Neotectonic uplift and mountain building in the Alpine-Himalayan Belt

V.G. Trifonov, S.Yu. Sokolov and D.M. Bachmanov

The book describes neotectonic uplifts producing mountain building in the Alpine-Himalayan Belt. This process began in the Oligocene as formation of local uplifts in zones of concentration of collision compression and accelerated in the Pliocene and Quaternary as the isostatic effect of decrease of density of the uppermost mantle and the lower crust by partial replacing of the lithospheric mantle by the asthenosphere material and retrograde metamorphism of high-metamorphosed rocks by asthenosphere fluids. These changes were initiated and kept up by the sub-lithosphere upper mantle flows that spread, according to the seismic tomography data, from the Ethiopian-Afar superplume and were enriched by fluids, reworking the transitional mantle layer beneath the future mountain belt. The upper mantle flows not only move lithosphere plates with all plate-tectonic consequences of this process, but also initiate transformations of the lithosphere that results in vertical movements producing mountain building. The book is intended for wide circle of geoscientists.

Keywods: Oligocene to Quaternary, neotectonics, uplift, mountain building, molasses, seismic tomography, lithosphere, asthenosphere, mantle flows

Content

Introduction

Obruchev (1948), introducing the terms *neotectonics* and *neotectonic* epoch, applied them to the process leading to the formation of the present-day topography that is distinguished by high-mountain systems, which did not exist earlier in the Mesozoic and Cenozoic geological history. In this book, we consider the tectonic movements, which gave rise to the contemporary topography of the central Alpine–Himalayan Orogenic Belt between the Carpathians and Balkan–Aegean region in the west and the Tien Shan, Kunlun, Tibet, and Himalayas in the east (Fig. 1)[1].

In the first part of the book, we describe the history of neotectonic (Oligocene–Quaternary) movements that produced uplift of orogenic structures of the belt. Analyzing neotectonic evolution of the Central Tien Shan, Pamirs, Great Caucasus, and finally the orogenic belt as a whole, we show that their evolution includes two main stages. During the first long-time stage that lasted from Oligocene till the end of Miocene and even Pliocene in some regions, local uplifts formed. They were usually not higher than middle-level mountains (< 1500 m) and formed under collision compression as the results of isostatic compensation of thickening of the Earth's crust in zones of concentrated deformation. During the second Pliocene–Quaternary stage, the height of the mountains increased 2–3 times. This intensification of tectonic uplift producing mountain building can not be explained by effects of the collision compression. It was caused by a decrease in the density of the crust and upper mantle under the effect of the asthenosphere, which was activated by fluids.

The second part of the book is devoted to deep-seated sources of the neotectonic processes mentioned above. The analyzing seismic tomography data demonstrate two important features of the mantle. First, in the eastern (Indonesian) part of the Alpine-Himalayan Belt, where subduction has continued till now, the higher-velocity subducted slabs became approximately horizontal at the depths of about 400–700 km and these sub-horizontal lenses spread beneath the adjacent continental upper mantle. The same continuations of the subducted slabs are known in the North-Western Pacific, where they were termed as stagnant slabs (Fukao et al., 2001), or big mantle wedges (BMW) (Zhao, 2009; Zhao et al., 2010). Second, in the more western mountain part of the Alpine-Himalayan Belt, sub-lithosphere low-velocity (hot and lower-dense) mantle flows were identified. They begin in the Ethiopian–Afar superplume rising from the lower mantle and spread beneath the orogenic belt.

[1] For the sake of brevity, these segments will be called further merely the Alpine–Himalayan Belt.

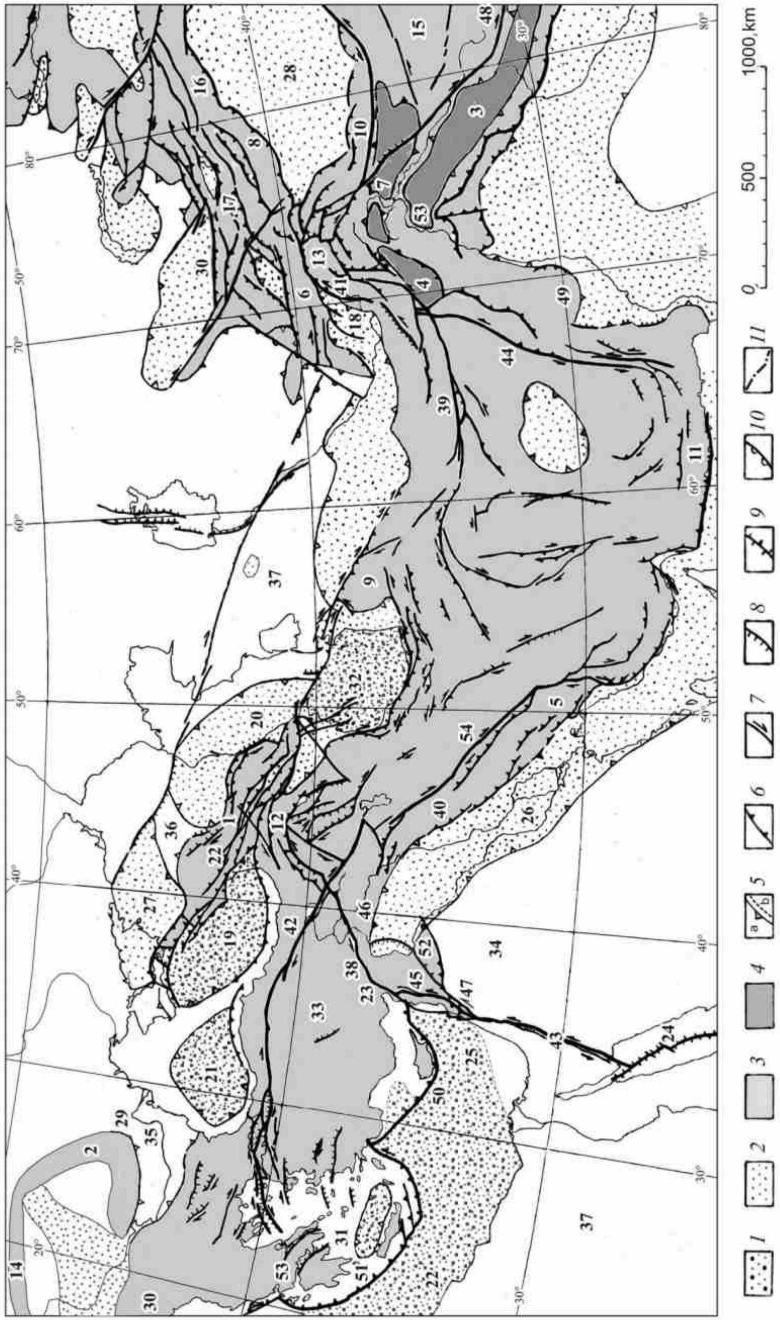


Fig. 1. Pliocene–Quaternary structure of the Alpine-Himalayan Belt from Eastern Mediterranean to Central Asia, modified after (Trifonov et al., 2012.)
1, basins with sub-oceanic crust; *2*, major basins on continental crust; *3*, mountains (in fig. 1) and uplands not higher than 1.5 km (in fig. 16,17, & 18); *4*, highest mountains; *5*, boundaries of mountains and basins: (a) in fig. 1 and (b) in fig. 16,17, and 18; *6–10*, faults (the largest faults are shown by thickened lines in fig. 1): *6*, thrust or reverse fault, *7*, strike-slip fault, *8*, normal fault, *9*, extension fault, *10*, flexure; *11*, the Tesseire-Tornquist lineament (in fig. 16,17, & 18)

We suppose that the elongated Ethiopian-Afar super-plume developed as a more or less stationary structure at least from the end of the Paleozoic. The portions of moving Gondwana, which turned out to lie above the superplume, underwent rifting that developed into spreading that formed the Tethys Ocean. A flow of heated asthenosphere material from the superplume caused moving of torn-off fragments of Gondwana to the north-east toward Eurasia. The oceanic Tethyan lithosphere subducted there, and the Gondwanan fragments accreted to Eurasia. As a result, series of microplates, separated by sutures, accretionary wedges, and magmatic bodies related to different stages of the Tethyan evolution, formed on the place of the future mountain belt. Probably, the mountain segments of the belt had previously the same structure as the south-eastern Indonesian segment, where subduction has continued till now, i.e., the subducted slabs transformed there at the depths of 400–700 km into the BMW that extended beneath the future mountain belt.

Closure of the Tethys and collision of the Eurasian and Gondwanan lithosphere plates decelerated their convergence, but the hot asthenosphere flows from the Ethiopian-Afar superplume probably prolonged the former movement and gradually spread under the entire orogenic belt. On moving, the sub-lithosphere flows were enriched in aqueous fluids that could derive from the former BMW lenses related to subduction zones. The asthenosphere, activated in this manner or its fluids penetrated into the lithosphere and produced its softening and detachment that facilitated deformational thickening of the Earth's crust and, correspondingly, the tectonic uplift in areas of maximum compression. During the first stage, it was the single or, at least, main source of the rise. During the second stage (the last 5–2 Ma), the deformational effect was supplemented by two other processes that were initiated by the sub-lithosphere flows and their fluids. The first process was the partial replacing of the lithosphere mantle by the lower-dense asthenosphere material and, as a result, decrease of density of the uppermost mantle.

The second process was the retrograde metamorphism of high-metamorphosed rocks of the crustal origin within the lower crust and near the crust-mantle boundary

In Fig. 1: *Mountains*: (1) Greater Caucasus, (2) Eastern Carpathians, (3) the Himalayas, (4) Hindu Kush, (5) Zagros, (6) Western Tien Shan, (7) Karakoram, (8) Kokshaal, (9) Kopet Dagh, (10) Kunlun, (11) Makran, (12) Lesser Caucasus, (13) Pamirs, (14) Northern Carpathians, (15) Tibet, (16) Khan Tengri, (17) Central and Eastern Tien Shan, (22) Elbrus, (30) Dinarides, (45) Coastal Range (in Syria), (47) Lebanon Range; *basins*: (18) Afghan-Tadjik, (19) Eastern Black Sea, (20) Terek-Derbent, (21) Western Black Sea, (23) Misis-Andirin (paleo-basin), (24) Red Sea (rift), (25) Levantine, (26) Mesopotamian, (27) Azov-Kuban, (28) Tarim (microplate), (29) Focsani (a part of the Carpathian Foredeep), (31) Aegean, (32) South Caspian; *platforms*: (33) Anatolian, (34) Arabian, (35) Moesian, (36) Scythian, (37) African; *fault zones*: (38) East-Anatolian Zone, (39) Herat, (40) Main Recent Fault of Zagros, (41) Darvaz, (42) North-Anatolian Zone, (43) Dead Sea Transform, (44) Chaman, (46) South Taurus; *tectonic zones*: (48) Indus-Tsang Po, (49) Quetta, (50) Cyprus Arc, (51) Crete-Hellen Arc, (52) Palmyrides, (53) Hellenides, (54) Sanandaj-Sirjan with participation of the asthenosphere fluids and, as a result, decrease of density of these rocks. The both processes produced additional rise of the land surface and caused the acceleration of total uplift of the belt during the Pliocene–Quaternary.

The determining role in this model of the Alpine-Himalayan Belt evolution belongs to the sub-lithosphere upper mantle flows that spread away from the Ethiopian-Afar superplume. We have analyzed the neotectonic and seismic tomography data on other territories and have found similar features. The other orogenic belts demonstrate acceleration of tectonic uplift in the Pliocene–Quaternary. Several super-plumes and upper mantle flows, which spread away from them, were found by the analysis of seismic tomography data. Comparison of the data gives a possibility to propose the following model of the global tectonic processes. Lithosphere plates are moved by the sub-lithosphere upper mantle flows because of viscous friction in the lithosphere-asthenosphere boundary. The flows spread away from the superplumes that represent the upwelling strands of the mantle convection. As a rule, zones of the lithosphere spreading do not correspond to the superplumes. The MORB volcanism is a result of adiabatic melting of the uppermost astenosphere and the lithosphere due to the extension. Because majority of the subducted slabs transforms into the BMW in the depths about 400–700 km, only a part of the subducted material penetrates into the lower mantle and is not enough to compensate a grow of the lithosphere in the spreading zones. The sinking strands of the mantle convection are represented also by volumes of dense and depleted upper mantle as well as high-metamorphosed basic rocks beneath cratons and collision zones. The plate tectonic mechanism is not the only result of the upper mantle flows. It is supplemented by tectonic processes that are caused by phase mineral transformations of the mantle and lower crustal rocks, by formation of the BMW and their fluid potential. The tectonic uplift producing mountain building is one of such processes.

In this paper, we use the new stratigraphic division of the Pliocene and Quaternary, confirmed in the 33rd IGC (www.stratigrahy.org). We use the following abbreviations in some tables and figures: E_1, Paleocene; E_2, Eocene; E_3, E_3^1, E_3^2, & E_3^3, Oligocene, correspondingly Lower, Middle, and Upper; N_1, N_1^1, N_1^2, & N_1^3, Miocene, correspondingly Lower, Middle, and Upper; N_2, N_2^1, & N_2^2, Pliocene, correspondingly Lower (Zanclean) and Upper (Piacenzian); Q_1, Q_1^1, & Q_1^2, Lower Pleistocene, Gelasian, and Calabrian, Q_2, Q_2^1, & Q_2^2, Middle Pleistocene, correspondingly lower and upper; Q_3, Upper Pleistocene; Q_4 – Holocene.

Part 1. Neotectonic development of the central part of the Alpine-Himalayan belt

1.1. Pre-history

The longitudinal tectonic zoning dominating in the orogenic belt is expressed in the progressive rejuvenation of the crust to the south and southwest in compliance with the evolution of the Tethys Ocean. Over its entire history, rifting passed into spreading at the south-western (in present-day coordinates) passive continental margin of the ocean, which was underlain by Precambrian basement. The continental fragments separated from Gondwana moved to the northeast, where the Tethyan oceanic lithosphere was subducted beneath the island arcs or active south-western (in present-day coordinates) margins of the northern plates and the fragments of Gondwana accreted to Eurasia. Because of this, the subduction zone jumped back to the rear sides of the accreted fragments. The recurrence of this process during formation of the Paleo-, Meso-, and Neotethys led to the consecutive attachment of new microplates (fragments of Gondwana) to the northern plates. These fragments were separated by sutures, accretionary wedges, and zones of subduction-related and collision-related magmatism and metamorphism. The poorly reworked fragments retained the platform tectonic regime. This process has developed since the breakdown of Pangea in the Carboniferous and became especially distinct in the Mesozoic and Cenozoic, when the northern plates merged to form the Eurasian plate.

In the present-day outlines, most of the orogenic belt is made up of tectonic units of the northern active margin of the Neotethys, whereas only a few mountain systems occur at its southern passive margin; the Himalayas and Zagros are the largest. The mountains of the northern margin of the belt (the Greater Caucasus, Kopet Dagh, Tien Shan, Northern Afghanistan, Northern Pamir, Kunlun, and Northern Tibet) are superposed on the Paleozoides, the participation of which diminishes westward. For example, the northern part of the Tien Shan is Caledonian, whereas the southern part is Hercynian. To the west, only part of the Hercynides entered into the belt, while their northern extensions formed the basement of the Turanian and Scythian post-Paleozoic platforms and Hercynides of Central Europe. The sutures and other structural elements of the Mesotethys are localized further to the south, while the Neotethys is situated in the extreme south. Relics of the backarc troughs either inheriting older Tethyan basins or partly superposed on other structural elements occur at the active margin. As a result of multiple closures of basins with oceanic and sub-oceanic crust, relics of the oceanic crust are retained in the lithosphere of the belt. They are detected as high-velocity bodies at various levels of the lithosphere and as xenoliths in igneous rocks.

From the Late Paleozoic to the Paleogene, the Tethys was a gulf of the Pacific extending to the north-west and narrowing in this direction. This is why the horizontal offsets during its closure and formation of the orogenic belt generally increased eastward. In the Late Cenozoic, this tendency was expressed in the magnitudes of lateral offsets increasing from the west-eastward both in particular structures, e.g., the greater magnitude of shear at the western flank of the Indian Plate as compared with the western flank of the Arabian Plate (Trifonov et al., 2002), and in variable dimensions of the belt's segments, which have shortened in the transverse direction to different magnitudes (Molnar, Chen, 1978; Bazhenov, Burtman, 1990).

The contemporary mountain edifices originated in different parts of the Alpine–Himalayan Belt asynchronously, but mostly in the Oligocene (Shultz, 1948; Trifonov, 1999). Therefore, the history of recent mountain building is considered below since the Oligocene; the Eocene is regarded as a preceding epoch. In the Eocene, the lithosphere of the future orogenic belt was a combination of microplates, sutures, accretionary wedges, and magmatic zones related to the earlier collision stages of the Eurasia and Gondwana plates. In the west and the center of the belt, the vast areas were territories of epicontinental and shallow-water marine sedimentation. Such seas covered the microplates and the former fold-thrust zones and spread over the neighboring Moesian, Scythian, Turanian, Arabian, and African platforms. In the east, marine sedimentation covered the Afghan–Tajik Basin and the western Tien Shan, extending up to the western parts of the Shu Basin and Tarim microplate (Dmitrieva, Nesmeyanov, 1982; Burtman, 1999).

