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Pliocene-Quaternary orogeny in the Central Tien Shan

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Abstract

Formation of mountains of the Central Tien Shan is usually explained by the isostatic response of the crust to its shortening caused by the India-Eurasia collision. The rise of the region in the period from Oligocene to Late Pliocene (2 myr ago) reached ~700 m on average, which corresponds to the isostatic response. For the last 2 myr (Late Pliocene–Quaternary), the rate of rise increased by an order of magnitude. This is proved by the coarsening of Cenozoic molasse up the sections, acceleration of cutting of drainage systems into ridges, and formation of new ridges within basins. In the Quaternary, most of intermontane basins underwent uplifting, though not so intense. The average rate of lateral crustal shortening increased ~2–2.5 times only, and the contribution of this process to the Late Pliocene-Quaternary orogeny was no more than 10%. The acceleration of rise was caused mainly by the convective replacement of the mantle lithosphere by the less dense asthenosphere. This was due to the quick softening of the mantle lithosphere as a result of the infiltration of active fluid from the lower mantle. Such accelerations of crustal uplifting took place in the Pliocene and Quaternary in many continental regions. This evidences that mantle processes, first of all, the full or partial replacement of the mantle lithosphere by the asthenosphere, played an important role in the formation of mountains.

Keywords: Orogeny (formation of mountains); asthenosphere rise; acceleration of rise; Pliocene; Quaternary; Tien Shan

Introduction

Long-term research gave an idea of the modern Tien Shan structure as a result of the 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 molasse-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 (Chediya, 1986; Dmitrieva and Nesmeyanov, 1982; Makarov, 1977; Mikolaichuk, 2000; Nikolaev, 1988; Shul'ts, 1948). 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 geodesic observations (Abdrakhmatov et al., 1996; Laverov, 2005; Nikonov, 1977; Zubovich et al., 2001). 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 grew with acceleration (Chediya, 1986; Krestnikov et al., 1979).

This paper 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), part of the mountains between the Talas-Fergana fault in the west and Khan-Tengri mountain plexus in the east (Fig. 1). In the north, the CTS borders the epi-Paleozoic Chu and Ili basins, and in the south, the Tarim basin, which originated on the Late Proterozoic basement probably as early as the Paleozoic.

Changes in the regime of vertical movements during the neotectonic evolution

The CTS is an epi-Paleozoic mountain system that passed

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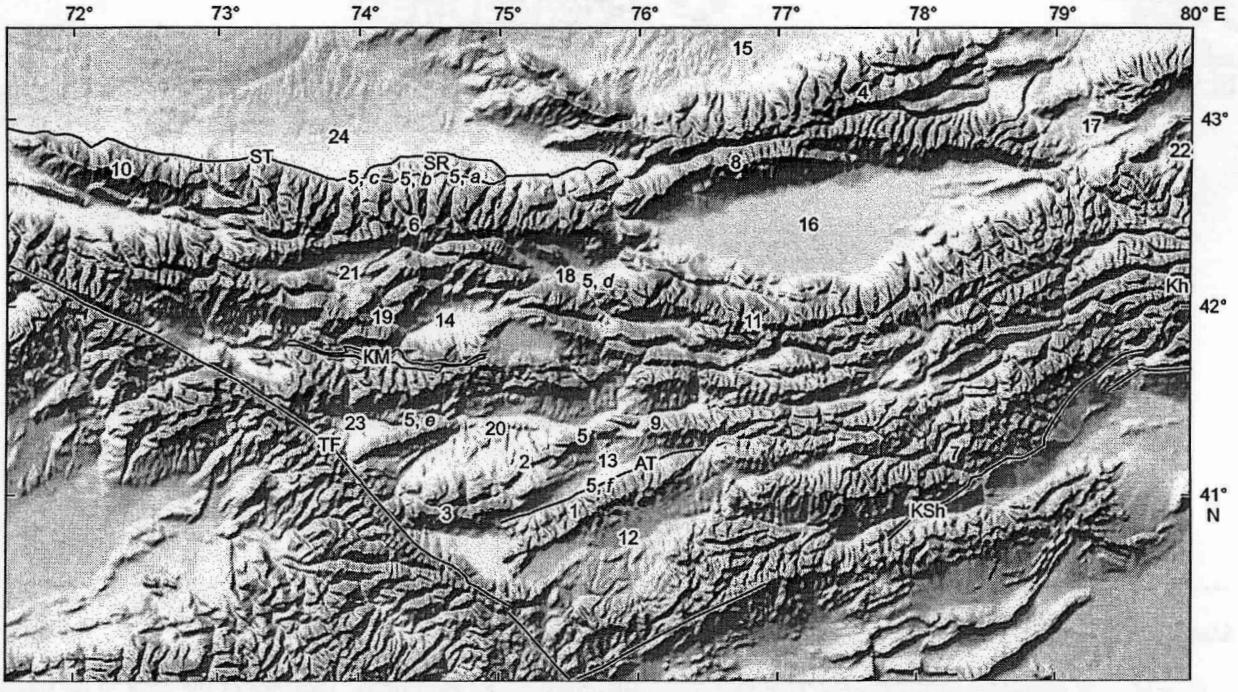

Fig. 1. Map of the Central Tien Shan relief. Ridges: 1 — Atbashi, 2 — Baibichetau, 3 — Dzhamantau, 4 — Trans-Ili Alatau, 5 — Karatau, 6 — Kyrgyz, 7 — Kokshaaltau, 8 — Kungei Alatau, 9 — Naryntau, 10 — Talas, 11 — Terskei Alatau. Foredeeps and intermontane basins: 12 — Aksai, 13 — Atbashi, 14 — Dzhumgol, 15 — Ili, 16 — Issyk-Kul', 17 — Karkara, 18 — Kochkor, 19 — Kyzyloi, 20 — Naryn, 21 — Susamyr, 22 — Tekes, 23 — Toguz-Torou, 24 — Chu. Largest fault zones: AT — Atbashi, KSh — Kokshaal, NT — North Tien Shan, TF — Talas-Fergana. SR — Serafimovka anticline. KM — Kökömeren-Minkush zone. Kh — Khan-Tengri mountain plexus. Geologic profiles drawn in Fig. 7.

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 pre-orogenic planed surface (Chediya, 1986; Shul'ts, 1948; Trofimov, 1973). Redeposited crust of weathering that formed on the peneplain by the late Mesozoic composes a continental red-colored, mainly finely clastic rock unit with Middle-Late Eocene and, probably, Early Oligocene fauna (Dmitriev and Nesmeyanov, 1982). Its lower part includes basalt lavas, whose total thickness reaches 20 m on the northwestern edge of the Issyk-Kul' basin and 80 m in the Aksai basin (Bazhenov and Mikolaichuk, 2002; Chediya et al., 1973). The K-Ar and Ar-Ar ages of the basalts are 54–70 Ma (Krylov, 1960; Nesmeyanov et al., 1977; Simonov et al., 2002). The unit thickness usually does not exceed few tens of meters, though locally reaches 300–500 m in the Chu, Ili, Issyk-Kul', and Aksai basins, thus probably reflecting their started downwarping.

Figure 2 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 pre-orogenic surface under conditions of intense linear erosion that accompanied their growth (Chediya, 1986; Makarov, 1977). Thus, besides rare cases of significant erosional decline of these relics, the map reflects recent deformations and shifts of the pre-orogenic surface. This is a generalized map of all recent 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.

The step-like structure of the ridge slopes, which is accounted for by most researchers by pulse rise, formed the basis for concepts of the CTS relief layering (Chediya, 1986; Krestnikov et al., 1979; Makarov, 1977; Shul'ts, 1948; Trofimov, 1973). According to these concepts, intensification of vertical movements promotes erosion, which results in an erosion-tectonic scarp (incision) resting upon the bottom of basin or valley serving as a local erosion basis 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 relief layer. This suggests that the incision is correlated with the lower coarse part of a particular molasse complex and that the steps in the incision basement are correlated with the upper finer part of the complex (Makarov, 1977). Comparison of the relief layer with a particular molasse complex helps to clarify the evolution of the mountain system.

