Baussollet, J.P., Fourcade, E. and Peybernes, B., 1985. Paléobiogéographie des grands Foraminiferas bentiques des marges néo-tethysiennes au Jurassique et au Cretace inférieurs. Bull. Soc. Géol. Fr., Ser. 8, 1: 699–713.

Bazhenov, M.L., 1987. Paleomagnetism of Cretaceous and Paleogene sedimentary rocks from the Kopetdagh and its tectonic implications. Tectonophysics, 136: 223–235.

Bazhenov, M.L. and Burtman, V.S., 1981. Formation of Pamir–Punjab syntaxis: implications from paleomagnetic investigations of Lower Cretaceous and Paleocene rocks of the Pamirs. In: A.K. Sinha (Editor), Contemporary Geoscientific Researches in Himalaya, Vol. 1. Pal Singh, Dehra Dun, pp. 71–81.

Bazhenov, M.L. and Burtman, V.S., 1982. The kinematics of the Pamir Arc. Geotectonics, 16(4): 288–301.

Bazhenov, M.L. and Burtman, V.S., 1986. Tectonics and paleomagnetism of structural arcs of the Pamir–Punjab syntaxis. J. Geodyn., 5: 383–396.

Bazhenov, M.L. and Burtman, V.S., 1989. Paleomagnetism of Upper Cretaceous rocks from the Caucasus and its implications for tectonics. In: A.M.C. Sengör (Editor), Tectonic Evolution of the Tethyan Region. Kluwer, Dordrecht, pp. 217–240.

Bazhenov, M.L. and Burtman, V.S., 1990. Structural Arcs of the Alpine belt: Carpayhians–Caucasus–Pamirs. Nauka, Moscow, 168 pp. (in Russian).

Bazhenov, M.L., Began, A., Birkenmajer, K. and Burtman, V.S., 1980. Paleomagnetic evidence for the tectonic origin of curvature of the West Carpathian arc. Acad. Pol. Sci. Bull., Ser. Sci. Terra, 28(4): 281–290.

Bazhenov, M.L., Burtman, V.S. and Tsiganova, I.V., 1991. To reconstruction of the Mesozoic Tethys on the Caucasus. Geotectonics, 25(1): 48–58.

Begougnan, H., 1976. Structure de la Chaine Pontique dans le Haut-Kelkit. Soc. Géol. Fr. Bull., Sér. 7, 18: 675–686.

Berberian, M., 1983. The Southern Caspian: a compressional depression floored by a trapped modified oceanic crust. Can. J. Earth. Sci., 20: 163–183.

Berberian, M. and King, G.C.P., 1981. Towards a paleogeography and tectonic evolution of Iran. Can. J. Earth Sci., 18: 210–265.

Birkenmajer, K., 1977. Jurassic and Cretaceous lithostratigraphic units of the Pieniny Klippen belt, Carpathians, Poland. Stud. Geol. Pol., 14: 1–158.

Bysova, S.L., Rudakov, S.G., Slavin, V.I. and Khain, V.E., 1971. Main tectonic futures of north part of the Marmorosh massif. Geotectonica, 6: 79–86 (in Russian).

Bysova, S.L., Maslakova, N.I. and Rudakov, S.G., 1983. Cretaceous folding and overthrusts in the East Carpathians. Geotectonics, 17: 133–137.

Bleahu, M., Lupu, M., Patrulius, D., Bordea, S., Stefan, A. and Panin, S., 1981. The structure of the Apuseni Mountains. 12th Congr. Carpatho-Balkan Geol. Assoc. Bucharest, Guide Excurs. 23, 106 pp.

Braud, J., 1987. La suture du Zagros au niveau de Kermanshah. These, Univ. de Paris-sud, Centre d'Orsay, 490 pp.

Burchfiel, B.C., 1976. Geology of Romania. Geol. Soc. Am. Spec. Pap. 158, 82 pp.