The rest of the future High Asia (the Central and Eastern Tien Shan, Tarim, Pamir–Hindu Kush–Karakoram, Kunlun, and Tibet) was a land. The widespread granitic batholiths, the first phases of which are dated at the Cretaceous and the final ones at the Miocene, was one of a cause of relatively high standing of the Pamir–Hindu Kush–Karakoram region (Shvolman, 1977; Searle, 1991). In the Paleocene and Eocene, the territory of the Central and Eastern Tien Shan was a peneplain with relative elevations of a few hundred meters accepted as a pre-orogenic planation surface (Shultz, 1948; Trofimov, 1973; Chediya, 1986). The Paleocene and Eocene sequence is composed of mainly fine-clastic red beds (the redeposited Late Mesozoic weathering mantle) with basaltic flows in the lower part (see section 1.2). The

thickness of this sequence commonly does not exceed a few tens of meters and is greater than 100 m in some basins.

Deeper troughs with thinned (sub-oceanic) crust stand out against the background of the land and areas of epicontinental marine sedimentation (Fig. 2). These were relics of the Neotethys and backarc basins (Golonka, 2004; Kazmin et al., 2010). The Eocene relics of the Neotethys existed in the Trans-Himalayas (Indus–Zangpo Zone); to the south of Makran, where a basin was retained which would later become the periphery of the Indian Ocean; and between the Arabian Plate and the Sanandaj–Sirjan Zone of Iran, where relics of the basin comprise accreted Paleogene sedimentary rocks and Mesozoic ophiolites.

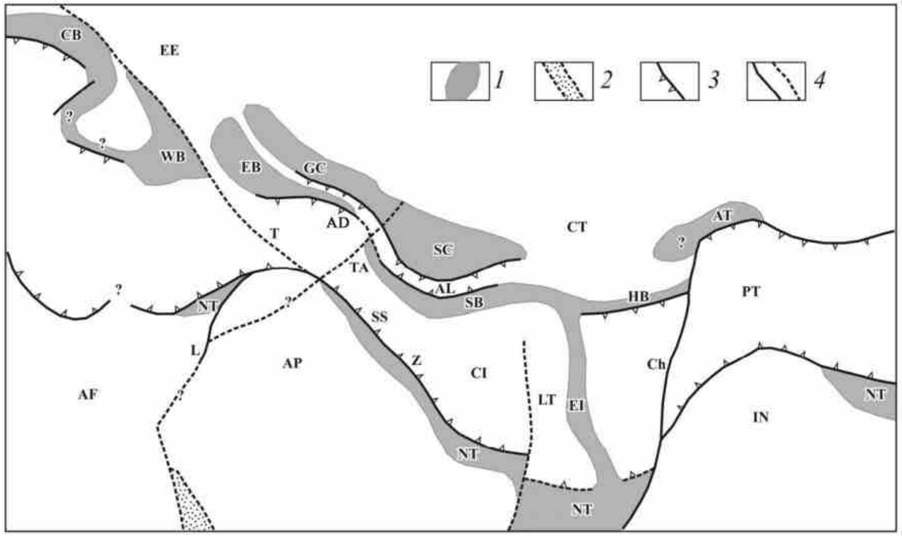


Fig. 2. Conceptual map of the basins with a thin (suboceanic?) crust in the Alpine-Himalayan Belt in the Eocene (~45 Ma ago), modified after (Golonka, 2004; Robertson et al., 2004; Alpine history…, 2007; Kaz'min et al., 2010; Trifonov et al., $2012_{1,2}$)

1, thin-crust basins; *2*, Red Sea proto-rift basin; *3*, major thrusts and subduction zones; *4*, major transform and other faults and their supposed continuations. AD, Adjaro-Trialetia; AF, African Plate; AL, Alborz; AP, Arabian Plate; AT, Afghan-Tadjik Basin; CB, Carpathian Basin; Ch, proto-Chaman Transform; CI, Central Iran Microplate; EB, Eastern Black Sea Basin; EE, East European Platform; EI, East Iranian Basin; GC, Greater Caucasus Basin; HR, Hari-Rud Basin; IN, Indian Plate; L, proto-Levant Transform; LT, Lut Microplate; NT, Neo-Tethys relics; PT, Pamirs, Tibet; SB, Sabzevar Basin; SC, South Caspian Basin; SS, Sanandaj-Sirjan zone; T, proto-North Anatolian Fault, extending into the Pechenega-Kamena Fault and the Tesseire-Tornquist Line; TL, Talysh; WB, Western Black Sea Basin; Z, Main Zagros Thrust

The region between the Taurus microplate and the northwestern margin of the Arabian part of the African Plate has a complex tectonic history (Krylov et al., 2005).

Before the end of the Cenomanian, this region was a part of the Tethys subducted to the north beneath the Taurides. In the Late Cenomanian–Early Turonian, the ensimatic proto-Cyprus arc arose, having separated the backarc trough from the Tethys, and the oceanic Trodoos Complex was formed there. In the Late Campanian–Early Maastrichtian, this complex was deformed and became a part of the arc as a paraautochthon. At the same time, the northeastern continuation of the arc and oceanic rocks in front of this arc were obducted on the margin of the Arabian Plate (Knipper et al., 1988). The retained Kilikia–Adana backarc basin deepened in the late Maastrichtian and continued to subduct beneath the Taurides. The pelagic cherty-clayey sedimentation therein is dated at the Paleocene–Middle Eocene in the Misis–Andirin melange complex of the South Taurus (Robetrton, 2000; Robertson et al., 2004). In the south, this basin connected with the Levantine and Ionian basins of the Mediterranean Sea, which developed at the southern passive margin of the Neotethys and probably were at that time shallower than now.

The earlier closure of the Neotethys in the south of the South Taurus Zone (northern flank of the Arabian Plate) as compared with its western part (East Mediterranean margin of the African Plate), as well as the greater thickness of the Upper Mesozoic and Cenozoic sediments in the Levantine Basin than in its eastern continental framework (Garfunkel, 1998) allow us to suggest that a transform-type structural boundary existed between them, at least beginning from the Late Mesozoic. This boundary followed from the South Taurus Zone along the present-day East Anatolian Fault Zone and extended southward along the continental slope of Eastern Mediterranean, where it is expressed in seismic sections as faults (Ben-Avraham et al., 2002; Ben-Gai et al., 2004). In the south, this zone probably merged with a proto-rift developing in the Late Cretaceous–Eocene partly on the place of the future Red Sea Rift (Almeida, 2010).

Among the backarc basins, the approximately W–E-trending Carpathian–Greater Caucasus system of troughs, extending from the outer Carpathian Zone to the proto-South Caspian Basin, was the largest (Kopp, Shcherba, 1993; Shcherba, 1994; Golonka, 2004; Alpine history…, 2007). The system was en echelon arranged: the troughs extended in the NW–SE direction in such a manner that the southeastern termination of each trough was situated to the south of the northwestern end of more eastern trough. The troughs were separated by relative uplifts (partly shoals) striking in the northwestern direction. The Sabzevar Trough, situated in the south and reaching Talysh in the west, extended to the east as the Herirud Trough and

connected with the pre-Makran relic of the Neotethys via the Central Iran Basin (Kazmin et al., 2010).

The origin of the backarc troughs remains conjectural. The Cretaceous ophiolite in the relics of the Sabzevar and East Iranian Troughs indicate spreading (Kazmin et al., 2010). As concerns the Carpathians–Greater Caucasus system, arguments have been stated about Cretaceous and even locally developed Late Jurassic rifting as a mechanism of sagging (Nikishin et al., 2001; Golonka, 2004; Alpine history…, 2007). These troughs inherited from the Cretaceous, however, did not show magmatic indications of spreading or deep rifting in the Eocene. In contrast, they underwent transverse shortening with deposition of flysch and volcanic activity in the adjacent territories. Therefore, deepening of troughs in the Carpathians–Greater Caucasus system in the Paleogene (Kopp, Shcherba, 1993) and similar deepening of the relict Kilikia–Adana Basin (Robertson et al., 2004) should be related to other causes, e.g., to compaction of the lower crust mafic rocks as a result of metamorphism rather than to ongoing extension.

The Paleogene troughs are not everywhere inherited from the Cretaceous. Indications of their superposition on the Upper Cretaceous shallow-water sedimentary rocks overlapping the destroyed island arcs of the Mesotethys were found in the Adjaria–Trioletia continuation of the East Black Sea Basin and in the Talysh continuation of the Sabzevar Trough (Shcherba, 1994). This gives grounds to suggest that they could have been parts of a single trough overthrust later by the Lesser Caucasus; i.e., the Sabzevar Basin continued or added to in an en echelon way the East Black Sea Basin.

At the end of the Middle Eocene and in the Late Eocene most of the belt except for its northern periphery underwent folding and thrusting (Bazhenov, Burtman, 1990). The Neotethyan backarc basins became narrower or partly closed (Khain, 2001; Golonka, 2004; Kazmin et al., 2010). In the regions where the Neotethys had been closed before the Eocene, for example in front of the Punjab indenter of the Indian Plate, a peak of high-pressure metamorphism fell on the Eocene (50–40 Ma) (Searle, 1996; Guillot et al., 2007). Intensive deformation did not lead to the formation of mountain topography. Fine clastic fractions dominated in the regions of sedimentation; large fragments and blocks occurred only in some accretionary wedges of the subduction zones.

1.2. Neotectonic evolution of the Central Tien Shan

Long-term research showed that the neotectonic (Oligocene-Quaternary) structure of Tien Shan is a result of deformation of the Cretaceous-Paleogene peneplain, which formed on the Paleozoic basement of the Turan Plate and the Kazakh Shield. The structure is a system of anticlinal and synclinal folds of the basement, expressed as ridges and molasses-filled intermontane basins, respectively. The folds are separated and complicated by large faults with reverse or thrust component of motion. It is generally recognized that starting from the Oligocene, this structure developed under transverse horizontal shortening (Shultz, 1948; Makarov, 1977; Dmitrieva, Nesmeyanov, 1982; Chediya, 1986; Nikolaev, 1988; Mikolaichuk, 2000). Its continuing development is confirmed by the inherited tectonic movements along active faults (Abdrakhmatov et al., 2001; Trifonov et al., 2002) and data of repeated geodetic observations (Nikonov, 1977; Abdrakhmatov et al., 1996; Recent geodynamics…, 2005; Zubovich et al., 2001, 2010). The average rate of tectonic uplifting in the period embracing the Oligocene to Quaternary was lower than that in the Quaternary alone and still lower than in the Late Pleistocene and Holocene. This evidences that the mountains rose with acceleration (Krestnikov et al., 1979; Chediya, 1986).

Our research was aimed at studying the change in the rate of rise of mountains in time and geodynamic processes that might have caused it. For study, we chose the Central Tien Shan (CTS), a part of the mountains between the Talas-Fergana Fault in the west and Khan-Tengri mountain plexus in the east (Fig. 3). In the north, the CTS borders the post-Paleozoic Chu and Ili basins, and in the south, the Tarim Basin, which originated on the late Proterozoic basement probably as early as in the Paleozoic.

1.2.1. Changes in the regime of vertical movements during the neotectonic evolution

The CTS Paleozoic orogen passed through a platform stage of evolution in the Mesozoic and Early Paleogene. In the Early Paleogene, the CTS area was a peneplain with uplifts reaching few hundreds of meters, which was earlier considered a preorogenic planed surface (Shultz, 1948; Trofimov, 1973; Chediya, 1986). Redeposited crust of weathering that formed on the peneplain by the late Mesozoic composes a continental red-colored, mainly fine clastic rock unit with the Middle-Late Eocene and, probably, Early Oligocene fossils (Dmitrieva, Nesmeyanov, 1982).

Its lower part includes basalt lavas, whose total thickness reaches 20 m on the north-western edge of the Issyk-Kul Basin and 80 m in the Aksai Basin (Chediya et al., 1973; Bazhenov, Mikolaichuk, 2002). The K-Ar and Ar-Ar ages of the basalts are 54–70 Ma (Krylov, 1960; Nesmeyanov et al., 1977; Simonov et al., 2005; Bachmanov et al., 2008). The unit thickness usually does not exceed few tens of meters, though locally is greater than 100 m in the Chu, Ili, Issyk-Kul, and Aksai basins, thus probably reflecting their started sinking.

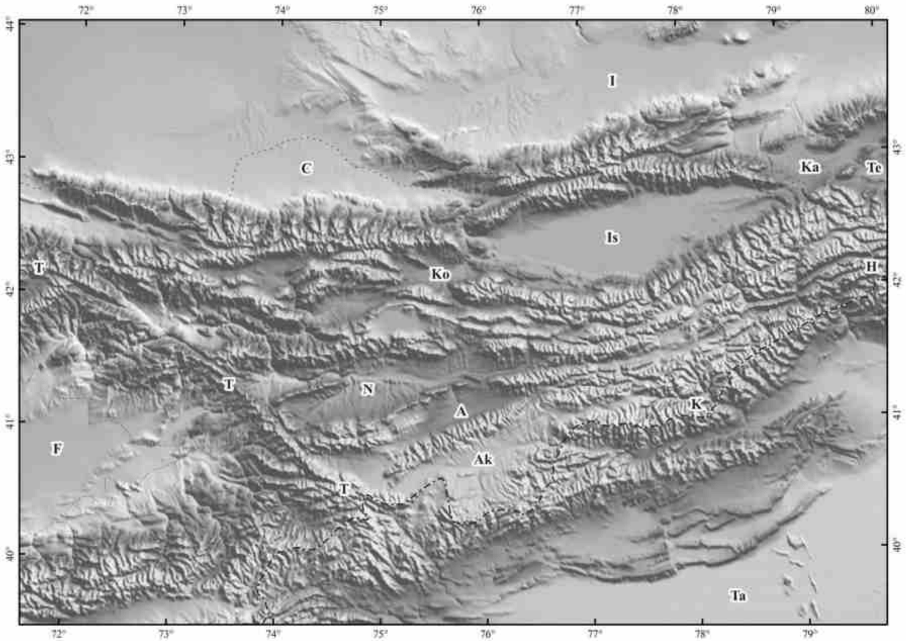


Fig. 3. The Cenrtal Tien Shan, modified after (Trifonov et al., 2008). A, Atbashi Basin; Ak, Aksai Basin; C, Chu Basin; F, Fergana Basin; H, Khan Tengri; I, Ili Basin; Is, Issyk-Köl Basin; Ka, Karkara Basin; Ko, Kochkor Basin; K, Kokshaal; N, Naryn Basin; T, Talas-Fergana Fault; Ta, Tarim Basin; Te, Tekes Basin. The China-Kyrghyzstan boundary is shown

Figure 4 shows a map of the CTS mountain peak surface combined with the position of the pre-orogenic surface in recent basins, which was determined from a complex of geological, drilling and geophysical data (Geological Map…, 1982). The central parts and slopes of ridges have preserved relics of preorogenic surface under conditions of intense linear erosion that accompanied their growth (Makarov, 1977; Chediya, 1986). Thus, besides rare cases of significant erosional decline of these relics, the map reflects recent deformation and uplift of the pre-orogenic surface. This is a generalized map of all neotectonic vertical movements. The axial parts of the ridges rise above the bottoms of neighboring basins to ~3–5 km, and the maximum magnitude of the surface relief is 10 km.

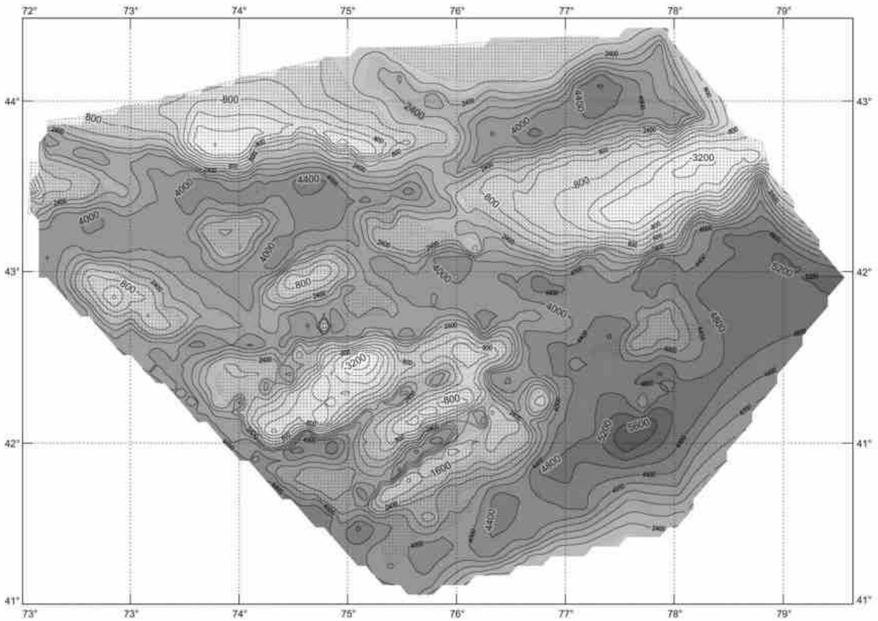


Fig. 4. Assumed position of the pre-orogenic surface drawn by peak surface isolines within the ridges and by basement surface isolines within basins (speckled sites), modified after (Trifonov et al., 2008). The position of the surface beneath recent molasses was designed from geological, drilling, and geophysical data (Geological Map of Kyrgyzstan, 1982)

The step-like structure of the ridge slopes, which is accounted for the most researchers by pulse rise, formed the basis for concepts of the step-like levels of the CTS topography (Shultz, 1948; Trofimov, 1973; Makarov, 1977; Krestnikov et al., 1979; Chediya, 1986). According to these concepts, intensification of vertical movements promotes erosion, which results in an erosional-tectonic scarp (incision) resting upon the bottom of basin or valley serving as a local basis of erosion and accumulating erosion products. The higher is the rate of rise, the coarser and thicker are accumulating deposits. The next pulse of rise leads to the uplifting of the scarp-bordering site of the depression, below which a younger incision forms. The uplifted site becomes a slope step. The steps, located at close hypsometric levels on slopes of different ridges and the incisions resting upon them form a regional topographic layer. This suggests that the incision is correlated with the lower coarse part of a corresponding molasse unit and that the steps are correlated with the upper finer part

of the unit (Makarov, 1977). Comparison of the topographic layer with related molasse unit helps to clarify the evolution of the mountain system.