Deposits of CTS basins were described elsewhere (Chediya, 1986; Dmitrieva and Nesmeyanov, 1982; Nesmeyanov and Makarov, 1974; Shul'ts, 1948). Our goal was to generalize these data and supplement them with results of our research (Dodonov, 2002; Mikolaichuk et al., 2003). The above-Eocene molasse is subdivided into four complexes: Kyrgyz, Tien Shan, Sharpyldak, and Pleistocene-Holocene, each being di-

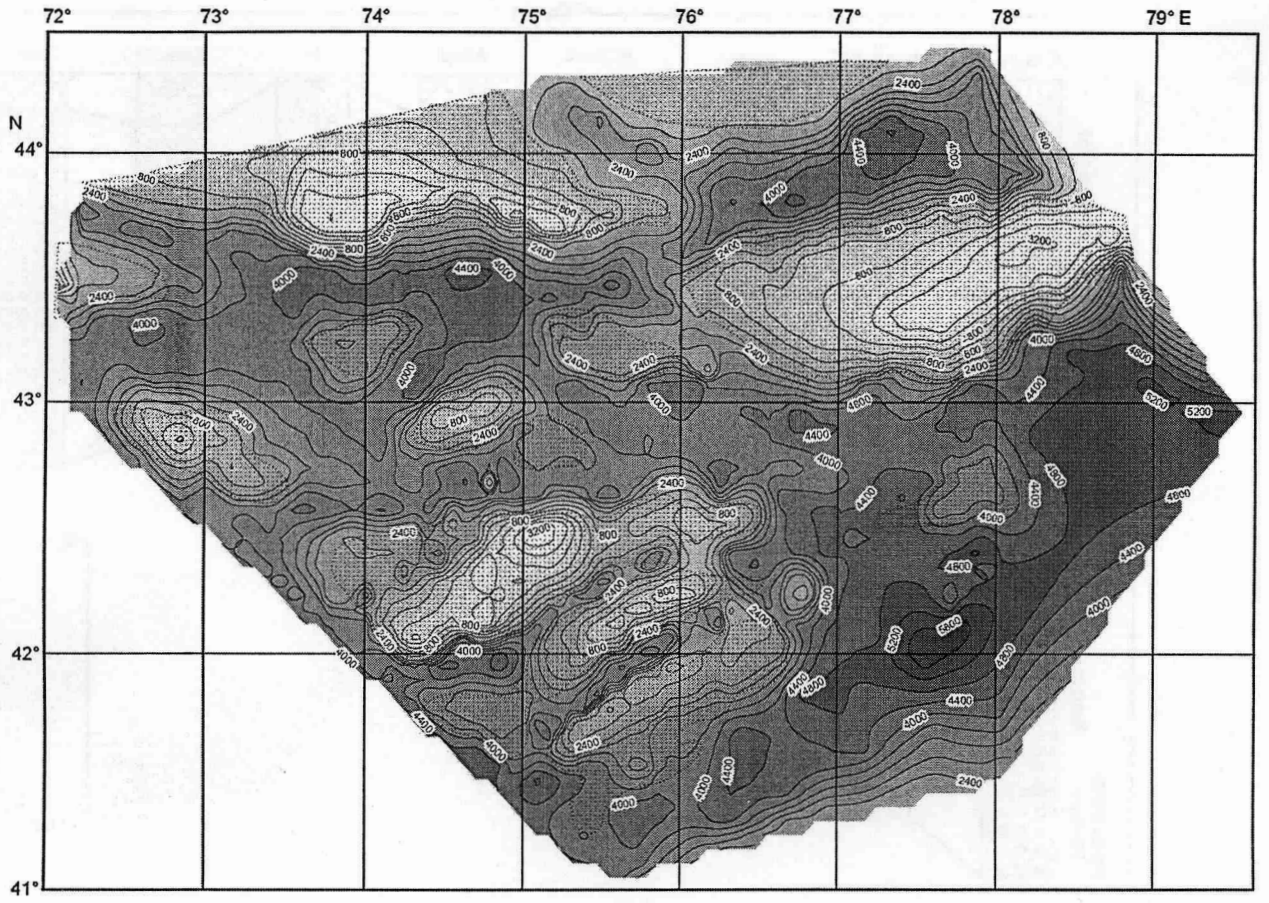

Fig. 2. Assumed position of pre-orogenic planed surface drawn by peak surface isolines within the ridges and by basement surface isolines within basins (speckled sites). The position of the surface beneath recent molasse was reconstructed from geological, drilling, and geophysical data (Geological Map..., 1982).

vided into local units (Figs. 3–5). Their comparison and dating (still debatable) are based on the composition and color of deposits, incomplete paleomagnetic data, and scarce fauna remains.

The lower Kyrgyz complex consists of two series — lower red and upper brown. The lower series is composed of proluvial and alluvial fine- and, seldom, medium-pebble conglomerates, gravelstones, and sandstones, which at the center of large basins give way to sand-clayey deposits with gypsum. Based on the fauna found in the Issyk-Kul' and Ili basins, the lower beds of the series were dated to the Middle Oligocene, and the upper ones, to the Late Oligocene-Early Miocene (Dmitrieva and 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) intercalates, 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 Middle and Late Pliocene (Dmitrieva and 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 beds of the Pliocene.

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

According to fauna remains and paleomagnetic data for the Issyk-Kul', Ili, and Chu basins, the Sharpyldak complex is of Eopleistocene and, possibly, late Pliocene age (~2–0.8 Ma) (Aleshinskaya et al., 1972; Dmitrieva and Nesmeyanov, 1982; Dodonov, 2002; Trofimov et al., 1976). It is composed of proluvial and alluvial gray coarse conglomerates and conglomerate-breccias (up to boulder ones) with gravel, sand, and silt intercalates.

The Pleistocene-Holocene complex, compositionally similar to the Sharpyldak one, is formed by fluvial sediments of seven cycle terraces (Shul'ts, 1948), floodplains, and riverbeds as well as glacial and basin (at the center of the Chu, Ili, and Issyk-Kul' basins) deposits. Three series of the complex are dated to: (1) Early-Middle Pleistocene, (2) Late Pleistocene, and (3) late Pleistocene-Holocene (Chediya, 1986; Dodonov, 2002).

