# 4.3.15. GEODYNAMICS OF THE BALKAN PENINSULA

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## 4.3.15.1. Introduction

**Geodynamics is a relatively new branch of geology. It emerged during the 60-ies of the last century and is closely related with the formation and development of the new global tectonics. In the course of their development geology and geomorphology in particular, were descriptive and explanatory sciences. The existing or consecutively created during different stages of their development particular theories did not provide a uniform view or general geological conception about the genesis and evolution of the Earth. Although in the investigation of the Earth development the whole complex of knowledge of all significant and reliable relevant facts was used objectively, in their greater part they remained in the sphere of suppositions and often failed in the test of the quantitative evaluations, experiments or observations. On the other hand, it must not be forgotten that the geological phenomena and events proceed in the course of millions of years and cannot be always reproduced in different laboratory experiments and observation. All phenomena and events related to the genesis, evolution and age of the Earth have physical essence and a consistent model of this evolution as a whole or of its single segments and localities could be created using the fundamental laws of physics.**

**On the basis of different reference sources, the main principles of geodynamics consist in the following: the Earth is a physical body, which is developed in accordance with fundamental physical laws; the Earth and its envelopes (crust, mantle and core) represent a uniform geodynamic system; the basic energy sources of the Earth and its tectonic activity are inside the Earth itself; the geodynamic evolution of the Earth has to be obligatorily considered in both time and space together; the chemical and mechanical processes realized in the Earth's bowels are also considered together; the geodynamic activity of the Earth is irregular in time and space and it is presumed that periodicity exists in the main stages of its evolution.**

**In the light of the above ideas and formulations, an attempt is made to consider the evolution of the Balkan Peninsula as a part of the Eastern Mediterranean area during the neotectonic stage but within its broader range from the Paleogene till now.**

## 4.1.15.2. General ideas about the subduction-collisional conditions in the Eastern Mediterranean area

**The assimilation of the ocean lithosphere in the subduction zones represents the main mechanism of space reduction between the continental plates in the course of their opposite movement. The observations in the Eastern Mediterranean area show that some segments of the subduction belt continue their development after the beginning of the intercontinental collision too. The subduction continues only in these sections, which turn out to be closed between the indenters (gibbosities) of the sinking continental plate, where relict oceanic basins are preserved. However, in these cases the subduction**

displays another type of development to a significant extent. The specific features of the subduction in the relict basins of the intercontinental orogenic belt are determined according to a number of researchers by three main reasons: rearrangement of the subduction kinematics, the impact of the adjacent collision structures and the restricted space where the subduction occurs. The consequences of these effects are expressed in tectonics, metamorphism and magmatism (Dewey, 1980; Lomize, 1988, Yilmaz, 1993, Robertson, 2000, Keay, Lister, 2002, Jolivet et al., 1998, Lomize, 2004 and others). Most generally the pre-collision history of the Mediterranean area during the Mesozoic and the Early Cainozoic is very well studied and well-known. It had been developed between the passive African-Arabian periphery to the south and the active Eurasian edge to the north, where the subduction was directed and vast accretion belt was created. The successive reduction of the ocean had led to the situation that in the end of the Middle Eocene the Arabian indenter (gibbosity) reached this accretion belt (Yilmaz, 1993) starting the intercontinental collision that separated the west part of the Tethys together with the subduction zones along its northern border. In this way the contemporary Mediterranean Sea was formed, which is a typical relict ocean basin (Fig. 4.3.15.1.).

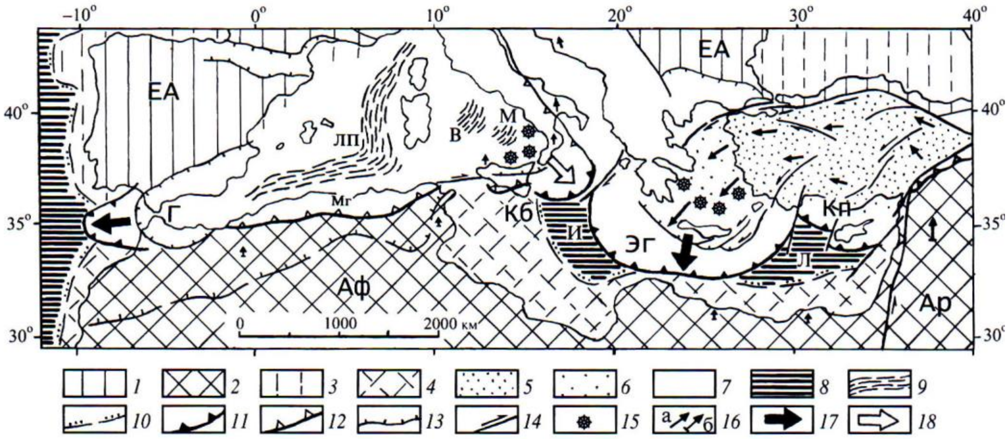


**Fig. 4.3.15.1. Contemporary tectonics of the Mediterranean Sea: conclusive stage of closing the Tethys ocean basin and forming the intercontinental orogenic belt. According to Lomize, 2004**

