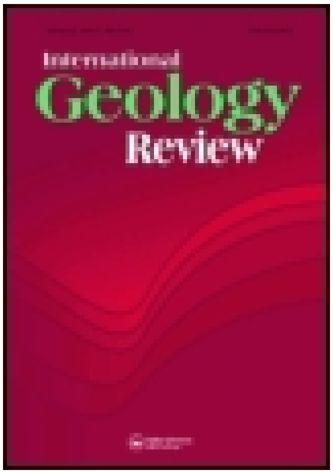

PROCESSES OF TECTONIC FLOW IN THE ALPINE BELT

V. S. Burtman [a]

[a] Institute of Geology , USSR Academy of Sciences , Moscow
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PROCESSES OF TECTONIC FLOW IN THE ALPINE BELT

V. S. Burtman
(Institute of Geology, USSR Academy of Sciences, Moscow)

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The tectonic structure of the Alpine belt was created as a result of the interaction of tectonic flowage directed toward the flanks of syntaxes. Structural arcs serve as good indicators of tectonic flowage. The patterns of surficial tectonic flowage and flowage within the deep portions of the lithosphere differ. Tectonic, seismic, and seismological data indicate that the principal level of structural disharmony within the continental crust, at which these patterns change, is located at a depth of 10-15 km, which often corresponds to the top of a low-velocity layer.

Tectonic flowage is the flow of rocks within a tectonic stress field.* Clear examples of tectonic flow of material with results visible at all levels from thin section to foldbelt are provided by serpentinite protrusions and melanges. In this article, we discuss tectonic flowage at the megalevel. New fold systems are created and earlier fold systems are deformed as a result of such flowage. Tectonic flowage is accompanied by displacements along strike-slip faults. Long-distance overthrusting of rock masses is often a consequence of tectonic flowage. The structural consequences of tectonic mass flowage of the crust have been described for many regions [2, 11, 17, 28, 22, etc.].

Structural arcs are good indicators of tectonic mass flowage. Among them, it is necessary to differentiate arcs of primary and secondary origin. If the velocity of tectonic flow falls from the center of the flow to the flow's boundaries, folds develop that have a primarily arcuate shape with the apex of the fold oriented in the direction of the flow. Secondary structural arcs are formed as a consequence of the deformation of previously created folds, fold systems, or tectonic zones which had a different shape and different trend of tectonic elements prior to the deformation. Studying the pre-fold magnetization of the rock is useful for differentiating primary and secondary structural arcs. Such studies allow us to decide whether the tectonic zones underwent rotational movements within the plane (secondary arc) or whether no such rotations took place (primary arc). Paleomagnetic investigations have been carried out for the majority of the structural arcs of the Alpine belt.

We will discuss the results of an investigation of the structural arcs located between the Alps and the Himalayas, using these arcs as indicators of directions of tectonic flowage.

*Tectonic flow is regulated by the rheological characteristics of rocks, which has the properties of a viscous, viscoelastic, or elasticoplastic body. When using these rheological concepts to describe tectonic processes, it should be kept in mind that the viscosity of the rock, in the majority of cases, is not an intrinsic viscosity; rather it characterizes the quasiviscosity of an heterogeneous medium governed by a combination of different physical processes including, for example, the propagation of dislocations through the crystal lattice, intergranular gliding and displacement along fissures. This applies to such concepts as the viscosity of the earth's crust or the plastic viscosity of tectonic flow within a foldbelt. Bodies with relatively high viscosities (median massifs, blocks of crystalline rock) also move within tectonic flows. They can be considered rigid inclusions carried by the flow.

Structural Arcs and Tectonic Flowage

Three syntaxes created as a result of the collision of the Indian, Arabian, and Adriatic plates with Eurasia are located within the portion of the Alpine belt under consideration (Fig. 1). The Hindustan continent converged on Eurasia during the Eocene, and the collision of these continents began at that time. Magnetic anomalies on the Indian Ocean floor indicate that the Indian plate has moved 2000-2500 km to the north after the collision began. Initially, the direction of movement of the Indian plate was northeast, corresponding approximately to the direction of the Chaman strike-slip fault. Counter-clockwise rotation of the Indian plate took place during the collision process. As a result, the direction of movement of the plate during the Late Miocene changed to north-northwest, similar to the strike of the Pamir-Karakorum strike-slip fault. A consequence of the rotation of the Indian plate was that the Assam salient of the Hindustan continent advanced to the north a significantly greater distance than did the Punjab salient. This probably caused a tectonic mass flow from the region of the Himalayas toward the Punjab syntaxis.

