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Geological-Geomechanical Simulation of the Late Cenozoic Geodynamics in the Alpine-Mediterranean Mobile Belt

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1. Introduction

Alpine-Mediterranean Mobile Belt, which currently ongoing to develop, is one of the best sites for studying of geodynamic mechanisms for formation of such regions. In this case we are dealing with "alive" tectonomagmatic processes, which are reflected in neotectonics, topography, geophysical fields, and the present-day magmatic activity. Last circumstance allows independent control of petrological processes in the underlying mantle of the belt. Comparison of geological-geophysical data available with the results of mechanic and mathematic simulation allows us to establish the relationships of all these processes and the character of their manifestation on the Earth's surface. This is the purpose of our study.

2. Geology and petrology of Alpine Belt

Alpine-Mediterranean Mobile Belt (Alpine Belt) represents the western part of the huge Alpine-Himalayan collision zone, which appeared in the late Cretaceous-early Paleogene after closure of the Tethys Ocean. The suture of this neotectonic zone is traced by chain of late Cenozoic andesite-latite volcanism, stretching across Eurasia from Mediterranean to the Indonesian Island Arc and back-arc seas of the Western Pacific as well as areas of continental rifting and areas of intraplate basaltic volcanism.

The most complicated structure of this belt is in its west, in the Alpine segment (Fig. 1), where there is a whole system of mountain ridges, andesite-latite volcanic arcs and back-arc basins with thinned crust of intermediate to oceanic-type (Alboran, Tyrrhenian, Aegean Sea, and Pannonian Basin) occurs. Despite the differences in the morphology of these structures, they have several common features: along their periphery volcanic arcs and fold-thrust belts which form arc-shaped mountain ridges are developed. Among their thrust slices are often observed deep-water sedimentary rocks of Tethys, ophiolitic complexes, and sometimes blocks of the lower crust and upper mantle. In general, the situation in many aspects is similar to that which takes place on the active margins of continents and oceans. Such structures are characterized mainly for the West

Mediterranean, while for the Eastern Mediterranean, as well as for the Black and Caspian seas typical passive margin. For this reason, we divide the Alpine Belt into two segments: the eastern, or the Aegean-Caucasian, and western, or proper Alpine, which will be considered separately.

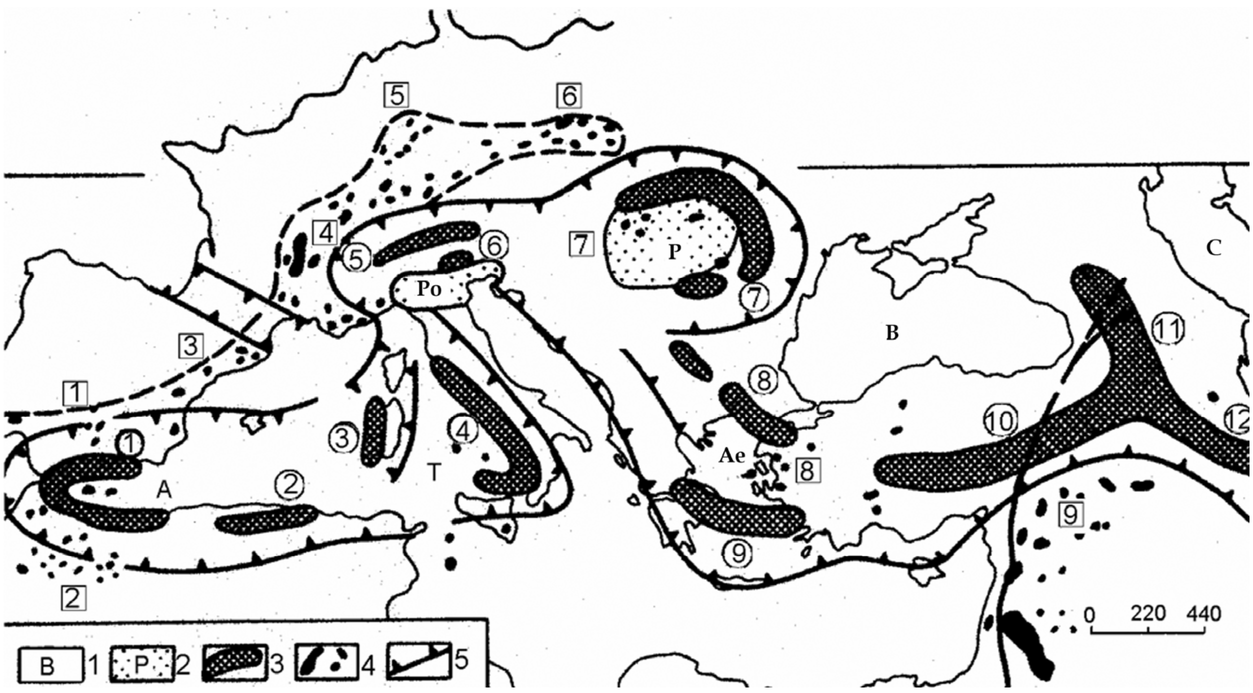


Fig. 1. Development of the Late Cenozoic igneous rocks within the Alpine Belt
1 – back-arc seas (A – Alboran, T – Tyrrhenian; Ae – Aegian) and “downfall” seas (B –Black, C – Caspian); 2 – back-arc sedimentary basins (P- Pannonian, Po – Po valley); 3 – Late Cenozoic andesite-latite volcanic arcs (in circles): 1 – Alboran, 2 – Cabil-Tell, 3 – Sardinian, 4 – South-Italian, 5 – Drava-Insubrian, 6 – Evganey, 7 – Carpatian, 8 – Balkanian, 9 – Aegian, 10-12 – Anatolia-Elbursian (10- Anatolia-Caucasian, 11 – zone of the Modern Caucasus volcanism, 12 – Caucasus-Elbursian); 4 – areas of flood basaltic volcanism (in square): 1 – South Spain and Portugal, 2 – Atlas, 3 – Eastern Spain, 4 – Central France massif, 5 – Rhine graben, 6 – Czech-Silesian, 7 – Pannonian, 8 – Western Turkey, 9 – northern Arabia; 5 – suture zones of major thrust structures interaction of a superplume head with mobile continental lithosphere. Good example of such situation is the TEB (Sharkov, this book), where processes of collision continue now. The main feature of this belt is wide spread of the Late Cenozoic-derived volcanism, which has displayed practically coeval conditions on all its length, presuming existence of a superplume (or asthenospheric rise) beneath it. The belt has the most complicated structure within the Alpine segment, where a system of andesite-latite volcanic arcs and back-arc basin, bordering by nappe-folded mountain ridges were observed. In front of these ridges in Western Europe, north-west Africa and Arabia, coeval rift systems and flood basaltic volcanism often occur

2.1 Aegean-Caucasian segment

Caucasian part of the segment is located to the north of the major suture zone traced by ophiolites of Cyprus, Syria, southeastern Turkey, Zagros, etc. (Periarabian ophiolitic arc). Late Cenozoic Anatolian-Elbursian andesite-latite volcanic arc is developed at its rear. It, in turn, is formed by two arcs – Anatolian-Caucasian and Caucasian-Elbursian, touching in the transverse (Transcaucasian) zone of the latest volcanism (Fig. 1). The Black and Caspian seas with oceanic crust, cut Pre-Pliocene structure of the Caucasus and the Kopet Dag; they are filled with Mesocenozoic sediments of 12-15 km thick. The nature of these deposits, until the late Miocene was generally similar to developed within the Caucasus and the Kopet Dag (Zonenshain, Le Pichon, 1986).

The Caucasus is located in the zone of the Arabian syntax (Burtman, 1989), where Arabian plate is subducted beneath Eurasian. Specific analog of deep-see trench is represented by the Mesopotamian trough here, which, beginning from the Eocene, is experiencing an active submergence, due to thick molasse accumulated (Ponikarov et al, 1969). The northern part of the Arabian plate began to rise above sea level in the late Oligocene and early Miocene, about 26-25 Ma when there began the development of basaltic volcanism (Sharkov, 2000) and the Red Sea rift opened up in southwest, separating Arabian plate from Africa. Rate of ascending movements sharply increased to the Miocene- Pliocene boundary, when Gulf of Aqaba opened approximately 5 Ma and Arabian plate began quickly shifted to the north along a large Levant Fault (Dead Sea transform) (Kopp, Leonov, 2000; Prilepin et al, 2001; Sharkov, this book). However, this displacement is hardly manifested in the Greater Caucasus shift and, moreover, GPS data indicate that the width of the central Caucasus is not decreasing but increasing (Shevchenko et al, 1999)

Specific structure occurred at the northern side of the Black Sea. Judging from the geological and geophysical data along the profile of Tuapse-Armavir, the Black Sea microplate is separated from the Eurasian by narrow subvertical zone of strong positive gravitational anomalies (Shempelev and et al, 2001). It concerned to large blocks of deformed and metamorphic rocks, close on the density to crust-mantle mixture. This zone can be traced to depth of 60-70 km and the Moho is not established here. The northern blocks move upward along their separating steeply-dipping faults, ensuring the existence of mountain relief of the Western Caucasus.