Burtman, V.S., 1982. Development of the Pamir–Punjab syntaxis. Geotectonics, 16: 382–388.

Burtman, V.S., 1986. Origin of structural arcs of the Carpathian–Balkan region. Tectonophysics, 127: 245–260.

Burtman, V.S., 1988. Kinematics of the Carpathian–Balcan region in Cretaceous and Cenozoic. Stud. Geol. Pol., 91: 39–60.

Burtman, V.S., 1989. Kinematics of the Arabian syntaxis. Geotectonics, 23: 139–145.

Burtman, V.S. and Molnar, P., 1993. Geological and geophysical evidence for deep subduction of continental crust beneath the Pamir. Geol. Soc. Am. Spec. Pap. (in press).

Chang, Ch. and Pan, Sh., 1984. Preliminary synthesis of the geological structure of the Tibetian Highland. 27th Int. Geol. Congr. Rep., 5: 159–176.

Danilovich, L.G., 1981. Fragments of oceanic crust in the Carpathians. Geol. Zh. (Ukr. Ed.), 41(4): 93–106 (in Russian).

Davies, R.G., Jones, C.R., Hamzepour, B. and Clark, G.C., 1972. Geology of the Masuleh sheet (1:100.000), north-west Iran. Geol. Surv. Iran Rep., 24: 1–110.

Davoudzadeh, M. and Schmidt, K., 1981. Contribution to the paleogeography and stratigraphy of the Upper Triassic to Middle Jurassic of Iran. Neues Jahrb. Geol. Palaontol. Abh., 162(2): 137–163.

Delaloye, M. and Desmons, J., 1980. Ophiolites and melange terranes in Iran: a geochronological study and its paleotectonic implications. Tectonophysics, 68: 83–111.

Dercourt, J., Zonenshain, L.P., Ricou, L.-E., Kazmin, V.G., Le Pichon, X., Knipper, A.L., Grandjaquet, 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. and Biju-Duval, B., 1986. Geological evolution of the Tethys belt from the Atlantic to the Pamirs since the Lias. Tectonophysics, 123: 241–315.

Desmons, J. and Beccaluva, L., 1983. Mid-ocean ridge and island-arc affinities in ophiolites from Iran: paleogeographic implications. Chem. Geol., 39: 39–63.

Dronov, V.I. (Editor), 1980. Geology and Commercial Minerals of the Afghanistan, Vol. 1. Nedra, Moscow, 536 pp. (in Russian and English).

Dronov, V.I., 1986. Stratigraphy of the Jurassic and older volcanogenic-sedimentary sequence in the Bashgumbez, Irikyak, Sedek and Tashdzhilga river basins (Southeastern Pamirs). Dokl. Akad. Nauk Tadzh. SSR, 29(9): 549–533 (in Russian).

Dronov, V.I., 1988. Triassic volcanic deposits in the South-East Pamirs. Dokl. Akad. Nauk SSSR, 303(2): 437–440 (in Russian).

Dronov, V.I. and Andreeva, T.F., 1962. Stratigraphy of the Jurassic deposits of Central and South-East Pamirs. In: Stratigraphy of the Jurassic System. Akad. Nauk Gruzinskoi SSR, Tbilisi, pp. 343–358 (in Russian).

Embey-Isztin, A., 1980. Major element patterns in Hungarian basaltic rocks: an approach to determine their tectonic setting. Ann. Hist. Nat. Mus. Nat. Hung., 72: 19–31.

Enay, R., 1974. Faunes anatoliennes (Amonitina, Jurassique) et domaines biogéographiques nord et sud tethysiens. Soc. Géol. Fr. Bull., Sér. 7, 18: 533–541.

Gamkrelidze, P.D., 1949. Geology of the Adzaria–Trialet Folded System. Akad. Nauk Gruzinskoi SSR, Tbilisi, 508 pp. (in Russian).

Gamkrelidze, P.D. (Editor), 1964. Georgian SSR. Geology of the USSR, Vol. 10. Nedra, Moscow, 656 pp. (in Russian).