The formation of the Hindukush-Karakorum structural arc, which envelops the Punjab salient of Hindustan, was a consequence of the movement of the Hindustan continent to the north (Fig. 1, *15*). Paleomagnetic investigations [35] in Kashmir on the eastern wing of this arc indicate that the arc is of secondary origin. These data suggest formation of the arc after the Eocene. The Hindukush-Karakorum arc probably formed shortly after the convergence of the Indian plate and Eurasia during the Eocene. The Punjab salient was subsequently underthrust beneath the Hindukush-Karakorum structural arc, which eventually ended up in an allochthonous position. This facilitated the preservation of the original shape of the structural arc. Folding ceased here during the Neogene [8]. Only the flanks of Hindukush-Karakorum arc were subsequently reworked by shearing.

The Punjab salient advancing toward the north broke up crustal blocks with rigid, Pre-Alpine basement. The Tibetan block moved to the east and the South Afghanistan block moved toward the west [24, 38]. The Pamir structural arc (Fig. 1, *14*), which is bordered by the tectonic zones of the Pamirs, western Kuenlun, and Badakhshan, is located at the apex of the Punjab syntaxis. This is a secondary arc [3, 6], disharmonious to, and tighter than, the Hindukush-Karakorum arc. The relationships between these arcs are probably a consequence of mass flow into the Pamir region from the Himalayan-Tibetan zone of plate convergence.

The tectonic zones of the Pamirs had an east-northeast-west-southwest strike during the Paleogene. This was probably the original strike of the Post-Alpine fold systems of the Pamirs [3]. These systems developed primarily during the Neogene. In the central Pamirs, a multi-layered system of nappes and recumbent folds were created during this time, and then compressed into vertical folds. Folds formed in the southeastern Pamirs at the same time [19]. The folds in the outer zones of the Pamirs were created during the Pliocene and were deformed into the Pamir structural arc at the end of the Pliocene and during the Quaternary. The formation of the arc probably took place by way of a tectonic outflow from the advancing salient of the Indian plate [6].

The displacement of the Pamir masses to the north was accompanied by spreading toward the sides. The flow of the Pamir masses toward the northeast overlapped the boundary of the Tarim massif, which had a relatively thin cover of Mesozoic and Cenozoic deposits. Thick Mesozoic and Tertiary deposits lay to the north and northwest of the advancing Pamirs. These deposits were partially overlapped during the advance of the Pamirs to the north and were, to a significant degree, merged with the Pre-Alpine basement, pinched in a western direction, and compressed, creating the folds of the Tajik structural arc which envelop the allochthonous mass of the Pamirs from the west (Fig. 1, *13*). This structural arc is outlined by the folds of the Tajik depression and southwestern Hissar.

To the west, the Pamirs arc is joined with the broad Darvaz-Kopetdag structural arc (Fig. 1, *12*), outlined by folds of southwestern Darvaz, northern Afghanistan, northern Khorasan, and Kopetdag. The tectonic zones within the arcs already had arcuate shapes during the Cretaceous. The apex of this arc was turned toward the south. The northern Pamirs were part of the eastern wing of this arc [3]. Most recently, the eastern wing of this arc has been curved into the Pamir arc. During the deformation process, the Darvaz-Kopetdag arc became tighter. The eastern