Formation of the Black Sea began, apparently, in the early Cretaceous, but significant deepening of the basin occurred at the Oligocene-Early Miocene boundary (Zonenshain, Le Pichon, 1986, Nikishin et al, 2001), followed in the Miocene by filling of the deep-water depressions by sediments and a gradual shallowing of the basin (Kazmin et al, 2000). Since the Pliocene-Quarternary new significant deepening of the Black Sea basin has occurred (Nikishin, Karataev, 2000), which occurred almost simultaneously with the uplift of the Caucasus and Crimea, which in the Oligocene-early Miocene were not expressed in the relief (Neotectonics…, 2000; Kostenko, Panina, 2001). The close sequence of events took place in the South Caspian Basin, which is a similar structure (Zonenshain, Le Pichon, 1986; Grachev, 2000).

Aegean part of the segment is characterized by island arc associated with subduction of the oceanic East-Mediterranean plate beneath the Eurasian (Papazachos et al., 1995). There are all typical structures of the active zones of transition from continent to ocean here: the deep-water Hellenic trough, Aegean volcanic arc and back-arc Aegean Sea with basaltic magmatism in its periphery (in the west of Asia Minor, near Izmir).

Numerous deformations of extension in subhorizontal submeridional direction (strike-slip and normal faults, nappe-thrusts, grabens, etc.) are known in the Aegean basin and in adjacent parts of Greece and Turkey (Prilepin et al, 2001). At that for Balkan Mountains in the north are characteristic north-vergentes imbricate nappes and thrusts whereas for

Hellenides-Aegides-Taurides they are south-vergent. Judging from the seismic data, the stress state of the subhorizontal N-S stretching is characteristic of only for the upper 50-60 km of the Aegean basin lithosphere, leading to its expansion. At greater depths within the mantle beneath the basin both in south and north are fixed compression conditions. The regions of extension and compression are in direct contact by subhorizontal section at depth of 70-80 km; probably, it is the boundary of lithosphere and plastic material of extended plume head.

Aegean volcanic arc is a Pleistocene in age, but the development of the Aegean Sea, started earlier, about 12 Ma (Evsyukov, 1998). Apparently, earlier the main subduction zone located north, and its residues were survived in Dinarides and western Asia Minor (Western Anatolian: Fig. 1).

Descent of the Eastern Mediterranean (Ionian Sea, underwater Medina Ridge, Levant Basin) began in the late Miocene (Evsyukov, 1999). Approximately at the middle Pliocene (about 3-3.5 Ma) processes reinforced on the east: judging by the results of the 5th Cruise of R/V "Akademik Strakhov", Sinai plate descended here beneath the sea level to depths of 2-2.5 km (Geological…, 1994). Fragments of this plate are preserved in the form of Eratosthenes Seamount and smaller rises.

The northern part of the Eastern Mediterranean is separated from the southern part by zone of powerful deformations, which runs along the base of the Cyprus arc (Zverev, 2002). As in the case of the northern side of the Black Sea, large sub-vertical faults and a sharp increase in the boundary velocities up to 7-7.6 km/sec is recorded here. In the northern zone is observed uplift of blocks of basement, represented by ophiolites of Cyprus. This deformation zone begins at the northern end of the Arabian plate, and following along Periarabian ophiolite belt, comes to Cyprus and then, continuing to the west, joins the Aegean subduction zone (see Figure 1).

The situation on the eastern passive margin of the Mediterranean Sea also looks like northern passive margin of the Black Sea. System of subparallel mountain ranges (Lebanon, Anti-Lebanon, Jabal al-Ansari, Amanus, etc.), separated by a system of steeply dipping faults, occur along the coast here. The largest of the latter is the aforementioned Levant (Dead Sea) Transform Fault. Parallel to it, already under water, there is a zone of Pelizium faults bounding the east Levant deep depression (Khair, Tsokas, 1999). This depression has the oceanic crust, overlapping by thick (approximately 10 km) sequence of Phanerozoic sediments, and in this respect no different from the Black and Caspian seas.

Origin of within-plate Palmirides folded-thrusted zone of deformed platform cover is related to development of Levant Fault. According to Kopp and Leonov (2000), the formation of this structure was caused by braking of the western edge of the Arabian plate in the zone above the bending S-like curve of this fault at its motion to the north during the Neogene-Quaternary. Palmirides were formed under compression of the crust by approximately 20-25 km, compensating the northern Arabian plate movement that started in the Middle-Late Miocene and continues today.

Areas of intraplate moderate alkaline basaltic volcanism are widely developed in the northern Arabian plate, indicating the presence of a mantle plume here. Judging from the isotopic dating, basalt eruption began at the Eocene-Miocene boundary, about 25-26 Ma, and almost without interruption continued until historic times. The most powerful eruption occurred in the late Miocene-early Pliocene and Late Pliocene-Quaternary (Sharkov, 2000; Lustrino, Sharkov, 2006).

The Caucasus-Aegean segment in geophysical terms is characterized by two strong positive isostatic anomalies, one of which is confined to the area of the Aegean Sea, and the second – to the Transcaucasian zone of modern volcanism on the north of Eurasian-Arabian syntax. It likely evidence about uncompensated excess of mass beneath these structures, presumably associated with the kinematics of the mantle plume ascending and extending of its head. This is supported by seismic tomography data (Gök et al., 2003) and consistent with the wide development of the Neogene-Quaternary platobasaltic volcanism in the north Arabian plate (Trifonov et al., 2011) and more rare – in Transcaucasia. In this regard, attention is drawn to the isotopic characteristics of lavas of Mount Elbrus, for which was establish the impurity plume material increases with time (Chernyshev et al., 2002) Another indication of the existence of the plume under the South Caucasus is a found of a mantle helium in the Lake Van in the north-eastern Turkey (Kipfer et al., 1994).

However, under the Eastern Mediterranean and the Caspian Sea, conversely, are a strong minimum isostatic anomalies, indicating the mass deficit beneath them, which is probably due to the presence beneath them descending flows in the mantle.

2.2 Structure and development of the western (Alpine) segment

Judging on geological data, the current structure of the Alpine segment was formed mainly in the late Cenozoic, largely on the continental crust of the African plate. Remnants of this plate commonly observed along the northern coast of the Mediterranean Sea, and formed south of Spain (Betic Cordillera), Balearic Islands, Corsica and Sicily and the Apennines, the southern part of the Alps, large parts of the Balkans and Asia Minor (Ricou, 1986).

For the Alpine segment, in contrast to the Caucasian, is characteristic of complex configurations of the major structures related to thrusting the African plate beneath the Eurasia. Andesite-latite volcanic arcs, which associated with compression zones, represented by ridges, partly bordering back-arc basins with thinned crust and with often well expressed basaltic volcanism (Fig. 1). The formation of these subduction-related arcs occurs mainly in middle-late Miocene and Pliocene, and South-Italian arc is active till now. The feature of these subduction zones is that they involved material continental crust (Pino, Helmberger, 1997: Morales et al., 1999; Marson et al., 1995). For these arcs is characteristic their distinct migration in space – Alboran arc has moved to westward (Morales et al., 1995), Carpathian – to the east (Royden, 1989), and the Tyrrhenian – to the southeast direction (Rehault al., 1987).

Back-arc basins of Alpine segment (Tyrrhenian and Alboran seas, Pannonian Basin), became formed around the same time (Marotta et al., 1995; Duggen et al., 2004). Originally the back-arc seas were developed as continental rifts, which submerged under the sea level approximately at the boundary of the Miocene and Pliocene. At the same time under the sea level began sink the South Balearic Basin, which became faster deeper in the late Miocene (Trifonov et al., 1999).