Deposits of CTS basins were described elsewhere (Shultz, 1948; Nesmeyanov, Makarov, 1974; Dmitrieva, Nesmeyanov, 1982; Chediya, 1986; Dodonov, 2002; Mikolaichuk et al., 2003). Our goal was to correlate these data and supplement them with results of our research (Trifonov et al., 2008). The post-Eocene molasses are subdivided into four complexes: Kyrgyz, Tien Shan, Sharpyldak, and Middle Pleistocene–Holocene, each being divided into local units (Fig. 5 and Table 1). Their comparison and dating (still debatable) are based on the composition and color of deposits, incomplete paleomagnetic data, and scarce fauna remains.

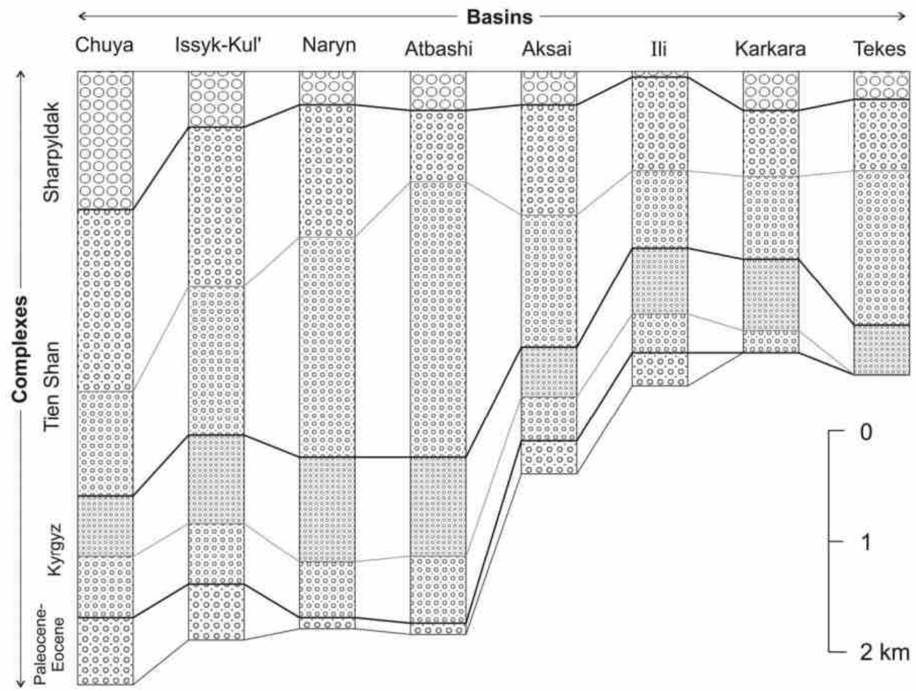


Fig. 5. Comparison of sections of the Cenozoic deposits in the Central Tien Shan basins, modified after (Trifonov et al., 2008). Different units are shown by different symbols

The lower Kyrgyz complex consists of two series, lower red and upper brown. The lower series is composed of fluvial fine- and, seldom, medium-pebble conglomerates, gravelstones, and sandstones, which give way to sand-clayey deposits with gypsum at the central parts of large basins. Based on the fossils found in the Issyk-Kul and Ili

Complex	Age	Chu (south)	◎	Kochkor	◎	Issyk-Kul	◎	Ili	◎	Naryn	◎	Atbashi
sh	N_3^2-Q_1	Noruz unit	N_2^3	Sharpyldak unit			Q_1	Khorog unit	Q_1	Sharpyldak unit		
Tien-Shan	N_2^2-N_2^3	Chu unit	N_2^3	Djuanaryk unit: Upper subunit	N_2^{2-3}	Issyk-Kul unit: Djoukin unit (Upper Issyk-Kul subunit)	N_2^3	Ili unit	N_2^3	Naryn unit: Upper Naryn subunit		Kulnei unit: Upper subunit
Tien-Shan	N_1^3-N_2^1	Saryagach unit: Djeldisuy unit	N_1^3-N_2^1	Djuanaryk unit: Lower subunit	N_1^3-N_2^1	Issyk-Kul unit: Sogutin unit (Lower Issyk-Kul subunit)	N_1^3-N_2^1	Santash unit	N_1^3-N_2^1	Naryn unit: Middle Naryn subunit	younger N_1^2?	Aktal unit: Middle, Lower subunit
Kyrgyz	N_1^1-N_1^2?	Kokomeren unit; Shamsa unit		Sharybkol unit		Djetyoguz unit: Upper subunit		Chuladyr unit		Naryn unit: Lower Naryn subunit	N_1^{1-2} ?	Kyzylbulak unit: Upper subunit
Kyrgyz	$E_3(E_3^{2-3})$ -N_1^1a	Kokomeren unit; Shamsa unit		Bizhin unit		Djetyoguz unit: Lower subunit	E_3	Aktal unit: Upper subunit; Lower subunit	E_3^3-N_1^1a; E_3^2	Kyrgyz unit		Kyzylbulak unit: Lower subunit
Paleocene -Eocene	E_{1-2} (-E_3^1?)	Suluterek unit		Kokturpak unit		Chonkurchak unit Ar-Ar and K-Ar age of basalts 54-70 Ma	E_2^3-E_2^2	Akbulak unit		Kokturpak unit		

Table 1. Comparison and age of the Paleocene – Lower Pleistocene units in intermontane and piedmont basins of the Central Tien Shan; data from (Abdrakhmatov et al., 2001; Aleshinskaya et al., 1972; Bullen et al., 2001; Chediya et al., 1973; Dmitrieva, Nesmeyanov, 1982; Dodonov, 2002; Makarov, 1977; Mikolaichuk et al., 2003; Nesmeyanov, Makarov, 1974; Nesmeyanov et al., 1977; Shultz, 1948; Simonov et al., 2005; Trifonov et al., 2008; Trofimov, 1973; Trofimov et al., 1976)

basins, the lower beds of the series were dated to the Middle Oligocene, and the upper ones, to the Late Oligocene – Early Miocene (Dmitrieva, Nesmeyanov, 1982). The upper series differs from the lower one in finer composition and the presence of carbonate and gypsum (and, locally, mirabilite and halite) interbeds, which points to plain landscape and the existence of drying lakes at the time of their formation. The position of the series in sections evidences its Early-Middle Miocene age.

The Tien Shan complex includes two series. Based on fauna remains, the lower series was dated to the Late (Middle?) Miocene – Early Pliocene, and the upper series, to the Late Pliocene (Dmitrieva, Nesmeyanov, 1982). According to paleomagnetic data, the series boundary in the south of the Chu Basin lies within 8–5 Ma (Bullen et al., 2001). Probably, it lies in the lowermost part of the Pliocene.

The above series differ in color (the lower series are variegated, and the upper ones are straw-gray) and are composed of terrigenous, mainly clay-silt-sandy rocks with interbeds of carbonate and, in the Naryn Basin, gypsum and, seldom, halite (Fig. 6). In the east of the region (Issyk-Kul, Tekes, and Karkar basins), conglomerates are predominant. Their amount increases in the upper parts of sections in the Naryn Basin and the southern Chu Basin. Upsection coarsening of deposits is also observed in the Ili Basin.

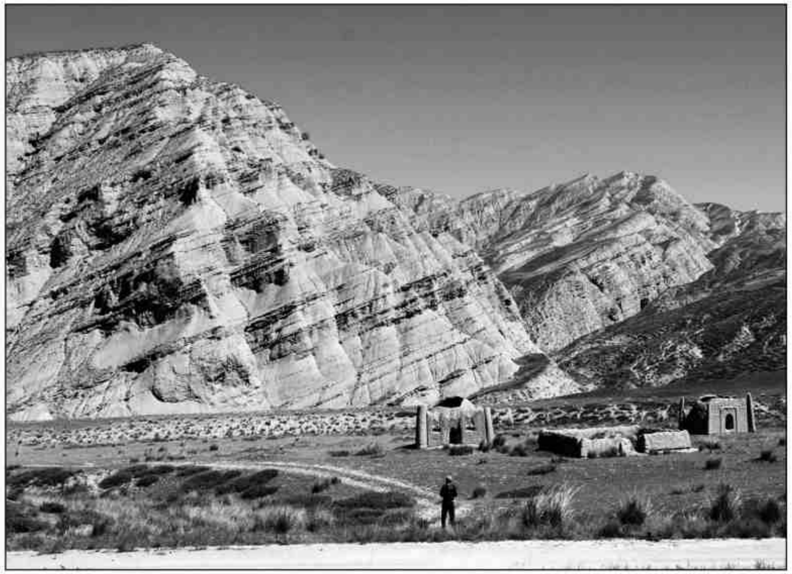


Fig. 6. The Naryn unit in the northern side of the Naryn Basin. Photo by D.M. Bachmanov

According to fauna remains and paleomagnetic data for the Issyk-Kul, Ili, and Chu basins, the Sharpyldak complex is of the Lower Pleistocene (Eopleistocene in the Russian terminology) age (~2–0.8 Ma) (Aleshinskaya et al., 1972; Trofimov et al., 1976; Dmitrieva, Nesmeyanov, 1982; Dodonov, 2002). It is composed of fluvial grey coarse comglomerates and conglomerate-breccias (up to boulder size) with gravel, sand and silt interbeds (Fig. 7).

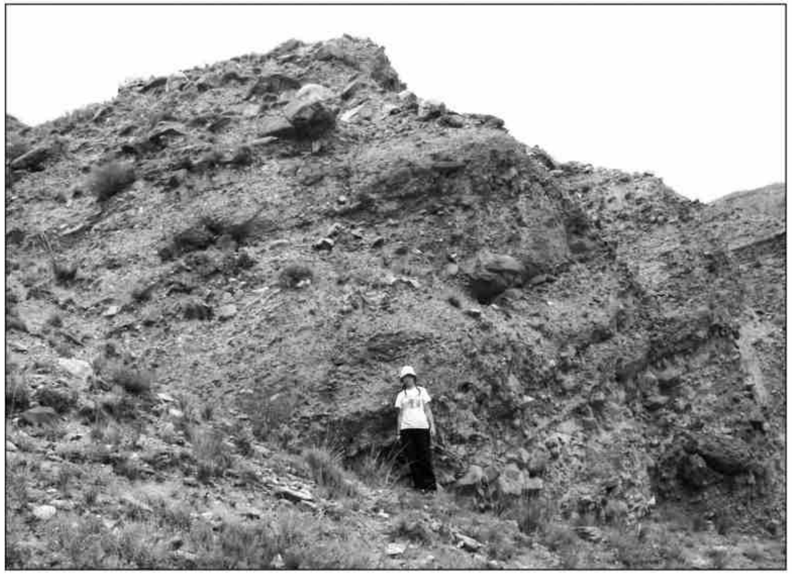


Fig. 7. The Sharpyldak unit in the Issyk-Kul Basin. Photo by D.M. Bachmanov

The Middle Pleistocene–Holocene complex, compositionally similar to the Sharpyldak one, is formed by fluvial sediments of seven cyclic terraces (Shultz, 1948), floodplains, and recent channels as well as glacial and basin (at the centers of the Chu, Ili and Issyk-Kul basins) deposits. Three series of the complex are dated to: (1) Middle Pleistocene, (2) Late Pleistocene, and (3) Late Pleistocene–Holocene (Chediya, 1986; Dodonov, 2002).

First uplifts serving as sourcelands of debris appeared in the Oligocene (Bachmanov et al., 2009), but the predominance of small-pebble conglomerates and gravelstones in the lower series of the Kyrgyz complex indicates a minor amplitude between uplifts and basins. According to the data on magnitude of the incision, it was not more than 1 km in most of the CTS area (Makarov, 1977; Chediya, 1986). Judging from the deposit composition, the topographic contrast was reduced during

the accumulation of the upper series of the Kyrgyz complex and the lower series of the Tien Shan complex (Bachmanov et al., 2009). The sediments covered a part of the Oligocene uplifts. Lake basins formed, which were isolated from the regional basis of erosion under highly arid conditions (presence of evaporates) and thus might rise together with uplifts supplying clastic material. Lake conditions also existed locally during the accumulation of the upper series of the Tien Shan complex, but evaporates became scarcer. The coarsening of clastic material began in the south-east of the region (near the Khan Tengri) during the accumulation of the lower series of this complex and penetrated later to the north-west (Trifonov et al., 2008). It points to an increase in the amplitude between the uplifts subjected to erosion and the basin bottoms. Strong coarsening of clastics occurred in the Early Pleistocene. It evidences a considerable increase in the above-mentioned amplitude, i.e., the formation of high mountain topography.

The change in the rate of vertical movements in the basins can be judged from the average rates of accumulation of various series of molasses in them (Fig. 8 and Table 2), though these estimates are only tentative because of the complex structure of the molasses and the incomplete sections of the basins.

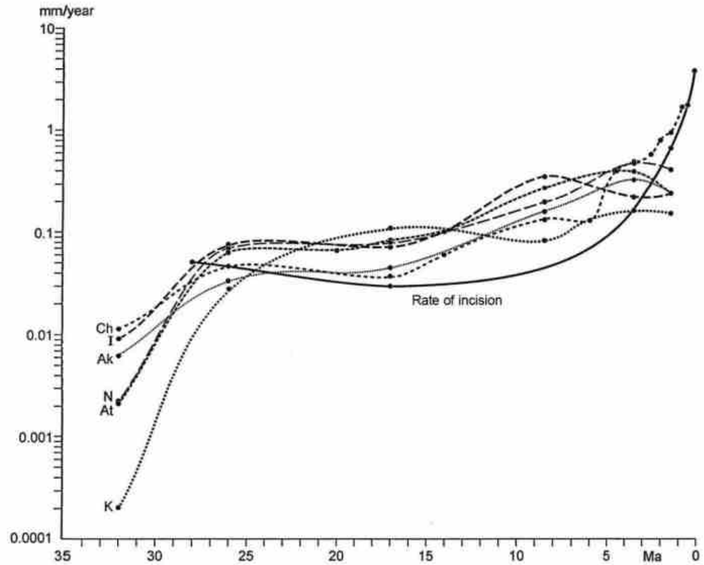


Fig. 8. Changes in the rates of sedimentation in the Cenozoic basins and in the rates of incision of drainage systems into ridges of the Central Tien Shan, modified after (Trifonov et al., 2008). Deceleration of deposit accumulation in the Aksai, Kochkor, Naryn basins in the Quaternary was caused by the transition from basin sedimentation to terrace regime. Basins: Ak, Aksai; At, Atbashi; Ch, Chu; I, Issyk-Köl; K, Kochkor; N, Naryn

Complex	Stage	Series	Duration, Myr	Boundaries, Ma	Chu (south)	Kochkor	Issyk-Kul'	Tekes	Karkara	Ili	Naryn	Atbashi	Aksai
					Basin								
					Thickness, m / Rate, mm/year								
sh	N_3^2–Q_E	sh	1.2	0.8 2	1000–1300 / 0.8–1.08	200 / 0.167	500 / 0.417	250 / 0.208	350 / 0.28	>20 / >0.017	300 / 0.25	>300 / >0.25	300 / 0.25
Tien Shan	$N_2^2-N_2^3$	ts_2	3	5	1250–1700 / 0.4–0.57	200–850 / 0.067–0.283	1450 / 0.483	650 / 0.217	600 / 0.2	380–880 / 0.127–0.293	1200 / 0.4	650 / 0.217	1000 / 0.333
Tien Shan	$N_1^3-N_2^1$	ts_1	7	12	950 / 0.136	150–1000 / 0.0214–0.143	1380 / 0.197	1400 / 0.2	230–800 / 0.033–0.144	125–760 / 0.018–0.109	2000 / 0.286	2500 / 0.357	1200 / 0.171
Kyrgyz	$N_1^1-N_1^2$?	kz_2	10	22	150–600 / 0.015–0.06	450–1700 / 0.045–0.17	800 / 0.08	170–500 / 0.017–0.05	150–700 / 0.015–0.07	160–670 / 0.016–0.067	670–1000 / 0.067–0.1	450–1000 / 0.045–0.1	450 / 0.045
Kyrgyz	$E_3(E_3^{2-3})$–N_1^1a	kz_1	8	30	150–600 / 0.019–0.075	200–250 / 0.025–0.031	500–600 / 0.063–0.075	No data	200 / 0.025	390 / 0.049	500 / 0.063	600 / 0.075	100–450 / 0.012–0.056
Paleocene–Eocene	E_{1-2} ($-E_3^1$?)	E	36	66	150–635 / 0.0041–0.0176	7 / 0.0002	100–550 / 0.0028–0.0153	No data	No data	320 / 0.0089	100 / 0.0022	80 / 0.0022	10–350 / 0.0003–0.0098

Table 2. Thicknesses and rates of accumulation of the Paleogene – Lower Quaternary molasse complexes in the Central Tien Shan, data from (Dmitrieva and Nesmeyanov, 1982; Makarov, 1977; Shults, 1948; Simonov et al., 2005; Trifonov et al., 2008)

Calculations showed that the average rates of accumulation did not exceed thousandths of a millimeter per year and reached ~0.02 mm/year only at some sites of the future basins at the Paleocene-Eocene platform stage. The accumulation rates increased to hundredths of a millimeter per year during the formation of the lower series of the Kyrgyz complex and stayed the same during the sedimentation of its upper series. The Tien Shan complex accumulated with acceleration of the rates up to 0.1–0.6 mm/year.