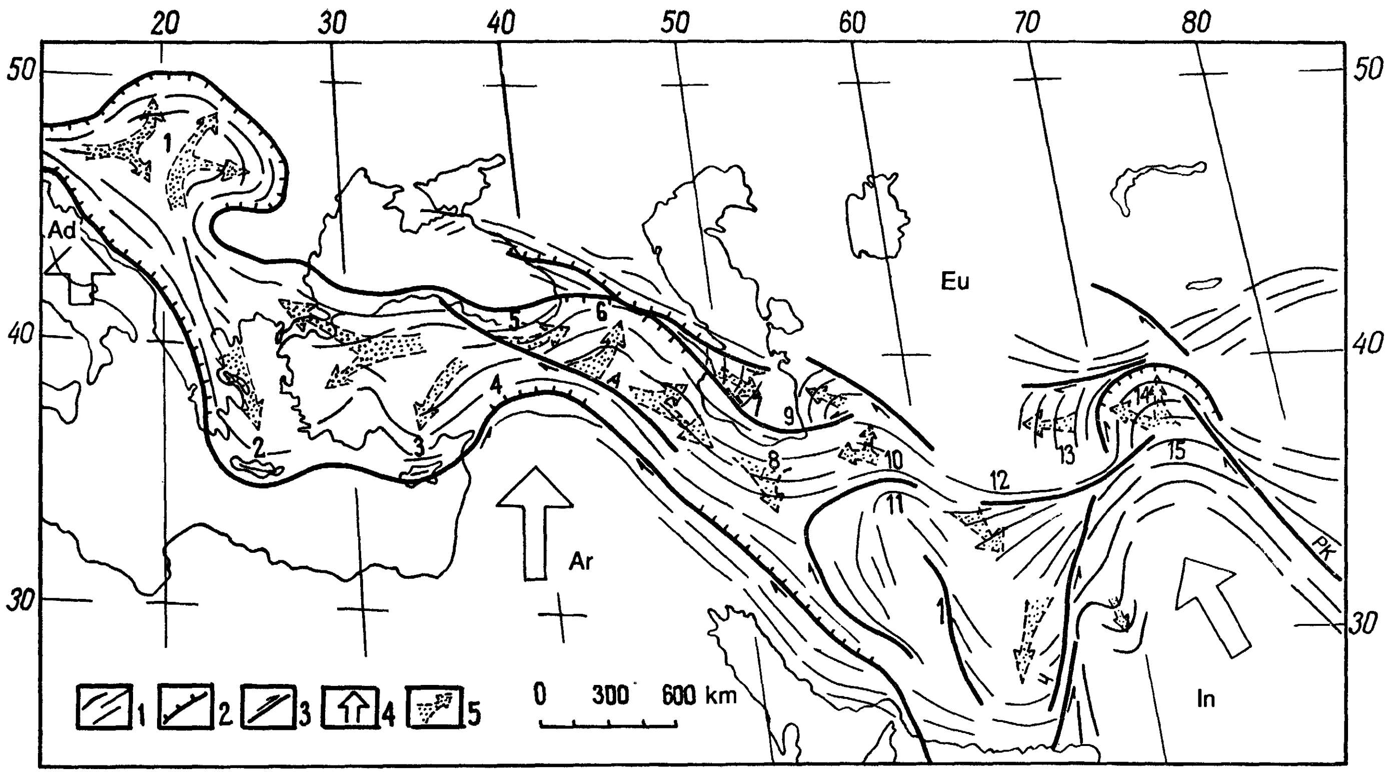

FIGURE 1. Structural arcs and tectonic flows: 1) trend of folds; 2) nappes, overthrust front; 3) strike-slip faults; 4) directions of movement of lithospheric plates relative to Eurasia during recent times. Structural arcs (numbers on maps): 1) Carpathian loop, 2) Cretan, 3) Cyprus, 4) Taurus, 5) Trabzon, 6) Lesser Caucasus, 7) southern Caspian, 8) Elbrus, 9) western Kopetdag, 10) Khorasan, 11) Lut, 12) Darvaz-Kopetdag, 13) Tajik, 14) Pamir, Hindukush-Karakorum. Lithospheric plates and microplates: Ad) Adriatic, Ar) Arabian, In) Indian. Strike-slip faults: A) Anatolian, PK) Pamir-Karakorum, Ch) Chaman.

(Darvaz) wing of the arc was reworked significantly more intensively than the western wing.

Turning now to the Arabian syntaxis, convergence of the Arabian continent and Eurasia took place during the Early Oligocene. The Arabian continent moved in a northwestern direction during the Oligocene. It forced back the crustal masses of Asia Minor to the northwest and unrolled them in a counter-clockwise direction. Tectonic flows were created which led to the formation of the Carpathian loop [7, 29].

The Carpathian loop (Fig. 1, *1*) was formed by fold systems of the western, eastern, and southern Carpathians. These fold systems belong to the Carpathian-Pontian branch of the Alpine belt. The Carpathian loop has a secondary origin. The structural and facial zones defining this loop trended in a southeastern direction during the Late Cretaceous. The principal deformation events resulting in the formation of the structural loop took place during the Late Oligocene and Early Miocene. The development of the Carpathian loop was accompanied by sinistral strike-slip displacement of the Carpatho-Pontian branch of the Alpine belt relative to the Dinaric-Taurian branch. The internal space of the loop was filled as a result of a tectonic mass flow of the crust along the foldbelt.

