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Recent Mountain Building in Geodynamic Evolution of the Central Alpine–Himalayan Belt

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Abstract—From the end of the Eocene through the Pliocene, the Alpine–Himalayan Belt underwent collisional compression induced by convergence of the Gondwana plates with the Eurasian Plate and varied in orientation from the north-northwestern to the northeastern directions. The collisional compression was expressed in folding, thrusting of continental crustal tectonic sheets over one another, and closure of the residual basins of Neotethys and its backarc seas; it resulted in local thickening of the crust and its isostatic uplifting. As a rule, the uplifts were not higher than ~1.5 km. In other words, before the Pliocene, the growth of local mountain edifices was caused by collisional compression of the belt. Isostatic uplifting of the thickened crust was continued in the Pliocene and Quaternary even more intensely than before, but the general rise of the mountain systems was superposed on this process. The rise substantially exceeded in amplitude the contribution of the uplift caused by compression and did not depend on the preceding Cenozoic history of either territory. Not only the mountain ridges but also most adjacent basins were involved in rising, which eventually led to the contemporary mountain topography of the belt. The spread of the hot and fluid-enriched asthenosphere of the closed Tethys beneath the orogenic belt could have been a cause of such additional rising. The uplift was an isostatic reaction to decompaction of the lithospheric mantle partly replaced with asthenosphere and of the lower crust subject to retrograde metamorphism under the effect of cooled asthenospheric fluids. The deep transformations are also probably responsible for deepening of some basins in the Pliocene–Quaternary and more contrasting transverse segmentation of the belt.

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INTRODUCTION

Obruchev [41], introducing the terms *geotectonics* and *neotectonic* (recent) stage, applied them to the process leading to the formation of the present-day topography, which is distinguished by high-mountain systems, which did not exist earlier in the Mesozoic and Cenozoic geological history. In this paper, 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], as well as relationships between mountain building and collisional compression induced by convergence of the southern (Gondwana) plates with the Eurasian Plate. From the Oligocene to the early Pliocene, the isostatic compensation of the thickened Earth's crust has been the main mechanism of local uplifting in zones of concentrated deformation. In the Pliocene–Quaternary, the height of the mountains increased 2–3 times. This intensification of mountain building was caused by a decrease in the density of the crust and upper mantle under the effect of the asthenosphere, rather than by increasing compression.

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 southwestern (in present-day coordinates) passive continental margin of the ocean, which was underlain by Precambrian basement. The continental fragments detached from Gondwana moved to the northeast, where the Tethyan oceanic lithosphere was subducted beneath the island arcs and active margins of the northern plates. 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- and collision-related magmatism and metamorphism. The poorly reworked fragments (median masses) retained the platform tectonic regime. This process has developed since the breakdown of Pangea in the Carboniferous and became especially distinct in the Meso- and Cen-

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

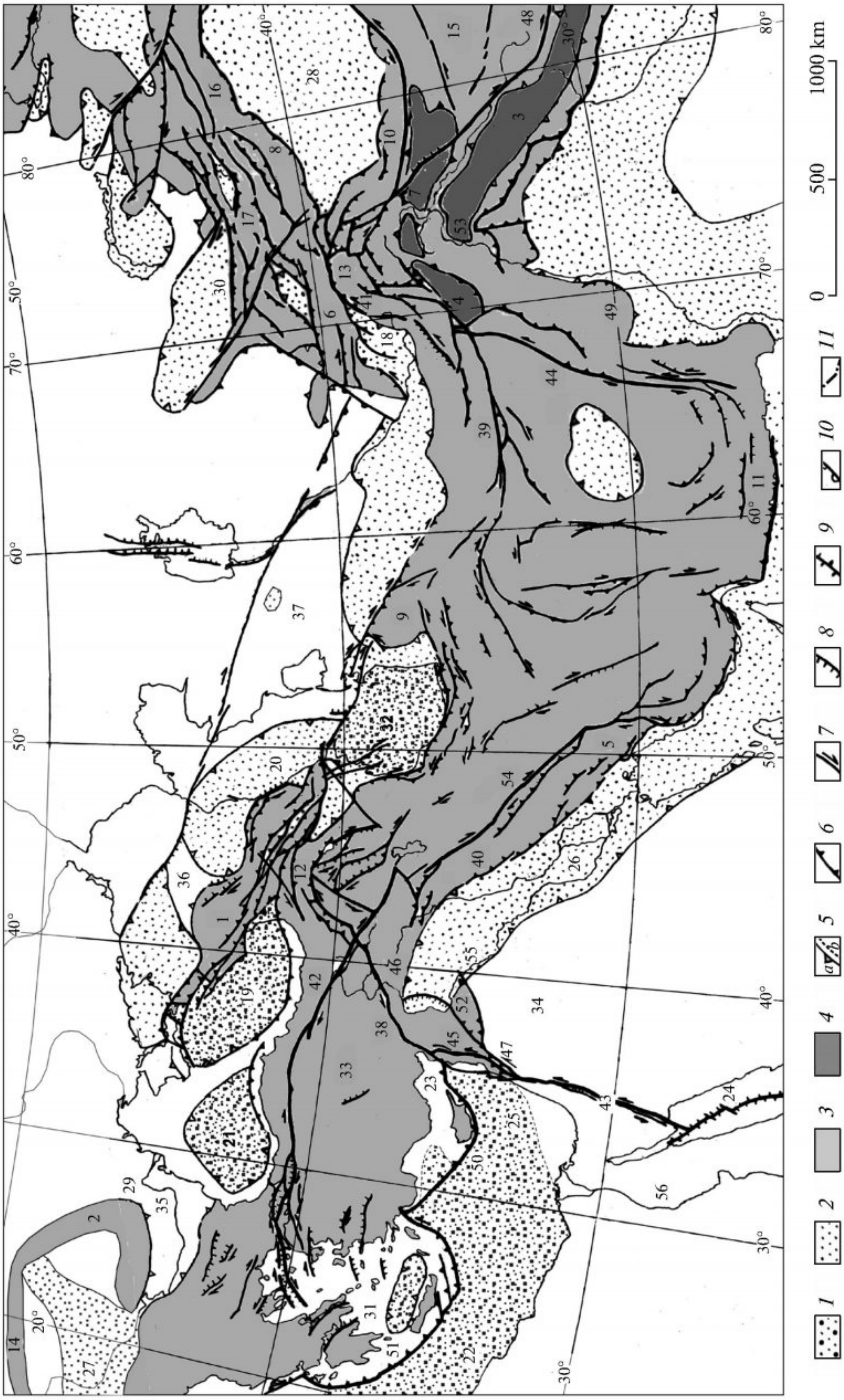
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Fig. 1. Schematic map of Pliocene–Quaternary structural elements of the Alpine–Himalayan Orogenic Belt from the Eastern Mediterranean to Central Asia. (*1, 2*) Basins: (*1*) with suboceanic crust and (*2*) domains of steady sagging of the continental crust (uplifted Tarim Basin in other figures); (*3*) mountain systems (low-mountain systems (<1.5 km) in Figs. 3–5); (*4*) the highest mountain edifices; (*5*) boundaries between mountains and intermontane basins: (*a*) in Fig. 1 and (*b*) in Figs. 3–5; (*6–10*) faults: (*6*) thrust or reverse, (*7*) strike-slip, (*8*) normal, (*9*) pull-apart, (*10*) flexure (master Pliocene–Quaternary faults shown in Fig. 1 are denoted by heavy lines); (*11*) Tornquist Line (in Figs. 3–4) and Palmyra–Apsheron Lineament in Fig. 5.