In the Sharpyldak time, the basin sedimentation regime was changed by the terrace regime, which evidences a drastic intensification of linear erosion as a result of the acceleration of uplift. In the Middle-Late Pleistocene, this led to the formation of terraces within the more ancient deposits. Because of the change in regimes, the average rates of the Sharpyldak sedimentation in most basins became commensurate with or lower than the rates of earlier sedimentation. The rate of the accumulation increased twice only at some sites of the Chu Basin, where the basin sedimentation lasted. A single estimate for the later Pleistocene sedimentation was obtained for the Chu Basin, where up to 500 m deposits accumulated for that epoch (Abdrakhmatov, 1988). Its rate is close to the rate of the Sharpyldak sedimentation. But the thickness of the Quaternary coarse-clastic deposits is obviously smaller than the amount of removed clastic material, because the formed river runoff and aeolian processes transported fine-clastic fractions from the mountain system.

Thus, analysis of Cenozoic deposits showed a significant acceleration of rise in the Quaternary. The same is evidenced from the data of analysis of the levels of topography of the CTS ridges. Three levels were recognized in the mountain system (Shultz, 1948; Trofimov, 1973; Makarov, 1977; Krestnikov et al., 1979; Chediya, 1986). The upper level that is formed by one or two steps, incised into the preorogenic surface is correlated with the Kyrgyz complex (E_3–N_1^2); the middle level, formed by two steps is correlated with the Tien Shan and Sharpyldak complexes (N_1^2–Q_1); and the lower level is correlated with the Middle Pleistocene to Holocene basin deposits of the Chu and Ili basins. The highest of seven cyclic terraces of the lower level has been incised into the roof of the Sharpyldak complex and is dated to the earlier Middle Pleistocene (Q_2^1). Makarov (1977) established that the magnitude of the Oligocene incision into the CTS ridges were no more than 200–400 m. According to Chediya (1986), this magnitude reaches 700 m; that of the middle level reaches 1500 m (more than a half of this value falls onto the lower incisions formed in the Sharpyldak time); and the magnitude of the lower level incision reaches 1500 m. Krestnikov et al. (1979) reported similar magnitudes: ~1000 m for the Sharpyldak incision in the CTS ridges and >1000 m in the north and up to 1500 m in the south-east of the CTS for the Middle Quaternary–Holocene incision.

Taking into account the duration of the epochs of sedimentation of correlated molasse complexes (Table 2) and using the above estimates, we can tentatively calculate (ignoring variations of the incision during each cycle) the average rates of incision for different levels and sublevels of the topography of the CTS anticlinal ridges, which, to a first approximation, reflect the rates of their rise (Trifonov et al., 2008). The rates of incision were 0.03–0.05 mm/year during the accumulation of the Kyrgyz unit and its analogs (E_3–N_1^1), ~0.04 mm/year during the deposition of the Kyrgyz complex (E_3–N_1^2), ~0.07 mm/year during the sedimentation of the Tien Shan complex (N_1^2–N_2), 0.6–0.7 mm/year in the Sharpyldak time (Q_1), and 1.6–2.5 mm/year in the Middle Pleistocene to Holocene (Q_2–Q_4). Thus, the rate of incision increased ~10 times in the Early Pleistocene and 20–30 times in the Middle Pleistocene to Holocene (Fig. 8).

The CTS intermontane basins, which developed for a long time in the regime of sinking down and basin sedimentation, also began to be involved in the rise in the Sharpyldak time. In the early Middle Pleistocene, linear incision covered all intermontane basins (probably, except for the center of the Issyk-Kul Basin), thus forming the lower level of their topography. The average rates of incision in the basins were 1.5–2 times lower than those in neighboring ridges; the rates were higher in the south and south-west of the region than in the north (Krestnikov et al., 1979). In the Middle and Late Pleistocene, zones of sinking and basin sedimentation still existed in the central parts of the Chu, Ili and, probably, Issyk-Kul basins, but their areas were reduced, giving way to terraces.

Thus, since the Early Pleistocene, the CTS has undergone an intense uplifting, which was maximum in the south and south-east. The mountain building was expressed as a drastic coursing of molasses and acceleration of incision. The uplifting was not dome-like, because the intermontane basins rose less intensively than neighboring ridges. The Early Pleistocene orogenic activation was also manifested by the form of contacts between molasses series. Ancient series usually overlap each other concordantly, and their “transgressive” relationships are observed only on the basin periphery, whereas the Sharpyldak series lie with an angular inconformity (few to >10°) almost ubiquitously (Trifonov et al., 2008).

In the same epoch, the uplifts widened at the expense of the basins. In the south of the Chu Basin, the rise of the Kyrgyz Ridge involved sites that were first covered with molasses and then were uplifted to several kilometers (Chediya, 1986;

Mikolaychuk et al., 2003). These sites formed a high piedmont there and on the periphery of the Issyk-Kul, Susamyr and Atbashi basins. Within the CTS mountain system, the basins began to be separate by sownecks, which appeared first in the south and then in the north. For example, the sowneck between the Tuyun and Aksai basins appeared as early as the Pliocene and was overlapped by the Sharpyldak deposits. The Baibichetau-Narynan sowneck, which separated the Naryn and Atbashi basins and was uplifted above their bottoms to $\geq$2 km, was formed by the left en echelon row of ridges combining transverse shortening (with extension of the Baibichetau dome) and longitudinal sinistral slip (Makarov, 1977). The late age of the sowneck is proved by the fact that between the Naryntau and Karatau ridges (the eastern part of the sowneck) and in the western slope of the sowneck, the fine-grained deposits of the Naryn unit do not contain the clastic material from these uplifts. The Sharpyldak deposits are absent in the sowneck slopes and do not contain its clastic material around. Therefore, the Baibichetau-Narynan sowneck formed finally as an topographic feature as late as the Middle Pleistocene (Trifonov et al., 2008). The same relationships were observed around the 2-km high sowneck between the Dzhumgol and Kyzyloi basins that are composed of the Kyrgyz and Tien Shan complexes deposits.

One more Quaternary formation is the Kökömeren-Minkush zone (Sadybakasov, 1972; Bachmanov et al., 2008), striking approximately along the Paleozoic Nikolaev’s structural line west of Lake Song-Kel. The zone forms an intensively deformed ramp, narrowed by a transverse shortening, with signs of a longitudinal sinistral slip. The Late Quaternary strike slip fault is expressed by shift of topographic features and inherits earlier displacement that is manifested by slickensides on the fault planes. Within the zone, the Lower Carboniferous red-colored deposits, Jurassic sediments, and the thin Kokturpak unit with basalts in the lowest part (68.4$\pm$2.3 Ma, dated by V.A. Lebedev in the Institute of geology of ore deposits, petrography, mineralogy and geochemistry, Moscow, Russian Academy of Sciences) underlie conglomerates of the lower series of the Kyrgyz complex; they are overlapped by deposits of the Naryn and Sharpyldak units (Bachmanov et al., 2008). All strata, from the Lower Carboniferous to Naryn unit change one another without serious angular inconformity. The latter appears only in the base of the Sharpyldak unit (locally) and the Middle Pleistocene deposits (ubiquitously). This points to the young age of the ramp and its deformation.

1.2.2. Horizontal shortening of the Earth's crust in the Oligocene-Quaternary

The CTS ridges and intermontane basins evolved in the Late Cenozoic under horizontal compression caused by the north-westward pressure of the Tarim microplate, which, in turn, was due to the movement of the more southern parts of the Alpine-Himalayan orogenic belt (Ivanova, Trifonov, 2005). Many researchers relate the mountain uplift to these lateral movements. Let us check the possibility of the relationship between the Quaternary acceleration of the uplift and the lateral movements by comparing the average rates of the CTS shortening in the Oligocene–Quaternary and the Late Pleistocene–Holocene.

The GPS measurements of lateral movements in the Kyrgyz part of the Tien Shan that had been carried out since 1992 showed that the rates of the CTS shortening reach 12–13 mm/a (mm per year), with the shift vectors in the east departing from the normal to the ridge strike, thus pointing to the presence of the longitudinal sinistral component of movements (Abdrakhmatov et al., 1996; Zubovich et al., 2001; Recent geodynamics…, 2005). Later Zubovich et al. (2010) recalculated the data, taking into account the recent underthrust of the Tarim Basin beneath the CTS with the rate of 4–7 mm/a, and estimated the total rate of the CTS shortening (convergence of the Tarim Basin and the Kazakh Shield) at 20±2 mm/a. The rates of recent vertical movements in places reach 10 mm/a (Nikonov, 1977).

Since the period of the GPS observations was too short for estimating the rates of lateral movements, the data on faults that were active in the Late Pleistocene and Holocene were used. Eight large active thrusts were studied, and the rates of movements along each of them were determined at 0.1 to 4 mm/a (Abdrakhmatov et al., 2001).The quoted authors suggested that most of recent transverse shortening of the Kyrgyz CTS falls just on these faults and estimated its total rate at 10 mm/a. But this estimate cannot be accepted because the studied faults die out along the strike and there are also active faults that were ignored by the quoted authors.

We carried out an analysis of the set of known active faults in the Kyrgyz CTS (Trifonov et al., 2002) and revealed that many longitudinal faults have not only a thrust, but also a substantial sinistral component of slips. Faults of the NW trend were recognized as dextral strike-slip ones. Based on the analytical results, we calculated tensors of deformation rates in the region. The total rate of the Late Quaternary horizontal shortening of the Kyrgyz CTS was estimated at ~6 mm/year. In the calculations, we ignored folded bends and shifts along fractures, whose contribution

to the total deformation is 10–20% in other active regions of this kind (Anatolia and Middle East). Therefore, the rates of the Kyrgyz CTS lateral shortening should be increased to ~7 mm/a, with the transverse-shortening component not exceeding ~6 mm/a. The farther studies showed that we had underestimated the rates of slip on same faults, and the transverse-shortening rate has to be increased as minimum to ~7 mm/a. We do not know the rate of shortening in the Chinese part of the CTS, but can suggest (by analogy with the results of GPS measurements) that it reaches about a half of the shortening in the Kyrgyz CTS. So, the estimate of the total Late Quaternary transverse-shortening rate of the CTS at ~10 mm/a seems to be reasonable.

Calculation of the total neotectonic (Oligocene-Quaternary) deformation of the CTS transverse shortening is based on the measurement of fold bends and shifts along the faults that deform the preorogenic peneplain and the Cenozoic molasses (Chediya, Utkina, 1975; Yunga, Yakovlev, 2000). Chediya (1986) reported that reverse faults became steeper with depth and ignored their gentle dip near the land surface in the calculations. He estimated the total transverse shortening of the Kyrgyz CTS at 4–5% of its width (14–18 km at the longitude of the Naryn Basin) and considered that the shortening is smaller at the longitude of the Khan-Tengri mountain plexus, where the CTS is the most uplifted. Yunga and Yakovlev (2000) made a similar calculation, not introducing corrections to observed fold bends, dip and offsets on faults. They estimated that the total shortening varied from 9–12% at the longitudes of the Naryn Basin and the city of Bishkek to 5–6% at the longitude of Khan-Tengri, i.e., from 40 to 20 km.

The representativeness of the obtained data depends on the model of the CTS neotectonics. The quoted calculations were based on the conventional model implying that the CTS mountain system is a combination of anticlinal ridges and synclinal basins complicated by reverse faults and thrusts. In recent years, a new model has been elaborated, which relates fold bends to movements along large thrusts flattening at depth (Abdrakhmatov et al., 2001). The new model admits the total shortening of 35–80 km, i.e., 10–20% of the CTS width. The validity of both models can be tested by three methods (Trifonov et al., 2008).

Structural method. Following the new model, thrust zones must exist throughout the whole length of the mountain system irrespectively of changes of fold forms. According to the conventional model, thrusts seldom extend beyond the folds that the thrusts complicate.

<u>Geomorphological method</u>. Following the new model, the preorogenic peneplain is eroded near thrusts, and if it were preserved, it would be uplifted abnormally highly. According to the conventional model, the peneplain reaches its maximum height in the axial part of anticlinal ridges and lowered to their near-thrust edges.

<u>Geological method</u>. Following the new model, ascent metamorphosed rocks should expose in the most uplifted and eroded near-thrust part of the ridge. According to the conventional model, the distribution of rocks of different metamorphism grades is determined by their pre-Mesozoic history; near thrusts, there might occur weakly metamorphosed rocks exposed during shallow erosion.

Study of marginal thrusts at the boundary of uplifts and basins shows that in most of the region including the North and Central Tien Shan megazones, the Cenozoic thrusts and reverse faults usually do not extend beyond ridges and their magnitude vary along the strike. Locally on the northern flank of the CTS (Fig. 9, *a–c*) and within it (Fig. 9, *d*), the thrusts transform along their strike into overturned folds of the basement surface or are complicated by similar folds of the lower molasse series, with molasses being present in both fold limbs (Fig. 9, *e*). The thrust magnitude or its deformation is relatively small in all observed cases. In ridges, the peneplain surface outlines an anticlinal bend, and the Devonian-Permian weakly metamorphosed rocks are often exposed in the near-thust marginal parts of the ridges. All this agrees with the conventional model.

A different pattern was observed in the zone of the Late Cenozoic Atbashi Fault at the boundary of the Atbashi Basin and the Atbashi Ridge that belongs to the South Tien Shan megazone (Trifonov et al., 2008). Where the Sarybulak stream flows into the Karakoyun River (western part of the structure), the fault zone is separated into two main strands. The northern strand borders the southern part of the basin. The strand runs along the Karakoyun River and is overlapped by the Late Quaternary alluvium. According to the geophysical data, the basement is subsided to 3–4 km north of the strand (Geological Map…, 1982). South of the strand, only thin Paleogene–Lower Miocene deposits are locally exposed. Near the Sarybulak stream, the Permian clay and silty slates and sandstones with cleavage cracks (dipped at 70º to the south) are stripped beneath glacial and fluvial deposits of the terrace cover. The southern strand of the fault zone dips at 60– 70º to the south and forms a scarp separating the terrace from the highly uplifted ridge slope that is composed of metaterrigenous quartz-sericite and, southward, quartz-mica schists tentatively dated to the Riphean (Fig. 9, *f*). Schistosity dips to the south; it dips at 70º just near the fault, and farther to the south, it is characterized by a steep dip in the upper part of the

slope, which decreases to 60º and then to 40º near the channel (Trifonov et al., 2008). Studies by apatite fission track thermochronology showed that the schists reached the subsurface as late as ~20 Ma (Sobel et al., 2000), though their presence in the upper horizons of the Earth's crust is related to the Hercynian nappe formation (Burtman, 2006). The near-fault ridge slope is strongly eroded and lacks relics of the preorogenic peneplain.