This flow also engulfed median massifs with Pre-Alpine basement (Bihar, Mechek) which were displaced northward into the region of the Carpathian loop. The fate of the Haemus-Tatra crustal block is interesting. The Mesozoic rocks of the Dinaric-Taurian branch of the Alpine belt were formed at the African margin, and the rocks of the Carpatho-Pontian branch were formed at the Eurasian margin of the Mesozoic Tethys. An exception to this rule is the Haemus-Tatra block, which is included within the Carpathian fold system but which is composed of rocks from the African margin of the Mesozoic Tethys. During the formation of the Carpathian branch, the Haemus-Tatra block underwent counter-clockwise rotation about an axis situated at the boundary of the western Carpathians with the Eastern Alps. As a result, the eastern margin of the block described an arc of more than 600 km and the block itself was displaced farther northward. During the formation of the Carpathian structural branch, the marginal structures of the Eurasian continent were compressed, twisted, partially crushed, and thrust far into this continent. The thickness of the allochthon is 10-15 km [7, 29].

During the Early Miocene, the direction of movement of the Arabian continent relative to Eurasia switched to north-northeast and formation of the Arabian syntaxis began. The vertical folds and overthrusts of the eastern Taurus, southern Kurdistan, and Zagros were formed first. These folds outline the eastern Taurian arc (Fig. 1, *4*), which envelops the northern projection of Arabia. The eastern Taurian arc was created during the Miocene. The formation of folds within the arc continued into Quaternary time [9]. Within the Lesser Caucasus and Iran to the north, northwest-southeast folds were created probably during the Miocene.

During the formation of the syntaxis, tectonic flows in the Alpine belt were directed toward the flanks from the apex of the syntaxis. To the east, the space between the Arabian and Hindustan platforms serves as a discharge region for the tectonic flow. Mass outflow took place by slip along the Anatolian strike-slip fault. To the north of this fault, Trabzon, Lesser Caucasian, and Elbrus conjugate secondary arcs were formed (Fig. 1, *5*, *6*, *8*).

The Lesser Caucasian and Trabzon arcs have a secondary origin (Fig. 1, *4*). The same is probably true of the Elbrus arc. The Lesser Caucasian, Trabzon, and Elbrus arcs are disharmonious relative to the Eastern Taurian arc. They formed as a consequence of colliding tectonic mass flows along the Alpine belt from regions of convergence of the Arabian plate and the Lut microplate with Eurasia. The beginning of the formation of these arcs can be assigned to the Miocene; the major stage in their development was during the Pliocene.

The formation of the arcs of Asia Minor was accompanied by thrusting of the apexes of the arcs onto the Rioni-Shirvan microplate, which has a Pre-Alpine basement. This microplate then began to move up under the tectonic zones of the Greater Caucasus, and folding of the latter began. These movements took place during the Pliocene and continued during the Quaternary. At the end of this process, the overthrusts of both margins of the Rioni-Shirvan microplate met and the Greater Caucasus thrust system partially overlapped the Lesser Caucasus system. The convergence of the mutually approaching thrusts was accompanied by the extrusion of rocks from the cover of the Rioni-Shirvan microplate in a southeastern direction. The southern Caspian structural arc was created as a result [11].

The tectonic mass flow within the belt captured median massifs (microplates) with rigid Pre-Alpine basement. This was the case with the Lut microplate, which is situated between the Arabian and Punjab syntaxes. During the formation of the syntaxes, this microplate advanced in a northward direction and rotated. The consequences of this movement included the formation of the Lut arc (Fig. 1, *1*), the emergence of the Khorasan arc (Fig. 1, *10*), and then the emergence of the western Kopetdag arc (Fig. 1, *9*). The origin of the latter was similar to the origin of the South Caspian arc [11].

The Khorasan arc envelops the northern margin of the Lut massif. As the massif advanced to the north, a mass discharge toward the western portion of the Khorasan arc took place. As a result, a disharmony arose in inner structure of the arc and its outer structure became tighter than the inner portion. The cause of the disharmony was the rotation of the Lut massif in a counter-clockwise direction [27, 30].