First uplifts serving as sourcelands of debris appeared in

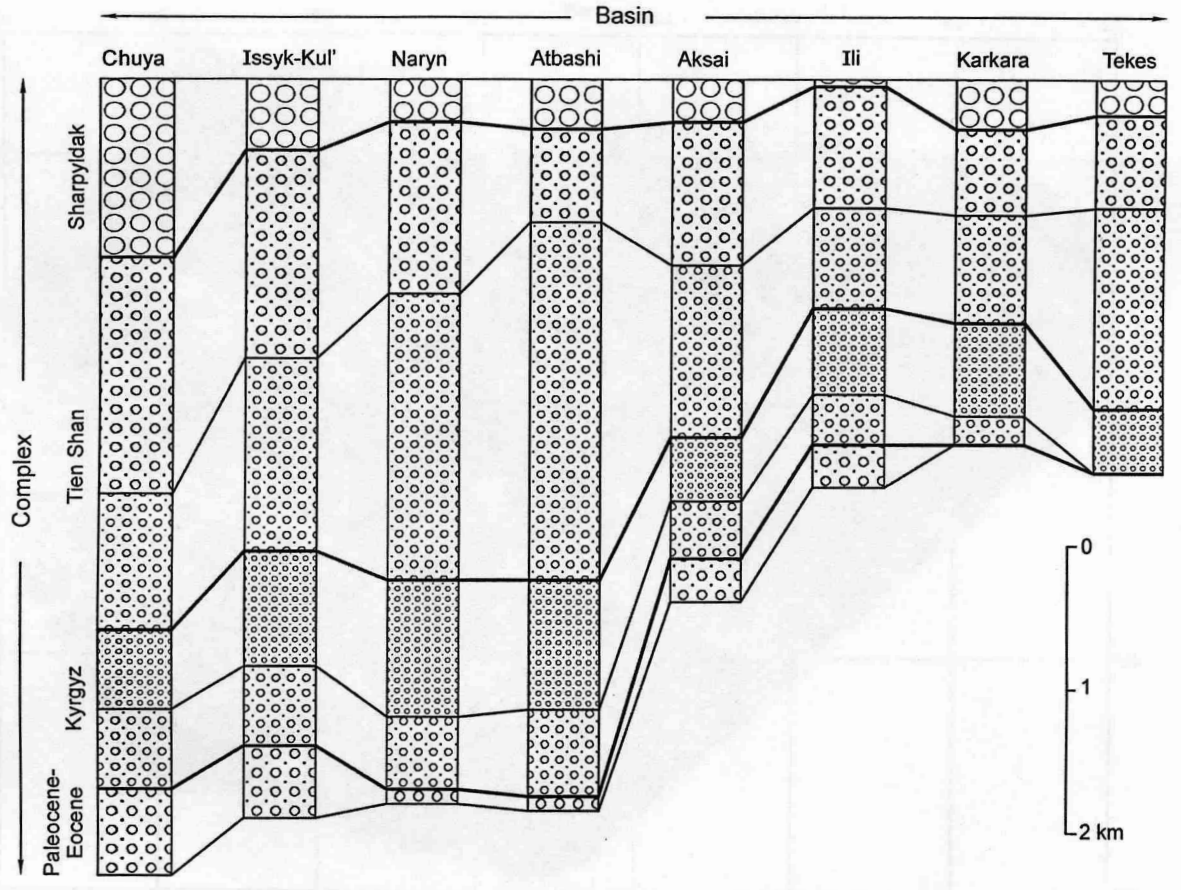

Fig. 3. Comparison of sections of Quaternary deposits in CTS basins. Series of deposits of different ages are marked by different symbols.

the Oligocene, 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, probably no more than 1 km in most of CTS areas. Judging from the deposit composition, the relief contrast was reduced during the accumulation of the upper series of the Kyrgyz complex and the lower series of the Tien Shan complex. 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 evaporites became scarcer. The coarsening of clastic material began in the southeast of the region during the accumulation of the lower series of this complex and penetrated later to the northwest. 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 latest Pliocene. It evidences a considerable increase in the above-mentioned amplitude, i.e., the formation of mountain relief.

The change in the rate of vertical movements in the basins can be judged from the average rates of accumulation of various series of recent molasse in them (Figs. 5 and 6), though these estimates are only tentative because of the complex structure of the molasse and the incomplete sections of the basins. Calculations showed that at the Paleocene-Eocene platform stage, the average rate of accumulation did not exceed thousandths of a millimeter a year and reached ~0.02 mm/year only at some sites of the future basins. The accumulation rate increased to hundredths of a millimeter a year during the formation of the lower series of the Kyrgyz complex and stayed the same during the deposition of its upper series. The Tien Shan complex accumulated with acceleration of up to 0.1–0.6 mm/year.

In 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 uplifting acceleration. In the Pleistocene, this led to the formation of terraces in 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 deposit accumulation increased twice only at some sites of the Chu basin. A single estimate for the Pleistocene sedimentation was also obtained for the Chu basin, where up to 500 m sediments accumulated for that epoch (Abdrakhmatov, 1988): Its rate is close to the rate of the Sharpyldak sedimentation. But the thickness of 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 late Pliocene and Quaternary. The

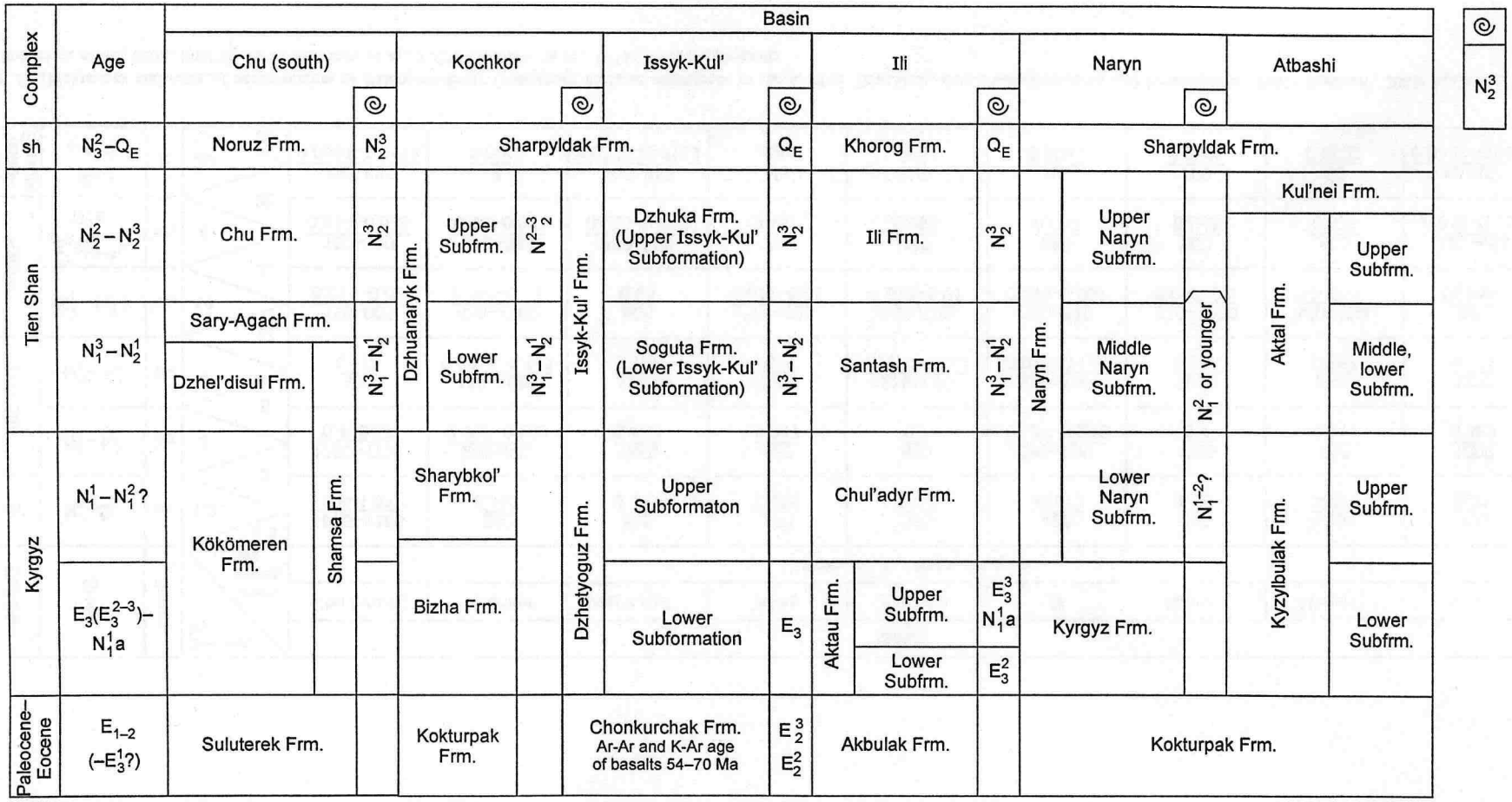

Fig. 4. Comparison and age of formations of Paleocene-Eopleistocene deposits in intermontane and piedmont basins of the Central Tien Shan, data from (Abdrakhmatov et al., 2001; Aleshkinskaya et al., 1972; Bullen et al., 2001; Chediya et al., 1973; Dmitrieva and Nesmeyanov, 1982; Dodonov, 2002; Makarov, 1977; Mikolaichuk et al., 2003; Nesmeyanov and Makarov, 1974; Nesmeyanov et al., 1977; Shul'ts, 1948; Simonov et al., 2005; Trofimov, 1973; Trofimov et al., 1976). *1* — age of fauna.

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	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	2	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	5	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	12	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	22	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
Paleo-cene–Eocene	E_{1-2} (–E_3^1?)	E	36	30 / 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

Fig. 5. Thicknesses and rates of accumulation of Paleogene-Early Quaternary molasse complexes in the Central Tien Shan, data from (Dmitrieva and Nesmeyanov, 1982; Dodonov, 2002; Makarov, 1977; Mikolaichuk et al., 2003; Shul'ts, 1948; Simonov et al., 2005; Trofimov et al., 1976), with supplements.