**Subduction zones: Г- Gibraltar, Кб – Calabrian, Ег – Aegean, Кр – Cyprus. Lithospheric plates: ЕА – Eurasian, Аф – African, Ар – Arabian. 1, 2 – continental borders of the Alpine orogenic belt: Eurasian (1) and African-Arabian (2); 3, 4 – the same but under the contemporary seas; 5, 7 – the orogenic belt, including the Aegean-Anatolian plate on the land (5) and under the sea (6); 8 – relict basins of the ocean crust of the Mesozoic Tethys (И – Ionic, Л – Levantian); 9 – newly formed ocean crust in the areas of the sub-arc spreading in the Cainozoic basins: Algerian-Provance (А–П), Vavilov (В) and Marseilles (М); 10 – passive continental peripheral parts; 11, 12 – subduction zones: active (11) and with completed development (12); Мг – Magrebian fragment of the Western Mediterranean paleozone of subduction;**
**13 – thrusts; 14 – reverse faults; 15 – active subduction volcanism; 16 – horizontal movements with respect to Eurasia according to GPS (a) and NUVEL1 (b), the vector of the Arabian plate (2.5 cm/yr) serves as a scale; 17, 18 – recoil of the subduction zone: continuing (17) and ceased (18).**

### 4.3.15.3. The subduction in the Eastern Mediterranean area and its connection with the Arabian indenter

**The subduction in the Eastern Mediterranean area was developed much more complicated and in close connection with the Arabian indenter, which exerted impact for realizing the collision compression and longitudinal movement of the Earth's crust (tectonic escape) in west direction from the Anatolian plate. Two subduction zones are known in the Eastern Mediterranean: the Cyprus and the Aegean one. According to a number of researchers the subduction zone is under a process of elimination. Under the conditions of a closed oceanic basin, with passive subduction, it is found in a wedged state. At present, tearing away of the slab is observed, followed by means of tomography at a depth of about 130 km (de Boorder et al., 1998,Papazachos, Papaioanou, 1999).**

**The seismic activity is also observed to such depth, i.e. the Cyprus zone is fading away, regardless of the fact that before it ocean lithosphere is still observed at the bottom of the Levantian depression (Ben-Avraham et al., 2002, Khair, Tsocas, 1999).**

**The development of the Aegean zone is much more complicated and full. Data about it are known from the Middle Eocene. Its bedding is preceded by fading away of the subduction zone, situated significantly further to the north. Its traces are found in the Sredna Gora – Pontian volcanic belt. The subduction there faded away and stopped by the gradual annihilation of the northern branch of the Tethys Ocean. The ocean annihilation was accompanied by adjoining of micro continents of Gondwane genesis (African ones). The following micro continents were consecutively adjoined from east to west: the Nahichevani one - after Lower Coniacian, Sakar – during the Turonian and the Coniacian, and the Rhodopean – during the Campanian and the Maastrichtian (Lomize, 1998, Lomize, 2004). Later the subduction was resumed to the south of these continental units by insertion in the end of the accretion belt. In this way the Aegean subduction zone was formed, which caused the Eocene (40-45 million years) metamorphites of the "blue schist unit of the Cyclades". The age was determined according to U-Pb datings of circonia (Keay, Lister, Keay et al., 2001, Ring et al., 2001).**

### 4.3.15.4. Brief characteristics of the paleogeodynamic circumstances during the Upper Cretaceous – Lower Eocene

**This stage was started for Bulgaria in the beginning of the Cennomanian with subduction of the ocean crust to the north under Eurasia, respectively under the Balkan micro continent, with the formation of Turonian-Campanian volcanic island-arch system with frontal Macedoniam-Rhodopean arc, internal arc basin (Sredna Gora one), Balkanide sub-arc, and sub-arc basins (Emine, Kula) and an epicontinental sea in the Moezian platform (Nachev, 2002, Dabovski et al., 2002). According to the basin analysis, on the base of grouping of various-type sediment circumstances, the paleogeodynamic development took place temporally during the Cenomanian under limnic circumstances with deposition of alluvial, lacustrine and boggy sediments. In the end of the Cenomanian and Lower Turonian, an epicontinental sea was formed with litoral and nerite zones, in which conglomerates, sandstones, limestones and delta material were deposited. Since the Upper Turonian the inter-arc trough formation was started and continued during the Coniacian, Santonian and Campanian with active effusive and intrusive magmatism. Some authors assume that about 50 volcanoes and about 52 plutons existed in the volcanic arc on Bulgarian territory (Nachev, 2004). The volcanism was andesite-basalt, basalt, trachibasalt to trachite one. The trough was annihilated by folding, collision of the arc-type (an arc and rising before the**