The Lut structural arc, outlined by folds developed within the Alpine cover of the Lut massif (Fig. 1, *11*), is interesting. These folds may have had primarily arcuate axes conforming to the western boundary of the massif. As a consequence of the rotation of the Lut massif, it is probable that these folds were subsequently reoriented, and that the gentle, primary arc was transformed into a tighter arc. Data on the neotectonics of Asia Minor [40] are evidence in favor of a tectonic flow from the apex of the Arabian syntaxis to the region of the Cyprus arc (Fig. 1, *3*).

Events within the Arabian and Apulian syntaxes influenced the formation of the younger structure of the Carptho-Balkan region. According to paleomagnetic data [35, 43], the Adriatic plate rotated in a counter-clockwise direction during Cretaceous times. Later, it was displaced to the north. During the Miocene, the collision of this microplate with Eurasia began and the Apulian syntaxis was created. The interaction of the colliding flows and counterflows directed from the Arabian and Apulian syntaxes had consequences for the formation of the Cretan structural arc and the expansion of the Carpathian loop. These processes were accompanied by the development of overthrusts at the outer margins of the arcs.

The Cretan structural arc (Fig. 1, *2*) was formed by rocks of the Helleno-Cretan fold system. This is a secondary arc which was created after the Carpathian loop. Judging from paleomagnetic data [34], the Cretan arc began forming during the Miocene and continued to develop during Pliocene-Quaternary times. The formation of the structural arc took place simultaneously with downwarping of the superimposed Aegean basin, expansion of the basin toward the south, and its deepening [26].

Lastly, expansion of the Carpathian loop took place and the Pannonian superimposed basin was formed. The expansion of the Carpathian loop in a northeastern direction, the advance of the Cretan arc to the southwest, and the formation of the Pannonian arc and the Aegean basin in the rear were all interrelated processes. They were probably the result of a left-lateral movement of the Carpatho-Balkan and Dinaric-Hellenic fold systems relative to one another, a movement which continued into the Miocene and Quaternary [7, 29]. This displacement created the conditions for extension of the Carpathian and Cretan structural arcs in the rear. This stimulated the rise of the mantle diapirs, the formation of basins and the development of vulcanism within the basins created.

Depth of Tectonic Flowage

The tectonic flows mentioned are consequences of interactions between lithospheric plates. It is therefore logical to infer that depths of flowage were commensurate with the thicknesses of these plates. It seems just as logical to conclude that surficial flows and flows in the deep regions of the lithosphere differ. The principal factors which determine the rheological properties of the rocks within the deep regions of the lithosphere are temperature and pressure. The influence of these factors result in relative uniform rheological properties for different rocks as a consequence of metamorphic transformations. Changes in *PT*-conditions within deep flows are gradual.

Surficial flowage is the flow of discrete geological bodies which preserve their individual rheological properties. These properties are determined

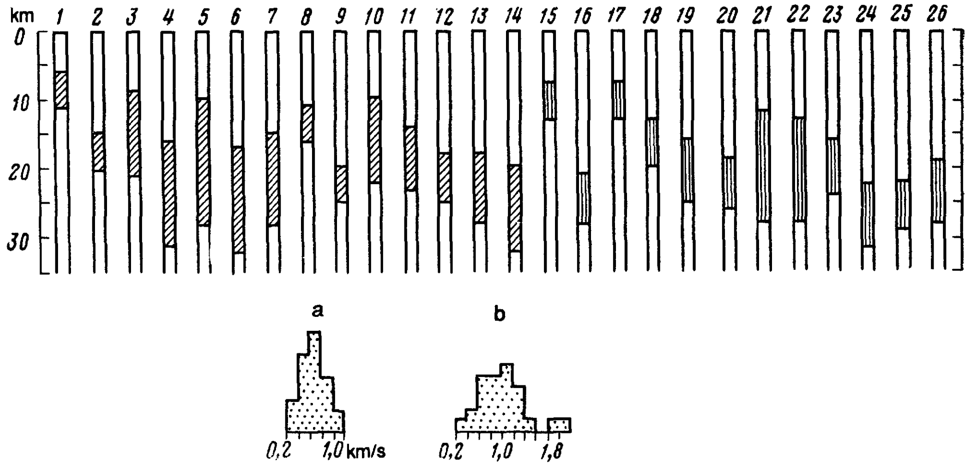

FIGURE 2. Position of low-velocity layers in the crust of the Alpine belt, according to data from [12]. Diagonal hachuring indicates layers with low velocity of seismic waves from seismic explosion data. Vertical hachuring) same from earthquake seismological data: 1-6) Alps, 7-8) Carpathians, 9) Crimea, 10-12) Greater Caucasus, 13-14) Kurin depression; a, b) histograms showing drop in velocities of seismic waves at the top (a) and at the bottom (b) of the low velocity layer.

by primary composition for the majority of rocks. Formations involved in surficial flowage often show contrasting rheological properties. These include carbonate and clastic strata, evaporites, crystalline rocks, etc. As a result, the rheological gradient in a surficial flow is significantly higher than in a deep flow. As a consequence of these differences, a surficial current should have a pattern different from that of a deep flow.