It is noteworthy that the region of the Alpine orogen, including all Western Mediterranean, is surrounded by a broad band of Late Cenozoic basaltic volcanism associated with rift structures of Central and Western Europe (grabens Rona, Rhine, Hessen, Polabian, etc.) as well as numerous basaltic plateaus, stretching to west from the French Massif Central via south of Spain to Portugal. It further extends beneath waters of the Atlantic Ocean (seamounts Amper, Josephine, and others), as well as on the islands of Madeira and Canary, boardering Alboran arc from the west. Powerful basaltic volcanism of Atlas occurs to the south of the arc. In the southwestern part of the Alpine segment basaltic volcanoes of islands Sicili (including Etna), Pantelleria, Lemos and seamount volcanoes on the Tunisia Threshold are occurred.

Together with platobasalts of Syria and southern Turkey, they form anorogenic circum-Mediterranean magmatism with common source – so-called Common Magmatic Reservoir (Lustrino, Wilson, 2007). It is, obviously, evidence about existence beneath the region a present-day mantle superplume; Alpine orogen with complicate combination of mountain ranges and basins is located in its inner part (Sharkov, this book). Earlier this entire basaltic volcanism was considered as the final, which appeared after the cessation of collision, but recent studies have shown that it is the beginning of a new distructive phase of the Europe development (Grachev, 2003).

The Alpine segment in geophysical terms represents a region with decreasing overall thickness of the earth's crust due to moving away of high-velocity layer of lower crust, characteristic of the East European Craton, and the high density of heat flow (Gize, Pavlenkova, 1988). The compression zones, represented by ridges, are characterized by deep roots (till 200 km in the Alps: Laubscher, 1988); part of lithoplastines, especially adjacent to the back-arc basins, has a steep attitude (Ricou, et al., 1986). In cases where the butt-end parts of lithoplastines come to the surface, within these contours are observed blocks of lower crust and even upper-mantle rocks lifted from 70-120 km, as it observed in Western and Eastern Alps and the Gibraltar arc (Magmatic ... 1988; Harley, Carswell, 1995).

At the same time, oceanic crust, composes the floor of the newly formed seas of the Western Mediterranean, appeared due to thinning and rupture of continental lithosphere of the African plate. The latter survived along the periphery of these seas, in particular, the Tyrrhenian, where seismic data from the periphery to its center set reduction of crustal thickness from 20-30 to 6-8 km to the extent to complete disappearance of the "granite" layer (Royden et al., 1986; Marson et al., 1995). The development of this basin began from continental rifting in the middle-upper Miocene, and the open of the sea has occurred at the boundary of the Late Miocene and Pliocene and continues today (Bartole, 1995; Story, 1996). In the Pannonian Basin reducing crust has the same trend - it is reduced from 30 to 18-20 km mainly at the expense of the "basaltic" layer (Nikolaev, 1988; Horváth et al., 2006). Development of the basin started in the Middle Miocene, 11-10.5 Ma, simultaneously with the appearance of basaltic volcanism here (Neotectonics…, 2000).

Attention is drawn to another feature of the back-arc basins of Alpine segment - they, as well as to the Aegean Sea, are associated with the maxima of isostatic anomalies of average (Alboran and Tyrrhenian seas) and large (Pannonian) intensity (Artemiev, 1971; Sparkman et al., 1993; Artemieva et al., 2006) (Fig. 2). As in the case of the Aegean-Caucasian segment, it may indicate uncompensated excess of mass beneath these structures associated with ascending of mantle plumes. These facts, along with materials on magmatism, show an essential deep mantle roots of observed geological processes here, as evidenced by extensive manifestations within the Alpine Belt epicenters of intermediate-focus earthquakes of the depths of 100 to 500 km (Tyrrhenian Sea, the Carpathians, Caucasus, etc.: Gize, Pavlenkova, 1988).

Thus, the formation of the major geological structures of the Alpine Belt began approximately at the boundary the Oligocene and Miocene and proceeded almost synchronously on all territory. At the first stage of development gentle uplift and subsidence of the relief of the region solid surface was dominated and began the formation of cavities of back-arc basins in the western Mediterranean, deepening of the Black and Caspian seas has occurred, platobasalt eruptions began in the north Arabian plate. An the second stage, which started in the late Miocene-early Pliocene, intense of tectonic activity increased sharply, began to form mountain ranges, as well as sharply increased basaltic volcanism along the periphery of the Alpine belt. All major features of the structure of the region were formed at that time (Trifonov et al, 1999), sharp deepening of the Black and Caspian seas, and the Eastern Mediterranean occurred; uplift of Caucasus and Crimea has begun as well as appearance of the Dead-Sea (Levant) Fault and Palmirides fold-thrusting structure. All of these processes has got impulse in the Pliocene-Quaternary, when finalized tendencies inherent in the current stage of the Alpine Belt development.

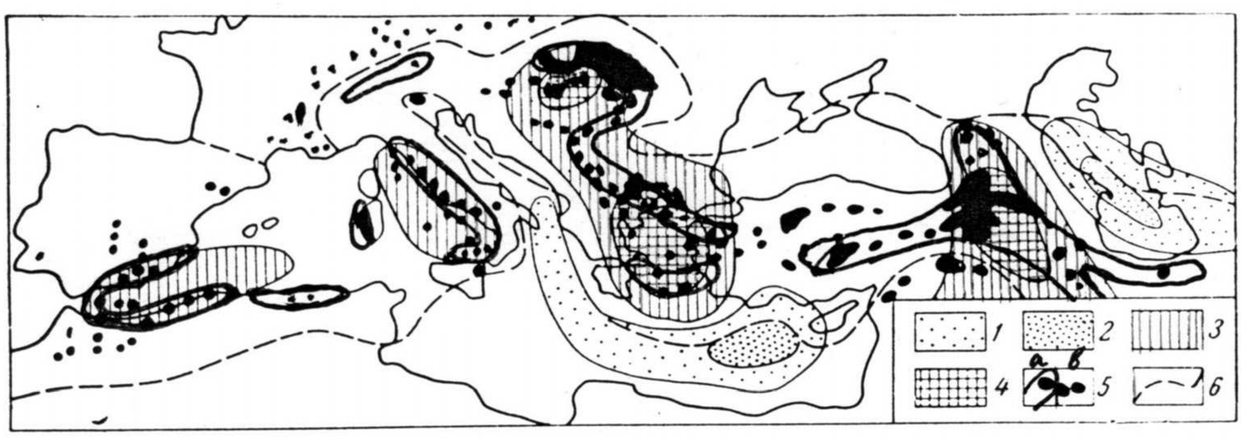


Fig. 2. Distribution of major regional isostatic anomalies and areas of Cenozoic volcanism in the Alpine Belt. After M. Artemiev (1971) with corrections.
1- regional lows of average intensity; 2 - of High intensity; 3 - regional highs of average intensity; 4 – of high intense; 5 - volcanic rocks (a - calc-alkaline series, b - basalts);
6 - boundaries of the Alpine Belt. Some depressions of the Alpine belt (Tyrrhenian, Aegian, Alboran, Pannonian) are characterized by positive anomalies, which evidence about excess of mass beneath them. Probably, they represent the present-day plume heads, which support basaltic volcanism and lead to onwards displacement of andesitic volcanic arcs in time. According to geophysical data, the plumes joint together in common layer at the depth 200-250 km, forming of circum-Mediterranean common magmatic reservoirs (Lustrino & Wilson, 2007). This rise starts in the North Atlantic, near Azores and extends to the east to Western Europe

3. Mechanical-mathematical simulations of deep-seated processes in the Alpine mobile belt

Almost simultaneous occurrence of all these tectonomagmatic processes on the vast territory assumes that in this place we are faced with combination of present-day continental collision zone and a mantle superplume (Sharkov, this book). The relief of the superplume head is complicated by numerous protuberances (local plumes), controlling the position of modern depressions in the Alpine Belt and are caused extended zones on the general context of compression. The presence of such superplume under Alpine Belt also supported by seismic tomography data (Anderson, Dzevonsky, 1984; Sparkman et al., 1993; Hearn et al., 1999). This uplift starts in the Eastern Atlantic, extending eastward into parts of Western and Central Europe (Hoerne et al., 1995).