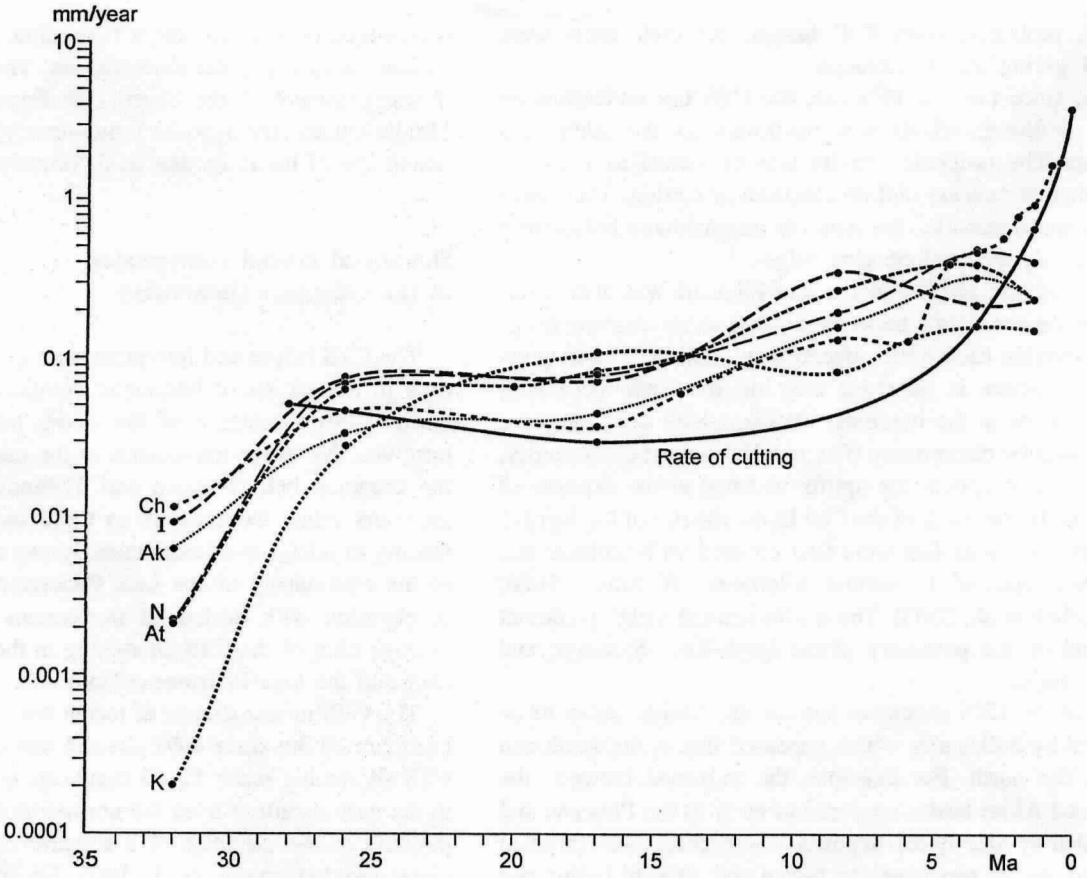

Fig. 6. Changes in the rates of sedimentation in recent basins and in the rates of cutting of drainage systems into the CTS ridges during the neotectonic stage. Deceleration of deposit accumulation in the Kochkor and Naryn basins in the Eopleistocene was related to the transition from basin sedimentation to terrace regime. Basins: Ch — Chu, K — Kochkor, I — Issyk-Kul', N — Naryn, At — Atbashi, Ak — Aksai.

same is evidenced from data of analysis of the relief layering. Three stages were recognized in the CTS ridges (Chediya, 1986; Krestnikov et al., 1979; Makarov, 1977; Shul'ts, 1948; Trofimov, 1973). The upper stage, formed by one or two steps cut into the pre-Orogenic surface, is correlated with the Kyrgyz complex (E_3–N_1^2); the middle one, formed by two steps, with the Tien Shan and Sharpyldak complexes (N_1^2–Q_E); and the Lower stage is correlated with the Pleistocene-Holocene basin deposits of the Chu and Ili basins. The highest of seven cycle terraces of the lower stage has been incised into the roof of the Sharpyldak complex and is dated to the late Early Pleistocene (Q_1^2). Makarov (1977) established that the amplitudes of Oligocene cuttings into the CTS ridges were no more than 200–400 m. According to Chediya (1986), the amplitude of the relief (depth of incisions) of the upper stage reaches 700 m; that of the middle stage, 1500 m (more than half of this depth falls onto the lower incisions formed in Sharpyldak time); and the relief amplitude in the lower stage reaches 1500 m. Similar values were reported by Krestnikov et al. (1979): ~1000 m for Sharpyldak incisions in the CTS ridges and >1000 m in the north and up to 1500 m in the southeast of the CTS for Quaternary cuttings (from Q_1^2).

Taking into account the duration of the epochs of deposition of correlated molasse complexes (Fig. 5) and using the above estimates, we can tentatively calculate (ignoring the irregular cutting during each cycle) the average rates of incision for different relief stages and substages of the CTS antliclinal ridges, which, to a first approximation, reflect the rates of their rise. The rates of incision were 0.03–0.05 mm/year during the accumulation of the Kyrgyz Formation 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 deposition of the Tien Shan complex (N_1^2–N_2^3), 0.6–0.7 mm/year in Sharpyldak time and the earliest Pleistocene (N_2^3–Q_1^1), and 1.6–2.5 mm/year in the Pleistocene. Thus, in the late Pliocene, the rate of cutting increased ~10 times, and it remained such in the Pleistocene (Fig. 6).

The CTS intermontane basins, which evolved for a long time in the regime of downwarping and basin sedimentation, also began to be involved in the rise in Sharpyldak time. In the early Pleistocene, linear cutting covered all intermontane basins (probably, except for the center of the Issyk-Kul' basin), thus forming the lower stage of their relief. The average rates of cutting in the basins were 1.5–2 times lower than those in neighboring ridges; in the south and southeast of the region they were higher than the rates in the north (Krestnikov et al., 1979). In the Pleistocene, zones of downwarping 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 late Pliocene, the CTS has undergone an intense uplifting, which was maximum in the south and southeast. The orogenic activity was expressed as a drastic coarsening of molasse and acceleration of cutting. The uplifting was not dome-like, because the intermontane basins rose less intensely than neighboring ridges.

The orogenic activity in the late Pliocene was also manifested in the correlation between molasse series. Ancient series usually overlap each other concordantly, and their "transgressive" occurrence is observed only on the basin periphery, whereas series in the basement of Sharpyldak deposits occur with an angular discordance (few to >10°) almost ubiquitously.

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 molasse and then were uplifted to several kilometers (Chediya, 1986; Mikolaichuk et al., 2003). These sites formed a high piedmont there and on the periphery of the Issyk-Kul', Sysamyr, and Atbashi basins.

Within the CTS mountain system, the basins began to be separated by bulkheads, which appeared first in the south and then in the north. For example, the bulkhead between the Tuyun and Aksai basins appeared as early as the Pliocene and was sealed by Sharpyldak deposits. The Baibichetau–Naryntau bulkhead, which separated the Naryn and Atbashi basins and was uplifted above their bottoms to ≥2 km, was formed by the left echelon series of ridges combining transverse shortening (with extension on the Baibichetau Ridge dome) and longitudinal sinistral fault (Makarov, 1977). The late age of the bulkhead is proved by the presence of fine-clastic deposits of the Naryn Formation between the Naryntau and Karatau Ridges. These deposits, as well as those of the Tien Shan complex, lack traces of clastics from the above uplifts. Sharpyldak deposits are absent from the slopes of the bulkhead and lack its clastic material. Thus, the Baibichetau–Naryntau bulkhead finally formed as an orographic element as late as the Pleistocene. The same correlation is observed near the up to 2 km high bulkhead between the Dzhumgol and Kyzyloi basins composed of alike deposits of the Kyrgyz and Tien Shan complexes.

One more example of a Pliocene-Quaternary formation is the Kökömeren-Minkush zone (Sadybakasov, 1972), stretching approximately along the Nikolaev line west of Lake Song-Kel'. The zone forms an intricately crumpled ramp, narrowed by a transverse shortening, with signs of a longitudinal sinistral fault. The Late Quaternary fault is expressed as relief shifts and inherits earlier formed shifts manifested as sliding grooves on the fault planes. Within this zone, Lower Carboniferous red-colored deposits, Jurassic sediments, and the thin Kokturpak Formation with basalts in the basement (68.4±2.3 Ma, data of the Institute of Geology of Ore Deposits, Petrography, Mineralogy and Geochemistry, Moscow) underlie pebble conglomerates of the lower series of the Kyrgyz complex, which are overlapped by deposits of the Naryn and Sharpyldak Formations. All strata, from Lower Carboniferous red-colored deposits to Naryn Formation, change one another without serious angular discordances. The latter appear only in the basement of the Sharpyldak Formation (locally) and Middle Quaternary deposits (ubiquitously). This points to the young age of the ramp and its deformation.