Another important difference in the case of surficial flowage is the presence of a free surface, as a result of which mass displacement is usually compensated for by the formation of mountain systems. One can conclude that a disharmony should exist within the lithosphere between surficial and deep flows. We will attempt to estimate the depth of this discontinuity. In other words, we will attempt to determine the depth to which one can extrapolate data on the mass flowage of the crust obtained by studying the structure of fold systems. Such an estimate can be made on the basis of studies of overthrusts, earthquake investigations, and seismic data on the structure of the crust.

Overthrusts suggest a tectonic stratification of the upper portions of the crust in folded regions [22]. As was pointed out above, the formation of overthrusts is a consequence of tectonic mass flowage. Consequently, the depth of occurrence of overthrust structures will be commensurate with or less than the depth of the surface of tectonic flow. The problem consists in determining the thicknesses of the allochthonous masses. It is clear that these thicknesses cannot be computed by adding up the thicknesses of the tectonic sheets. Overthrusts often adjoin one another; in accretionary prisms, this is the rule. Tectonic erosion, as well as variable thicknesses of tectonic sheets, also interfere with estimation of the thickness of an allochthon. However, the thicknesses of overthrusts exposed within a tectonic window can be used to estimate *minimal* depths of tectonic flow.

According to structural-geological data, the thicknesses of the Helvetic and the lower and upper Pennine nappe ensembles in the Alps are 8 or 9 km. However, the total thickness of allochthonous masses in the Alps is estimated to be 12-15 km. Estimates of the depth to the bottom of the Dinaride and Hellenide allochthon are similar [5]. The bottom of the overthrusts of the Carpathian flysch also lies at a depth of more than 10 km [28].

In the last decade, the method of multichannel vibroseismic profiling has found widespread application in the study of foldbelts. In the western Carpathians, the bottom of the Carpathian overthrusts into the European platform was found using this method. The top of the autochthon lies at a depth of 10-15 km and can be traced under the fold system

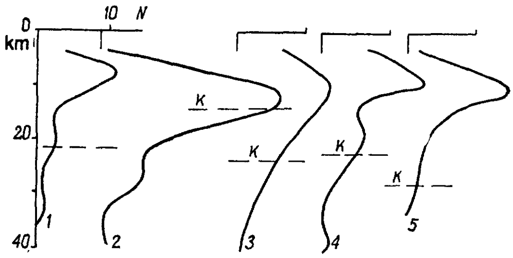

FIGURE 3. Distribution with depth of crustal earthquake foci in the Alpine belt over 100 years of observations (according to data from [23]). 1) Eastern Alps; 2) Dinarides; 3) Eastern Carpathians; 4) Balkanides; 5) Caucasus. K) Conrad discontinuity.

for a 100 km from the overthrust front [42]. The results of magnetotelluric investigations [20] allow us to conclude that the East Carpathians were thrust 60 km onto the East-European platform and that the bottom of the allochthon lies at a depth of 10 km. A similar pattern can be seen in other foldbelts. In the Canadian Rocky Mountains, an allochthonous sheet is overthrust 300 km; the thickness of the allochthonous mass is approximately 10 km [39]. In the southern Appalachians, the bottom of an allochthon can be traced by vibroseismic profiling for a distance of 250 km from the nappe front. The top of the autochthon lies as deep as 15 km [31, 33].

Low-velocity layers, characterized by lower seismic wave velocities than the layers above and below, have been discovered in the upper portion of the crust by seismic and seismological investigations in many continental regions. The positions of the low-velocity layers in the portion of the Alpine belt under consideration are shown in Figure 2. The drops in the velocities of longitudinal seismic waves range from 0.2 to 1.0 km/s at the top and from 0.2 to 2.0 km/s at the bottom of such a layer (Fig. 2a, b). The thicknesses of low-velocity layers within the crust of the Alpine belt range from 5-15 km. They usually lie within a depth interval between 10 and 25 km and are located between the Forsh (upper) and Conrad discontinuities [15]. In the Pamirs and in Tibet, where the crust is anomalously thick (60-70 km), a low-velocity layer occurs below 30 km [14, 41]. In the Himalayas and Karakorum, where the thickness of the crust also reaches 70 km, two low-velocity layers have been found, one in the depth interval of 3-22 km, the other in a lower portion of the crust [32, 37].