**Maastrichtian as a result of the sub-Hercinian orogenesis). According to other authors the folding was realized in the end of the Maastrichtian during the Laramian tectonic phase (Dabovski et al., 2002). It has been established from the last investigations on the basis of calcareous nano plankton in the East Stara Planina, that the folding there was realized in the end of the Middle Eocene as a result of the Illyrian folding (Sinyovski & Sinyovska, 1988; Dzhuranov, Pimpirev, 1989; Sinyovski, Sultanov, 1994, Sinyovski, 2004, Sinyovski, 2005). The latest data of these authors for the East Stara Planina show that the Emine Flysch Formation was gradually transformed from the Dragonovo Volcanic Formation and its lowest boundary is of Upper Campanian age, while its upper boundary represents a rapid transition to the rocks of the Dvoynitsa Formation and is of Upper Paleocene age. The lower levels of the Dvoynitsa Formation are with the same age as that of the Emine Formation and lateral transitions exist between them at some places, this overlapping being within the range of the whole Paleocene series. Except for the Paleocene, the Dvoynitsa Formation comprises also Lower Eocene series and the base of the Middle Eocene. The Obzor Formation, which is an Upper Eocene one, is situated on top of it with washing away. The data from other investigations, mostly concerning the age of different sediments in South Bulgaria and Central Stara Planina, show that after the drying up of the Sredna Gora trough before the Maastrichtian, already under continental conditions, alluvial, deluvial-proluvial sands and conglomerates of gneisses, amphibolites, granites, pegmatites and Upper Cretaceous (Coniacian-Campanian) sediments, volcanites and plutonites were deposited in various lowerings (Goranov, Atanasov, 1992, Zagorchev et al., 2001a, Zagorchev et al., 2001b, Dimitrova et al., 2001, Boyanov, Goranov, 2001, Okay, 2001, Pavlishina, 2002, Boyanov et al., 2003) and only in the SE Rhodopes foraminifers were encountered in marine coarse terrigenic deposits, which are included in the range of the Upper Paleocene – Middle Eocene of the Southern Tethys. This information is complemented by data from the Kraishte magmatic-tectonic zone for the presence of magmatites of radiogeological age between 47.4 and 42.2 + 1.60-1.80 Ma, which fall within the range of the Lutecian and Bartonian (Lower-Middle Eocene) (Harkovska et al., 2004). The Kraishte magmatic-tectonic zone represents the southern end of a regional linear, more than 250 km long, analogous arc formed due to Late Alpine (Late Cretaceous – Early Tertiary) subduction processes on the Balkan Peninsula. This arc is fixed to the erosion boundary of the Middle Cretaceous thrusts described in East Serbia as "Ridan-Krepolinski" belt, and in SW Romania as "Western Banatite zone". In addition, according to the discrimination diagrams the volcanites of the Kraishte magmatic-tectonic zone (KMTZ), it is proved that they are not typical collision magmatites but are rather the products of orogenic magmatism. On the basis of the above described specific features the authors consider that KMTZ does not belong to the regional Macedonian-Rhodopean-North Aegean Magmatic Zone (MRNAMZ), which is presumed to have more probable collisional genesis. It may be pointed out that such volcanites are also observed on Greek territory as far as the Halkidiki Peninsula (Falalakis et al., 1995), the arc reaching a length of up to 350 km. The Romanian geologists have arrived to similar conclusions, distinguishing also four main stages of magmatic activity in the interval Upper Cretaceous – Eocene in the Southern Carpathian Mts. (Cioflica et al., 1995; Ciobanu et al., 2002). The first stage ranges from the Turonian till Early Maastrichtian (91-68 Ma, K-Ar age), the second stage is of Late Maastrichtian age, the third ranges from the end of the Maastrichtian to the Paleocene (68-60 Ma, K-Ar age) and the fourth – from the Paleocene to the Lower Eocene (65-43 Ma, К-Ar age). The Upper Cretaceous – Lower Eocene magmatites in the western part of the South Carpathians were generated by the Alpine multi-stage subduction processes of the Vardar-Izmir-Ankara**

zone. The data shown above prove that the liquidation of the troughs related with the Vardar-Izmir-Ankara subduction zone occurred gradually within the interval Upper Cretaceous – Lower Eocene. If we assume that the subduction was completed in the end of the Senonian, after the Campanian, with the liquidation of the Sredna Gora volcanic arc, then the continuing convergence of the Eurasian and African plates, according to the existing data, was completed as early as in the Lower Eocene with the known Illyrian radical change with closing the Emine trough by folding, thrusting and respective rising.