Surface of the superplume head is highly variable, apparently reflecting the development of gravitational Rayleigh-Taylor instability at the boundary of the rigid lithosphere and the heated the superplume head. Judging by the foregoing isostatic anomalies, beneath the back-arc depressions of the Alpine orogen (Tyrrhenian, Aegean, and Alboran seas, Pannonian Basin, etc.) an excess of mass is occurred, obviously connected with the existence beneath them heads of local plumes (protuberances). Extending of these heads led to displacement of subduction zones and their andesite-latite volcanic arcs (Harangi et al., 2006). Judging from the observations in the Aegean region, the thickness of an extended plume head does not exceed 40-50 km, and its spreading leads to appearance of field strong subhorizontal strength in the lithosphere in its front (see above). These plumes at depths of 200-250 km are merged into a single asthenospere layer, corresponding, apparently, to body of Alpine superplume head, which is the major source of geodynamic activity in the region.

The exception to this general rule is the North-Arabian-Transcaucasian plume, where while there were no basin, but there is a clear shift of the Anatolian-Elbursian subduction zone to the north. Perhaps this is due to the spread of the plume head to the north and its relative youth. Obviously this is due to the current increase in the width of the Central Caucasus, to which attention was drawn above.

In contrast to these structures, the Eastern Mediterranean, as well as the Black and Caspian seas are characterized by negative isostatic anomalies, which evidence of downward mantle flows beneath them (“cool plumes”) located between ascended “hot plumes” (Sharkov, this book). Unlike the back-arc seas, all of them have typical passive margins and significant thickness of Meso-Cenozoic sediments. They look like "downfall", which cut-off earlier geological structures of the continent. Origin of such “cool plumes” obviously linked with appearance of excess of mantle material between extended plume heads.

Apparently, in these basins survived oceanic crust of Tethys. Judging from the northern sides of the Black Sea and Eastern Mediterranean, belts of strong positive gravity anomalies occurred along the sides of these basins. They formed by blocks of high-density rocks, separated by subvertical faults, receding into the mantle to the depths 60-70 km. These belts are reminiscent of similar zone of strong magnetic and gravitational anomalies developed along the passive margins of the Atlantic Ocean and are known as "seaward-dipping reflectors» (SDR) (Bogdanov, 2001; Larsen, 2002).

Very likely that such structures appear along the boundary between simultaneously active ascending (plumes) and descending currents in the mantle, due to rocks here are undergone by powerful deformation and metamorphism. As seen from the presented data, blocks, adjacent to the descending currents in the mantle are penetrated into its, and adjacent to the plume - ascended upstairs. Obviously, to the same circumstance, i.e., with the ascending of mantle plumes through the thickness of the lithosphere, can be related processes of exhumation of deep-seated rocks, large slices of which, as shown above, are observed in the mountain ridges on the periphery of plumes.

Thus, we can assume that the situation within the Alpine Belt defined by the presence beneath it the large (approximately 2000 x 5000 km) superplume head with a complex relief of surface. At the sites of the large uprising of this relief are usually placed back-arc basins now. The highest elevation of the superplume material is observed near the Tyrrhenian Sea, which probably can be interpreted as a modern center of activity of the whole zone. It is here very thin lithosphere (up to 30 km) and the maximum heat flux (above 3 E.T.P.) occurred. On the other hand, the most powerful isostatic anomalies are observed in areas of the Pannonian Basin and the Aegean Sea, which are reminiscent of the Tyrrhenian Sea at the early stages of its development, whereas the Caucasus syntax can be the very beginning of the process. Apparently, this implies that the center of activity will shift in the future.

However, it is possible to describe some of the characteristics of these structures, in particular areas of back-arc spreading over regional uplifts of the superplume surface relief even now. They exactly are the centers of deep-seated activity, which are very largely determined all the tectono-magmatic processes. For this analysis, we used the general model of high-viscosity incompressible fluid, the parameters of which vary from layer to layer (Zanemonets et al, 1974).

3.1 Development of structured above regional uplifts asthenosphere (plume heads): mechanical-mathematical simulations

As it was shown above, depth of asthenosphere (superplume) surface under Alpine Belt changes from 30 km in the centre of Tyrrhenian sea up to 70-100 km in depressions of East Mediterranean, strongly changing on lateral. The characteristic size of depressions is 500-1000 km and more, distance between them is 1000-1500 km.

Hence we have characteristic parameters of a problem: $h_3 \sim 10$ km - thickness of sedimentary cover, $h_2 \sim 100$ km - thickness of lithosphere, $L \sim 1000$ km - horizontal scale, $\varepsilon = h_3/L = 10^{-2}$-small parameter.

Decomposing velocities and pressure in line on $\sqrt{\varepsilon}$ and considering the boundaries between layers as material ones, it is possible to receive in zero approximation the equation of a day surface ζ_3 and a surface of the basement ζ_2 in dependence of dynamics of mantle plume $U_0, W_0\big|_{\zeta_1}$:

$$\begin{cases} \dfrac{\partial^2 \zeta_3}{\partial X^2} = \beta\left[h_2 \dfrac{\partial U_0}{\partial X} - W_0\right] \\ S\dfrac{\partial \zeta_2}{\partial t} + U_0 \dfrac{\partial \zeta_2}{\partial X} + \alpha\left[h_2 \dfrac{\partial U_0}{\partial X} - W_0\right] = 0 \end{cases} \tag{1}$$

$$\alpha = \frac{(h_3)^3}{(h_3)^3 + \dfrac{\mu_3}{\mu_2}(h_2)^3}, \ \beta = \frac{1}{\dfrac{\rho_3}{3}\left[\dfrac{(h_3)^3}{\mu_3} + \dfrac{(h_2)^3}{\mu_2}\right]}$$

$S = \dfrac{L}{u_l t_0}$ - Strukhal Number, u_l - characteristic velocity of the lithosphere matter, t_o - characteristic time of the processes, , μ_i - viscosity, ρ_i - density.

Let us set a field of velocities and morphology of boundary ζ_1 as:

$$U_0 = a\ \mathrm{th}\ kX, \quad \zeta_1(X,t) = -\gamma\ \mathrm{sh}^2\ kX - (h_2 + h_3) + \frac{D}{S}t \tag{2}$$

where k, a characterize intensity of rifting: k-in the centre of structure, a - far from the centre; γ - allows to vary the form of rising plume; D - velocity of the plume rise $D = S\frac{\partial \zeta_1}{\partial t}$ (fig. 4a).

The given field of velocities qualitatively enough well reflects the basic features of a considered class of movements: rise of plume, rifting above it and lowering of substance on sufficient distance from the centre. Quantitative conformity at comparison with the available geological-geophysical data is achieved with the help of a variation of coefficients in a modelling field of velocities and their change during considered process at preservation of the general structure of movements. From the decision of system (I) we shall receive for enough big t:

$$\begin{cases} \zeta_2 = -h_3 - \alpha\gamma\ \mathrm{sh}^2\ kX + \alpha h_2 \ln(\mathrm{ch}\ kX) + \alpha(D - h_2 ak)\dfrac{t}{S} \\ \zeta_3 = \beta\left[\dfrac{h_2 a}{k}\ln(\mathrm{ch}\ kX) + \dfrac{\gamma a}{(2k)^2}\mathrm{ch}\ 2kX - \left(\dfrac{\gamma a + D}{2}\right)X^2\right] + C_1(t) \end{cases}$$

C1 (t) is determined from balance of mass.

The analysis of the received equations shows, that there is a critical depth of rise of mantle plume $h_2 = 2\gamma$, when the characteristic form of the lithosphere layers changes. If $h_2 > 2\gamma$, there is a deflection of a surface of the basement in the centre of rifting, that really takes place in considered back-arc seas. If $h_2 < 2\gamma$ (depth of plume is insignificant) or velocity of its rise is essential ($D > h_2 ak$) then the swelling of the basement surface corresponds to swelling and rise of a surface of the mantle plume (fig. 3).

When on periphery of basin there are the conditions interfering free rifting of the lithosphere of region, for example, caused by collision of the Arabian-African and Eurasian plates, the field of velocities on the bottom boundary of layers can be modelled as:

$$U_0 = \frac{\mathrm{th}\ X}{\mathrm{ch}^2 X}, \quad \zeta_1 = -\mathrm{sh}^2 X - (h_2 + h_3) \tag{3}$$

For the greater presentation of result the coefficients in a modelling problem are omitted. Then:

$$\begin{cases} \zeta_3 = -\dfrac{\beta h_2}{2}\dfrac{1}{\mathrm{ch}^2 X} + \dfrac{\beta}{2}X^2 - \beta\ln(\mathrm{ch}\,X) + C(t) \\ \zeta_2 \cong -h_3 - \dfrac{\alpha h_2}{S}t - \alpha(1-2h_2)\mathrm{sh}^2 X + \alpha(1-2h_2)(\mathrm{sh}\,X)^{\frac{2(h_2-1)}{1-2h_2}}\exp\left[\dfrac{2-3h_2}{1-2h_2}\left(\dfrac{t}{S} - \mathrm{sh}^2 X\right)\right] \end{cases}$$

C (t) it is determined from balance of mass.