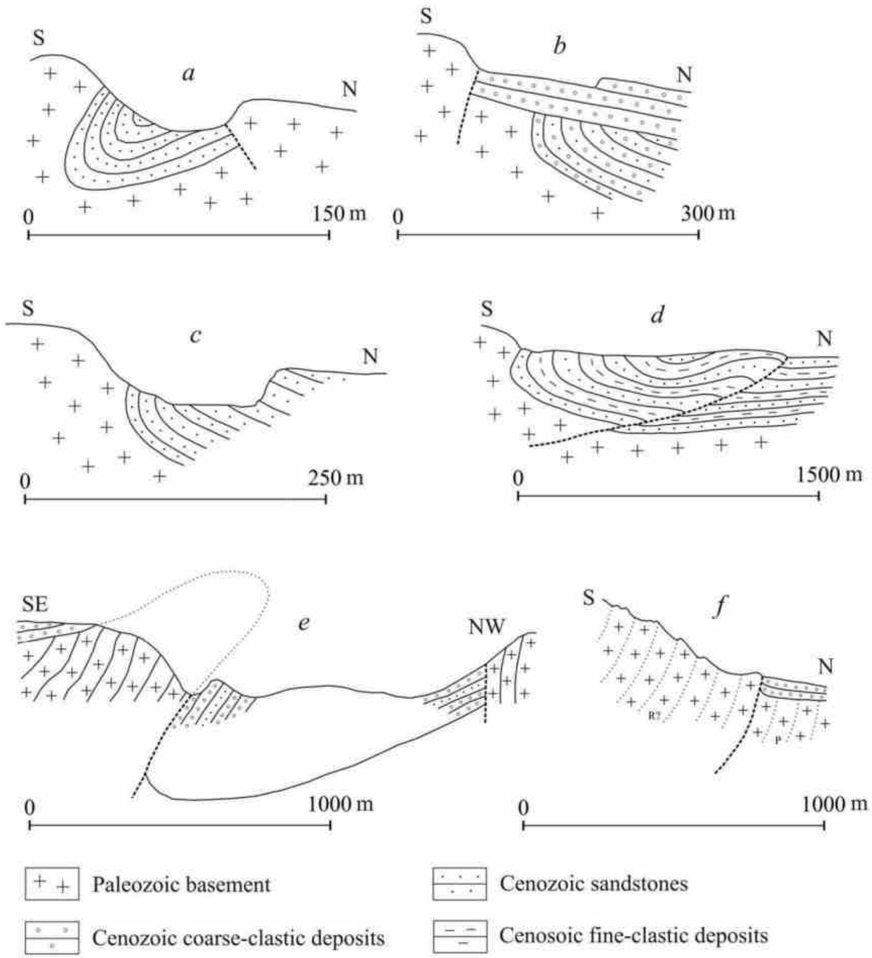


Fig. 9. Geological profiles across the boundaries of basins and uplifts in the Central Tien Shan, modified after (Trifonov et al., 2008): *a-c*, the northern flank of the Tien Shan: *a*, Chonkurchak, *b*, Dzhalamysh sai, *c*, Aksu River; *d*, southern edge of the Kochkor Basin; *e*, southern edge of the Toguz-Torou Basin (Shultz, 1948); *f*, southern edge of the Atbashi Basin along the Sarybulak Stream. The horizontal and vertical scales are equal

The reported data fit the new model better as compared with the conventional one and suggest that the Atbashi Fault is flattened at depth. Since the rocks exposed in its southern side might have been earlier situated at depths reaching 5 km (judging from the degree of their metamorphism) and the basement north of the fault zone is subsided to 3–4 km, the magnitude of the Late Cenozoic thrust is estimated at ~10 km. The thrust is listric; the increase in the uplift of its southern side toward the Khan-Tengri mountain plexus points to an increase in its magnitude and, correspondingly, shortening.

Large Late Cenozoic south-vergent thrusts were revealed in China along the southern flank of the Tien Shan (Deng Qidong, 2000; Recent geodynamics…, 2005) In the southern piedmont at the longitudes of Issyk-Kul and Khan-Tengri, the total magnitude of offsets on listric thrusts was estimated at 12–15 km (Yin et al., 1998). East of them, at the longitude of Lake Lobnor, the total shortening of the South Tien Shan is 20–40 km (Yin et al., 1998). Burtman (2012) reported similar estimates for the Eastern (Chinese) Tien Shan. With regard to the thrusts in the South Tien Shan megazone, the total Oligocene-Quaternary transverse shortening in the CTS is estimated at 50–70 km during ~30 Ma that yields the average rate of shortening of ~2 mm/a (Trifonov et al., 2008).

The obtained estimate ignores the strike slip along faults. The sinistral shift along the CTS is inferred from the en echelon mutual location of neotectonic structures (Makarov, 1990; Recent geodynamics…, 2005). We discover the Late Cenozoic longitudinal sinistral shifts in the Kökömeren-Minkush zone (Bachmanov et al., 2008) and on the northern slope of the Baibichetau Ridge (Trifonov et al., 2008). Dextral offsets and shear zones of the NW strike were revealed earlier (Bogachkin et al., 1997; Trifonov et al., 2002). Lacking detailed data on the magnitudes of slip along the faults, we admit that their contribution to the total Oligocene-Quaternary deformation is proportional to the contribution of active strike-slip faults to the total Late Quaternary deformation. With regard to the strike slip, the average rate of the Oligocene-Quaternary lateral CTS shortening might reach 2.5–3 mm/a that is 3–4 times lower than the rates of the Late Pleistocene–Holocene shortening. Thus the acceleration of horizontal movements during the Late Quaternary was lower than the acceneration of the Quaternary uplift.

Let us analyze the physical mechanisms responsible for the crustal uplift in the region.

1.2.3. Contribution of compression to crustal thickening and uplift in the Central Tien Shan

The represented data showed the following. In the Oligocene, the CTS paleotopography was at average heights of ~300 m. From Oligocene to the Early Pleistocene (beginning of the formation of the Sharpyldak unit), the average height of uplifts did not exceed 1.5 km, and the difference between the heights of uplifts and surfaces of adjacent basins was no more than 1 km. Based on these data and distribution of the uplifts and basins, we accepted that the average height of the CTS reached ~1 km by the Early Pleistocene. At present, the average height is ~3 km. Thus, the CTS was uplifted by on average ~700 m (up to ~1000 m in the NE of the region) from Oligocene to Early Pleistocene, i.e., for ~28 Myr, and by ~2 km for the last 2 Myr. In the SE and the east of the region, the height of the Quaternary uplift has been no less than 3 km. Based on the above data, Artyushkov estimated the role of compression of the Earth's crust in the Oligocene-Quaternary CTS uplift (Trifonov et al., 2008).

Let us denote the initial and final values of the width of compressed area and its thickness of the Earth's crust as L_0, L_1 and h_0, h_1, respectively. Then, an increase in crustal thickness due to compression, Δh_{comp}, and crustal uplift $\Delta\zeta_{comp}$ under local isostasy are:

$$\Delta h_{comp} = h_1 - h_0 = [(L_0 - L_1) / L_1]\, h_0; \quad (1)$$

$$\Delta\zeta_{comp} = [(\rho_m - \rho_c) / \rho_m]\, \Delta h_{comp}, \quad (2)$$

where $\rho_m = 3330$ kg/m^3 is the mantle density and ρ_c is the average crustal density. At present, the average width of the CTS is $L_1 \approx 400$ km. The Oligocene-Quaternary CTS shortening is $L_0 - L_1 = 50$–70 km, i.e. the initial width of the CTS was $L_0 = 450$–470 km. The average rate of the Late Quaternary shortening was ~10 mm/a. Taking this value for the last ~2 Myr, we obtain the CTS shortening $L'_1 - L_1 \approx 20$ km and the CTS width equal to $L'_1 \approx 420$ km ~2 Ma.

From Jurassic to Eocene, the CTS together with the southern part of Kazakhstan were a young platform. In the southern Kazakhstan, the thickness of the Earth's crust is ~42 km (Fig. 10). We accept that the same crustal thickness $h_0 = h_{pl} = 42$ km was also in the CTS in the Eocene. The average density of the platform crust is $\rho_c = 2830$ kg/m^3 (Artyushkov, 1993; Christensen, Mooney, 1995). With these values of ρ_c and h_0, we obtain that 2 Ma, when L_1 was equal to ~420 km, Δh_{comp1} was 4.7–6.5 km and $\Delta\zeta_{comp1}$ was 0.7–1.0 km. The latter value is close to the above-given geological-geomorphological estimate of the CTS uplift that occurred by ~2 Ma (0.7–1.0 km).

Therefore, it is most likely that the CTS uplift that proceeded from Oligocene to the beginning of Pleistocene was mainly due to the compression of the Earth's crust.

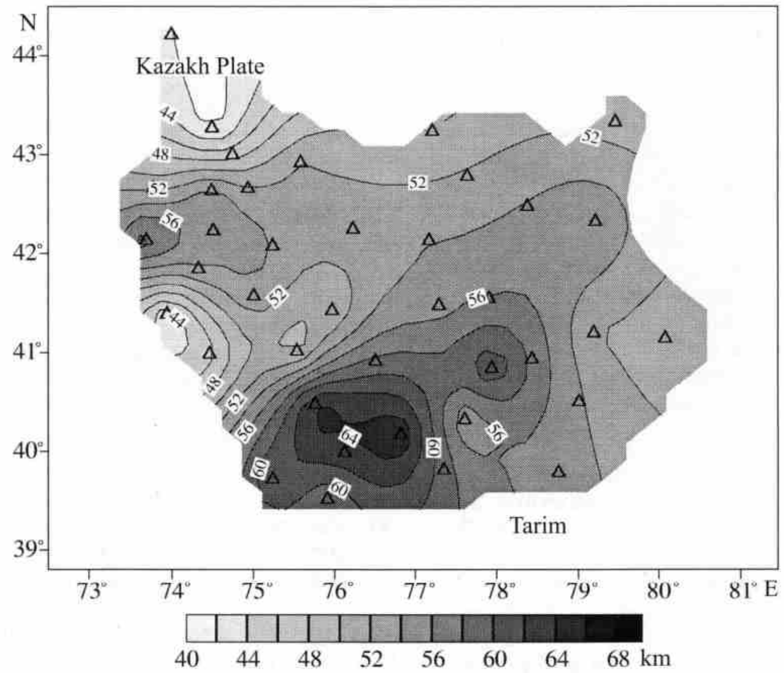


Fig. 10. The Earth's crust thickess in the Central Tien Shan, modified after (Vinnik et al., 2006). Triangles mark seismic stations

By the beginning of the accelerated uplift of the CTS ~2 Ma, the 4.7–6.5 km thickening of the crust must have increased its thickness to h'_0 = 46.7–48.5 km. Introducing this value as h_0 into (1) and assuming that $L_0 = L'_1$ = 420 km, from (1) and (2) we obtain the following values of crustal thickening and uplift for the last ~2 Myr: Δh_{comp2} = 2.2–2.3 km and $\Delta\zeta_{comp2}$ = 330–350 m. The latter value is 6–9 times less than the actual 2–3 km uplift occurring for the last ~2 Ma. Even if accepting the average rate of shortening for the last ~2 Ma equal to the GPS rate of ~20 mm/a (Zubovich et al., 2010), the uplift due to the crustal compression will not exceed 650 m. This is only 22–32% of the actual uplift. The other portion requires other mechanisms for explanation.

The total Oligocene-Quaternary CTS uplift due to the crustal compression is $\Delta\zeta_{comp} = \Delta\zeta_{comp1} + \Delta\zeta_{comp2} \approx$ 1000–1300 m.

1.2.4. The rise of the asthenosphere roof beneath the Central Tien Shan

Compared with the platform adjacent in the north, the velocities of transversal (V_S) and compression (V_P) waves beneath the Moho are significantly lower in the CTS mountains (Yudakhin, 1983; Lithosphere of the Tien Shan, 1986; Vinnik et al., 2004, 2006; Recent geodynamics…, 2005). This points to the ascent of the asthenosphere roof to the Earth's crust. According to the gravimetrical data, deconsolidation of the mantle beneath the CTS reaches ~0.1 g/cm^3 (Artemjev, Kaban, 1994). The rise of the asthenosphere roof is nonuniform throughout the region. Seismic data evidence that the asthenosphere roof reaches the Moho beneath high ridges and is separated from the crust by thick lenses of the lithosphere mantle beneath intermontane basins.

The replacement of the dense lithosphere mantle by the less compact asthenosphere matter (density ρ_a) must be accompanied by uplift of the crust. The uplift value is proportional to the squared thickness of the lithosphere mantle layer that has been replaced (Artyushkov, Hofmann, 1998). This value is unknown for the CTS, although negative isostatic gravity anomalies of up to –150 mGal were detected (Artemjev, Kaban, 1994). According to the Artyushkov calculation (Trifonov et al., 2008), these anomalies would correspond to anomalous masses $\Delta m \approx -3.6 \cdot 10^6$ kg/m^2. The mantle deconsolidation leads to isostatic uplift of the crust by

$$\Delta\zeta_a = -\Delta m/\rho_a. \qquad (3)$$

with $\Delta m \approx -3.6 \cdot 10^6$ kg/m^2, $\Delta\zeta_a \approx$ 1.1 km. Since the width of areas with the above-mentioned anomaly intensity does not usually exceed 100 km, the anomalous masses in the mantle beneath them and the corresponding uplift of the crust can be greater. Assuming that $\Delta m \approx -7 \cdot 10^6$ kg/m^2, we obtain from (3) that $\Delta\zeta_a \approx$ 2 km. In the most uplifted parts of the CTS, the deconsolidated mantle occurs in places immediately beneath the crust. This suggests that the uplift of the crust due to the asthenosphere rise might reach $\Delta\zeta_a$ = 1.5–2 km there.

The complete or partial replacement of the mantle lithosphere by the asthenosphere takes place during a drastic softening of the former (Artyushkov, 2003). The softening is caused by the infiltration of active fluids from the underlying mantle into the lithosphere. This leads to a drastic decrease in viscosity and strength of rocks as a result of the Rebinder effect (Rebinder, Venstrem, 1937; Salnikov, Traskin, 1987).

1.2.5. Great thickness and density of the Central Tien Shan crust before its Cenozoic compression and possible transformation of the lower crust during the Quaternary uplift

As shown above, the compression of the CTS crust resulted in the increase in its thickness by $\Delta h_{comp} = \Delta h_{comp1} + \Delta h_{comp2} \approx 7–9$ km. The present-day thickness of the CTS crust varies from 40–52 km beneath a foredeep and largest intermontane basins to h = 52–64 km beneath ridges (Lithosphere of the Tien Shan, 1986; Recent geodynamics…, 2005; Vinnik et al., 2006) (Fig. 10). The crustal thickness beneath ridges is 10–22 km greater than that in the southern Kazakhstan ($h_0 = h_{pl}$ = 42 km), which we accepted as the pre-orogenic thickness of the CTS crust by the Oligocene. The value of 10–22 km is 1.5–2.5 times greater than the calculated crustal thickening of the CTS crust due to the Cenozoic compression. Artyushkov paid attention to the following discrepancy (Trifonov et al., 2008). If the Cenozoic crustal thickening have been related to the compression only, the CTS crustal thickness by the Oligocene would have been $h_0 = h - \Delta h_{comp} \approx 45–55$ km, i.e., higher than h_{pl} by $\Delta h_0 \approx 3–13$ km. Introducing this value as Δh_{comp} into (2), we obtain that with the average crustal density ρ_c = 2830 kg/m^3, typical of platforms, the CTS peneplain would have been localized at heights of ~0.5–2 km. In fact, its height was close to ~0.3 km. Artyushkov (Trifonov et al., 2008) suggested that the average crustal density was at that time higher than the usual platform density. Under isostatic equilibrium, it must have been 2900–3000 kg/m^3. If this average density might have existed in the Eocene, the CTS lower crust had a layer of deeply metamorphosed basic rocks close in density to the mantle.

Heavy metabasites, garnet granulites and eclogites form in the lower crust of the fold belts as a result of phase transitions during strong compression (Artyushkov, 1993). In the CTS, they could have resulted from the Caledonian and Hercynian collision, when huge volumes of paleo-oceanic crustal matter got into the crust of the Northern and Southern Tien Shan, respectively (Kurenkov, 1983; Lomize et al., 1997; Burtman, 2006). Artyushkov (Trifonov et al., 2008) proposed a simplified model for the high-density crust that can be taken as a first approximation for the late Mesozoic and Eocene CTS. The upper layer of the crust is 42 km thick and has a density of 2830 kg/m^3; beneath it, there is the 3–13 km thick layer of garnet granulites and eclogites that is close in density to the upper mantle.

The ascent of the hot lower-density asthenosphere to the uppermost mantle might have been accompanied by two processes in the lower crust (Artyushkov, 1993).

Eclogites and basic garnet granulites, denser than the asthenosphere matter, were replaced by it, being detached from the crust and submerged together with the lithosphere mantle. Under influence of the asthenosphere fluids, garnet granulites, less dense than the lithosphere mantle and asthenosphere, might have undergone the retrograde metamorphism accompanied by the serpentinization of neighboring peridotites. This led to the deconsolidation and, correspondingly, additional uplift of the land surface. Since the Oligocene, the CTS has uplifted as a result of compression to ~1–1.3 km, and the uplift due to the asthenosphere ascent and replacing of the lithosphere mantle might have reached 1.1–2 km. In total, this yields the uplift of ~2.5–3 km that is commensurate with the actual average uplift of $\geq$3 km. Therefore, the possible additional uplift due to the lower-crust deconsolidation did not probably exceed 0.5 km.