The Rhodopes massif occupies a special place in the Upper Cretaceous – Lower Eocene development of the Balkan Peninsula. It is most often considered as a micro continent of Paleozoic or pre-Cambrian type but recently a number of scientists have expressed doubts concerning this opinion. According to (Ivanov, 1989; Arnaudov et al., 1988, 1989; Arnaudov et al., 1990a, 1990b) it is supposed that the petrological-geochemical and lead-isotope data for the migmatic formations in the migmatite biotite gneisses in the Central Rhodopes and the U-Pb age according to circonia is within the range of 62-32 Ma. A bit later, again on the basis of data for circonia and ortites from migmatic pegmatites, the age of 58-49 Ma was confirmed by two laboratories. These data provide evidence about the existence of a uniform ultrametamorphic in Late Alpine time. The metamorphism was accompanied by synmetamorphic south-vergent thrusting and linearity (Ivanov, 1989, Burg et al., 1990, Dimov, et al., 1996, Dimov, Damyanova, 1996). Other authors have also arrived to similar conclusions (Peycheva et al., 1992a, 1992b, Mposkos, Wawrzenitz, 1995, Peycheva, Quadt, 1995 and many others). Later, blue schists were found in the central Rhodopes on the North Rhodopean anti-form (Guiraud et al., 1992) in metapelites of the Narechen Formation near the Byala Cherkva village. The authors consider that the found parageneses in the mica schists were formed at T = 550+25 Co and P = 13+2 kbar. The mica schists are rocks, formed under high pressure, and are related with the obduction of the Parvenets metaphiolites, observed along the northern slope of the Rhodopes and in many places in the east Rhodopes. This fact is an additional argument supporting the synmetamorphic thickening of the lithosphere as well as that the depth of the mica schists formation is at least 30 km. Undoubtedly the numerous evidences provided by a number of authors show that the main compression was realized in the end of the Campanian or during the Lower Maastrichtian and led to the annihilation of the Sredna Gora trough. According to Nozharov et al. (1984), as a result of this compression, the vast sea was probably shortened and moved 300-400 km to the north. According to some authors (Lomize, 1998; Lomize, 2004) the fading away of the Tethys Ocean subduction was preceded by obduction of ophiolites. For example, the obduction in the Small Caucasus ophiolitic suture occurred between the Lower and Upper Coniacian, forming the Vedin and Sevan ophiolitic alochtone zones, during the Turonian and the Coniacian – in the Anatolian area, and during the Campanian-Maastrichtian – in the southern part of the Balkan micro continent. It may be pointed out that subsequently from the east to the west the Small Caucasus – Internal Pontide northern branch of the Tethys was gradually closed, leading to the annexation of micro continents of Gondwane origin to the active Eurasian plate – first of the Nahichevani, then the Sakar, and finally the Rhodopean. The continuation of the Internal Pontide suture should be sought for in the East Rhodopes and on the North Rhodopes slope and finally – provably in the southern strip of the West Sredna Gora area (Gochev et al., 1970, Ivanov, 1989, Guiraud, 1992), (Fig. 4.3.15.2.).

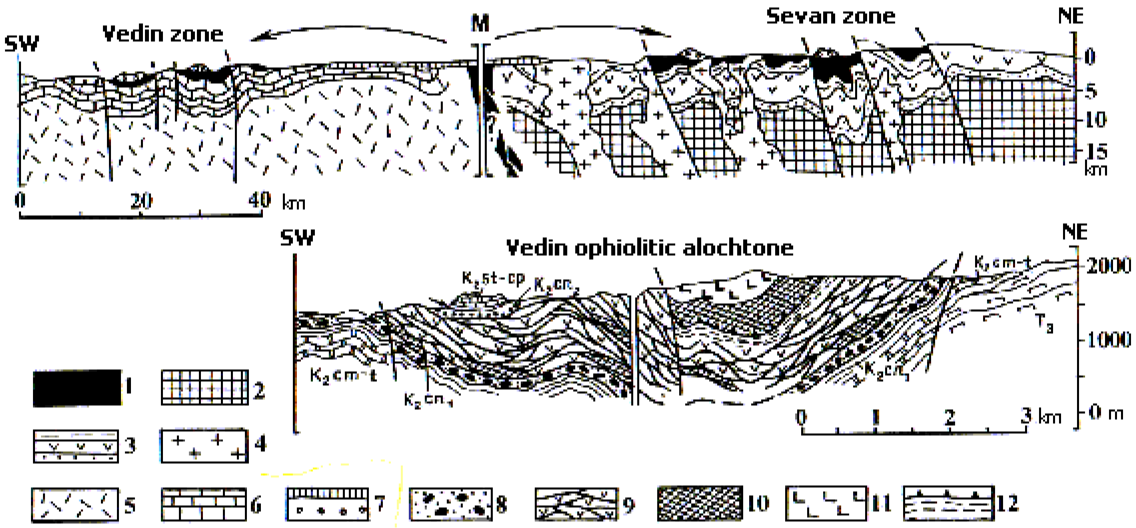


**Fig. 4.3.15.2. Ophiolites in the Small Caucasus suture. In the top profile the Small Caucasus suture (M) is observed together with the thrusted ophiolites of the Vedin zone (shown in the lower profile) and the Sevan zone. According to Lomize (1988)**