Numerals in map. *Mountain systems*: 1, Greater Caucasus; 2, Eastern Carpathians; 3, Himalayans; 4, Hindu Kush; 5, Zagros; 6, Western Tien Shan; 7, Karakoram; 8, Kakshaal; 9, Kunlun; 10, Makran; 11, Kopet Dagh; 12, Lesser Caucasus; 13, Pamir; 14, Northern Carpathians; 15, Tibet; 16, Khan Tengri; 17, Central and Eastern Tien Shan. *Basins*: 18, Afghan–Tajik Depression, 19, East Chernomorian Basin; 20, Derbent Trough; 21, West Chernomorian Basin; 22, Ionian Basin; 23, Kilikia–Adana Trough; 24, Red Sea Rift; 25, Levantine Basin; 26, Mesopotamian Trough; 27, Pannonian Basin; 28, Tarim Basin (microplate); 29, Fokshany Basin (a part of Carpathian Foredeep); 30, Shu Basin; 31, Aegean Sea; 32, South Caspian Basin. *Platforms*: 33, Anatolian; 34, Arabian; 35, Moesian; 36, Scythian; 37, Turan. *Faults and fault zones*: East Anatolian; 39, Herath; 40, Contemporary Main Zagros Fault; 41, Darwaza; 42, North Anatolian; 43, Dead Sea (Levantine) Transform; 44, Chaman; 45, El Gharb; 46, South Taurus Zone; 47, Yammunneh. *Tectonic zones*: 48, Indus–Zangpo; 49, Quetta; 50, Cyprus arc; 51, Crete–Hellenic arc; 52, Palmyrides; 53, Punjab Syntaxis; 54, Sanandaj–Sirjan. *Rivers*: 55, Euphrates; 56, Nile.

ozoic, 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 edifices 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. Thus, 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 Turan and Scythian epi-Paleozoic platforms and Hercynides of the 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 suboceanic 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 northwest 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 amplitudes of lateral offsets increasing from the west eastward both in particular structures, e.g., the greater amplitude of shear at the western flank of the Indian Plate as compared with the western flank of the Arabian Plate, and in variable dimensions of the belt's segments, which have shortened in the transverse direction to different extents [3, 46, 80].

The contemporary mountain edifices originated in different parts of the Alpine–Himalayan Belt asynchronously, but mostly in the Oligocene [46, 55]. Therefore, the history of recent mountain building is considered below since the Oligocene; the Eocene is regarded as a preceding epoch.

EOCENE

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 median masses, tectonic nappes obducted on the margins, and relict island arcs of the Mesozoic Tethys and spread over the neighboring Moesian, Scythian, Turan, 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 [7, 11].

The rest of the future High Asia—the Central and Eastern Tien Shan, Tarim, Pamir–Hindu Kush–Karakoram, Kunlun, and Tibet—were a land. The widespread granitic batholiths, the first phases of which are dated at the Cretaceous and the final ones at the Miocene, was a cause of relatively high standing of the Pamir–Hindu Kush–Karakoram region [53, 89]. The territory of the Central and Eastern Tien Shan in the Paleocene and Eocene was a peneplain with relative elevations of a few hundred meters accepted as a preorogenic planation surface [49, 52, 55]. The redeposited Late Mesozoic weathering mantle is mainly fineclastic red beds with fauna of the middle–late Eocene and probably early Oligocene [11]. Basaltic flows occur in its lower part. Their K–Ar and Ar/Ar age is 54–70 Ma [4, 28, 40, 43]. The thickness of this sequence commonly does not exceed a few tens of meters, but in some basins is greater than 100 m, pointing to initial sagging. A similar Lower Paleogene continental silty–clayey sequence up to 30 m thick is known in the Chuya Basin of the Gorny Altai [16].