Let us turn our attention to the locations of earthquake foci within the earth's crust (Fig. 3). A comparison of Figures 2 and 3 permits us to conclude that the depth interval at which low-velocity layers occur is characterized by the greatest density of earthquake foci. This conclusion has been verified for the Pamirs, where a low-velocity layer occurs at an anomalously low level, within the interval of 30-40 km. The maximum density of crustal earthquakes in Pamir coincides with just this interval [23]. The maximum seismic energy released, coinciding as it does with the position of the low-velocity layer probably indicates an elevated rate of tectonic deformation at this level.

Interesting results were obtained from observations of weak crustal earthquakes (K = 7-12) within a geographical network situated in the zone of convergence of the Pamirs and Tien Shan in the Garm region [21]. Investigation of the distribution of weak earthquakes in different levels of the crust have shown that variations in earthquake activity over time in the upper 10-km layer differ from area to area within the region, while the earthquake regime at a depth of 10 km is homogeneous over the entire area. This pattern probably reflects relative uniform properties of the medium at depths greater than 10 km.

Many explanations for the low-velocity layer phenomenon have been proposed. The creation of such a layer in the Central Alps is thought to be a consequence of the overthrusting of mantle rocks upon low-velocity formations [22, 26]. Plastic intrusions have also been used to explain them. Meanwhile, low-velocity layers have been discovered not only in fold regions but also in platform, including shield, sections [12]. There should therefore be a common physical principle behind the creation of low-velocity layers in the crust.

The decrease in the velocity of the layer could be either a consequence of partial melting of the layer or of shattering with filling of the fissures and pores with fluids. The conditions in the upper portion of the crust would not support widespread occurrence of melting processes. The variant involving a water-saturated, highly porous and highly fissured layer is the most plausible. In the Nikolayevskiy-Sharov model [15], the low-velocity layer is a zone of dilatant (open) fissures. In this theory, the phenomenon of dilational opening of fissures takes on appreciable importance at depths exceeding 10 km. An increase in the fissuring of rock with depth was observed during the drilling of the Kola superdeep hole [10]. A strong rarefaction of the medium and the emergence of low-velocity layers also occurs in the focal zones of earthquakes [23]. According to the calculations of Kropotkin and Valyayev [13], the fluid pressure is close to geostatic beginning at depths of 10-15 km, making conditions favorable for hydrofracturing of the strata. The widespread occurrence of cataclasis in combination with elevated pressures of pore and fissure fluids is probably the case of the lowered viscosity, more precisely, pseudoviscosity, of the geomaterial in the low-velocity layers.

The data presented suggests that there is an horizon within the continental crust at a depth interval of 10-25 km in which the viscosity of the geomaterial is decreased and in which conditions for the flow of material differ significantly from those associated with the movement of masses in the surficial layer. Data on the thickness of tectonic allochthons and the results of seismic and seismological investigations of the structure of the earth's crust give consistent results. They allow us to conclude that a change in the style of deformation characteristic of the surficial layer of crust occurs at a depth of 10-15 km. Regional geophysical investigations on the continents confirm this conclusion [16]: At the indicated depth, the heterogeneity of the geophysical fields decreases, and there is an obliteration of the density and magnetic inhomogeneities characteristic of the surficial structures. All of this allows us to assume that the 10-15-km level (the top of the low-velocity layer) is the principal level of structural disharmony within the continental crust. Below this level, tectonic inhomogeneities due to the formation of internal structure within fold systems probably disappear (although the system themselves have isostatic "roots of mountains" at the Mohorovicic discontinuity).

The considerations cited above provide a basis for estimating the depth of surficial tectonic flows within the Alpine belt. The fold structure of the belt expressed at the earth's surface is a consequence of such flows. It can be assumed that the pattern of tectonic flowage found in surficial structures is maintained down to a *major level of structural disharmony* which occurs at a depth of 10-15 km. Below this level, the overall direction of tectonic mass flow is maintained, but the pattern of deep currents is different.

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