Gamkrelidze, P.D. and Gamkrelidze, I.P., 1977. Napes of southern slope of the Great Caucasus. Metsniereba, Tbilisi, 82 pp. (in Russian).

Gasanov, T.A., 1967. Lower Jurassic of the Azerbaidzan (Lesser Caucasus). Akad. Nauk. Azerbaidjanskoi SSR, Baku, 200 pp. (in Russian).

Gasanov, T.A., 1985. Ophiolites of the Lesser Caucasus. Nedra, Moscow, 240 pp. (in Russian).

Geszy, B., 1973. The origin of the Jurassic faunal provinces and the Mediterranean plate tectonics. Ann. Univ. Sci. Budapest, Sect. Geol., 16: 99–114.

Geszy, B., 1984. Provincialism of Jurassic ammonites: examples from Hungarian faunas. Acta Geol. Hung., 27(3–4): 379–389.

Girardeau, J., Marcoux, J. and Montenat, C., 1989. The neo-Cimmerian ophiolite belt in Afghanistan and Tibet: comparison and evolution. In: A.M.C. Sengör (Editor), Tectonic Evolution of the Tethyan Region. Kluwer, Dordrecht, pp. 477–504.

Gocev, P., Kostadinov, V., Savov, S. and Zagorcev, I., 1974. Srednegorie. In: M. Mahel (Editor), Tectonics of the Carpathian–Balkan Region. Geol. Inst., Bratislava, pp. 322–330.

Görür, N., Oktay, F.Y., Seymen, I. and Sengör, A.M.C., 1985. Paleotectonic evolution of the Tozgülü basin complex, Central Turkey: sedimentary record of a Neo-Tethyan closure. In: J.E. Dixon and A.H.F. Robertson (Editors), The Geological Evolution of the Eastern Mediterranean.?, Oxford, pp. 467–482.

Hallam, A., 1975. Jurassic Environments. Cambridge Univ. Press, Cambridge, 269 pp.

Horvath, F., Vörös, A. and Onuoha, K.M., 1979. Plate-tectonics of the western Carpatho-Pannonian region: a working Hypothesis. Acta Geol. Acad. Sci. Hung., 21(4): 207–221.

Hristov, R.N., 1960. Geology of the region between the Luda Jana and the Topolnitsa rivers. God. Minno-Geol. Inst. (Sofia), 6(1–2): 181–202 (in Bulgarian).

Kazmin, V.G. and Knipper, A.L., 1989. Accretionary tectonic of the Caucasus. In: V.E. Khain (Editor), Tectonic Processes. Nauka, Moscow, pp. 75–89 (in Russian).

Knipper, A.L., 1975. Oceanic Crust in the Alpine Folded Belt. Nauka, Moscow, 208 pp. (in Russian).

Knipper, A.L., 1985. Movements of the Tethys oceanic crust during the Austrian phase: their character, results and possible causis. Geotectonics, 19(2): 85–94.

Knipper, A.L. and Khain, E.V., 1980. Structural position of ophiolites of the Caucasus. Ofioliti, 2: 297–314.

Kocyigit, A., 1990. Structural relationships of three suture zones to the west of Erzincan (NE Turkey): Karakaya, Inner Tauride and Erzincan sutures. 8th Pet. Congr. Turkish Ass. Pet. Geol. Proc., General Geology, Ankara, pp. 152–160 (in Turkish).

Kopp, M.L., 1979. Origin of transverse folded zones in eastern part of the Alpine belt. Geotectonics, 13(2): 94–107.

Kozur, H., 1984. New biostratigraphical data from the Bükk, Uppony and Mecsek mountines and their tectonic implications. Acta Geol. Hung., 27(3–4): 307–319.

Krs, M., Muska, P., Orlicky, O. and Pagac, P., 1979. Paleomagnetic investigations in the West Carpathians. In: Geodynamic Investigations in Czechoslovakia. Bratislava, pp. 207–214.