Deeper troughs with thinned (suboceanic) crust stand out against the background of the land and areas

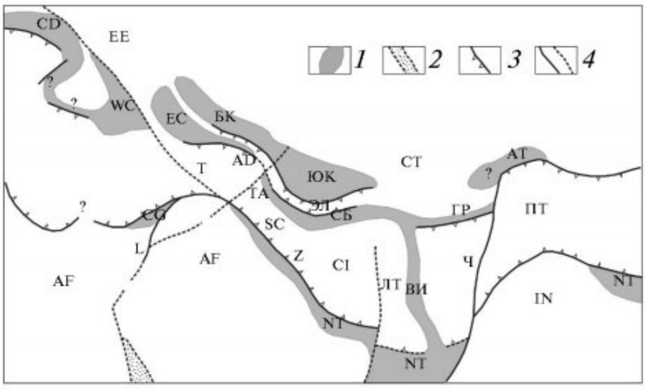

Fig. 2. Conceptual scheme of the main structural elements of the Alpine–Himalayan Belt in the Eocene (~45 Ma ago), modified after [21, 30, 67, 84]. (*1*) Basins with thinned crust; (*2*) Red Sea prorift trough; (*3*) main thrust faults and subduction zones; (*4*) main transform and other faults and their inferred continuations. Structural elements (abbreviations in figure): AD, Adjaria–Trialetia; AP, Arabian Plate; AT, Afghan–Tajik Depression; AF, African Plate; GK, Greater Caucasus Basin; EE, East European Platform; EI, East Iranian Basin; EC, East Chernomorian Basin; HR, Harirud Basin; Z, Main Zagros Fault; WC, West Chernomorian Basin; IN, Indian Plate; CB, Carpathian Basin; L, proto-Levantine Transform Zone; LT, Lut microplate; NT, Neotethyan relics; PT, Pamir and Tibet; SB, Sabzevar Basin; SS, Sanandaj–Sirjan Zone; T, proto-North Anatolian Fault Zone extending as Pechenega–Kamena Fault and Tornquist Line; TA, Talysh; CI, Central Iran microplate; CH, proto-Chaman Transform Zone; EL. Elbrus; SC, South Caspian Basin.

of epicontinental marine sedimentation (Fig. 2). These were relics of the Neotethys and backarc basins [21, 68]. The Eocene relics of the Neotethys existed in the Transhimalayas (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.

The region between the Taurus microplate and the northwestern margin of the Arabian prominence of the African Plate has a complex tectonic history [76]. 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, where the oceanic Trodoos Complex was formed. 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 [22]. 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 [84, 85]. 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 more 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 [66] 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 the eastern Mediterranean, where it is expressed in seismic sections as faults [61, 62]. In the south, this zone probably merged with a protorift developing in the Late Cretaceous–Eocene partly on the place of the future Red Sea Rift [58].

Among the backarc basins, the near-latitudinal Carpathian–Greater Caucasus system of troughs, extending from the outer Carpathian Zone to the proto-South Caspian Basin, was the largest [24, 31, 56, 68]. The system was en echelon arranged: the troughs extended in the NW–SE direction in such a manner that the northwestern end of each trough shifted to the east started to the north of the southeastern end of the previous trough shifted to the west. The troughs were separated by 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 Gerirud Trough and connected with the pre-Makran relic of the Neotethys via the Central Iran Basin [21].

The origin of the backarc troughs remains conjectural. The Cretaceous ophiolotes in the relics of the Sabzevar and East Iranian Troughs indicate spreading [21]. 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 [31, 68, 82]. 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 [24] and similar deepening of the relict Kilikia–Adana Basin should be related to other causes, e.g., to compaction of the lower crustal 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 super-

Fig. 3. Schematic map of the main structural elements of the Alpine–Himalayan Belt at the end of the Oligocene (~25 Ma ago), modified after [5, 12, 17, 21, 23, 3,67, 85, 87]. See Fig. 1 for legend.

position 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 Chernomorian Basin and the Talysh Trough as an extension of the Sabzevar Trough [56]. 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 Chernomorian 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 [3, 68]. The Neotethyan backarc basins became narrower or partly closed [21, 51, 68]. In the regions where the Neotethys had been closed before the Eocene, for example in front of the Punjab jut of the Indian Plate, a peak of high-pressure metamorphism fell on the Eocene (50–40 Ma) [70, 90]. Intense 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.

TECTONIC EVOLUTION OF THE BELT FROM THE OLIGOCENE TO THE PRESENT

Four stages of deformation and metamorphism differing in direction of compression of the orogenic belt related to the motion of the Gondwana plates are distinguished [17, 46, 87]. These stages correspond to Oligocene–early Miocene, middle Miocene, late Miocene–early Pliocene, and middle Pliocene–Quaternary.

Oligocene–Early Miocene (35–17 Ma)

Compressive deformations which started in the east of the region at the end of the middle Eocene continued in the Oligocene and resulted in closure of the suboceanic Sabzevar Trough [21] and the Indus–Zangpo Zone [57] (Fig. 3). The intense compressive deformations which took place in the Herath Zone in northern Afghanistan and in the northwestern Pamir–Hindu Kush brought about squeezing of the southwestern Pamir to the east and its thrusting over the zone of the southeastern Pamir [17]. Transverse compression in the northern part of the Quetta Zone was expressed in folding of the Eocene Katawaz Trough and formation of the NE-trending thrust faults in the

Khost, Tarnaka, and Khash Rud ophiolite zones [12, 91].