Horizontal crustal compression in the Oligocene-Quaternary

The CTS ridges and intermontane basins evolved in recent time in the setting of horizontal compression caused by the northwestward pressure of the Tarim microplate, which, in turn, was due to the movements of the more southern parts of the orogenic belt (Ivanova and Trifonov, 2005). Many researchers relate the orogeny to these horizontal movements leading to piling-up of mountains. Let us check the possibility of the relationship of the Late Pliocene-Quaternary orogeny acceleration with horizontal movements by comparing the average rates of the CTS shortening in the Oligocene-Quaternary and the Late Pleistocene-Holocene.

The GPS measurements of recent horizontal shifts that had been carried out since 1992 showed that the total rates of the CTS shortening reach 12–13 mm/year, with the shift vectors in the east departing from the normal to the ridge strike, thus pointing to the presence of a sinistral component of movements (Abdrakhmatov et al., 1996; Laverov, 2005; Zubovich et al., 2001). The rates of recent vertical movements in places reach the same values (Nikonov, 1977).

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

We carried out analysis of the set of known active faults in the CTS (Trifonov et al., 2002) and revealed that many longitudinal faults have not only a thrust but also a substantial sinistral component of shifts. Faults of NW strike with dextral shifts were recognized. Based on the analytical results, tensors of deformation rates were calculated. The total rates of Late Quaternary horizontal shortening were estimated at 4–6 mm/year. In the calculation we ignored folded bends and shifts along fractures, whose contribution to the total deformation is 10–20% in other active areas of this kind (Anatolia and Middle East). Therefore, the rates of the CTS horizontal shortening should be increased to 5–7 mm/year, with the transverse-shortening component not exceeding ~5 mm/year.

Calculation of the total neotectonic deformation of transverse shortening is based on the measurement of folded bends and shifts along the faults of the pre-orogenic planed surface and molasse complexes (Chediya and Utkina, 1975; Yunga

and Yakovlev, 2000). Chediya (1986) reported that reverse faults became steep with depth and ignored their gentle dip near the land surface in the calculations. He estimated the total transverse shortening of the CTS at 4–5% of its width, i.e., 14–18 km at the longitude of the Naryn basin. At the longitude of the Khan-Tengri mountain plexus, where the CTS is the most uplifted, the shortening is smaller. Yunga and Yakovlev (2000) made a similar calculation, not introducing corrections to observed folded bends and shifts along faults. They estimated that the total shortening varies from 9–12% at the longitudes of the Naryn basin and Bishkek to 5–6% at the longitude of the Khan-Tengri plexus, i.e., from 40 to 20 km.

The representativeness of the obtained data depends on the model of the CTS neotectonics. The calculations were based on the conventional model implying that the CST 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 folded bends to movements along large thrusts flattened at depth (Abdrakhmatov et al., 2001). The model admits a 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.

Structural method. Following the new model, thrust zones must exist throughout the mountain system irrespectively of changes of folded forms. According to the conventional model, thrusts seldom run beyond the folds they complicate.

Geomorphological method. Following the new model, the pre-orogenic surface near thrusts is eroded, and if it were preserved, it would be localized abnormally highly. According to the conventional model, the surface reaches its maximum height in the axial part of anticlinal ridges and lowers to their near-thrust edges.

Geological method. 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 altered rocks exposed during shallow erosion.

Study of marginal thrusts at the boundary of uplifts and basins showed that in most of the area including the North and Central Tien Shan megazones, recent thrusts and reverse faults usually do not run beyond ridges and vary seriously along their strike. Locally on the northern flank of the Tien Shan (Fig. 7, *a–c*) and within it (Fig. 7, *d*), thrusts are changed (along their strike) by overturned folds of the basement or are complicated by similar folds of the lower nappe horizons, with molasse cover being present in both fold limbs (Fig. 7, *e*). In all cases, the thrust or thrust deformation amplitude is small. The pre-orogenic ridge surface outlines an anticlinal bend, and the near-thrust marginal parts of the ridges often bear exposed Devonian-Permian rocks not subjected to metamorhism. All this agrees with the conventional model.

A different pattern was observed in the zone of the recent Atbashi fault at the boundary of the Atbashi basin and the Atbashi ridge (the latter lies in the South Tien Shan). Where the Sarybulak Brook flows into the Karakoyun River (western part of the structure), the fault zone is separated into two main branches. The northern branch runs along the Karakoyun River and is overlapped by Late Quaternary alluvium. According to geophysical data, in the south it borders the part of the basin where the basement is subsided to 3–4 km (Geological Map..., 1982). South of the northern branch, thin Paleogene-Lower Miocene deposits locally expose. Near the Sarybulak Brook, Permian argillaceous and silty slates and sandstones with cleavage cracks (dipped at 70° to the south) were stripped beneath glacial and proluvial-alluvial deposits of the terrace nappe. The southern branch of the fault dips at 60–70° to the south and forms a scarp separating the terrace from the highly uplifted ridge slope composed of metaterrigenous quartz-sericite and, in the south, quartz-mica schists tentatively dated to the Riphean (Fig. 7, *f*). Schistosity dips southward; near the fault it dips at 70°, and south of the fault it is characterized by a steep dip in the upper part of the slope, which decreases to 60° near the brook bottom and then to 40°. 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 section is related to the Hercynian nappe formation (Burtman, 2006). The near-fault ridge slope is strongly eroded and lacks relics of pre-orogenic surface.

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 limb might have been earlier localized 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 amplitude of the recent thrust is estimated at 8–10 km. The thrust is listric; the increase in the uplift of its southern limb toward the Khan-Tengri mountain plexus points to an increase in its amplitude and, correspondingly, shortening.

Large recent south-vergent thrusts were revealed in China along the southern flank of the Tien Shan (Deng Qidong, 2000; Laverov, 2005). In the southern piedmont at the longitudes of Issyk-Kul' and Khan-Tengri, the total amplitude of shift along 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). With regard to the thrusts in the South Tien Shan, the total recent transverse shortening in the CTS was estimated at 50–70 km; with the recent-stage duration of ~30 Ma, this yields the average rate of shortening of ~2 mm/year.

The obtained estimate ignores the strike slip along faults. The sinistral shift along the CTS is inferred from the echelon-like recent structures and their arrangement (Makarov, 1990; Laverov, 2005). We discovered recent longitudinal sinistral shifts in the Kökömeren-Minkush zone and on the northern slope of the Baibichetau uplift. Dextral shifts and shift zones of NW strike were also revealed earlier (Bogachkin et al., 1997; Trifonov et al., 2002). Lacking detailed data on the amplitudes of shifts along recent faults, we admit that their contribution to the total Cenozoic deformation is proportional to the contribution of active strike-slip faults to the total Late Quaternary deformation. With regard to strike slip, the average