Krumsiek, K., 1976. Zur bewegung der Iranish–Afganischen platte. Geol. Rundsch., 65: 909–929.

Lensch, G., Mihm, A. and Alavi-Tehrani, N., 1977. Petrography and geology of the ophiolite belt north of Sabzever, Khorasan (Iran). Neues Jahrb. Mineral. Abh., 131(2): 156–178

Lomize, M.G., 1968. Late Jurassic volcanism of the East Carpathians. Vestn. Moscow Univ., Ser. Geol., 6: 42–59 (in Russian).

Lordkipanidze, M.B., 1980. Alpine volcanism and geodynamics of central part of the Mediterranean folded belt. Metsnierba, Tbilisi, 160 pp. (in Russian).

Marschalko and Pagac, P., 1980. Preliminary results of paleomagnetic study of varied upper Cretaceous sediments of the Pienin, Manin and Central blocks of the West Carpathians. Contrib. Geophys. Inst. Slovak. Acad. Sci. (Bratislava), 10: 77–83.

Nastaseanu S., Bercia, J., Jancu, V., Vlad, S. and Hartopanu, J., 1981. The structure of the South Carpathians. 12th Congr. Carpato-Balkan. Geol. Assoc., Bucharest, Guideb. 22, 100 pp.

Neumayer, M., 1878. Ueber unvermittelt auftretende Cephalopodentypen in Jura Mittel-Europas. Jahrb. Geol. Reichsanst. (Wien)., 28: 38–80.

Norton, I.O. and Sclater, J.G., 1979. A model for the evolution of the Indian ocean and the breakup of the Gondwanaland. J. Geophys. Res., B84: 6803–6830.

Okay, A., 1991. Geology and tectonic evolution of the Biga Peninsula, Northern Turkey. Istanbul Tech. Univ. Bull., 44: 191–256.

Onuoha, K.M., 1977. Tectonic significance of some geochemical data associated with the ophiolitic complexes of the Darno megatectonic line, NE Hungary. Acta Geol. Acad. Sci. Hung., 21: 133–141.

Orbay, N. and Bayburdi, A., 1979. Paleomagnetism of dikes and tuffs from the Mesudiye region and rotation of Turkey. Geophys. J.R. Astron. Soc., 59: 437–444.

Paffengolts, K.N. (Editor), 1964. Geology of the Armenian SSR, Vol. 2. Akad. Nauk Armianskoi SSR, Erevan, 432 pp. (in Russian).

Panov, D.J., 1978. Regional and stratigraphic description of Lower Jurassic and Aalen deposits of the Lesser Caucasus. In: E.E. Milanovskii and I.A. Dobruskina (Editors), Problems of the Stratigraphy and Historical Geology. Moscowsky Univ., Moscow, pp. 153–164 (in Russian).

Pashkov, B.R. and Shvolman, V.A., 1979. Riftogenic margins of the Tethys in the Pamirs. Geotectonics, 13(6): 447–456.

Pearce, J.A. and Deng, W., 1988. The ophiolites of the Tibetan geotraverses, Lhasa to Golmud (1985) and Lhasa to Kathmahdu (1986). Philos. Trans. R. Soc. London, Ser. A, 327: 215–238.

Pechersky, D.M. and Hguen, T.K.T., 1979. Paleomagnetism of volcanic rocks from ophiolites and Late Cretaceous extrusions from Armenia. Izv. Akad. Nauk SSSR, Ser. Fiz. Zemli, 3: 48–63 (in Russian).

Peive, A.V., Burtman, V.S., Ruzhentsev, S.V. and Suvorov, A.I., 1964. Tectonics of the Pamir–Himalayan sector of Asia. In: Himalayan and Alpine Orogeny. 22nd Sess. Int. Geol. Congr., New Delhi, XI: 441–464.

Poliansky, B.V., 1980. Coal-bearing formations of the Iran and Afghanistan. Litol Polezn. Iskop., 2: 82–98 (in Russian).