**1-7 in the top profile: 1 – ophiolites and ophiolite mélange, fragments of the oceanic lithosphere of the northern branch of the Mesozoic Tethys; 2 – end of the Eurasian continent, pre-Mesozoic base; 3 – sediment formations and peripheral-materic volcanic belt, Jurassic-Cretaceous (to Lower Coniacian); 4 – granitoids; 5 – the end of the Nahichevani micro continent with Gondwane genesis, pre-Cambrian base; 6 – sediment cover of the Paleozoic-Mesozoic (to Lower Coniacian); 7 – an alochtone deposited on the Upper Coniacian-Paleogene and Neogene-Pleistocene volcanites; 8-12 – ophiolitic alochtone of the lower profile: 8 – ophiolitic clastic olistostromes of the Lower Coniacian in the base of the alochtone; 9 – ophiolitic mélange, mainly basalts and radiolarites; 10 – serpentinized peridotites; 11 – gabroids; 12.-«basal metamorphic halo» in the base of the peridotite-gabro alochtone unit.**

**The above figure is presented to show that the subduction was not transformed to collision simultaneous along the entire convergence boundary because the ends of the colliding continents were rheologically inhomogeneous and had irregular outlines in plan. For this reason the collision is always asynchronous. While on one segment of the convergence boundary where the continents collided with their gibbosities, collision had already started, on other segments, where relicts from oceanic lithosphere from previous oceans were found, the subduction continued. Such circumstances – intermediate between subduction and collision, are combined in the name of accretion. Usually heterogeneous tectonic elements (micro continents, island arcs, oceanic crust and peripheral seas) of different age fell into the space between the colliding continents. All these relatively heterogeneous blocks and terrains, emerged at different places and were united only in the time of the collision. For this reason each collision zone is heterogeneous and heterochronous tectonic mosaic of tectonic elements with different genesis. Such circumstances of accretion and collision on contemporary Earth were displayed within the range of the Alpine-Himalayas folded belt stretching from the Atlantic to the Pacific Ocean. The subduction on the Balkan micro continent continued from the Upper Turonian till the end of the Campanian or Lower Maastrichtian, with manifestation of sedimentation with thickness generally exceeding 4000 m. It was also accompanied by effusive-explosive ca- to K-alkaline subduction island-arc volcanism. The duration of this active stage continued from 90, 0 -73 Ma or about 17 million years and in the beginning of the Maastrichtian the Sredna Gora trough was annihilated by**

**the closing and folding and for this reason the Maastrichtian sediments everywhere in South Bulgaria lie in angular and structural discordance on the Coniacian-Campanian rocks. This tectonic fact occurred after the heavier oceanic lithosphere on the plate convergence boundaries had been already assimilated during the subduction and the continental parts of the lithospheric plates meet directly on the convergent boundaries. Then their mechanical interaction – the collision, really began. The velocity of the opposing movement of the convergence boundary in the transition process from subduction to collision was reduced but it was not immediately stopped. According to some calculations this transition lasted up to 20 million years (Aplonov, 2001) and beneath the collision zone continued the existence of the cool and dense end of the subducing oceanic plate or slab. This intraplate subduction lasted till the slab had been entirely melted and sunk in the astenosphere due to its higher weight. From this moment the subduction magmatism in the collision zone was completely stopped and the terrain above had started its sharp rising due to the fact that it was released from the depth loading of the slab oceanic crust, forming mountain structures. In the considered case the detachment of the slab occurred during the Lower Eocene if we use the data of Harkovska et al. (2004) at the age of 47.4- 42.2 Ma or from 65 to 43 Ma according to Cioflica et al. (1995).**

### 4.3.15.5. Paleogenic subduction, recoil and opening of sub-arc basin

**During the Late Eocene – Early Oligocene the Aegean subduction zone was formed in the end of the Eurasian peripheral parts and in connection with this zone a volcanic-pluton belt was created with lateral petrochemical composition that spread from the Rhodopes to West Thrace (37-30 million years, Lilov et al., 1987). Andesites of in-depth genesis were predominant among the volcanites and they were accompanied by high-alkaline riolite ignimbrites in secondary intracore focies (Yanev el al., 1995). From the beginning of the Upper Oligocene recoil of the continental periphery subduction zone was started and island arc formation began with an opened sub-arc basin in its rare. According to Sapundzhieva and Dragomanov (1991) on Bulgarian territory between the Lower and Middle Oligocene (the three-part division of the Oligocene for the Tethys has been accepted for the case) interruption of the connection between the Upper Thracian Paleogenic basin and the Tethys was realized, because only fauna typical for the Paratethys was established in the Ezerovo Formation. The latter lies with washing away above the limestone group of late Eocene – Lower Oligocene age. The Ezerovo Formation is referred to Middle-Upper Oligocene according to faunistic data. Or if we use the new international stratigraphic scheme of the Tethys, this refers to the Lower – Upper Oligocene boundary (Gradstein et al., 2004). The authors accept that the lack of karst phenomena and ferrization between the Early and Middle Oligocene with well-expressed washing away is due to a clearly exhibited tectonic event and freshening of the basin. Other researchers (Dimitrova et al., 2001, Boyanov, Goranov, 2001) have also arrived to similar conclusions.**