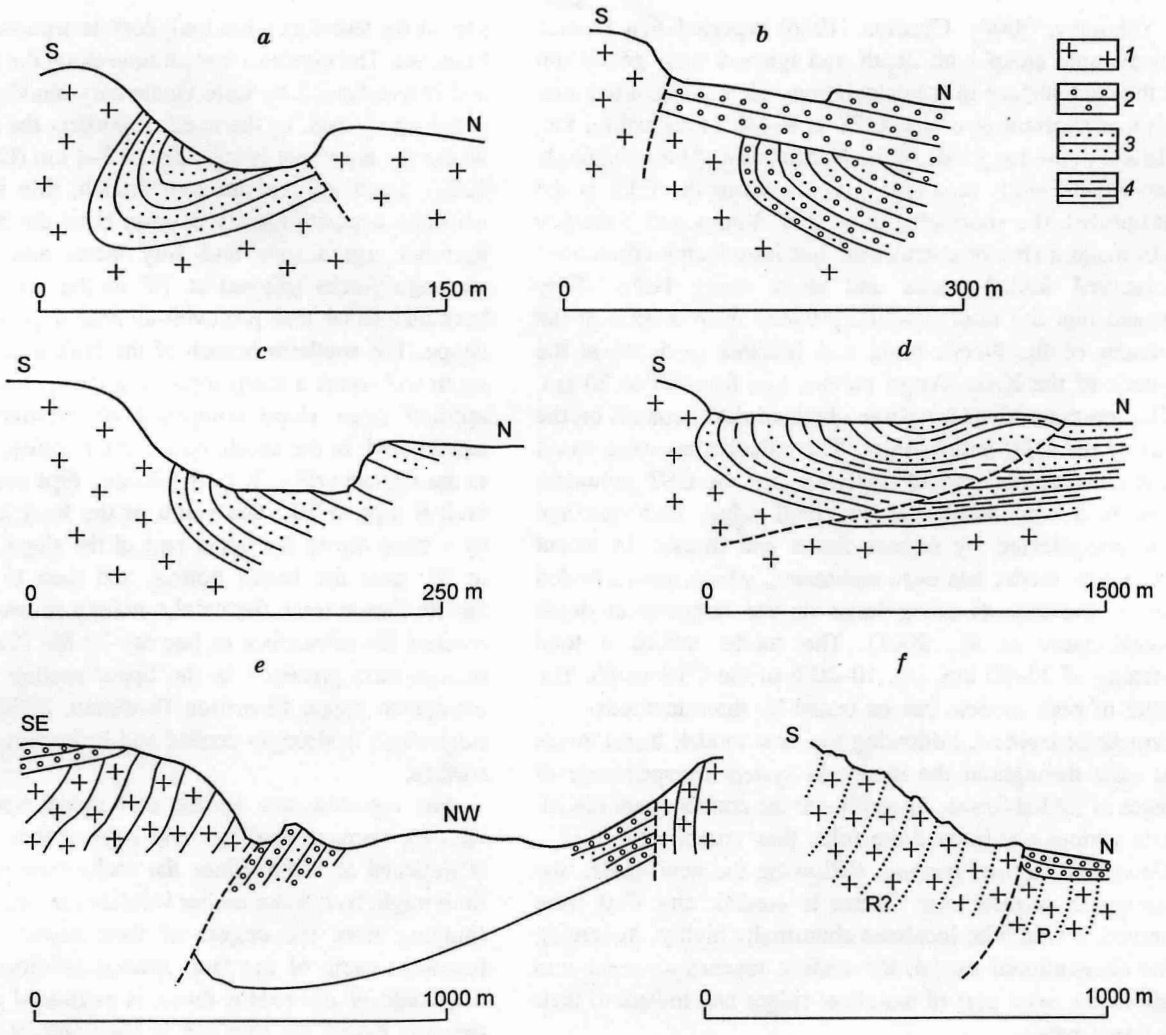

Fig. 7. Geologic profiles across the boundaries of basins and uplifts. *a–c* — Northern flank of the Tien Shan: *a* — Chonkurchak, *b* — Dzhalamysh sai (gully), *c* — Aksu River; *d* — southern edge of the Kochkor basin along the Dzhoanaryk River; *e* — southern edge of the Toguz-Torou basin (Shul'ts, 1948); *f* — southern edge of the Atbashi basin along the Sarybulak Brook. Horizontal and vertical scales are equal. *1* — Paleozoic basement, *2* — Cenozoic coarse-clastic deposits, *3* — Cenozoic sandstones, *4* — Cenozoic fine-clastic deposits.

rate of the recent horizontal CTS shortening might be higher, up to 2.5–3 mm/year, which is 2–2.5 times lower than the rates of the Late Pleistocene-Holocene shortening. Thus, the acceleration of horizontal movements by the late Quaternary was much lower than the acceleration of orogeny.

Let us consider the physical mechanisms responsible for the crustal uplifting in the study area.

Contribution of compression to the crustal thickening and uplifting

Based on the above data, we estimated the role of crustal compression in the recent CTS uplifting. Till the beginning of recent time, the CTS paleorelief was at heights of ~300 m. From Oligocene to Late Pliocene (beginning of the formation of the Sharpyldak complex), the average height of uplifts did not exceed 1.5 km, and the difference in the heights of uplifts and basin surface was no more than 1 km. Based on these data and the area of uplifts and basins, we accepted the average height of the CTS reached by the Late Pliocene as ~1 km. At present, the average height is ~3 km. Thus, from Oligocene to Late Pliocene (i.e., for ~28 Myr), the CTS was uplifted by on average ~700 m, and for the last 2 Myr, by 2 km. In the southeast and northeast of the area, the height of the Late Pliocene-Quaternary uplift is no less than 3 km.

Let us denote the initial and final values of the width of compressed area and crustal thickness in it as L_0, L_1 and h_0, h_1, respectively. Then, an increase in crustal thickness due to compression, Δh_{comp}, and crustal uplifting $\Delta\zeta_{comp}$ 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 ρ_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 CTS shortening starting from the Oligocene is $L_0 - L_1 \approx 50$–70 km, i.e., the initial width of the

area was $L_0 \approx 450$–470 km. The average rate of Late Quaternary compression was ~5 mm/year. Taking this value for the last ~2 Myr, we obtain the CTS shortening $L'_1 - L_1 \approx 10$ km and the area width ~2 Myr ago equal to $L'_1 \approx 410$ km.

From Jurassic to Eocene, the CTS, along with Kazakhstan adjacent in the north, was a young platform area. In the south of Kazakhstan, the crustal thickness is ~42 km (Fig. 8). Let us accept that the same crustal thickness $h_0 = h_{pl} = 42$ km was also in the CTS in the Eocene. The average crustal density on platforms is $\rho_c = 2830$ kg/m^3 (Artyushkov, 1993; Christensen and Mooney, 1995). With these values of ρ_c and h_0, from (1) and (2) we obtain that 2 Myr ago, when L_1 was equal to 410 km, Δh_{comp1} was 4–6 km and $\Delta\zeta_{comp1}$ was 0.6–0.9 km.

The latter value is close to the above-given estimate of the CTS uplifting that occurred by ~2 Myr ago (~0.7 km). Therefore, it is most likely that the CTS uplifting that proceeded from Oligocene to Late Pliocene was mainly due to the crustal compression.

By the beginning of the accelerated CTS uplifting ~2 Myr ago, the 4–6 km thickening of crust must have increased its thickness to $h'_0 = 46$–48 km. Introducing this value as h_0 into (1) and assuming that $L_0 = L'_1 = 410$ km, from (1) and (2) we obtain the following values of crustal thickening and uplifting for the last ~2 Myr: $\Delta h_{comp2} = 1.2$ km and $\Delta\zeta_{comp2} = 180$ m. The latter value is an order of magnitude lower than the ~2 km uplifting that occurred for the last ~2 Myr. Even if accepting the average rate of compression for the last ~2 Myr equal to the modern rate of ~10 mm/year (GPS data from (Zubovich et al., 2001)), the uplifting due to crustal compression will be only ~360 m. This is ~18% of the actual uplifting, which can be explained by other mechanisms.

The total CTS uplifting due to crustal compression from early Oligocene to present time is $\Delta\zeta_{comp} = \Delta\zeta_{comp1} + \Delta\zeta_{comp2} \approx 1$ km.

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 in the CTS mountains are significantly lower (Laverov, 2005; Gubin, 1986; Vinnik et al., 2004, 2006; Yudakhin, 1983). This points to the ascent of the asthenosphere layer to the roof crust. According to gravimetrical data, deconsolidation of the mantle beneath the CTS reaches ~0.1 g/cm^3 (Artemjev and Kaban, 1994). The rise of the asthenosphere roof is nonuniform throughout the study area. Seismic data evidence that the Moho reaches the crustal base beneath high ridges and is separated from the crust by thick lenses of lithospheric mantle beneath intermontane basins.

The replacement of mantle lithosphere by less compact asthenosphere matter (density ρ_a) must be accompanied by crustal uplifting. The value of uplifting is proportional to the squared thickness of the layer of replaced mantle lithosphere (Artyushkov and Hofmann, 1998). For the CTS this value is unknown, though negative isostatic gravity anomalies of up to –150 mGal were detected (Artemjev and Kaban, 1994). In a wide area they would correspond to anomalous masses $\Delta m \approx -3.6 \cdot 10^6$ kg/m^2. The mantle deconsolidation leads to an isostatic crustal uplifting by

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

With $\Delta m \approx -3.6 \cdot 10^6$ kg/m^2, we obtain: $\Delta\zeta_a \approx 1.1$ km. Since the width of areas with the above anomaly intensity does not exceed 100 km, the anomalous masses in the mantle beneath them and the corresponding crustal uplifting must be greater. Assuming that $\Delta m \approx -7 \cdot 10^6$ kg/m^2, from (3) we obtain: $\Delta\zeta_a \approx 2$ km.