Now there are two critical depths of the asthenosphere upwelling, when cross-section of layers changes its structure. At $h_2 > 2/3$ in the centre of structure the deflection is formed. At

$1/2 < h_2 < 2/3$ the surface of the basement is inclined, and at $h_2 < 1/2$ it reflects the morphology of plume in the centre of rifting and forms concavity of the basement on periphery of basin (fig. 3b).

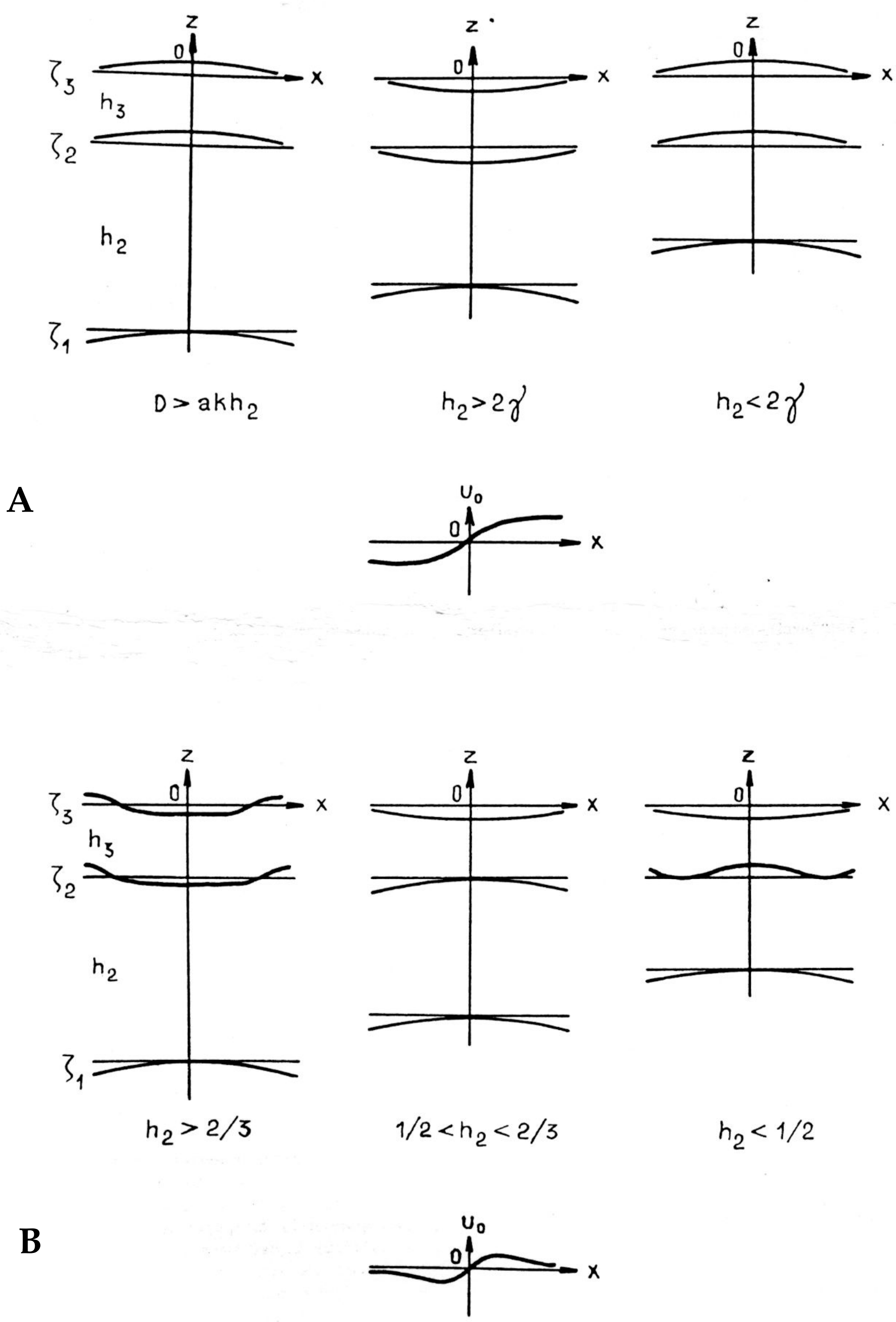


Fig. 3. Characteristic cross-sections of layers of earth crust and subjacent lithosphere above ascending mantle plume: a – without lateral limitations (U_0 = th kX), b – with lateral limitations (U_0 = th kX / ch^2 kX; D = 0)

The first type of velocity (2) can simulate the early stages of structures development, and the second type (3) - Alboran, Tyrrhenian and Aegean seas, as well as the Pannonian depression. The second type of activity, according to geological data, often accompanied by a extending of the local plume heads. Under conditions of the collision zone, at the boundary of the extended plume head and limiting its blocks of the continental lithosphere arise powerful strengths. It lead to formation of zones of deformations and metamorphism, which can later develop into zone of downward flowage of the material (subduction zone), to which involves an excess of crustal material, appeared as a result of displacing of the material. It is often activated already existed subduction zones also, as evidenced by the data on the Western Mediterranean (Morales et al., 1999). Change of regime of upwarping to structure of deep-water basin is confirmed by a number of geological factors: the change in the regime of sedimentation and the removal of terrigenous sediments, wedge-out of layers of sedimentary cover, changing the direction of flow paleorivers, evolution of paleodepths of basins, etc. (The crust ..., 1982).

An example of the interaction of ascended plumes with the earth's crust in the absence of side limitations, apparently are intracontinental rifts such as Baikal Rift. As Grachev (1987) shown, during pre-rift stage of their development an overall rise of the territory occurs, and the on the rift stage itself – its descent with the formation of extensional structures – grabens and sedimentary basins. In contrast to the of collision zones, deep-water basins with oceanic crust are not formed. The case is usually limited to thinning of the crust and the relatively small submergence of the crystalline basement for some kilometers. Only in exceptional cases, when a powerful inflow of plume material may be breaking the crust with formation of structures such as the Red Sea.

Thus, the proposed model seems to adequately describe the mechanism of formation of geological structures associated with the plume tectonics. From this it follows that the formation of depressions over the ascending mantle plumes depends on the geodynamics of the deep-seated layers. Determinants are mechanical processes that reflect the general direction of movements from formation of upwarping to formation of basin with unidirectional motion (plume ascending). At that formation of deep basins do not require much stretching of the layers. Morphology of the deep-seated boundaries determined by the shape of the plume, the rate of its ascensing and the intensity of moving apart a material over it, i.e., effective viscosity of the lithosphere layers. For sufficiently large gradients of plume heads surface that ascending in areas of collision, above them are formed deep-water depressions such as the Tyrrhenian Sea.

3.2 Development of volcanic arc – backarc systems

The principal components of such systems a subduction zone accompanied by andesite-dacite magmatism and a newly formed back-arc basin with transitional to oceanic crust, formed originally on the continental crust (Bogatikov et al., 2009). Both structures generally seem to have been initiated and developed about the same time, possibly implying a common reason for their formation. However, the nature of back-arc basins and their role in geodynamic processes were largely overlooked until recently, even though these settings are likely to have played the most crucial role here. High heat flows, positive isostatic anomalies, extensive basaltic magmatism, and the presence of basalt-hosted mantle xenoliths suggest that the back-arc basins have developed above the asthenoshere rises or extended plume heads. This is in good agreement with seismic tomography results, which revealed that the back-arc basins are underlain by a hot mantle as deep as 400 km (Anderson et al., 2002). Some of Mediterranean basins (Alboran, Tyrrhenian, Aegean, and Pannonian) are also marked with large positive isostatic anomalies mostly confined to back-arc basins (Fig. 2). This suggests uncompensated excess mass which can be related to mantle plumes upwelling beneath these structures. Like the Pacific back-arc basins, Mediterranean's back-arc basins are also characterized by extensive basaltic magmatism and intermediate- to deep-focus earthquakes of 100-500 km in depth.