Syn- and postfolding uplifts arose in the compressed zones. The Oligocene–Miocene conglomerate unconformably overlapped the deformed rocks of the Indus–Zangpo Zone [57, 92] and was found in the frameworks of the Pamir [53] and Kunlun [37]. Differential vertical movements spread over the Tien Shan. In the Central Tien Shan, Oligocene fine-pebble conglomerate and fine clastic sediments are known [11, 36, 52, 55]. Local clastic material occurs and occasionally dominates in the pebbles [5]. This implies that the ridges as provenances and the basins as depocenters originated in the Oligocene. In the early Miocene, vertical movements became more sluggish; hillside and lacustrine clayey sediments locally with evaporites were deposited at that time. In the Chuya Basin of the Gorny Altai, the Oligocene and early Miocene are composed of lacustrine and bog sand-shale sediments with interbeds of brown coal. The alluvial sandy–gravely–pebble sediments in the marginal parts of the basin provide evidence for origination of neighboring uplifts [10].

Judging from the relatively fine clastic material and shallow (a few hundreds of meters) downcutting of valleys formed at that time [36, 52], the vertical range of the Oligocene topography in the Central Tien Shan did not exceed a kilometer. The anomalously coarse conglomerate of the Ming-Kush–Kökömeren ramp is a product of destruction of reactivated Late Paleozoic nappes and unrelated to significant hypsometric contrast [4]. The fine clastic molasse in the foothills of the Pamir and Hindu Kush (Afghan–Tajik Depression), Kunlun (south of the Tarim Basin), and Gorny Altai (Chuya Basin) indicates that no high mountains existed at that time. Nevertheless, the isotopic data on the paleosoil in Central Tibet show that high mountains existed there ~26 Ma ago [63]. A high isostatic uplift at that time is also suggested in the Southwestern Pamir, the upper crustal sheet of which, reaching 25 km in thickness, is thrust off the continental crust of the Southeastern Pamir [15]. These high mountains were not extensive and were later subjected to erosion.

In the Arabian–Caucasus segment of the orogenic belt, the subduction in front of the South Taurus led to the formation of the accretionary wedge on the northern slope of the Kilikia–Aadana Trough in late Eocene–early Oligocene. This wedge is composed of fragments of the Mesozoic oceanic crust and its Lower Paleogene cover. The blocks of carbonate cover of the Taurides slid over them. The process completed with the collision of the Taurides with the Arabian Plate in the northeast of this region and overlapping of the accretionary wedge by lower Miocene sediments [84, 85]. A relic of the southern margin of this basin was retained in the southwest. In the early Miocene (~17 Ma ago), it was detached by the renewed Cyprus arc, and the Levantine Basin at the southern margin of Tethys began to subduct beneath this arc. Deformations reached a culmination at that time. The sharp angular unconformity between the Eocene and Helvetian is documented in the northwest of Syria [87]. Deformations developed in other zones of the Arabian–Caucasus Orogenic Belt up to the southern flank of the Caucasus part of the Carpathian–Caucasus system of troughs. Their underthrusting beneath the Lesser Caucasus was accompanied by formation of flysch along with tectonic and gravity mixtites [29, 31, 56]. The troughs themselves did not undergo deformation. In the Oligocene, they even locally deepened despite a global regression, especially intense in the very beginning of the late Oligocene [95], whereas an epicontinental sea spread over the entire Greater Caucasus and the adjacent Scythian Plate adjacent to the Caucasus and the Carpathians [24]. The supply of clastic material into the sedimentary basin was reduced in the early Miocene.

The origination of a graben on the spot of the future Aden–Red Sea Rift was the most important event in the Oligocene, which initiated pulling apart of Arabia from the African Plate. In this connection, the transform of the Dead Sea arose in the early Miocene (~20 Ma ago) [67]. Its northern segment extended along the continental slope of the Levantine Basin [87], inheriting an earlier transform zone. In the Balkan Mountains, the late Eocene phase of thrusting was followed by development of a foredeep, where flysch sedimentation gave way to deposition of molasse [68]. The displacement of the Carpathian inner zones gave birth to the Carpathian arc, which completed by thrusting of the detached nappes of the Northern Carpathians over the foredeep at the end of the early Miocene.

The Oligocene uplifts (mainly low-mountain as judged from the composition of the piedmont molasse) were confined to compression zones in the west of the belt. Except for the Caucasus troughs of the Paratethys, the uplifts grew in area, while the sediments in the epicontinental basins, e.g., in northern Arabia, were related to the regressive phase of the Paleogene sedimentation cycle. This was probably caused by increase in collisional compression, though it can be partially explained by a global drop of ocean level.

All structural units of the orogenic belt which underwent compressive deformation in the Eocene and early Miocene extend in the latitudinal or northeastern directions. This implies that the principal compression axis was oriented in the north-northwestern direction, which coincides with the directions of movement of the Gondwana plates.

Middle Miocene (16–11 Ma)

During the second stage (end of the early Miocene and the middle Miocene), the most intense lateral displacements and deformations of crustal blocks took place in the east of the belt corresponding to the region