Prosorovskaya, E.L. and Vörös, A., 1988. Pliensbachian, Bajocian and Callovian Brachiopoda. In: A.E.M. Nairn (Editor), Evolution of the Northern Margin of Tethys, Vol. 1. Mem. Géol. Soc. Fr., N.S., 154: 61–70.

Rakus, M., 1988. Paleobiogeography of Pliensbachian ammonites in the West Carpathyan mountains. In: A.E.M. Nairn (Editor), Evolution of the Northern Margin of Tethys, Vol. 1. Mem. Soc. Géol. Fr., N.S., 154: 41–46.

Rassios, A., Beccaluva, L., Bortolotti, V., Mavrides, A. and Moores, E.M., 1983. The Vourinos ophiolitic complex. Ofioliti, 8: 275–291.

Ricou, L.-E., 1976. Evolution structurale des Zagrides la region Clef de Neyriz. Mem. Soc. Géol. Fr., N.S. LV(125), 140 pp.

Ron, H., Freud, R., Garfunkel, Z. and Nur, A., 1984. Block rotation by strike-slip faulting: structural and paleomagnetic evidence. J. Geophys. Res., 89(B7): 6256–6270.

Rostovtsev, K.O., 1978. Paleobiogeography of the Caucasus basins in Lower and Middle Jurassic. In: V.A. Maslov and K.M. Hudolei (Editors), Problems of the Paleobiogeography. Ufa, pp. 97–102 (in Russian).

Rostovtsev, K.O. and Azarjan, N.R., 1971. Jurassic deposits of the Nahithevan and South-Western Armenian. Izv. Akad. Nauk SSSR, Ser. Geol., 7: 123–127 (in Russian).

Russo-Sandulescu, D., Sandulescu, M., Udrescu, C., Bratosin, J. and Medesan, A., 1984. Le magmatisme d'age mésozoique dans les Carpathians Orientales. Ann. Inst. Geol. Geofiz. (Bucuresti), 61: 245–252.

Sandulescu, M., Krautner, H.G., Balintoni, I., Russo-Sandulescu, D. and Micu, M., 1981. The structure of the East Carpathians. XII Congr. Carpatho-Balkan. Geol. Assoc. Guide Excurs. 21, 92 pp.

Saribudak, M., 1989. New results and the paleomagnetic overview of the Pontides, Northern Turkey. Geophys. J. Int., 99: 521–531.

Satian, M.A., 1979. Late Cretaceous lithogenesis of the ophiolite zones of the Armenian SSR. Akad. Nauk Armianskoi SSR, Erevan, 168 pp. (in Russian).

Satian, M.A., 1984. Ophiolitic depressions of the meso-Tethys. Akad. Nauk Armianskoi SSR, Erevan, 196 pp. (in Russian).

Sengör, A.M.C., 1984. The Cimmeride orogenic system and the tectonics of Eurasia. Geol. Soc. Am., Spec. Pap. 195, 82 pp.

Sengör, A.M.C. and Yalmaz, V., 1981. Tethyan evolution of Turkey: a plate tectonic approach. Tectonophysics, 75: 181–241.

Sengör, A.M.C., Yalmaz, V. and Ketin, J., 1980. Remnants of a pre-Late Jurassic ocean in northern Turkey: fragments of Permian–Triassic Paleo-Tethys? Geol. Soc. Am. Bull., 91: 599–609.

Sengör, A.M.C., Altiner, D., Cin, A., Ustaömer, T. and Hsü, K.J., 1988. Origin and assembly of the Tethyside orogenic collage at the expense of Gondwana Land. In: M.G. Audley-Charles and A. Hallam (Editors), Gondwana and Tethys. Geol. Soc. London, Spec. Publ., 37: 119–181.

Shvolman, V.A., 1978. Relicts of the Mesotethys in the Pamirs. Himalayan Geol, 8: 369–378.