During stretching of the relatively soft oceanic upper mantle (asthenosphere), its edges exert mechanical pressure on the continental block. The rotational motions of the continental block cause some part of the asthenospheric material to move beneath the softened base of the latter and then begin to ascend as an independent "asthenospheric" plume (Sharkov, Svalova, 2005). In our case, the role of asthenosphere plays material of superplume head and protuberances on its surface (local plumes). Above a certain depth, after reaching the buoyancy level, the plume head begins to spread out laterally to form an extensional zone, such as a continental rift zone developed above the plume, as was the case for the initial stages of the development of the Sea of Japan and Sea of Okhotsk (Lelikov, Emel'yanova, 2007). However, unlike the ordinary rift zones, where the situation is symmetric on either side of the spreading axis, here it is sharply asymmetric: on one side is the massive cold continental lithosphere, and on the other side is a less dense oceanic plate. Under such conditions, the spreading may have occurred in a different way, mainly oceanward, to the side of mechanical downdragging. Accordingly, the continental material transported by the spreading plume head is expected to move in the same direction, where both material flows, migrating from the ocean and continental side consequently, will eventually collide.

We consider the sequence of events on the basis of the mechanical-mathematical model for a multilayer viscous incompressible fluid, describing the dynamics of "granite" and "basalt" layers of the earth's crust, lithosphere, and asthenosphere (Sharkov, Svalova, 1991; 2005). At the initial stages, when deflection of the layer boundaries from their original position is still insignificant, the base of the lithosphere always dives into the asthenosphere. In other words, during the early stages of structural evolution in zones where interaction between the plates is the most active, the lithosphere sinks into the less dense asthenospheric material to form a subduction zone. The calculations imply that the most dense rocks of the ancient continental lithosphere, its mantle and lower crust, made up of garnet granulites, which are much denser than rocks of oceanic crust, were first to begin descending. Much sialic material of the upper continental crust (its granitic layer) from the back-arc region, which was sandwiched between two subsiding plates (oceanic and lower continental crust), may have also been involved into the overall motion. Having a subduction rate of 7-10 cm/year, this motion causes rocks of this layer to be sucked in the subduction zone. As a result, a MORB-type oceanic lithosphere in a back-arc setting will develop when the old continental lithosphere is partially or fully removed to be involved in this subduction zone and buried in the deep mantle (Fig. 4).

As the subducting plate (slab) sinks, its rocks transform to high-dense amphibolites, garnet granulites, and eclogites as well as relatively light sedimentary rocks, volcanics, and gneisses (including granite-gneisses). Meanwhile the latter, being metamorphosed under ultrahigh pressure and moderate temperature (P> 2.8–4, possibly, up to 8.5 GPa, T = 600–900°C), often retain their structural and textural characteristics in a metastable state and are thus easily recognizable (Ernst, 2001).

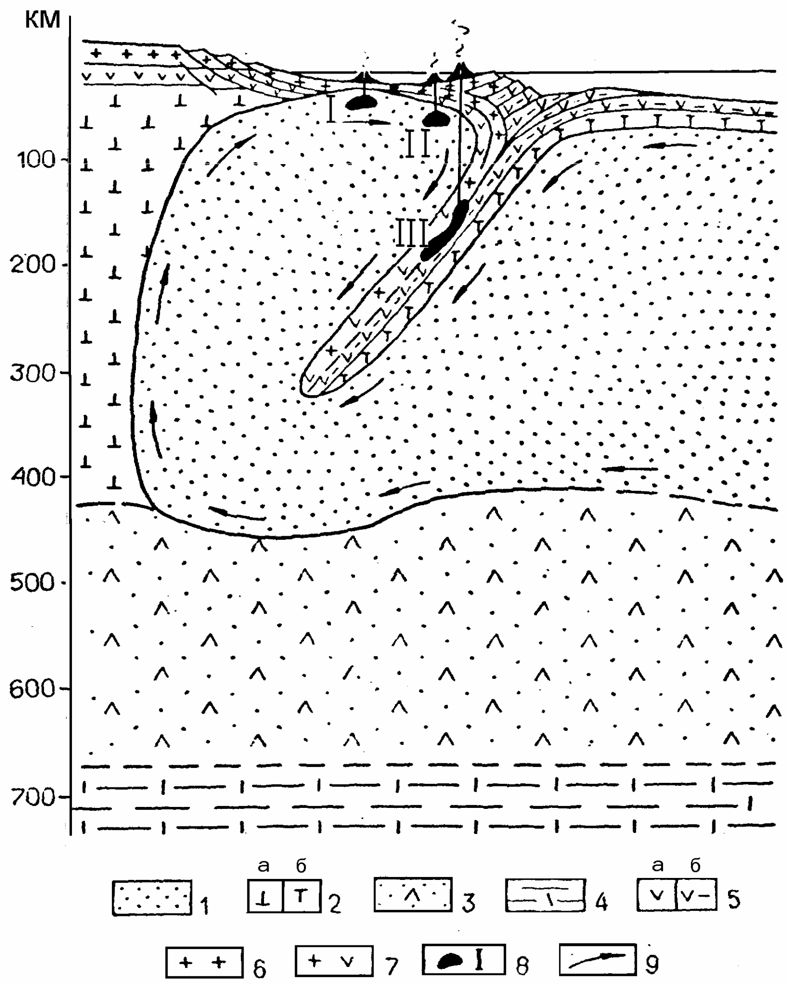


Fig. 4. Schematic view showing the evolution of back-arc spreading
1) Young soft upper mantle (asthenospheric) beneath oceans; 2) lithospheric mantle beneath: (*a*) continent, (*b*) ocean; 3) upper mantle beneath the 410 km discontinuity; 4) lower mantle; 5) lower crust of (*a*) continent, (*b*) ocean; 6) sialic continental crust; 7) mixture of the sialic and basite crustal materials within a subduction zone; 8) magma chambers (I - tholeiite series; II - boninite series; III – calc-alkaline series); 9) flow direction of the soft oceanic upper mantle

Thus, the development of Late Cenozoic volcanic arcs and their adjacent back-arc basins, both in the Pacific and Mediterranean, have evolved in similar way. Their principal component is a back-arc mantle (asthenospheric rise or plume head) that possibly spreads out laterally towards the less dense lithosphere. At the place where the mantle flows collide, new subduction zones and back-arc basins at their rear have been formed. This results in the gradual deepening of the back-arc sea, thinning of its parental continental crust, and formation of transitional and oceanic type crust. At early stages, these systems might have looked like active continental margins, which then evolved in a complex arc-backarc system. Through continuous involvement in subduction processes, crustal material from the back-arc region is removed from the tectonosphere and stored in the "slab cemetery," revealed in the mantle by seismic tomography (Karason, van der Hilst, 2000). Only a minor portion of crustal materials is returned to the surface in form of subduction-related magmatism.

So, asymptotic models provide an opportunity to explore the main features of the formation and development of geological structures. Ascending of mantle plumes determines the depth of the deep-water basins initiation. The lithospheric material above them (the uppermost mantle and crust) is moved apart to make room for an oceanic crust. As already mentioned, the resulted excess of this lithospheric material is involved in subduction zones with formation of volcanic arc-backarc basin system. The process of subduction depends on the difference in density of the lithospheric and plume material, and can only contribute to spreading, freeing up space for moved apart crust from the back-arc basin. However, existence of mantle plume itself, which occurred against the background zone of lithospheric plate collision, is the result of collision of deep-seated mantle flows, contributing to pumping and ascending of local protuberances in the form of plumes with the formation structures of extension over its extended heads. Relationship between the area of collision of mantle flows at the depth and the zone of collision of lithospheric plates, along with relative velocities and ratios of densities, determines the dippig of subduction slab also.

Thus, complex processes in the zone of lithospheric plates collision are interrelated and interdependent. The relative rapidity of the observed processes evidence about defining role of mechanical motions in the formation of structures, forced by the influence of thermal factors. The Alpine Belt is an area of elevated heat flows due to removal of abyssal heat by ascending of plumes caused by deep-seated (up to the core) mantle activity and complex interaction with the thinned lithosphere. Analysis of times of formation of the Mediterranean's deep-water basins shows that the process of activation occurred from the periphery to the center, which was caused by compression and ascending of superplume material between lithospheric plates.