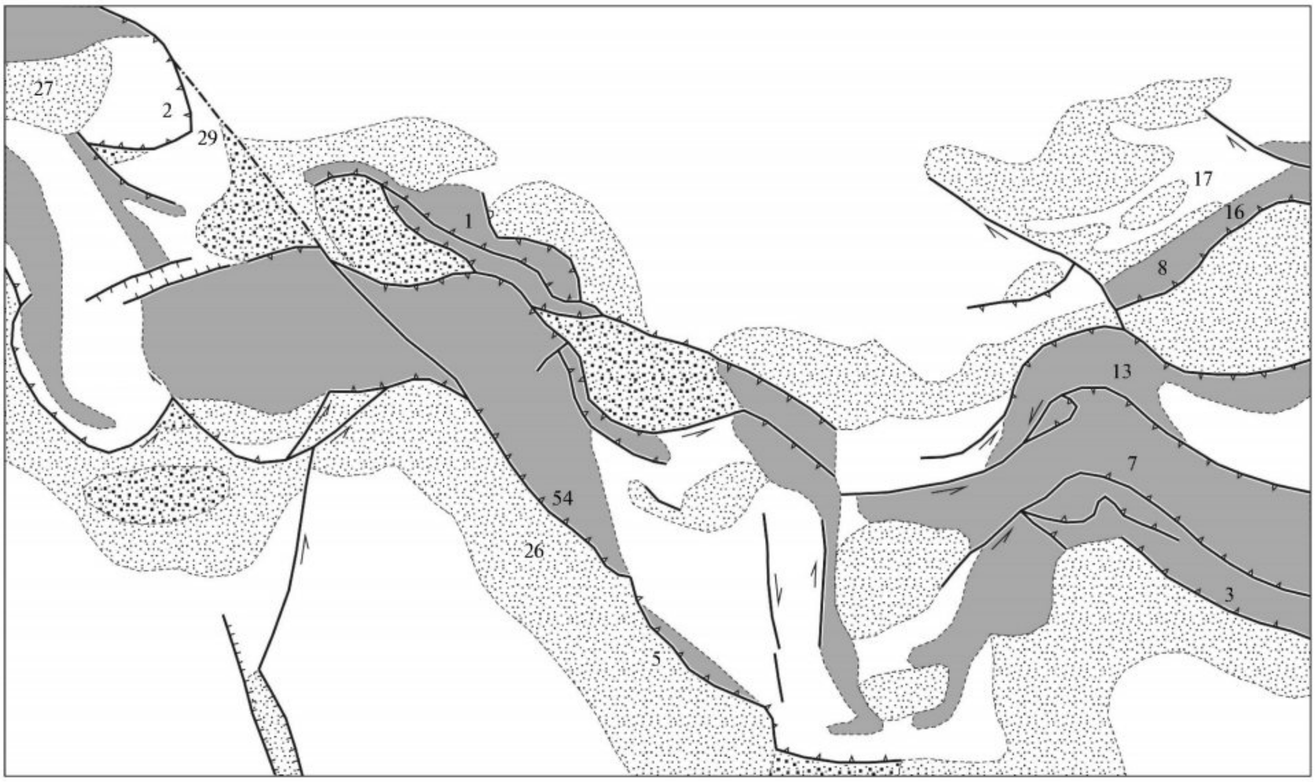

Fig. 4. Schematic map of the main structural elements of the Alpine–Himalayan Belt at the onset of the middle Miocene (~18 Ma ago), modified after [5, 17, 23, 30, 58, 67, 89, 90]. See Fig. 1 for legend.

of the Indian–Eurasian collision. The Himalayas, Karakoram, and NW-trending Pamir zones were involved in deformation and thrusting accompanied by a peak of metamorphism and granite formation [17, 89, 90] (Fig. 4). At the same time, the intensity of tectonic movements decreased in the Central Tien Shan, where Oligocene uplifts extended in the NEN direction. The average rate of erosion became lower than in the Oligocene [52]. The Miocene lacustrine fine clastic sandshale sediments are predominant, whereas alluvial and proluvial sediments are second in abundance. The areas of sedimentation expanded, having overlapped some Oligocene uplifts [5]. Each sedimentary basin was a chain of lakes connected by permanent or intermittent channels. The basins were separated by flat uplifts, which act as additional provenances. To the south and the east, approaching the present-day Kakshaal-Too Range and the Khan Tengri Massif, the clastic material becomes coarser, indicating higher elevations and more intense erosion. They were the main sources of removed clastic material. Carbonate interlayers were replaced with evaporites moving away from the uplifts. A similar setting is established in the Chuya Basin in the Gorny Altai from the middle Eocene to the early Pliocene [6]. The lacustrine fine clastic sandshale sediments were deposited in the center of the basin. At the margins, these sediments are replaced with coarse clastic proluvial–alluvial–deltaic deposits varying from inequigranular sand to pebbles in clast size.

Evidence for rearrangement of the principal compression direction with its shift to the northeast in the early–middle Miocene in the western segments of the belt was also documented. A tectonic quiescence in the northwest of the Arabian Plate came with development of the Helvetian–Tortonian sedimentation cycle. The intense movements along the Main Thrust Fault of the Zagros led to the closure of the Neotethys relict basin between the Arabian Plate and the Sanandaj-Sirjan Zone [68]. This event initiated onset of the development of the Mesopotamian Foredeep, which inherited the formerly sagging northeastern part of the plate. Folding started to develop at the northeastern flank of the trough in the middle–late Miocene.

The Caucasus troughs of the Paratethys were shoaled and then closed; at the end of the second stage, their sedimentary fill underwent folding [24, 31]. Thrusting of the outer zone of the Eastern Carpathians over the Fokshany Basin of the Carpathian Foredeep in the middle–late Miocene was probably also related to the rearrangement of compressive stresses [60]. The thickness of the sedimentary cover in the Eastern Carpathians is now estimated at 8–12 km and initially could have been 10–14 km. Such an increase in the thickness of the sedimentary cover did not bring about a rise of the surface to the calculated