In the Pliocene-Quaternary, uplifts resulted from the rise of the asthenosphere roof that existed in many platform and folded areas (Artyushkov, 2003; Artyushkov and Hofmann, 1998; Briem, 1989; Map..., 1977; Milanovsky, 1974; Zorin et al., 1990). For example, uplifts with amplitudes of up to 1–2 km were in South and East Africa, on the Arabian Platform, and in Transbaikalia, Mongolia, and West Siberia. In most of these areas, the asthenosphere roof did not reach the crustal base and remained separated from it by a thick layer of cold mantle lithosphere. In the most uplifted parts of the CTS, the deconsolidated mantle occurs in places immediately beneath the crust. This suggests that the crustal uplifting due to the asthenosphere ascent might reach $\Delta\zeta_a \approx 1.5$–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; Artyushkov et al., 2000). The softening is caused by the infiltration of active fluids from the underlying mantle into the lithosphere, which leads to a drastic decrease in viscosity and strength of rocks as a result of the Rehbinder effect (Rehbinder and Venstrem, 1937; Sal'nikov and Traskin, 1987).

Great thickness and density of crust before its recent compression

As shown above, the compression of the CTS crust resulted in an increase in its thickness by

$$\Delta h_{comp} = \Delta h_{comp1} + \Delta h_{comp2} \approx 5\text{–}7 \text{ km}. \quad (4)$$

The present-day thickness of the CTS crust varies from 40–52 km beneath foredeeps and largest intermontane basins to 52–64 km beneath ridges (Gubin, 1986; Laverov, 2005; Vinnik et al., 2006) (Fig. 8). The crustal thickness beneath ridges is 10–22 km greater than that in southern Kazakhstan ($h_0 = h_{pl} = 42$ km), which we accepted as the pre-orogenic thickness of the CTS crust formed by the Oligocene, and two to three times greater than the average crustal thickening due to compression. If the modern crustal thickening have been related to compression only, the crustal thickness by recent

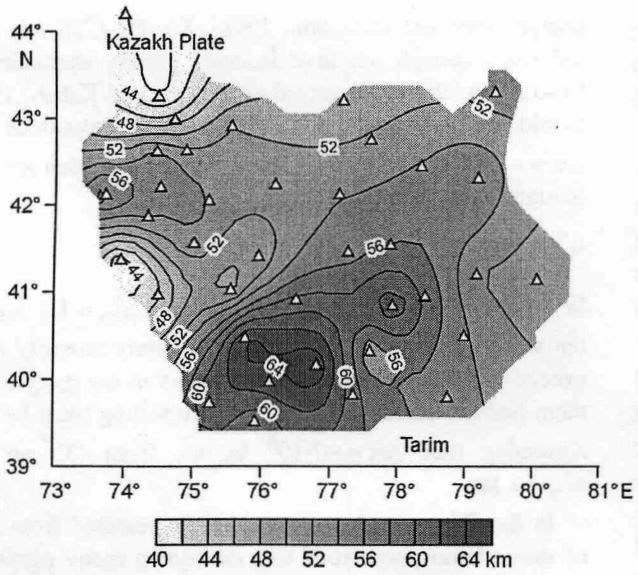

Fig. 8. Crustal thickness in the Central Tien Shan (Vinnik et al., 2006). Triangles mark seismic stations.

time would have been $h_0 = h - \Delta h_{comp} \sim 47–57$ km, i.e., higher than h_{pl} = 42 km by $\Delta h_0 \sim 5–15$ km. Introducing this value as Δh_{comp} into (2), we obtain that with the average crustal density $\rho_c = 2830$ kg/m^3, typical of stable platforms, the CTS surface would have been localized by the Oligocene ~0.8–2.3 km higher than the one on stable platforms. In fact, the elevation was in ~300 m. Hence, the average crustal density at that time was higher than the density of common platforms. Under isostatic equilibrium it must have been 2900–3000 kg/m^3. This density might have existed if in the late Eocene the lower crust had a layer of deeply metamorphosed basic rocks close in density to the mantle.

A similar structure was typical of some ancient and epi-Paleozoic platform areas. For example, in the southwest of the Ukrainian Shield, the crustal thickness varies from 40 to 55–60 km (Chekunov, 1988) (see also Fig. 5.8 in (Artyushkov, 1993)). Nevertheless, the crustal surface in this area is almost flat, which points to lateral variations in the average crustal density. In places with $h \sim 55–60$ km, at depths more than ~35 km, the velocities of compression waves increase with depth from V_P = 6.8–7.0 km/s, typical of basaltic layer, to V_P = 7.5–7.7 km/s. The latter values are typical of deeply metamorphosed basic rocks — garnet granulites with a density close to the mantle one, ρ_m (Sobolev and Babeiko, 1984). In Transbaikalia, a 10–30 km thick rock layer with V_P = 7.7–7.8 km/s occurs beneath 40 km thick crust (Krylov et al., 1981). The crustal uplifting began a few million years ago (Map..., 1977). The average relief height reached 1–1.5 km. The uplifting is of minor amplitude; therefore, the present-day crust is, most likely, of the same structure as before the process, when its surface was at a height of about 0.5 km.

Heavy metabasites, garnet granulites, and eclogites form in the lower crust of folded belts as a result of phase transitions during strong compression (Artyushkov, 1993; Artyushkov et al., 2000). In the CTS they seem to 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 (Burtman, 2006; Kurenkov, 1983; Lomize et al., 1997). Figure 9 shows a simplified model for high-density crust, which can be taken as a first approximation for the Early Oligocene CTS. The upper layer of the crust is 42 km thick and has a density of 2830 kg/m^3; beneath it, there is a 5–15 km thick layer of garnet granulites and eclogites, which is close in density to the mantle.

Possible processes in the lower crust during the Pliocene-Quaternary uplifting

The ascent of the hot low-density asthenosphere to the lower crust including deeply metamorphosed rocks might have been accompanied by two processes (Artyushkov, 1993). Eclogites or basic garnet granulites, more consolidated than the asthenosphere, were replaced by it, being detached from the crust and submerging with the mantle lithosphere. On the contact with the hot asthenosphere and during the supply of active fluid, garnet granulites, less consolidated than the mantle, might have undergone retrograde metamorphism accompanied by the serpentinization of neighboring peridotites. This led to the deconsolidation and, correspondingly, additional uplifting of the surface. Since the Oligocene, the CTS has uplifted (as a result of compression) to ~1 km, and the uplifting due to the asthenosphere ascent might have reached 1.5–2 km. In total, this yields an uplifting of ~2.5–3 km, which is commensurate with the average uplifting of ~3 km. Therefore, the probable additional uplifting due to the lower-crust deconsolidation did not exceed 0.5 km.

Discussion

The performed analysis of data permitted us to describe the neotectonic evolution of the CTS. By the early Paleogene, the crustal surface was located at a small height above the sea level (Fig. 9, *a*). The density of its upper layer, ~42 km thick, was close to the average density of crust in platform areas. Beneath this layer, there was a layer of garnet granulites and eclogites with an average density close to the mantle one. The Early Cenozoic effusions of basalts indicate a possible ascent of small volumes of deep-seated mantle matter containing a surface-active fluid to the lithosphere (Fig. 9, *b*), which reduced the lithosphere strength. In the Oligocene, under the compression caused by the India-Eurasia collision, the lithosphere basement was subjected to folding and, as a consequence, faulting. From Oligocene to Middle Pliocene (i.e., over a period of ~28 Myr), the average rate of crustal compression was 2–2.5 mm/year. The compression led to a slow isostatic uplifting of the crust, and by the Late Pliocene its average height reached ~1 km.