It is possible that a complex dynamic pattern of interaction between a plume and lithosphere is not limited to the examples above. As seen in Fig. 1, volcanic arcs are lenticular in shape and are distributed in space fairly chaotic, which differ them from the Western Pacific island arcs or active margins of the both Americas. However, as it know, active continental margin occurred along the northern margin of northern Tethys in this place in the Mesozoic, where zone of subduction existed and there were a powerful eruptions of lavas of calc-alkaline series (Khain, 1984). In this regard, it has been suggested that this ancient subduction zone during early-middle Miocene split into two fragments, one of them roll back, forming the Tyrrhenian Sea and the modern Calabrian arc, and another rolled to west, forming Alboran Sea and Betik Reef arc (Lonengran, White, 1997).

Continuing this logic, one might think that one more fragment rolled to the north-east, forming the Carpathian arc. In this case the Anatolian-Caucasus-Elbursian arc may represent the eastern fragment of this old subduction zone. As already stated, under the central part of this arc the northern end of a mantle plume occurs with which probably connected the bend in place of Transcaucasian structure, where the area of modern magmatism of the Caucasus occurs. From this perspective, we can assume that in the near future two composing arcs (Anatolian-Caucasian and Caucasian-Elbursian) under the effect of the plume extending will be separated and will exist independently, like the most of Late Cenozoic arc of Alpine Belt. From this obviously implies that extended plume heads under conditions of large collision zone can impact not to only ancient lithosphere, but also to shift and even break off into individual pieces of existing subduction zone.

4. Conclusions

1. The Late Cenozoic Alpine-Mediterranean Mobile Belt (Alpine Belt) has appeared under condition of collision of lithospheric plates above superplume head. It's surface is complicated by number of protuberances (local plumes), which are a cause of emergence of extending zones on the background of the overall structure of compression. Geological situation in the belt is considered with complex interaction of converging lithospheric plates with plastic plume material.
2. It is shown that two types of depression with a predominance of the oceanic crust occur within the belt: (1) newly-formed back-arc basins above extended heads of local plumes (Western Mediterranean, Aegean Sea, Pannonian Basin), and (2) fragments of ancient oceanic crust of the Tethys, which has descended under the influence of downward movements in the mantle between plumes (Eastern Mediterranean, Black and Caspian seas). The second type is characterized by basins with passive margins, along which are developed steep deep faults; these areas in their structure resembles the structure characteristic of Atlantic passive margins.
3. The exception is the region of the North-Arabian - Caucasus plume without depression above it. This plume, apparently bends to the north the surface of the subduction zone, ensuring the existence of Anatolian-Elbursian andesite-latite volcanic arc to form a transverse area of modern volcanism of the Caucasus.
4. Geodynamics of the Alpine Belt has developed from the periphery of this structure to its center, in other words to the central part of the superplume head. Maximum geodynamic activity is now in the Tyrrhenian Sea, where thickness of the lithosphere is minimal; in the future it will probably be moved into the region of the Aegean, Pannonian, and Caucasus, where the most powerful positive isostatic anomalies occur.
5. Asymptotic models provide an opportunity to explore the major features of formation and development of geological structures. Ascending of mantle plumes determines the depth of initiation of deep-water basins. The lithospheric material above them (the uppermost mantle and crust) is moved apart to make room for an oceanic crust. The resulted excess of this lithospheric material is involved in subduction zones with formation of volcanic arc-backarc basin system.
6. The process of subduction depends on the difference in density of the lithospheric and plume material. It can only contribute to back-arc spreading, freeing up space for moving apart ancient crust of the back-arc basin which involved in subduction and further buried in the deep mantle. Existence here of the mantle plume itself is a result of collision of deep-seated mantle flows, contributing to pumping and ascending of local protuberances in the form of plumes with the formation of extensional structures over their extended heads. Relationship between the area of collision of mantle flows at the depth and the zone of collision of lithospheric plates, along with relative velocities and ratios of densities, determines the dip of subduction slab also.
7. Complex processes in the zone of lithospheric plates collision are interrelated and interdependent. The relative rapidity of the observed processes evidence about defining role of mechanical motions in the formation of structures, forced by the influence of thermal factors. The Alpine Belt is an area of elevated heat flows due to removal of

abyssal heat by ascending of plumes caused by deep-seated mantle activity (up to the core) in the complex interaction with the thinned lithosphere. Analysis of times of formation of the Mediterranean's deep-water basins shows that the process of activation occurred from the periphery to the center, which was caused by compression and ascending of superplume material between lithospheric plates.

5. References

Artemiev, M.E., 1971, Some peculiarities of deep-seated structure of depressions of Mediterranean type: evidence from data on isostatic gravity anomalies. *Bull. Soc. of the Nature Investigators of Moscow*. Geol. Dept., 4: 5-10 (in Russian)

Artemieva, I.M., Thybo, H. & Kaban, M.K., 2006, Deep Europe today: geophysical synthesis of the upper mantle structure and lithospheric processes over 3.5 Ga. In: *European Lithosphere Dynamics*. D.G. Gee and R.A. Stephenson (eds). Geol. Soc. London Mem. 32, 11-42.

Bartole, R., 1995. The North Tyrrhenian - Northern Apennines Post-Collisional System - Constraints for a Geodynamic Model. *Terra Nova*, 7(1), 7-30.

Bogatikov, O.A., Sharkov, E.V., Vesselovsky, A.V. & Mehscheryakova, V.B., 2009. Synchronism of development of Cenozoic volcanic arcs and back-arc basins: Reasons and consequences. *Doklady Earth Sciences*, 427A (6), 907-911.

Bogdanov, N.A., 2001. Continental margins: general problems of structure and tectonic evolution. In: *Fundamental problems of general geotectonics*. Yu.M. Puscharovsky (Ed.). Moscow, Nauchnyi Mir Publ., 231-249.

Burtman, V.S., 1989. Geodynamics of Arabian syntax. Geotectonics, № 2, 67-75.

Chernyshev, I.V., Lebedev, V.A., Bubnov, S.N. et al., 2002. Isotopic geochronology of the Quaternary volcanoes eruptions of the Great Caucasus. *Geochim. Intern*., 40, 1151-1166.

Duggen, S., Hoernle, K., van den Bogaard, P. & Harris, C., 2004. Magmatic evolution of the Alboran region: The role of subduction in forming the western Mediterranean and causing the Messinian Salinity Crisis. *Earth and Planetary Science Letters*, 218, 91-108.

Geological structure of the Eastern Mediterranean. V.A. Krasheninnikov and J.K. Hall (Eds.). Jerusalem, 1994.

Girbacea, R., Frisch, W., 1998. Slab in the wrong place: Lower lithospheric mantle delamination in the last stage of the eastern Carpatian subduction retreat. *Geology*, 26 (7). 615-618.

Gize, P. & Pavlenkova, N.I., 1988, Structural maps of the Earth's crust of Europe. *Physics of the Earth*, N 10, 3-14.

Gobarenko, V.S., Nikolova, S.B., Yanovskaya, T.B., 1986. Structure of the upper mantle of South-Eastern Europe, Asia Minor and Easterm Mediterranean based on data of discrepancy time of race P-wave. *Physics of the Earth*, № 8, 15-23.

Gök, R., Sandvol, E., Türkelli, N., Seber, D., Barazangi, M., 2003. Sn attenuation in the Anatolian and Iranian plateau and surrounding. *Geophys, Research Letters*, 30 (24), 8042, doi: 10.1029/2003GL018020.

Grachev, A.F., 2003, Final volcanism of Europe and it's geodynamic nature. *Physics of the Earth*, N 5, 11-46.

Grachev, A.F., 1987, Rift zones of the Earth. Moscow, Nedra Publ., 247 p. (in Russian)

Dobretsov, NL, Kirdyashkin, AG, & Kirdyashkin, AA (2001) Deep Geodynamics. GEO Publ., Novosibirsk (in Russian).

Evsyukov, Yu.D.,1998. Origin and development of morphostructuree of northern Aegian Sea. *Oceanology*, 38 (2), 286-292.

Evsyukov, Yu.D., 1999. Origin and main stages of development of Medina Ridge (Central Mediterranean, *Doklady Earth Sciences*, 366 (3), 364-368.

Earth crust and history of development of the Mediterranean Sea, 1982. Moscow, *Nauka Publ.*, 207 p. (in Russian).

Harangi, S., Downes, H, & Seghedi, I., 2006. Tetriary-Quaternary subduction-related magmatism in the Alpine-Mediterranean region // In: *European lithosphere dynamics.* Edited by D.G. Gee and R.A. Stephenson. Geol. Soc. Mem. 32. London, Geol.Soc. London Publ., 167-190.