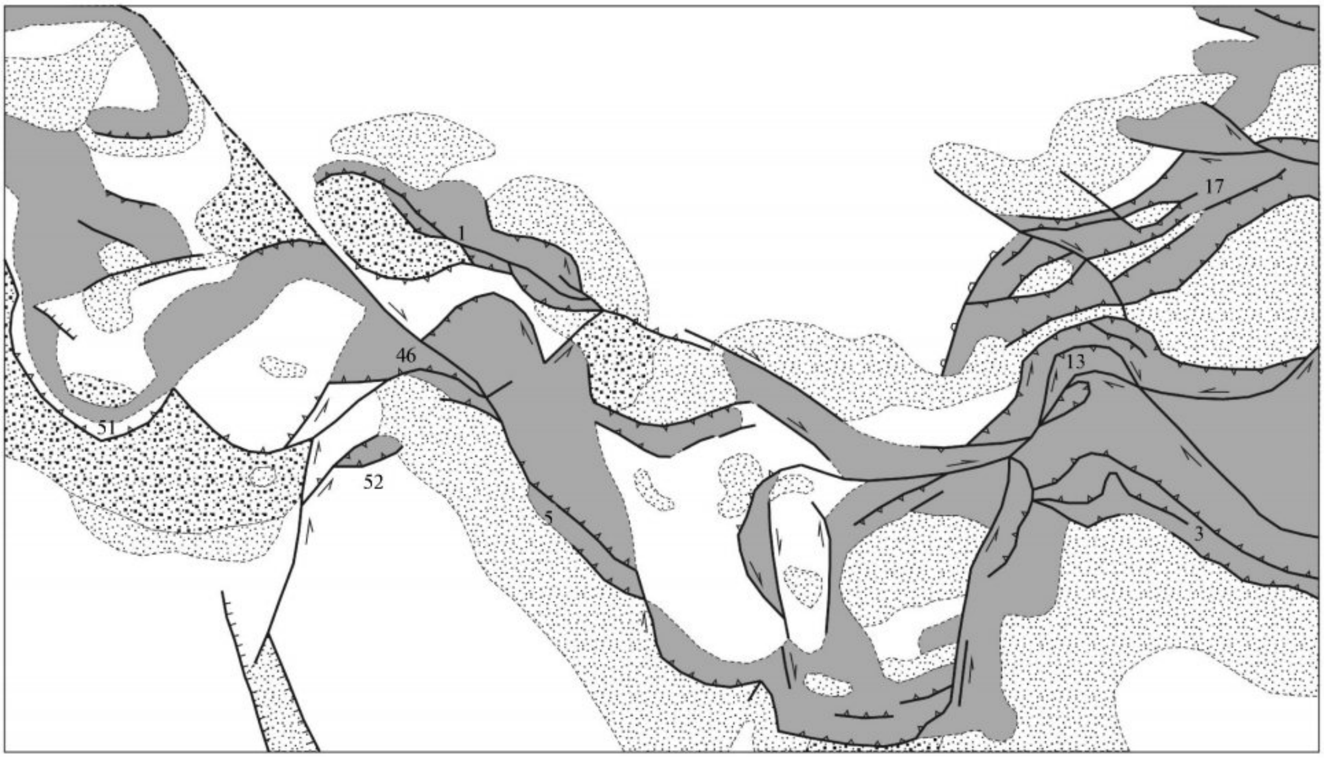

Fig. 5. Schematic map of the main structural elements of the Alpine–Himalayan Belt in the Messinian (~6 Ma ago), modified after [5, 17, 30, 67, 87]. See Fig. 1 for legend.

value of 1.5–2.4 km, and the surface remained at a height of ~0.5 km. Thus, the uplift to 1–2 km was compensated by compaction of matter at a deeper level of the lithosphere. A similar phenomenon probably took place at the southern slope of the Greater Caucasus, where intense folding and stacking of sedimentary sequences also did not lead to the formation of high mountains. Judging from the composition of the clastic complexes, highlands were not formed in other regions of the belt either. Moreover, the Pannonian Basin was formed on the place of the deformed inner zones of the Carpathians.

Late Miocene–Early Pliocene (10.0–3.6 Ma)

During the third, late Miocene–early Pliocene stage, the prevalent orientation of compression again became north-northwestern or nearly meridional. The peak of diastrophism fell on the Messinian. A system of south-verging thrust faults developed on the southern slope of the Greater Caucasus (Fig. 5). At the southern flank of the region of interaction of the Arabian and Eurasian plates, the main phase of folding and thrusting took place in the Palmyrides. Folding in the Hellenides and thrusting in the Pamir were resumed. The fold–nappe zones expressed in the topography shifted to the south from the Main Thrust Faults of the Zagros and the Taurus. In the Himalayas, such a progradation was marked by shift of the maximum displacements and deformation to the zone of the Frontal Fault.

In some intermontane basins of the Central Tien Shan, the middle Miocene sequences were enriched in coarse clastic rocks, which gave way to fine clastic rocks upsection. The coarse clastic rocks are products of destruction of the Late Paleozoic tectonic nappes; i.e, of activization of horizontal rather than vertical movements [5]. In the Gorny Altai, the content of coarse clastic rocks increases in the fill of the Chuya Basin. In some places, for example, in the Greater Caucasus [24], the late Miocene displacements and folding resulted in formation of dissected topography; however, the composition of clastic material in the intermontane basins and foredeeps indicates that the uplifts were characterized here, as in other segments of the belt, by moderate height of mountains.