1.2.6. Relative significance of different processes producing neotectonic uplift of the Central Tien Shan

The performed analysis of the data permitted to describe the neotectonic evolution of the CTS as the following (Trifonov et al., 2008). By the Early Paleogene, the surface of the Earth's crust (Paleozoic basement) was situated at a small height above the sea level. The density of the upper layer of the crust, ~42 km thick, was close to the average density of the crust in platform regions. Beneath this layer, there was a layer of garnet granulites and eclogites with an average density close to the mantle one. The boundary between the lower crustal layer and the mantle lithosphere was probably very uneven and uncertain. The Early Cenozoic basaltic eruptions indicate a possible ascent of small volumes of deep-seated mantle matter containing active fluids, to the lithosphere. Later the infiltration of fluids increased that reduced the lithosphere strength. In the Oligocene, under the compression caused ultimately by the India-Eurasia collision, the lithosphere including the Earth's crust was subjected to folding and faulting. From the Oligocene to the Pliocene, i.e., over the period of ~28 Myr, the average rate of crustal lateral shortening was ~2 mm/year. The compression led to a slow isostatic uplift of the crust, and its average height reached 1–1.3 km by the beginning of Pleistocene.

By the Late Pliocene or Early Pleistocene, large portions of the asthenosphere matter that were enriched by fluids, penetrated beneath the CTS. Infiltration of the fluids into the lithosphere drastically reduced the viscosity of the latter. The deconsolidated lithosphere was detached into layers along the surfaces with the highest gradient of the deformational properties. The detached lithosphere mantle began rapidly to destruct, submerge, and be convectively replaced by the matter of the hot and less dense asthenosphere that resulted in the rapid uplift of the CTS in the last ~2 Myr. This process was most intensive beneath ridge zones, where the asthenosphere closely approached the base of the crust (Lithosphere of the Tien Shan, 1986; Vinnik et al., 2006). Beneath the crust of large intermontane basins, lenses of the lithospheric mantle had been preserved; therefore, the basins rose to a smaller height than the ridges. Unlike the subsided heavy lithospheric mantle including the garnet granulites and eclogites of the crustal origin, the high-pressure metabasites, less consolidated than the asthenosphere, remained near the Moho. As the crust was uplifted, they transferred into the stability fields of less consolidated basites. After the supply of the cooled asthenosphere fluids, these rocks underwent the retrograde metamorphism that led to their partial deconsolidation and, as a consequence, the additional uplift of the crust during the last ~2 Myr.

In general, the Oligocene-Quaternary collision compression of the Earth's crust led to ~1–1.3 km uplift of the CTS; the replacement of the lithospheric mantle by the asthenosphere produced 1.1 to 2 km of the uplift in different parts of the region; and the probable metamorphic deconsolidation of metabasites near the crust-mantle boundary gave ~0.5 km of the uplift (Trifonov et al., 2008). These three processes gave rise to the mountain system of $\geq$3 km in average height. The proportion between influence of each of the processes is about 7/10/3.

Relative significance of three mentioned processes changed in time. During the long period from the Oligocene to the beginning of Pleistocene (~28 Myr), the collision compression was the only factor responsible for the slow rise of neotectonic structures. Its average magnitude reached 0.7–1 km. In the Quaternary, during the last ~2 Myr, the compression intensified and the average rate of the shortening increased probably 3–4 times. Nevertheless, the contribution of the compression into the total rise was limited only by 11–17%. Two other factors that had not acted before, were mainly responsible for the rapid uplift of the CTS during the last ~2 Myr. They were the replacement of the lithospheric mantle by the asthenosphere and the metamorphic deconsolidation of metabasites near the crust-mantle boundary. Their contribution to the total uplift in the last ~2 Myr is estimated approximately at 60–70% and 20–25%, respectively.

1.3. Neotectonic evolution of the Pamirs and surrounding

The mountain system of Pamirs consists of topographic features convex to the north. The eastern Pamirs is the high-elevated plateau that represents the axial part of this arc. Ridges of the western Pamirs strike to the NE–SW and ridges of its eastern (Chinese) termination strike to the NW–SE. More southern mountain systems, the Hindu Kush and Karakorum and the Kohistan and Ladakh, have the same arc-type pattern, which reflects the Mesozoic tectonic zonation of the region. All these tectonic zones and topographic features form the Pamir-Penjab syntaxis. This is an area of intense tectonic deformation related to Neotethys closure. Collision at its northern flank was accompanied by volcanic activity and large-scale granite formation that testifies to heating of the Earth's crust. This heating could promote delamination of the crust along surfaces with the highest gradient of mechanical properties. Such delamination provides differentiated displacements of crustal sheets and blocks under variously oriented horizontal compression. By the end of Miocene, this resulted in substantial disturbance of isostatic equilibrium, which was produced by compressive thickening of the Earth's crust and decrease of density of the lithospheric mantle and stimulated intense and contrasting vertical movements in the Pliocene and Quaternary.

The present-day tectonic zoning of the syntaxis [Desio, 1976; Shvolman, 1980; Shvolman, Pashkov, 1986; Geological Map of the Tajik SSR…, 1989; Ruzhentsev, 1990; Searle, 1991; Gaetani, 1997, Burtman, Samygin, 2001; Pashkov, Budanov, 2003) (Fig. 11 & 12) reflects its crustal structure, which was formed as a result of multifold deformational events during the stage-by-stage closure of the Tethys. The existing structural pattern was eventually formed at the late collision stage following the closure of the Neotethys. This span of time corresponds to the neotectonic epoch lasting from the Oligocene to the Recent (Trifonov, 1999). This period is subdivided into the early stage (Oligocene–Miocene), when heating and tectonic delamination of the crust were the most important factors of tectogenesis; the late stage commenced in the Pliocene–Quaternary, when the role of these processes decreased and intense vertical movements were occurring.

1.3.1. Mesozoic zoning and its deformation due to neotectonics

In the present-day structure of the eastern Pamirs, the consecutive series of tectonic zones exhibits the evolution of the early Mesotethys. The Hercynides of the Northern Pamirs, where the main structure-forming processes ceased by the end of the

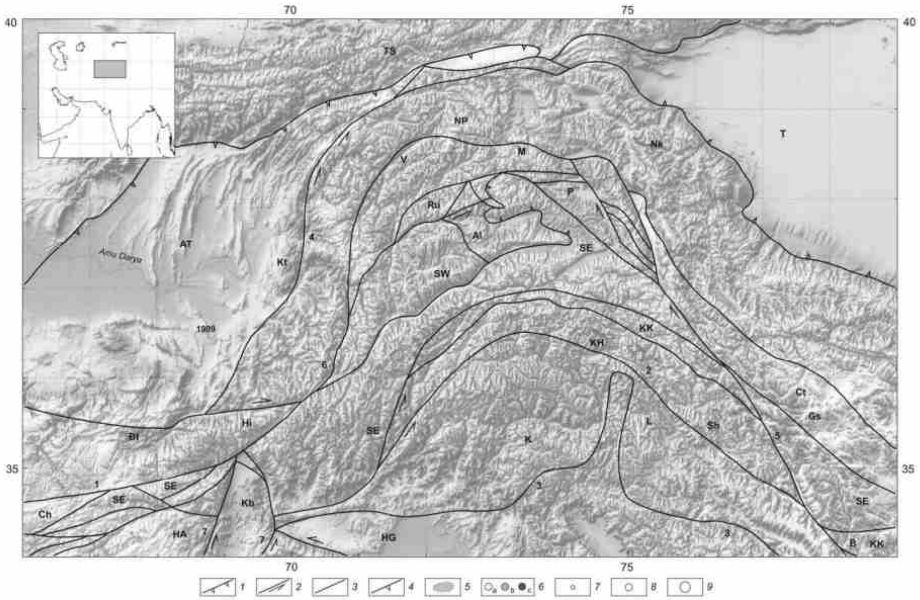


Fig. 11. The orographic map of the Pamir–Karakorum region and its surrounding with contours of tectonic zones that are shown in fig. 12, modified after (Ivanova, Trifonov, 2005). Symbols of the tectonic zones are the same as in fig 12

Paleozoic, developed in the Triassic as a volcanic arc at the active northern flank of the basin underlain by oceanic crust. The arc is marked by Triassic subduction-related granites and calc-alkaline volcanics. The nonvolcanic part of the arc that comprised continental blocks of the Central Pamirs, heterogeneous in their geological history and structure, accreted to the Hercynides during the Permian after the closure of the Paleotethys. In the considered part of the region, the Central Pamirs, represented by the Muzkol Zone (Ruzhentsev, 1990), is underlain by crust 60–65 km thick; its lower part (approximately 35 km) is seismically homogeneous (Seismic models…, 1980; Pamirs–Himalayas…, 1982). The basin itself is designated by the Pshart Suture, where the Upper Permian–Triassic sequence is largely composed of clayey and cherty slates, basalts, and basaltic andesites; volcanics prevail in its Upper Triassic portion (Pashkov, Budanov, 2003). This sequence is overlain with unconformity by Norian (?) volcanogenic and terrigenous rocks with olistoliths of Paleozoic limestone. Northward, in the western Pshart and the northern Dunkeldin blocks, the Permian–Triassic calcareous–terrigenous sequences with sporadic volcanics mark the northern periphery of the basin (Pashkov, Budanov, 2003). Its southern periphery is made up of an allochthon of the South-Eastern Pamirs, where relatively deep-water flyschoid

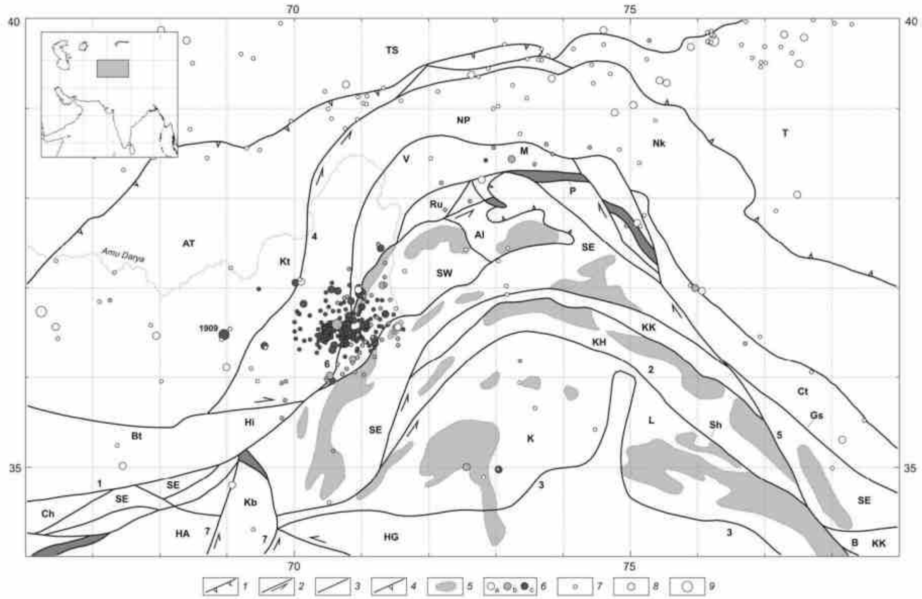


Fig. 12. The Pamir–Karakorum region, modified after (Ivanova, Trifonov, 2005): tectonic zones after (Burtman, Samygin, 2001; Desio, 1976; Gaetani, 1997; Pashkov, Budanov, 2003; Ruzhentsev, 1990; Searle, 1991; Shvolman, 1980), granitic nagmatism after (Desio, 1976; Geological Map of the Tajik S.S.R., 1989), and epicenters of strong earthquakes ($M_S \geq 5.7$)

1, nappes and thrusts; *2*, strike-slip faults; *3*, other major faults; *4*, boundaries of basins; *5*, granitic batholiths that continued to develop in the Miocene; *6*, epicenters of earthquakes with different depths (*h*) of hypocenters: *a*, $h \leq 70$ km, *b*, $70 < h < 150$ km, *c*, $h \geq 150$ km; *7–9*, magnitudes of earthquakes: *7*, $M_S = 5.7$–6.5, *8*, $M_S = 6.6$–7.4, *9*, $M_S = 7.4$–8.3.

Tectonic zones: AT, Afghan-Tajik Basin with the Kulyab trough (Kt); T, Tarim Basin; NP, North Pamir zone and its continuations: Nk, NW Kunlun, Hi, Western Hindu Kush, and Bt, Bandi-Turkestan; zones of the Central Pamir type: M, Muzlol, V, Vanch, SW, South-Western Pamir and Badakhshan, Al, Alichur Block, Ru, Rushan zone, Kb, Kabul Block, Ct, continuation of the Central Pamir zone in Tibet, and Ch, fragments of the Central Pamir type in the Herat fault zone in Afghanistan; P, Pshart suture and its continuations (shown by dark-grey color): Db, Dunkeldin Block, Gs, its Tibetan continuation that continues to the SE as the Ganmatso-Shuanhu suture, Vf, Vatasaif fragment, Ar, Altimur ophiolites, and H, Khashrud zone; SE, the South-Eastern Pamir and Nuristan zone and its continuations; KK, the North Karakorum zone and its Tibetan continuation; HA, Helmand-Argandab Block; KH, the Southern Karakorum and Eastern Hindu Kush zone; Sh, Shyok suture, and B, Bangun suture; K, Kohistan, and L, Ladakh; HG, the Khazar segment of the Himalayas. Batholiths: 1, Bagarak; 2, Karakorum; 3, Kohistan; 4, Lagman; and 5, Shugnan. Faults: 6, Alichur Thrust; 7, Andarab strike-slip fault; 8, Herat (Main Herirud) fault zone; 9, Main Karakorum Thrust; 10, Main Mantle Thrust; 11, Gunt Fault; 12, Darvaz reverse-sinistral fault; 13, Zebak Fault; 14, Kunar-Tashkupruk zone; 15, Pamir-Karakorum strike-slip fault; 16, Central Pamir Fault; 17, Chaman strike-slip fault

facies of the passive slope give way to the carbonate platform facies (Ruzhentsev, 1968, 1990). Both of these facies extends toward Nuristan (Geology and Mineral Resources…, 1980). The similarity in the early collision evolution of the Pshart and

South-Eastern Pamir–Nuristan regions is expressed in the pre-Jurassic unconformity (Pashkov, Budanov, 2003) and in the occurrence of Cretaceous orogenic complex (Shvolman, 1977).

Farther southward, there is a succession of tectonic zones related to the late Mesotethys and Neotethys:

(1) The Northern Karakorum is underlain by the Proterozoic–Cambrian continental basement overlapped by the polycyclic Ordovician–Jurassic cover, with carbonate rocks prevailing over terrigenous sediments and with signatures of the mid-Cretaceous orogeny (Gaetani, 1997);

(2) The Southern Karakorum and the Eastern Hindu Kush that reveal intense regional metamorphism, enclose an axial batholith in the north, and are bordered by the Main Karakorum Thrust Fault in the south (Gaetani, 1997); the eastern part of this fault controls the Shyok Suture, a relict of the backarc (?) basin of the late Mesotethys that closed in the mid-Cretaceous, which is represented now by ophiolite melange (Searle, 1991);

(3) The Kohistan and Ladakh volcanic arc of the Neotethys with large granitic batholiths; the base of this section (ultramafics and garnet granulites overlain by amphibolites and gabbronorites) is exposed in the southern part of the zone, where it is bordered by the Main Mantle Thrust Fault (Khain, 2001).

This zonal pattern of the Pamir–Karakorum region likely indicates that the relative location of zones has remained principally unchanged since the late Mesozoic (the neotectonic period included). The following tectonic units among the Afghan zones serve as the most definite analogs of the Pamirs–Karakorum zones: the volcanic arc of the early Mesotethys inheriting the Hercynides in the Hindu Kush and Bandi-Turkestan and the Quetta ophiolite zone (the Neotethys suture). The latter soundly correlated with ophiolite of the Indus–Zangbo Zone as the southeastern extension of the Ladakh Zone (Gansser, 1966). The Altimur allochthonous ophiolite melange in the northern Kabul Block composed of peridotites, pillow lavas, tuffs, and cherts that are overlain by limestone with poorly preserved Jurassic (?) fauna (Tapponnier et al., 1981) is probably an analog of the Pshart Suture.

Westward, in central Afghanistan, the SW-trending ophiolitic Khashrud Zone branches out the Herat (Main Herirud) Fault. The Upper Jurassic–Hauterivian sequence of this zone is composed of basic and intermediate volcanics replaced upward by sandy–clayish sediments; ultramafics and gabbrodiorite intrusions are widespread (Geology and Mineral Resources…, 1980). It is assumed that this section

accumulated in a trough underlain by oceanic crust (Sborshchikov, 1988). The fact that, at the northwestern periphery of the trough, the Upper Jurassic volcanics are underlain by Rhaetian–Liassic sandstones and slates, as well as by Upper Permian–Norian calcareous–terrigenous rocks alternating with basic and intermediate volcanics (Geology and Mineral Resources…, 1980), points to almost coeval origination of the Khashrud and Pshart basins. It can be assumed that the Khashrud ophiolite is a fragment of the Pshart Basin extension, which continued to evolve, in contrast to the Pamirs, in the Jurassic and Early Cretaceous. Its evolution terminated by the mid-Cretaceous, as is evident from the unconformity at the base of calcareous–terrigenous partly red-colored Aptian–Upper Cretaceous sequence.