**To the south of this strip only volcanites of Early and Middle Miocene age (from 23 to 11 million years, Jolivet et al., 1998) are observed. The geochemical and isotope characteristics of the calcium-alkaline rocks of the Biga and Kozak peninsulas provide evidence for mantle origin, crust contamination of the magma and island-arc specialization (Yilmaz et al., 2001). At the same time in front of the subduction belt volcanism and the subduction front of the Atika-Cyclades zone of tectonic-metamorphic broadening under the conditions of high pressure and low temperature (10 kbars, 350-400 $^{o}$C, 24-21 million/yr according to 40 Ar/39 Ar and Rb/Sr), which at present outcrop at the surface of the Crete and Cyclades islands (Ring et al., 2001). In this way the**

**volcanic arc had been removed still further from its subduction focus, which had led to its complete fading away in the end of the Miocene.**

**Under these new paleotectonic circumstances the magmatism shifted still further to the south form the Atika-Cyclades area and the youngest volcanic arc was established, which had been active till present days (10.5- 0 million/yr). The northern Crete trough – an active structure of extension restricted by faults, with sharply reduced thickness of the continental crust (about 15 km) was formed in front of it. The North Crete trough predetermines the possibility of subsequent jumping over a volcanic arc, since nowadays it is at a distance of 400 km from the front of the accretion belt and 250 km from its internal end (Lomize, 2004). It could be also added that the vast accretion prism of the Eurasian plate, expressed in the relief of the Mediterranean bank continues its growth and retaining up to 60 % of the sediment material, it has already reached the Kirenaika protrusion of the African plate (Limonov, 1999; Kopf et al., 2003). The migration of the southern front on the basis of the island arc recoil average velocity from the Late Oligocene till now is about 2 cm/yr. According to GPS data, at present the central part of the Aegean arc continues its movement with an average velocity of about 3 cm/yr (Kahle et al., 2000).**

**Although the total recoil does not exceed 500 m, it has not been accompanied by sub-arc spreading and according to this specific feature the Aegean subduction zone differs form the west Sredna Gora area. The continental lithosphere is preserved for all sub-arc basins and its thickness never exceeded 25-30 km. Certain explanation of this phenomenon might be sought for only in the specific conditions of relict, closed basins with taking into account the collisional circumstances of the Aegean zone development during the Tertiary. According to a number of researchers, the shifting along viscous normal faults was the principal mechanism of extension of the continental crust under the Aegean basin till the end of the Miocene. Such movements were initially started at a relatively big depth of about 50 km as early as during the Upper Eocene. The explorations carried out on the Syros and Syphnos islands form the Cyclades group using Rb-Sr, Ar-Ar dating and track thermo chronology provide evidence that already in the late Oligocene (30 million yr) movement was realized at a depth of 20 km and later at a bit smaller depth (Trotet et al., 2001). During the Late Miocene and the Pliocene-Pleistocene the main extension structures were the normal faults and numerous grabens were formed with sub-latitude direction, situated along the basic arc length and along its periphery (Burchfiel et al., 2000, Jolivet et al., 1998, Walcott, White, 1998). The analysis of the internal crust seismic foci in the Aegean basin shows meridional extension and the evaluation of the present stressed state of the Earth crust according to the GPS horizontal velocities distribution exhibits contemporary extension only exclusively on the Asia Minor side of the basin (Kahle et al., 2000). It is pointed out that both the Earth's crust and the mantle have equal participation in the case of the basin bottom extension. It may be additionally mentioned that during all the stages of the growth of the Aegean basin till now, the meridional tensile stresses have stable prevalence, which is in agreement with the gravitational recoil of the subduction zone (Hatzfeld et al., 2001).**

**According to tomographic data the slab of the Aegean zone in the beginning is with a slanting slope (about 12 $^{o}$) to a depth of 100-150 km and then is sharply bended at an angle of 45$^{o}$, then 50$^{o}$ and after crossing the boundary with the lower mantle – to about 65$^{o}$ (Bijward et al., 1998). Under the western flank of the arc at a depth of about 200 km, rupture in the slab is observed (Spakman et al., 1989). The Aegean slab is distinctly separated from the Cyprus one (de Boorder et al., 1988). Despite of the great thickness**

**of the subducing lithosphere (mainly Early Mesozoic) the seismic foci of the Beneuf zone end at a depth of about 180 km (Papazachos et al., 2000). It is presumed that this is a consequence from the specific kinematics in subduction in the closed Mediterranean basin. The subduction was realized mainly on the account of the gravitational recoil of the slab (Fig. 4.3.15.3.).**