Middle Pliocene–Quaternary (the last 3.6 Ma)

The contemporary network of large active faults of the belt was formed by the middle Pliocene. The displacements along these faults (mainly strike-slip) indicate near-meridional orientation of the principal compression axis. In the northwest of Arabia, the onset of the fourth stage (4.0–3.5 Ma) was accompanied by rearrangement of the northern segment of the Dead Sea Transform. While in the Miocene, its main branch extended along the continental slope, now the

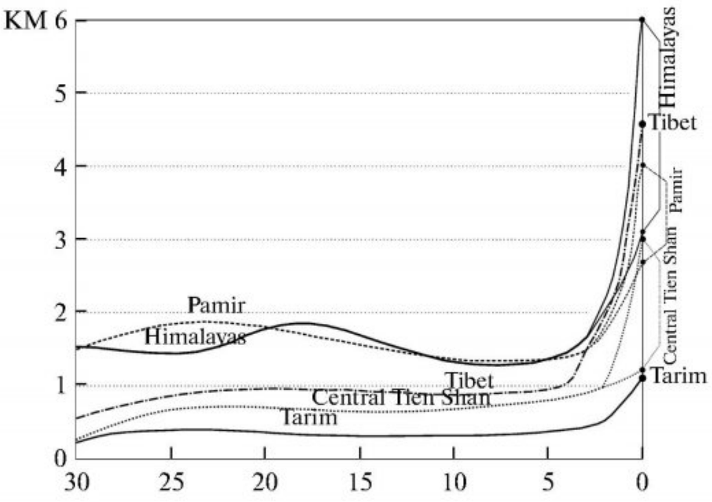

Fig. 6. Acceleration of Pliocene–Quaternary mountain growth in Central Asia. The lower calculated data showing what height would be reached by the Pliocene–Quaternary rise only as a result of increased compression are compared with the actual heights of the Himalayans, Pamir, and Central Tien Shan.

main displacements concentrate along the Yammunneh and El Gharb segments [87] (Fig. 1). At the same time, the East Anatolian and North Anatolian Fault Zone, as well as the Main Zagros Fault demarcating the present-day plate boundaries [46, 88, 97, 98], eventually formed.

The velocities of vertical tectonic movements sharply increased over the last 4–2 Ma. The vertical amplitude of displacements at least doubled or tripled. The contemporary mountain systems and high plateaus were formed during this time, when coarse molasse was deposited in the foredeeps and intermontane basins. The most significant increase in uplifting is established in Central Asia (Fig. 6). The onset of acceleration of vertical movements was not synchronous. The average height of the Himalayas has increased by over 3 km [81] and the Central Tien Shan by ~2 km [27, 47, 52] since the end of the Pliocene (~2 Ma). The rapid rise of Tibet started 2.8–2.4 Ma ago and reached 2500–3600 m; the Kunlun and Tarim have grown simultaneously for 2600–3100 and ~1200 m, respectively [37, 77, 81]. This yields an average rate of growth of Tibet and the Kunlun as 1.0–1.5 and 1.0–1.2 mm/yr, respectively.

The particular stages of rapid uplift are outlined; thus, the rate of uplifting increased with time. The last stage began at the end of the middle Pleistocene and the velocity of uplifting during this stage locally reached 10 mm/yr. According to the data of recurrent leveling, the velocity of contemporary uplift of Tibet is 6.8 mm/yr, on average, and increases from the Kunlun and northeastern Tibet to the Himalayas [99]. Over the last 3–5 Ma, the Pamir has grown ~2 km on average. The accelerated uplift of Altai over the last ~3.5 Ma has been revealed from the fission-track timing [64]. At the same time, the topography of the Transbaikal region became more contrasting. The mollase in the Tunka, South Baikal, and other basins became coarser, owing not only to activization of rifting but also to growth of high mountain ranges on the place of former low mountains.

The intense mountain growth in the Pliocene and Quaternary has been established in the Greater Caucasus [38], the Carpathians [60], and the Alps [1]. At the northwestern end of the Mesopotamian Trough (middle reaches of the Euphrates River), the lagoonal and lacustrine sedimentation continued into the early Pliocene, but then it gave way to coarse clastic alluvium fed by anticline uplifts prograding to the south. At the Syrian shore of the Mediterranean Sea, a rapid growth of the Coastal Anticline has been established. The anticline began to evolve in the Miocene, when its crest was eroded 500 m deeper than the eastern limb. The eroded surface was flooded by basalts, the K–Ar of which was estimated at 6.3 ± 0.3 to 4.3 ± 0.2 Ma [93]. The basaltic hyaloclastites formed 5.4 ± 0.2 Ma ago under the effect of seawater are now located 260–300 m above sea level [30]. In the axial part of the anticline, the basalts dated at 5.4–4.8 Ma have been raised to a height of 800 m. At the eastern limb of the fold, basalts are located 400 m lower. The coastal Lebanon Anticline underwent intense Pliocene–Quaternary uplift, as well [69].

Although the uplift of the mountain systems in the Pliocene and Quaternary involved most conjugate intermontane basins and foredeeps, some large negative structural elements in the western part of the belt underwent intense subsidence. Signs of this were found in the Black Sea, South Caspian Basin, and in the southeastern part of the Tersky Trough continuing into the Central Caspian as the Derbent Trough. The maximal thickness of the sedimentary cover here exceeds 14 km and the 5 km fall on the Pliocene–Quaternary complex. The most intense sagging began at the end of the Pliocene and continues at present, providing uncompensated sedimentation [32]. The western part of the South Caspian is a starved basin down to 1 km deep with thinned (8–10 km) consolidated crust. Up to 20 km of sediments have been deposited here, no less than half of them being Pliocene–Quaternary sediments. The thickness of the middle Pliocene–Quaternary complex locally exceeds 6 km [1, 32].

The subsidence of the Aegean Sea started in the late Miocene and became more intense in the Pliocene and Quaternary [68]. At the same time, from the Tortonian and especially in the Pliocene–Quaternary, the Ionian and Levantine basins of the Mediterranean Sea also deepened. Increase of sagging of the latter from the Tortonian to Pliocene–Quaternary is confirmed by the growth of the sedimentation rate by 2–6 times in various parts of the basin [19].