Tectonic blocks with structural features similar to those of the southeastern Pamirs and Nuristan are indicated in the zone of the Herat (Main Herirud) Fault. Gaetani (1997) notes the similarity in sedimentary covers of the northern Karakorum and the Helmand–Argandab continental massif bordered by the Khashrud ophiolite in the northwest. The Shyok Suture appears to be coeval with the Tarnak Suture at the southeastern flank of this massif (Sborshchikov, 1988).

Thus, the systems of the Mesozoic tectonic zones in the Pamirs and Afghanistan are similar, although there is no complete identity between them. However, most of the zones, which can be regarded as analogs, are tectonically separated by faults that extend along the western flank of the Pamirs and Badakhshan (Geology and Mineral Resources…, 1980; Geological Map of the Tadjik SSR…, 1989). Here, in the Vanch Zone of the central Pamirs and tectonic nappes of the Rushan Zone corresponding to the northern margin of the Pshart Basin, the Earth's crust is thinned to 50–55 km and its granitic–gneissic portion (approximately 35 km) rests upon the layer defined by seismic velocities like a mantle–crust mixture (Khamrabaev, 1980). This layer can be a relic of the early Mesotethys oceanic crust. The Vanch and Rushan zones pinch out southeastward, and the extension of the northern Pamirs borders along the steep Central Pamir Fault on the Archaean metamorphic massif of the West Pamir–Badakhshan zone. Thrusting of the Shakhdara Group over the Goran Group in the Precambrian resulted in a doubled section of the massif. Tectonic sheets at the contact are composed of the rocks pertaining to the Khorog unit and are formed in the lower crust close to the Moho discontinuity (The Earth's crust and upper mantle…, 1981; Budanova, Budanov, 1983; Ruzhentsev, 1990). Contacts of the massif with neighboring zones are either tectonic or sealed by Cenozoic granites. Its margins experienced maximum Cenozoic tectono-metamorphic reworking (Budanova, Budanov, 1983). The Kabul Block separates the northern Karakorum and the

Helmand–Argandab Massif, as well as Nuristan and its probable extension in the Herat Fault Zone. The Precambrian basement of the Kabul Block is overlain by the Upper Precambrian–Lower Paleozoic metaterrigenous complex and by the Upper Paleozoic complex, including the Upper Permian–Norian carbonate rocks. The Kabul Block is similar in this respect to the Muzkol Zone of the Central Pamirs (Geology and Mineral Resources…, 1980; Pashkov, Budanov, 2003).

The southwestern Pamir–Badakhshan Massif has been studied better as compared with the poorly explored Kabul Block. The southeastern tectonic boundary of the massif with Nuristan is marked by the Lagman Batholith that dates to the Oligocene–Miocene up to 16.5 Ma (Geology and Mineral Resources…, 1980). Northward, the Bagarak Batholith, 32–19.5 Ma in age (Geology and Mineral Resources…, 1980), extends along the boundary with the Central Pamirs. The batholith contacts, sharp intrusive in the northwest and complicated by numerous local injections in the southeast, suggest that the batholith (and, correspondingly, the boundary of the zones) plunges beneath the Archaean complexes (The Earth's crust and upper mantle…, 1981). To the east, at the boundary of Precambrian rocks with the Rushan Zone, a similar plunge of the Alichur Thrust is confirmed by geologic observations (Ruzhentsev, 1968). South of this thrust, the Precambrian–Paleozoic Alichur Group of metamorphic rocks crops out between Archean rocks and allochthon of the South-Eastern Pamirs. The Vatasaif fragment of the Pshart Suture, where Triassic volcanogenic rocks are overlain with unconformity by Jurassic strata, is retained farther to the east (Pashkov, Budanov, 1990, 2003). Boundaries between all these complexes are either tectonic or concealed by granites. The isotopic age of the largest Shugnan Batholith is estimated as 32–21 Ma; the recurrent metamorphism of older sequences took place approximately at the same time, 32–9 Ma ago (Shvolman, 1977).

The relationships described above suggest that the South-Western Pamir Massif has occupied its present-day location only recently, and the age of the boundary batholith emplacement corresponds to tectonic convergence of the South-Western and South-Eastern Pamirs. We suggest that the Triassic–Jurassic facial zones of the South-Eastern Pamirs that initially extend parallel to the Pshart Suture were curved during this convergence and formed an arc with the western margin that trends parallel to the boundary of the South-Western Pamirs. Judging from the bend configuration, the amplitude of the eastward or northeastward offset of the South-Western Pamirs could exceed 150 km. Thereby, the sedimentary sequences of the South-Eastern Pamirs were involved in the thrusting, and later, in the Pliocene and

Quaternary, they were subjected to strike-slip movements (Ruzhentsev, 1968). The Pshart Suture was also involved in bending that is evident from the location of its Vatasaif fragment. The area with the exposed Alichur Group, which is probably a subsided continuation of the South-Western Pamirs, also changed its location and was deformed.

According to the geophysical data, the granitic-gneissic complex of the South-Western Pamirs is 25 km thick, while the total thickness of the crust reaches approximately 60 km (The Earth's crust and upper mantle…, 1981). A part of the displaced complex likely overlapped the crystalline basement of the South-Eastern Pamirs that reaches a thickness of 30 km. To determine the initial structural setting of the complex, it is important to note that it could not be an element of the northern Pamirs, because no indications of Paleozoic and Early Mesozoic magmatism of this zone are known. Thus, it was probably an element of the Central Pamirs.

The Precambrian clastic material derived from the South-Western Pamirs is missing in the Upper Mesozoic and Lower Cenozoic sequences of adjacent zones and first appears in the immediate vicinity of the massif only in Oligocene sediments (Shvolman, 1977). This implies that the Precambrian complex was initially covered by sediments, fragments of which are represented by the Permian–Triassic sequence of the Central-Pamir type in the Zebak Fault Zone at the southern flank of the massif (Geology and Mineral Resources…, 1980). This might be responsible for the formation of the allochthonous series of the Vanch–Muzkol segment of the Central Pamirs, the nappe structure of which is a result of neotectonic movements, because it involves Upper Cretaceous and Paleogene strata (Ruzhentsev, 1990). In the opinion of Ruzhentsev (1971), the recumbent folds characteristic of the early deformation stage began to form here in the mid-Cretaceous or in the Paleogene and continued to develop until the Neogene, because they involve Paleogene sediments. Later, already in more recent times, structures of sedimentary cover of the Vanch Zone, where the root belts of nappes have been formed, were thrust over southerly and easterly areas of the Central Pamirs, the Muzkol Zone inclusive. Other authors (Leonov, Sigachev, 1984; Leonov, Nikonov, 1988; Sigachev, 1990) provided persuasive structural arguments in favor of thrusting from the south. Pashkov and Budanov (2003) assumed that the nappe rocks originated in the Kunar-Tashkupruk zone between the South-Eastern Pamirs and the Karakorum. We suppose that they originated nearer to their present-day location and are a detached cover of the displaced South-Western Pamir–Badakhshan Zone. The detachment was stimulated by heating and delamination of the massif that is reflected in intense generation of Cenozoic granites (the Shugnan Batholith) and by the uplift that followed the crust thickening.

Thus, the most evident deformation of the Mesozoic tectonic zoning caused by neotectonic deformation and offsets is confined to the transition between the Pamirs and Afghanistan; it is manifested first of all by the displacement of the massif of the South-Western Pamirs and Badakhshan.

1.3.2. The Pamirs and Afghan–Tajik Basin

The Afghan–Tajik and Tarim basins filled with Upper Cenozoic molasses are located on both sides of the northward-convex zone of the northern Pamirs. The Tarim Basin rests largely upon the Precambrian basement. The Afghan–Tajik Depression is a sedimentary basin with a heterogeneous basement that was consolidated by the end of the Paleozoic and probably inherited an ancient crystalline massif. The basin is filled with a thick (up to 18 km) sequence of alternating shallow-water and continental deposits or only continental (since the Oligocene) sediments. Compositionally similar Cretaceous and Cenozoic sequences extend along the northern periphery of the Pamirs and form its outer zone. In the northeast, the northern Pamirs is thrust over molasses of the Tarim Basin (Ding Guoyu, 1984), and this probably resulted in crust thickening to 75–80 km (Seismic models…, 1980; Pamirs–Himalayas…, 1982). To the west, the Northern Pamirs is thrust over the outer zone that determined its present-day structure (Neotectonics and recent geodynamics…, 1988). A waveguide with *Vp* of 6.0–6.3 km/s (Khamrabaev, 1980; Pamirs–Himalayas…, 1982; Makarov et al., 1982) at a depth of 5–10 km under crystalline rocks of the northern Pamirs favored this process.

The thrusting was accompanied by development of the fold structure in the Afghan–Tajik Basin, the formation of which was strongly influenced by the detachment of the 5–6-km-thick Cretaceous–Miocene cover along the Malm salt-bearing sequence (Zakharov, 1958; Becker, 1996). This growth of folds fell mainly in the late neotectonic stage, and the first regional unconformity in the molasse section that reflects this event is dated as the Late Miocene. During the folding, the sedimentation basin experienced differentiation, and the Kulyab Trough located in its eastern part accumulated 11 km of Pliocene–Quaternary sediments of the 17-km total of sedimentary cover. The folding and accumulation of young molasses transformed the crust beneath the depression. Its Cretaceous and Paleogene structure can be judged from the weakly deformed section in the Kurgan-Tyube area. The crust is

approximately 35 km thick there, and the thickness of its crystalline part is less than 20 km (The Earth's crust and upper mantle…, 1981).

The magnitude of the Northern Pamirs thrusting over the neighboring depressions is critical for estimating neotectonic deformations. Based on paleomagnetic studies of Cretaceous–Paleogene sediments in the Afghan–Tajik Basin (Bazhenov, Burtman, 1990) and on facies distribution, Burtman (1999) arrived at the conclusion that the Northern Pamirs was thrust over the eastern part of the Cretaceous–Paleogene trough approximately at 300 km. The sedimentary cover was detached and folded in the retained part of the basin (Becker, 1996; Burtman, 1999). We assume that the magnitude of overthrusting could have been less, particularly in the eastern part of the Pamirs. There are two reasons for this. First, the Cretaceous–Paleogene trough might have become narrower eastward prior to the neotectonic stage due to framing of ancient massifs by the Hercynides. Second, in the western Pamirs, conditions of the Hercynian complex thrusting over the thinned crust of the central part of the basin were more favorable than in the east, where the crust was normal. As concerns the folding controlled by the general detachment and displacement of sedimentary cover, this mechanism is acceptable only for the northern part of the basin and becomes doubtful in its southern part, where the detached anticlinal zones are separated by sizeable depressions that remain almost undeformed. That is why a more complex mechanism of folding has been proposed. This mechanism takes into account the change of sedimentary rock volume in response to its chemical alteration (Zakharov, 1958).

Thus, the fact that the northern Pamirs is thrust over the Afghan–Tajik Basin and partially overlaps its eastern part is beyond doubt, although the amplitude of overthrusting remains debatable. In any event, it is at least 100 km large.

Comparative analysis of the Oligocene-Quaternary sections of the closest to the Pamirs eastern part (near the Darvaz Ridge) and the central part of the Afghan–Tajik Basin is important to understand a history of neotectonic uplift in the Pamirs. The base of these sections is represented by the lower Sumsar and upper Shurysay beds that demonstrate a change of the Eocene mainly marine sedimentation to the younger continental one. The Shurysay Beds contain the Upper Oligocene fossils. All younger deposits belong to the fluvial or subaerial continental types of sedimentation. They are divided to several units (International Symposium…, 1977).

The Baljuan unit is represented in the eastern part of the basin by red sandstones and siltstones with gravelstone and conglomerate interbeds (Fig. 13). Their thickness is up to 1200 m. In the central part of the basin, they are replaced by sandstones, siltstones, and shales.

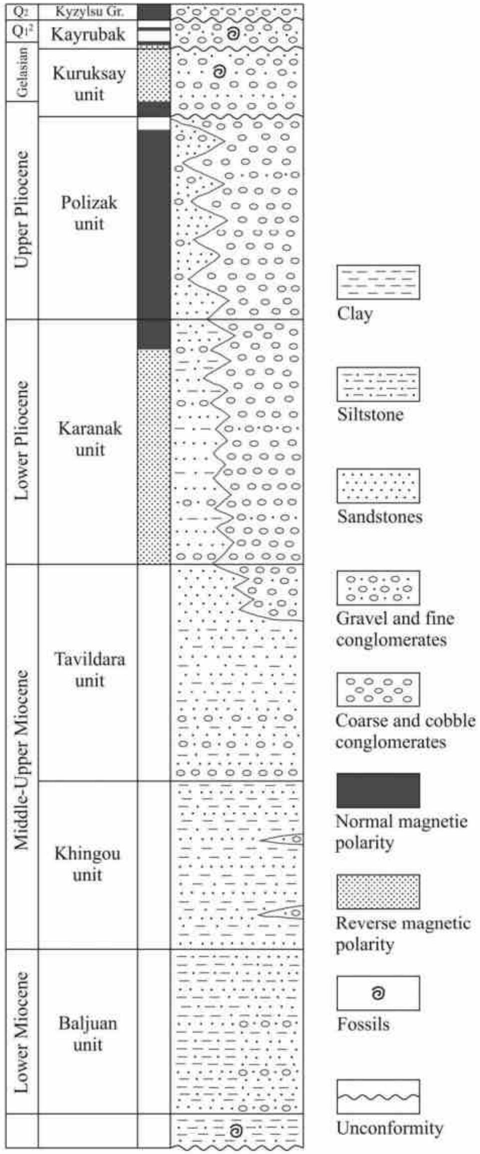


Fig. 13. Composite section of the Oligocene-Quaternary molasses in the eastern Afghan-Tajik Basin, compiled after (International Symposium…, 1977)

The alluvial sectins of the Khingou unit are composed near the Darvaz Ridge of alternating variegated (mainly brown-colored, red-colored and grey) sandstones, siltstones and shales with congromerate lenses that were sedimented by temporal streams. Thickness varies from 400 m to 1700 m. The following Tavildara unit consists of three members in the Darvaz foothills. A member of alternating layers of red-brown and brownish-grey siltstones, sandstones, and conglomerates occurs at the base of the unit. Higher up, there occur brownish-grey sandstones with siltstone interbeds. The unit is crowned with strata of grey conglomerates that disappear to the south and west of the town of Kulyab. Total thickness of the Tavildara unit is up to 1600 m. In the central basin, two mentioned units are replaced by the Kafirnigan unit that is composed of grey sandstones with interbeds of brown siltstones.

The up to 1800 m thick Karanak unit consists of coarse and cobble conglomerates near the Darvaz Ridge. They contain rare lenses of brown siltstones. In the central basin, the unit is mainly composed of fine-grained sediments. The next Polizak unit is represented in the Darvaz foothills by over 1500 m sequence of grey coarse

and cobble conglomerates. Southwestwards, conglomerates gradually change to sandstones. In the central basin, the Polizak deposits are partially denuded and cannot often be distinguished in the younger strata.

The younger units demonstrate change of the basin regime of sedimentation to the terrace one. Correspondingly, the younger units are distributed locally and have smaller thickness. These deposits were primary recognized as the Kulyab group. Later it was divided into the Kuruksay and Kayrubak units (International Symposium…, 1977). The Kuruksay unit covers the older molasses with angular inconformity. The unit is represented in the Kuruksay River by 180–200-meter thick sequence of coarse clastic material with predominance of boulder gravels. Other sections of the unit contain lenses and layers of sand and silt. The maximum registered thickness of the unit is 500 m, The Kayrubak unit covers the older deposits also with inconformity and consists of rhythmically altenating pebble and sand layers up to 200 m thick totally. This type of Kuruksay and Kayrubak deposits represents alluvial sedimentation. At the same time, alternation of loess-like silts and carbonate paleosoils were accumulated on watershed slopes. There were found signs of a synchronism between coarser members of fluvial sections and paleo-soil horizons within the Kayrubak unit (International Symposium…, 1977). The younger units form the Kyzylsu group that is represented by loess-soil and fluvial deposits. The fluvial deposits compose series of terraces set into the older molasse formations. Their total thickness is up to 120 m.

The units mentioned above are dated by combination of paleomagnetic and paleontological data together with geological correlation of the units (International Symposium…, 1977). The Baljuan unit is assigned to the Lower Miocene by its occurrence: according to paleontological data, the underlying Shurysay beds are correlated with Upper Oligocene and the overlying Kafirnigan unit corresponds to the Middle to Upper Miocene. The last estimate is based on the find of a scull of *Mastodon cf. angustidens* in the upper part of the Kafirnigan unit and gives also the age of the Khingou and Tavildara units that are correlated with the Kafirnigan one.