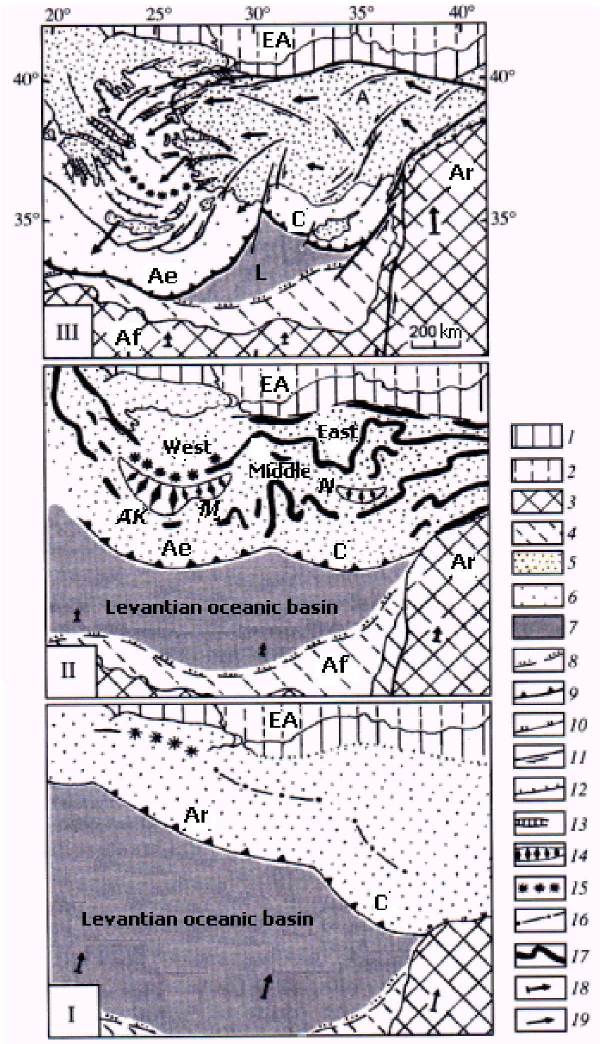


**Fig. 4.3.15.3. Development of the subduction zones in the East Mediterranean area: recoil of the island arcs combined with collisional shifting of the Earth's crust under the action of the Arabian indenter (according to Lomize, 2003)**

**I – active continental periphery of the closed basin, separated from the Tethys ocean in the beginning of the collision (end of early Oligocene, 30 million yr); II – Aegean (Ee) and Cyprus (C) island arcs in the completed stage of mega collision (end of Middle Miocene, 10 million yr); III – the same, for stages of solid collision (at present); Lithosperic plates: EA – Eurasian, Af – African, Ar – Arabian, A – Anatolian. Oroclines of the orogenic belt: East, Middle, West. 1, 3 – continental border of the Alpine orogenic belt: Eurasian (1), African-Arabian (3); 2, 4 – the same but under the contemporary seas; 5 – Late Mesozoic – Cainozoic accretion belt of Eurasia; 6 – the same but under the contemporary seas, 7 – basins on the oceanic crust (L –Levantian depression); 8 – passive continental periphery; 9 – subduction zones, 10 – thrusts, 11 – reverse thrusts, 12 – normal faults, 13 – grabens of extension, including of the pull-apart type; 14 – areas of continental crust growth during its metamorphic and tectonic transformation above the subduction zones /AK – Atika-Cyclades, M – Menderes, N – Nide; 15,16 – belt of subduction volcanism (15) and its supposed continuation (16); 17 – ophiolites; 18 – vectors of the horizontal movements according to the rotation parameters of the lithosperic plates (De Mets et al., 1990,**

**Savostin et al., 1986), the vector of the Arabian plate in map III (2.5 cm/yr) serves as a scale; 19 – contemporary movements according to GPS data (Kahle et al., 2000), the scale is the same.**