Shvolman, V.A., 1980. Mesozoic ophiolites in the Pamirs. Geotectonics, 14(6): 465–470.

Shvolman, V.A. and Pashkov, B.R., 1986. Early Cretaceous tectonic zonation of Central Asia. Dokl. Akad. Nauk SSSR, 286(4): 951–954 (in Russian).

Sokolov, S.D., 1974. Tectonic melange in the Amasia region (Lesser Caucasus). Geotektonika, 1: 69–77 (in Russian).

Sokolov, S.D., 1977. Olistostrome and ophiolite nappes of the Lesser Caucasus. Nauka, Moscow, 94 pp. (in Russian).

Stampfli, G.M., 1978. Etude géologique génerale de l'Elburz oriental au S de Gonbed-e-Qabus (Iran NE). Thesis, Dr. Sci. Univ. Geneva 1868, 329 pp.

Stöcklin, J., 1974. Ancient continental margin in the Iran. In: C.A. Burk and C.L. Drace (Editors), The Geology of Continental Margins. Springer, Berlin, Vol. 3, pp. 230–247.

Szepeshazy, K., 1979. Tiszentul es az Erdelyi Közephegyseg (Muntii Apuseni) naguszerkezeti es retegtani kapcsolatai. Altalanos Föld. Szemle, 12: 121–198 (in Hungarian).

Teng, J.W., Sun, K.Z., Xiang, Sh.B., Yun, Z.X., Yao, H. and Chen, L.F., 1983. Deep seismic reflection waves and structure of the crust from Dangxung to Yadan on the Xizang plateau (Tibet). Phys. Earth Planet. Inter., 31(4): 293–306.

Tirrul, R., Bell, J.R., Griffis, R.J. and Camp, V.E., 1983. The Sisten suture zone of Eastern Iran. Geol Soc. Am. Bull., 94: 134–150.

Tüysüz, O., 1990. Tectonic evolution of a part of the Tethyside orogenic collage: the Kargi massif, Northern Turkey. Tectonics, 9: 140–160.

Tüysüz, O., 1993. A geo-traverse from the Black Sea to the Central Anatolia: tectonic evolution of Northern Tethys. Turk. Assoc. Pet. Geol. Bull. (in press) (in Turkish).

Uhlig, V., 1907. Über die Tektonic der Karpaten. Sitzungsber K. Akad. Wiss. Math. Naturwiss. K1, 116: 871–972.

Valbe, S.P., 1967. Stratigraphy, tectonics and prospects for oil and gas of the Western Kopeh Dagh and Kopeh Dagh foredeep. Nedra, Moscow, 164 pp.

Van der Voo, R., 1968. Jurassic, Cretaceous and Eocene pole position from northeastern Turkey. Tectonophysics, 6: 251–269.

Vörös, A., 1977. Provinciality of the Mediterranean Lower Jurassic brachiopod fauna: causes and plate tectonic implications. Palaeogeogr., Palaeoclimatol., Palaeoecol., 21: 1–16.

Vörös, A., 1984. Comparison of Jurassic bentonic mollusc and brachiopod faunas of the Transdanubian mountains. Acta Geol. Hung., 27: 391–401.

Vörös, A., 1988. Conclusion on Brachiopoda. In: A.E.M. Nairn (Editor), Evolution of the Northern margin of Tethys, Vol. 1. Mem. Soc. Geol. Fr., N.S., 154: 79–84.

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

Yilmaz, Y., 1990. Allochthonous terranes in the Tethyan Middle East: Anatolia and the surrounding regions. Philos. Trans. R. Soc. London, Ser. A, 331: 611–624.

Yilmaz, Y., Gözübol, A.M. and Tüysüz, O., 1982. Geology of an area in and around the Northern Anatolian transform fault zone between Bolu and Akyazi. In: A.M. Isicara and A. Vogel (Editors), Multidisciplinary apart to Earthquake Prediction. Viewed, Wiesbaden, pp. 45–65.