In the late Miocene or Pliocene, large portions of deep-seated mantle matter (plume?) ascended to the lithosphere base

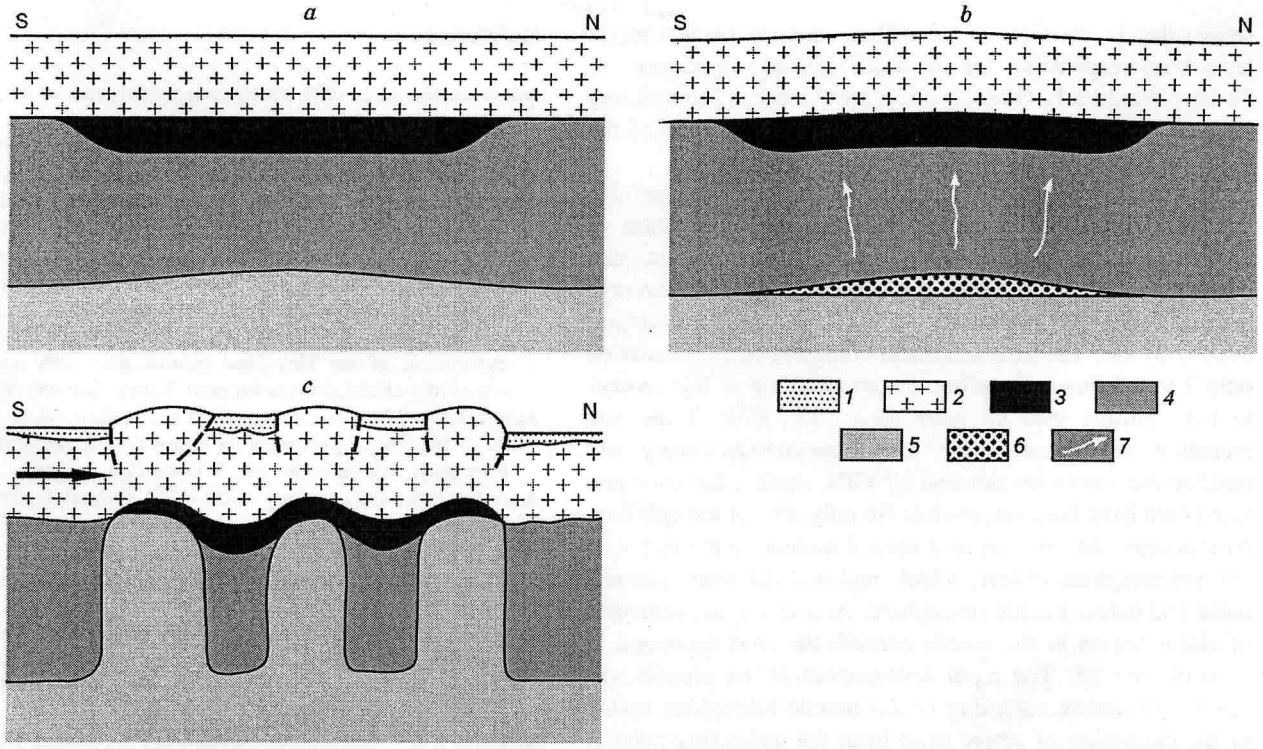

Fig. 9. Tentative scheme of the lithosphere transformation and uplifting of the Central Tien Shan at the neotectonic stage: *a* — late Mesozoic, *b* — Eocene, *c* — modern epoch. *1* — recent molasse; *2* — granite-metamorphic and basaltic layers of the Earth's crust; *3* — high-pressure metabasites with a near-mantle density; *4* — mantle lithosphere; *5* — asthenosphere; *6* — deep-seated mantle matter (plume matter?) with active fluid; *7* — fluid infiltration into lithosphere.

in Central Asia. Spreading along the base, this matter rich in active fluid reached the CTS in the Pliocene. Infiltration of the fluid into the mantle lithosphere drastically reduced the viscosity of the latter. The deconsolidated lithosphere was separated into layers along the surfaces with the highest gradients of deformation. The separated mantle lithosphere began to rapidly break down, submerge, and be convectively replaced by the matter of the hot and less dense asthenosphere, which resulted in the rapid uplifting of the CTS in the last ~2 Myr. This process was most intensive beneath ridge zones, where the asthenosphere closely approached the crustal base (Gubin, 1986; Vinnik et al., 2006). Beneath the crust of large depressions relic lenses of the lithospheric mantle had been preserved; therefore, the depressions rose to a smaller height than the ridges (Fig. 9, *c*). After the subsidence of the heavy mantle lithosphere, 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 fluids, the rocks underwent transformations, which led to their partial deconsolidation and, as a consequence, the intense crustal uplifting.

The collisional crustal compression led to a ~1 km uplifting of the CTS; the replacement of the lithospheric mantle by the asthenosphere, to 1.5–2 km; and the probable deconsolidation of metabasites in the lower crust, to 0.5 km. All this gave rise to a mountain system ~3 km in average height. In the late Pliocene and Quaternary, when the two latter factors took place, the contribution of compression to the mountain rise was close to 10% and anyhow was not more than 18%.

Similar acceleration of uplifting in the Pliocene-Quaternary took place in other areas of Central Asia beneath which mantle deconsolidation was established: the Pamirs, Tarim, Kunlun, Tibet, and Himalayas (Gubin, 1986; Laverov, 2005). For example, the rapid uplifting of the Tibet started ~3 Myr ago and reached 2500–3600 m. At the same time, the Kunlun area was uplifted by 2600–3100 m, and the Tarim area, by 1200 m (Laverov, 2005; Li Jijun, 1995; Mørner, 1991). In all these areas, the crustal compression contributed little to the uplifting. The Pliocene-Quaternary acceleration of orogeny was also established in other areas of the Alpine-Himalayan belt, in southern Siberia, northeastern Asia, some regions of the East Siberian Platform, in most areas of Africa, in the east of North and South America, in Antarctica, etc. (Artyushkov and Hofmann, 1998; Map..., 1977). In southern and eastern Africa and East Siberia, the uplifting was not accompanied by crustal compression. In Central Asia, western Mongolia, and the Andes, the compression might also have only slightly contributed to the recent uplifting. Hence, deep-level processes played a significant role in the uplifting acceleration, and the softened mantle lithosphere in most of continental areas was partly or completely replaced by the asthenosphere.

Conclusions

According to modern concepts, the uplifting of the Central Asian crust was related to its compression as a result of the India-Eurasia collision. Our analysis of available data has

shown that in the Central Tien Shan, this mechanism might have been responsible for the slow uplifting developed for 28 Myr, from Early Oligocene to Late Pliocene. The uplifting formed a relief 1 km in average height at the place of the platform peneplain.

The rate of crustal rise increased by an order of magnitude 2 Myr ago, which manifested itself in the acceleration of erosion cutting, formation of new ridges in basins, and coarsening of Cenozoic molasse. Since that time, the average height of the CTS has increased by 2 (and, in some ridges, even 4–5) km. The rate of crustal compression has increased only 2 to 2.5 times; therefore, the contribution of this process to the uplifting was no more than 10%. Even if the rate increased 4–5 times in the Late Pliocene-Quaternary and reached the values established by GPS studies, the compression could have been responsible for only 18% of the uplifting. At that time, the orogeny was caused mainly by the ascent of the asthenosphere matter, which replaced the more consolidated and colder mantle lithosphere. As a result, the velocities of elastic waves in the mantle beneath the crust decreased in most of the CTS. The rapid development of this process was due to the drastic softening of the mantle lithosphere owing to the infiltration of active fluid from the underlying mantle.

The crustal thickness beneath the CTS varies from 40–45 to 64 km. This great thickness could not have been the result of the compression only. By the beginning of the recent orogeny, the CTS crust, having a thickness and average density typical of platform areas, was probably underlain by deeply metamorphosed basic rocks with a density close to the mantle one. Under the action of the ascended asthenospheric mantle, part of these metabasites became deconsolidated as a result of phase transitions, and this intensified the surface uplifting.

Evidence for a drastic acceleration of orogeny under the effect of deep-level processes in the Pliocene-Quaternary is observed in many areas of neotectonic activity, irrespective of the compression and preceding evolution of the crust.

Some researchers think that the neotectonic stage was the last of the repeated 20–40 Myr long orogeny stages in the Phanerozoic (Leonov, 1980; Shul'ts, 1948). Orogeny covered regions with different evolution histories and was independent of regional processes of the plate interaction. Data from the present paper indicate that the orogeny drastically accelerated as a result of mantle processes by the end of the recent orogeny stage (in the Pliocene-Quaternary).

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