Harley, S.L., Carswell, D.A., 1995. Ultradeep crustal metamorphism: A prospective view. *Journal of Geophysical Research*, 100 (B5), 8367-8380.

Hearn, T.M., 1999. Uppermost mantle velocities and anisotropy beneath Europe. *Jour. Geophys. Res.*, 104 (B7), 15,123-15,139.

Hoernle, K., Zhang, Y.S., Graham, D., 1995. Seismic and geochemical evidence for large-scale mantle upwelling beneath the eastern Atlantic and western and central Europe. *Nature*, 374 (6517), 34-39.

Kazmin, V.G., Shreider, A.A., Finetti, I. et al., 2000. Early stages of the Black Sea development according seismic data. *Geotectonics*, № 1, 46-60.

Khain, V.E., 1984. Regional Geotectonics. Alpine-Mediterranean Belt. Moscow, Nedra Publ., 344 p.

Khair, K., Tsokas, G.N., 1999. Nature of the Levantine (eastern Mediterranean) crust from multiple-source Werner deconvolution of Bouguer gravity anomalies. *Jour. Geophys. Res.*, 104 (B11), 12469-12478.

Kipfer, R., Aeschbachhertig, W., Baur, H. et al., 1994. Injection of Mantle Type Helium into Lake Van (Turkey) - The Clue for Quantifying Deep Water Renewal. *Earth and Planetary Science Letters*, 125 (1-4), 357-370.

Kopp, M.L. & Leonov, Yu.G., 2000. Tectonics. In: Outline of geology of Syria. Yu.G. Leonov (Ed.). Moscow, Nauka Publ., 7-104 (in Russian).

Kostenko, N.P., Panina, L.V. 2001. Some pecularities of neotectonic structure and development of Caucasus on conerosional stage. In: Tectonic of the Neogea: Common and regional aspects. *Contributions of Tectonic meeting*, Vol. 1. Moscow, 248-252 (in Russian).

Lasen, H.S. 2002. Investigations of rifted margins. In: Achievements and Opportunities of Scientific Ocean Drilling. *Spec. Issue of Joides Journa*l, 28 (1), 85-90.

Laubscher, H., 1988. Material balance in Alpine orogeny. *Geol. Soc. Amer. Bull.,*100, 1313-1328.

Lelikov, E.P., Emel'yanova, T.A., 2007. Volcanogenic complexes of The Okhotsk and Japan seas (comparative analysis). *Marine geology*, 42 (2), 294-303

Lonergan, L., White, N., 1997. Origin of the Betic-Rif mountains belt. *Tectonics*, 16 (3), 504-522.

Magmatic rocks. Vol. 5. Ultramafic rocks, 1988. E.E. Laz'ko and E.V. Sharkov (Eds.). Moscow, Nauka Publ., 500 p. (in Russian)

Marotta, A.M., Sabadini, R., 1995. The Style of the Tyrrhenian Subduction. *Geophysical Research Letters,* 22 (7), 747-750

Marson, I., Panza, G.F., Suhadolc, P., 1995. Crust and upper mantle models along the active Tyrrhenian rim. *Terra Nova*, 7 (3), 348-357.

Morales, J., Serran,o I., Jaboloy, A. et al., 1999. Active continental subduction beneath the Betic Cordillera and the Alboran Sea. *Geology*, 27 (8), 735-738.

Neotectonics, geodynamics and seismicity of the Northern Eurasian, 2000. A.F. Grachev (ed.). PROBEL, Moscow, 487 p. (in Russian).

Nikishin, A.M., Korotaev, M.V., 2000. History of the Black Sea basin formation. In: *General Problems of Tectonics*. Tectonics of Russia. Moscow, GEOS, 360-363 (in Russian).

Nikolaev, V.G., 1986. Pannonian Basin. Moscow, Nauka Publ., 120 c. (in Russian)

Papazachos, C.B., Hatzidimitriou, P.M., Panagiotopoulos, D.G. & Tsokas, G.N., 1995. Tomography of the crust and upper mantle in southeast *Europe. Jour. Geophys. Res.*, 100 (B7), 12405-12422.

Pino, N.A., Helmberger, D.V., 1997. Upper mantle compressional velocity structure beneath the west Mediterranean Basin. *Jour. Geophys. Res*., 102 (B2), 2953-2967.

Ponikarov, B.П., Kazmin, V.G., Kozlov, V.V. et al., 1969. Syria. Leningrad, Nedra Publ., 216 p.

Prilepin, M.T., Guseva, T.V., Lukk, A.A., Shaevcheko, V.I., 2001. Modern geodesic measuring and major geodynamic conceptions. In: *Tectonic of the Neogea: Common and regional aspects.* Contributions of Tectonic meeting, Vol. 1. Moscow, 135-138 (in Russian).

Rehault, T.Г., Moussat, E., Fabri, A., 1987. Structural evolution of Tyrrhenian back-arc basin. Marine Geology, 74 (1-2), 123-150.

Ricou, L.E., Dercourt, J., Geyssant, J. et al., 1986. Geological constrain on the Alpine evolution of the Mediterranean Tethys. *Tectonophysics*, 123(1-4), 83-122

Royden, L.H. Late Cenozoic tectonics of the Pannonian Basin System// *Tectonics*, 1989. V.8. N 1. P. 51-61.

Sharkov, E.V., 2000. Mesozoic and Cenozoic basaltic magmatism. In: *Outline of geology of Syria*. Yu.G. Leonov (Ed.). Moscow, Nauka Publ., 177-200 (in Russian).

Sharkov, E.V. & Svalova, V.B., 1991. Possibility that continental lithosphere is involved in subduction in back-arc spreading. *Intern.Geol. Review*, 33 (12), 1184-1198.

Sharkov, E.V. & Svalova, V.B., 2005, Late Cenozoic geodynamic of Alpine Folded Belt in connection to formation of withincontinental seas (petrologo-geomechanical aspects). Proc. of Higher Educational Establishments. *Geology and Exploration*, № 1, 3-11 (in Russian)

Shempelev, A,G., Prutsky, N.I., Feldman, I.s. & Kuchmasov S.U., 2001. Geologo-geophysical model along profile Tuapse-Armavir. In: T*ectonic of the Neogea: Common and regional aspects*. Contributions of Tectonic meeting, Vol. 2. Moscow, 316-320 (in Russian).

Shevchenko, V.I., Guseva, T.V., Lukk, A.A. et al., 1999. Present-day geodynamics of the Caucasus (results of GPS measurements and seismological data. *Physics of the Earth*, № 9, 3-18.

Spakman, W., van der Lee, S., van der Hilst, R., 1993. Travel-time tomography of the European-Mediterranean mantle down to 1400 km. *Phys. Earth Planet Inter*., 79, 3-74.

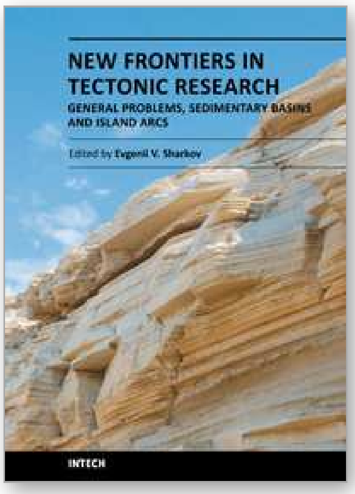
Storti, F., 1996. Tectonics of the Punta Bianca promontory: Insights for the evolution of the Northern Apennines Northern Tyrrhenian Sea basin. *Tectonics*, 14 (4), 832-847.

Trifonov, V.G., Dodonov, A.E., Sharkov, E.V. et al., 2011. New data on the Late Cenozoic basaltic volcanism in Syria, applied to its origin. *Journal of Volcanology and Geothermal Res*., 199. 177–192.

Zanemonents, V.B., Kotelkin, V.D., Myasnikov, V.P., 1974. About dynamics of lithospheric movements. *Physics of the Earth,* № 4, 43-54.

Zonenshain, L.r., Le Pichon, X. 1986. Deep basins of the Black Sea and Caspian Sea as remnants of Mesozoic back-arc basin. *Tectonophysics*, 123 (1-4), 181-212.

Zverev, S.M., 2002. Features of structure of sedimentary sequence and basement in frontal zone of the Cyprian arc (Eastern Mediterranean). *Oceanology*, 42 (3) 416-428.



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