The lower part of the Karanak unit, as well as the upper part of the Tavildara unit are characterized by reverse magnetic polarity. The uppermost part of the Karanak unit, the Polizak unit, and the lowest part of the Kuruksay unit demonstrate normal polarity (the Gauss chron). The upper and main part of the Kuruksay unit as well as the Kayrubak unit and the upper terrace deposits of the Kyzylsu group show reverse polarity (the Matuyama chron). Two episodes of normal polarity were found within the Kayrubak unit and identified as the Jaramillo and Olduvai subchrons. All younger deposits of the Kyzylsu group have normal polarity (the Brunhes chron). Mammal remnants from the upper Kuruksay deposits were attributed to the middle Villafrancian and the fauna from the upper Kayrubak bone beds was identified as post-Villafrancian (International Symposium…, 1977). These data give a possibility to assign the Karanak unit to the Lower Pliocene, the Polizak unit to the Upper Pliocene, the Kuruksay unit to the lower part of the Lower Pleistocene (Gelasian), and the Kayrubak unit to the upper part of the Lower Pleistocene (Calabrian, probably including the uppermost Gelasian). The Kyzylsu group is attributed to the Middle and Upper Pleistocene.

The analysis of composition and age of different molasse units show that the Pamir region did not highly rise till the Upper Miocene. Signs of local mountain uplifts arrived in the Late Miocene, but the total uplift with erosion and transportation of coarse clastic material including boulders occurred only in the Pliocene–Quaternary.

1.3.3. Recent geodynamics of the Pamir–Hindu Kush region

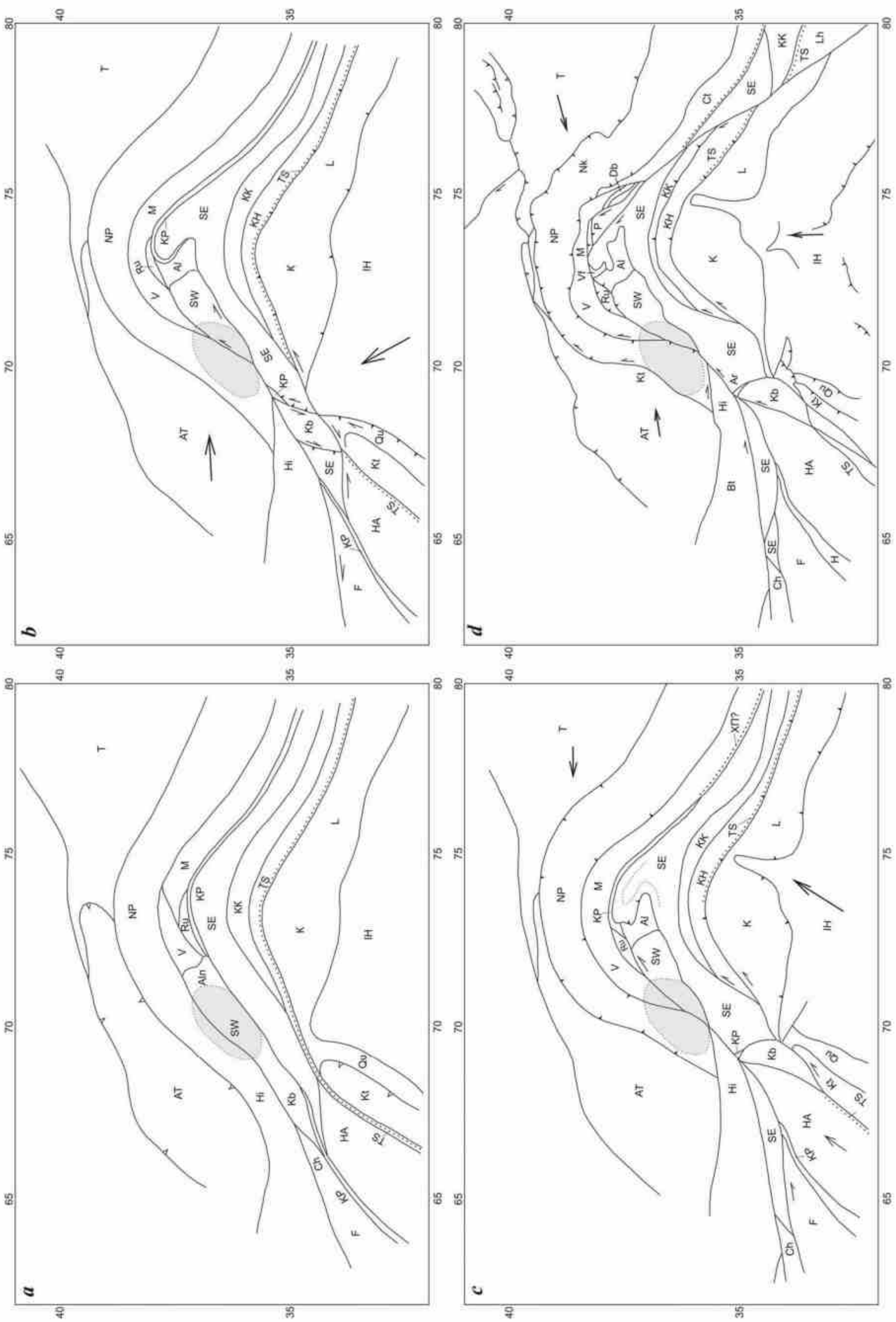
The most recent structure of the Pamirs was formed under horizontal compression commonly interpreted as a result of the pressure from the Punjab indenter of the Indian Plate. This assumption is consistent with the arcuate bend of the Pamir Zone: in particular, a bend of the northern Pamirs for 350–400 km with indications of N-trending compression and shortening in the W–E-trending thrusts and folds, the conjugated left-lateral slip along the Darvaz Fault, and right-lateral displacements in the southeastern Pamirs. However, magnitude of the arcuate bend in the southerly located Karakorum and Kohistan–Ladakh tectonic zones is only 200 km. This bend is conformable to the northern margin of the Indian Plate, and probably was formed immediately after the Neotethys closure, that is preceding, at least partly, the neotectonic epoch.

At the same time, the western and eastern flanks of the Pamirs bear indications of nearly W–E-trending recent compression and shortening. In the west, where the Hindu Kush and the North Afghan Hercynides join the Southwest Pamir–Badakhshan and Central Pamir zones, this deformation is expressed in the N–S-trending steep wedges, slices, and compressed folds with signs of transverse rock flattening, while the northwestern Kunlun demonstrates signs of the Hercynides thrusting over the Tarim Basin. Thus, the neotectonic structure of the Pamirs was formed under differently oriented compression.

Such an intricate structure of the Pamirs could result from variation of geodynamic settings during the neotectonic period. Its early stage (Late Eocene, Oligocene, and Early Miocene) was characterized by significant, although irregular, heating of the Earth's crust that gave rise to the emplacement of numerous large batholiths both along the fault-related boundaries and in axial zones of tectonic uplifts. These batholiths (Kohistan, Ladakh, Karakorum, Shugnan, and others) began forming in the Cretaceous or Paleogene at the onset of collision in the respective tectonic zones and continued to form until the Miocene. In some batholiths, the main phases of granite formation are related to the late collision (neotectonic) stage. Heating stimulated delamination of crustal rocks along the surfaces with the highest gradients of mechanical properties and differentiated lateral displacements. The heating of the thinned crust of the Afghan–Tajik Basin likely resulted in extension and volcanic activity at its southern flank, which was intensive at the early stage of the neotectonic period and lasted until the Mid-Pleistocene (Geology and Mineral Resources…, 1980). Since the Late Miocene, the heating of the Earth's crust waned, and the crust became more homogeneous in its physical properties and less favorable for tectonic delamination.

This background was complicated by changes in direction of maximal lateral compression in the orogenic belt (Fig. 14) that were similar to the changes in other regions of the Alpine-Himalayan Belt (Trifonov, 1999). Since the Late Eocene and until the Early Miocene (approximately 40–20 Myrs ago), the axis of maximal compression at the northern and western flanks of the Indian plate was probably oriented in the NW–SE direction. Intense transverse shortening was also recorded in the northern Quetta Zone, where the Eocene Katavaz Trough was deformed and the NE-trending tectonic nappes and thrust sheets were formed in the Khost, Tarnak, and Khashrud ophiolitic zones (Geology and Mineral Resources…, 1980; Tapponnier et al., 1981; Sborshchikov, 1988). They were conjugated with right-lateral movements along the nearly W–E-trending Gerat Fault Zone, along which the Khashrud Zone was displaced for 150 km relative to the Altimur ophiolite. The dextral displacement could also result from extension in the Afghan–Tajik Basin that was brought about by movement at 40 km along the W–E-trending Andarab Fault (Geology and Mineral Resources…, 1980).

Intense heating of the crust and its rheological delamination in the narrowest tract of the orogenic belt between the western Khazar Massif of the Himalayas and the salient southeastern margin of the Turan Plate could result in destruction and squeezing of crustal blocks away from this area. The South-West Pamir–Badakhshan



Block, which was formerly a part of the Central Pamir zone, moved eastward; this led to the detachment and sigmoid bend of the Pshart Suture and lithotectonic zones of the South-Eastern Pamirs, where tectonic nappes began forming. The sedimentary cover of the South-Western Pamirs became detached and formed the Vanch–Muzkol nappes of the Central Pamirs. It appears likely that the Kabul Block moved southward at that time dividing Nuristan from its western continuation and separating the Karakorum and Helmand–Argandab massifs.

Since the Early Miocene and until the Late Miocene (20–8 Ma ago), the Indian Plate moved to the northeast, and, correspondingly, the maximal compression and lateral shortening were oriented in the north-eastern direction. This was expressed in thrusting, granitization, and metamorphism in the Himalayas and the Karakorum (Gansser, 1964; Desio, 1976; Ratschbacher et al., 1993) and volcanism in Tibet. Involved in intense deformations, Tibet and Qaidam might have, in turn, exerted influence upon the Tarim Massif. The left-lateral Altyn Tagh strike-slip fault zone arose along its southeastern boundary; as a result, the Tarim drift acquired a substantial western component and compressed the Pamirs. Central Afghanistan was also involved in the northeastward drift. The sinistral slip occurred along the Herat Fault and continued with the Gunt Fault as its extension. These movements enhanced the displacement of the South-Western Pamir–Badakhshan Block and reinforced deformation of neighboring zones. In particular, the nappe structure of the South-Eastern Pamirs was eventually formed, and the northward-bent North Pamir Zone began to thrust over the Afghan–Tajik Basin.

Since the Late Miocene and at the Pliocene–Quaternary stage of the neotectonic epoch, when the Earth's crust was homogenized in its physical properties, direction of the Indian Plate pressure in the Pamir segment of the orogenic belt became close to the N–S, giving rise to the W–E-trending thrusts and folds and related strike-slip fault zones at the eastern and western flanks of the region. These zones have remained active till now (Trifonov et al., 2002). Simultaneously, W–E-trending compression of

Fig. 14. Conceptual maps of geodynamics and tectonic zonation of the Pamir-Karakorum region in the different substages of neotectonic epoch, modified after (Ivanova, Trifonov, 2005): a, the tectonic zonation at the Late Eocene; b, the geodynamics from the end of Eocene till the Early Miocene and the tectonic zonation at the late Early Moicene; c, the geodynamics from the end of Miocene till the Late Miocene and the tectonic zonation at the Late Miocene; d, the geodynamics from the end of Miocene till present time and the recent tectonic zonation

IH, Indian Platform and the Himalayas; KP, Khashrud-Pshart zone indifferentiated; Kt, the Katavaz Trench; Qu, the Quetta zone including the Host ophiolites; Lh, Lhassa Block; TS, the Tarnak-Shyok-Bangun suture; F, Farakhrud zone. Other symbols are the same as in fig. 12

the Pamirs continued. In the east, it experienced pressure from the Tarim Block, the drift of which had a western component due to the left-lateral movements along the Altyn Tagh Fault with a rate that reached 1 cm/yr in the Quaternary. The Tajik–Karakum Block of the Turan Plate could move in the opposite direction because of the right-lateral displacement along the Main Kopet Dagh Fault (>2 mm/year in the Quaternary). The countermovement of flanks shortened the Pamirs in the W–E direction and stretched it in the N–S direction, so that the North Pamir Zone was thrust over the Afghan-Tajik Basin. The convergence of the Pamirs and Tien Shan was caused by this process, and the westward removal of sedimentary sequences from the area of maximal shortening has been occurring until now, as follows from geodetic and geologic evidence (Guseva et al., 1993; Trifonov et al., 2002).

Intensive Pliocene–Quaternary vertical movements, the magnitude of which during only the Quaternary exceeded 6 km, were the most important process at the late stage of neotectonic period. Uplifting was driven by ongoing stacking of crustal blocks and by decrease of density of the upper mantle and the lower crust. The deformation was most large-scale in the western Pamirs; therefore, its uplift rate in the Pliocene–Quaternary was higher than in the eastern Pamirs (Krestnikov et al., 1979). The Quaternary rise was accompanied by gravity-driven overthrusting at the northern, western and eastern flanks of the Pamirs and by extension in its axial zone (Lake Karakul Depression).

The decrease of density of the upper mantle in the Pamir–Kindu Kush–Karakorum region was justified by the seismological data on lowered seismic wave velocities to 0.1–0.2 km/s relative to their worldwide background (Vinnik, Lukk, 1974; Lukk, Vinnik, 1975; Vostrikov, 1994) and by analysis of the gravimetrical data (Artemjev, Kaban, 1994). The decrease of density could be a result of the partial replacement of the lithospheric mantle by the asthenosphere matter. In the process of collision deformation, big volumes of the former oceanic crust within the lithosphere were pressed into the mantle to depths of 40–70 km, where they underwent the high-pressure metamorphism, being partly transformed into garnet granulites and eclogites (see section 2.2.3.1). A part of them that had the lesser density than the surrounding mantle and did not subside because of this, could undergo the retrograde metamorphism with participation of the cooled asthenosphere fluids during uplift at the Pliocene–Quaternary stage of neotectonic deformation. This decreased the density of the lower crust rocks and produced additional uplift in the region.

1.4. Neotectonic evolution of the Greater Caucasus

To understand sources of uplift of the Greater Caucasus (GC), we estimated consequently the following characteristics: (1) thickness and composition of the Earth's crust before its deformation by compression; (2) values of transverse shortening, thickening and uplift of the crust because of the compression; (3) transformation of the deformed crust into the recent mountain system. The used data on the Mesozoic-Cenozoic geology of the GC are based on the publications (Milanovsky, Khain, 1963; Panov, 1988; Shcherba, 1993; Alpine history…, 2007; Marinin, Rastsvetaev, 2008). The main part of the GC formed in the margin of the post-Paleozoic Scythian Plate. Its part, weakly deformed in Mesozoic and Cenozoic is separated from the GC by the foredeeps, Azov-Kuban in the west and Terek-Derbent in the east (Fig. 15).

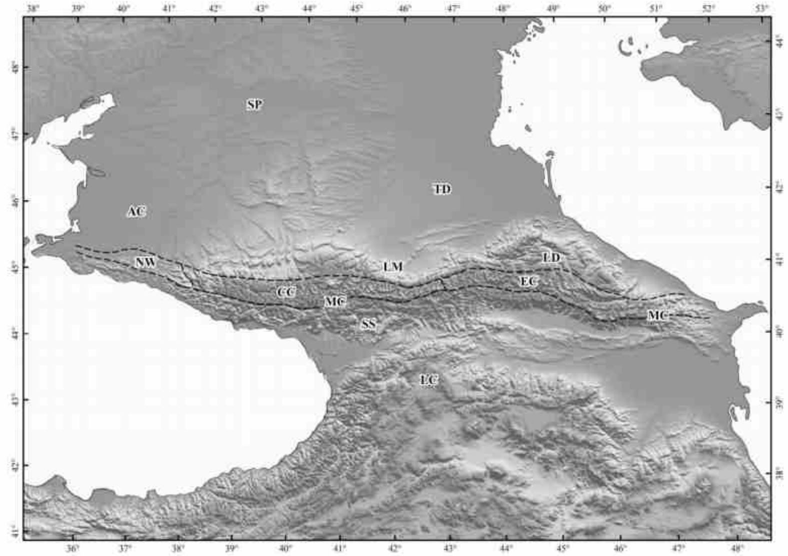


Fig. 15. The Greater Caucasus. AC, Azov-Kuban Basin; CC, Central Caucasus; EC, Eastern Caucasus; LC, Lesser Caucasus; LD, Limestone Daghestan; LM, Laba-Malka Zone; MC, Main Caucasus Fault; NW, North-western Caucasus; SP, Scythian Platform; SS, Southern Slope Zone; TD, Terek-Derbent Basin

The transitional Laba-Malka Zone (LMZ), including the East Balkar subzone and Limestone Daghestan, extends along the northern slope of the GC (Alpine history…, 2007). The thickness of the LMZ sedimentary cover varies from 5–5.5 km in the central part to 6.5–7 km in the east and ~10 km in the north-west. The GC itself consists of the northwestern, central and eastern segments. The North-Western