**In their simultaneous development with the Arabian synthax, the subduction zones of the East Mediterranean area were subjected to its significant influence. The contemporary kinematics, confirmed by GPS measurements is very well known and concerns the horizontal movements of the Asia Minor block (Kahle et al., 2000, McClusky et al., 2000). The latter shifts to the west and rotates to the south under the action of the Arabian indenter and the small fault tectonics accompanying it. To the west the Anatolian plate is directed to the area of the sub-arc extension of the retreating to the south Aegean subduction zone, compensating to a certain degree this extension. This model is based on the investigations of many researchers, starting with Dui, Chengore, Le Pisson, developed 20 years ago. It has to be noted that this kinematics starts from the Middle Miocene and continues to the present day and is controlled by the North Anatolian and other regional reverse faults (Barka et al., 1996, Hubert-Ferrari et al., 2002). The beginning of the collision with the Arabian indenter was started as early as during the Late Eocene (Meijer, Wortel, 1999, Yilmaz, 1993) and a number of researchers study the collision deformations and movements for these 30 million years starting at this point. The collisional movement of the Arabian indenter is more than 500 km, half of it being realized during the Upper Miocene (Savostin et al., 1986, Jolivet, Faccenna, 2000). An exact answer to all these issues is provided by the internal collisional structure of the Anatolian block, where enormous horizontal folds have been established on the flexures of the ophiolitic sutures, which were formed during the different stages of the collision during longitudinal shifting (pushing) of the Anatolian plate to the west of the Arabian indenter (Lomize, 2000). The most expressive flexure is that of the Izmir-Ankara-Erzidzhan ophiolite suture. In its central segment the left-sided Ankara sigmoid is observed. The ophiolitic sutures in the Anatolian mark generally three big oroclinal bendings. The east orocline is convex to the south in the Tavar folded mountain system. The Ankara sigmoid is observed in its core. The west orocline is convex to the south and is situated in the Aegean region, only its eastern part being included in the range of the Anatolian plate. The middle orocline in fact connects these two folds and is directed to the north, and the “bending of Isparta" is observed in the core. These three flexures were noticed as early as 50 years ago first by Brunn (1960), who supposed even then their flexure was due to the compression of the Arabian plate. The spatial connection of the main oroclines of the Anatolian with the subduction arcs is obvious – of the eastern with the Cyprus one, and of the western with the Aegean one. It is supposed that the orocline formation in the Anatolian was determined by the Anatolian block movement under the pressure of the Arabian plate, facilitated also by the lack of collision compression in the fore-arc volcanic arcs. The enormous accretion belt of the Tethys, composed by several inhomogeneous and non-consolidated continental terrains, was really moved. There are not many data about the time of the horizontal fold formation. The investigations on the structures around the Chankar basin, to the west of Ankara, provided certain information (Kaymakci, 2000, Kaymakci et al., 2003). The authors studied the orientation of the vectors of residual magnetization of the rocks with Lower-Middle Eocene age, for the Upper Eocene – Oligocene, for the Middle Miocene and Upper Miocene. As a result of the investigation they arrived to the conclusion that the flexuring of the structures around the Chankar basin occurred under the effect of the neighbouring Kirheshir crystalline block. Lomize (2004) supposes that the flexure was related to the general movement of the whole Anatolian block under the pressure of the Arabian indenter. Paleomagnetic data exist also for the “Isparta curvature”, where full remagnitization of rocks occurred 20-15 million years**

**ago (Burdigal-Langian) and in consequence this structure was rotated at $30^o$ counter-clockwise (Morris, Robertson, 1993).**

**There are grounds to presume that after the start of the intercontinental collision in the late Eocene – Oligocene, the formation of the west orocline was closely connected with the development of the Aegean subduction zone and of the east orocline – with the Cyprus one. Their spatial position shows that each ofthem may be regarded as an integral system, which was formed during the joint action of the gravitational recoil of the subduction zone with the formation of an arc and its combination with the orocline flexuring, determined by the width shortening of the Anatolian plate under the pressure of the Arabian indenter.**

**For the Aegean arc (west orocline) the width shortening of the space in meridional direction, measured by the flexure of the Vardar-Izmir-Ankara ophiolite suture, exceeds 250 km. The shortening is commensurate with the velocity of southern migration of the volcanic front (1.5-2 cm). The paleomagnetic data show flexure in clockwise direction to the south in the NW part of the Aegean basin and counter-clockwise movement in the NE part. The magnitude is respectively $30^o$ and $19^o$ for the Miocene (25-3 million yr) and $10^o$ and $15^o$ for the Pliocene and the Pleistocene (Walcott, White, 1998). It could be assumed that the collision compression with bringing closer the flanks of the Aegean arc had shortened the sub-arc space during the time of its meridional opening, so contributing to the continental crust and its thickness preservation. The orocline deformation could explain the curvature to the south in the distribution of the Early-Middle Miocene volcanites in the Atika-Cyclades volcanic arc. The eastern flank is even oriented in meridional direction (Muhin, 1999, Lomize, 2004).**

**The orocline curvature of the Aegean arc under pressure from the east was probably also determined by the linear synmetamorphic folding of the Atika-Cyclades area. As already mentioned, this area spreading before the volcanic front of the island arc was formed during the Early-Middle Miocene as a result of heating and metamorphism of the Earth's crust. It represents mechanically weakened zone between the volcanic and non-volcanic island arc. For this reason, under the lateral compression the plastic material could be moved from the flanks toward the apex of the Aegean orocline, where linear folds were formed drawn in meridional direction (Lomize, 2003).**

**Later, from the beginning of the Upper Miocene, the compensating movements from the east with respect to the opening of the sub-arc basin and under the pressure of the Arabian indenter became fault ones. They were expressed in the already consolidated Anatolian plate at that time by the formation of numerous sub-latitude grabens along its western periphery, but already under circumstances of sub-arc extension.**

**Even the first researchers of the Mediterranean area had noticed some specific features of the subduction zones, which made them significantly more different compared to these typical for the analogous oceanic zones. They have smaller sized and more complicated structure and many other geological and geophysical specificities. Nowadays these differences are studied in much greater detail and they can be better understood and more thoroughly interpreted. The newest data for the development of the subduction zones of the Mediterranean Sea show that their specifics is determined by three major reasons: a) rearrangement of the subduction kinematics due to the abruptly decreased convergence rate of the subjected to collision lithosphere plates; b) effect of collision deformations occurring in the neighbouring sections of the intercontinental plicative belts; c) restrictions in the space, in which subduction was developed and hence restricted resources of the assimilated oceanic lithosphere.**

### 4.3.15.6. References


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