The Levantine Basin, uncompensated by sediments, is up to 2500 m deep (3200 m at the Herodotus abyssal plain). In the north of the basin, a trough in front of the Cyprus arc is expressed in the west as a deep bathymetyric depression between the Cyprus and the submarine Eratosthenes Uplift and in the east as a submarine extension of the Nahr al-Kabir Trough, where the thickness of the Pliocene–Quaternary sediments is greater than 1800 m. The Levantine Basin proper is a relic of the southern margin of the Tethys, which now has suboceanic crust with thick (up to 10–14 km) sedimentary cover and the Moho surface at a depth of 20–25 km [61]. The lower and the upper parts of the Neogene–Quaternary section are separated by the Messinian evaporites, which are replaced in the south by alluvial and deltaic sediments of the pra-Nile. The level of the hypersaline Messinian Basin was lower than the present-day level of the Mediterranean Sea. This is proved by the overdeepening of the Messinian channels of the pra-Nile and the other rivers influent in the sea at that time. Currently evaporites occur at a depth of 2 km and deeper.

In the early Pliocene, the breaching waters of the Black Sea and Atlantic Ocean drowned out the Mediterranean Sea, including the Levantine Basin. The depth of its bottom decreases toward its eastern shore and especially southward, where a vast shoal is occupied by the Nile delta, in the underwater part of which the thickness of the Pliocene–Quaternary sequence reaches 3.0–3.5 km [86]. At the boundary of the basin the bottom and continental slope between Tel-Aviv and Beirut, the thickness is 1.3 km [62] and the sole of the Pliocene sediments is subsided to 2.2–2.4 km [19]. At the same time, in the west of Syria, in the subaerial part of the Nahr al-Kabir Trough, the 30-m section of the Messinian gypsum is exposed at a height of ~50 masl. The Pliocene marine clay overlaps it with scouring and basal breccia containing fragments of gypsum and pre-Messinian carbonates and covers the slopes of the adjacent uplifts at a height up to 250 m. No indications of ingressive attitude of the Pliocene were revealed. The submarine part of the Nahr al-Kabir Trough and the neighboring part of the continental slope are disturbed by faults, along which the trough stepwise plunges to the west [74]. The seismic profiles across the wall of the Levantine Basin between Tel-Aviv and Beirut demonstrate that horizontally lying Pliocene–Quaternary sediments are thinned landward, forming a flexure on the continental slope with dip angles up to 10° and complicated by faults [62]. The amplitude of displacement of the Pliocene bottom reaches 1.5–1.7 km. The slope of the beds decreases from the Pliocene to Quaternary, however, even the late Pleistocene (Tyrrhenian) terraces are locally tilted seaward at an angle of 3°.

The relationships described above show that the sea level in the Messinian was lower than the contemporary sea level by several hundred meters. The Tortonian carbonate rocks deposited in the very shallow-water sea occur now in the Nahr al-Kabir River valley at a height no more than a few hundred meters. Thus, the Pliocene–Quaternary uplift of the shore was not great and became significant only in the coastal anticlinal ranges [69]. Thus, increase in the vertical contrast between the early Pliocene surface on the present-day land and in the sea is determined largely by deepening of the Levantine Basin, which underwent tectonic subsidence with an amplitude no less than 1.5 km in the post-Messinian time. An additional isostatic subsidence related to load of thick sediments took place in the Nile delta.

Thus, the Pliocene–Quaternary was the time of activization of not only rising but also subsiding tectonic movements, that is, the time of general increase in their contrast.

RELATIONSHIPS BETWEEN MOUNTAIN BUILDING AND COLLISION

The signs of the first orogeny in the Alpine–Himalayan belt are referred to the Oligocene. The mountain system became widespread in the Pliocene and Quaternary. The neotectonic stage immediately followed, partly coinciding in time with the epoch of collisional closure of the Neotethys and its backarc basins, which began at the end of the Cretaceous and completed in the time interval from the late Eocene to middle Miocene. The region of mountain building is juxtaposed, to a great extent, with the domain of collision, though it expands beyond its limits in the east. This provides grounds to regard recent mountain building as a result of collisional compression, and this view is generally accepted now. Let us consider to what extent this opinion is valid.

Methodical Approach

The occurrence and height of the uplifts, which towered above the sea level or the surface of the subaerial peneplain which existed earlier and is retained nearby, can be judged from the composition of clastic material removed from the eroded uplift and the depth of the related downcutting into the peneplain. When analyzing clastic material, it should be kept in mind that coarse facies could have been accumulated as a result of destruction of the overriding allochthonous sheets, which did not undergo substantial uplifting [4, 29]. In some basins, clastic material was deposited as a product of remote transportation by water and does not characterize the height of the adjacent rises. All this requires ascertainment of the paleotectonic setting of sedimentation.

As concerns the depth of downcutting, in the case of intense linear erosion accompanying growth of mountain ranges, the remnants of the preorogenic surface can be retained on their summits and slopes, allowing judgment about the amplitude of uplift. The stepwise slopes of mountain ranges are commonly interpreted as evidence for pulsatory uplift and serve as the basis of the concept of staged topography. Acceleration of vertical movements reactivates erosion so that an erosion–tectonic scarp (cutting) is formed on the slope of the uplift, leaning on the bottom of the basin or valley, which serves as a local base level regulating deposition of erosion products. The higher the rate of uplifting, the coarser and thicker the accumulation of sedimentary material. The next pulse of rising leads to the uplift of an adjacent site of the basin and the formation of a younger cutting below it. The uplifted site becomes a step on the slope. The steps located at similar hypsometric levels on the slopes of different ranges make up, together with the cuttings leaning on them, a regional stage of topography formed at the same time. This suggests that the cutting is correlated with the coarse lower part of the molasse complex, whereas the steps at the base of cutting are correlated with the fine clastic upper part of the molasse section [36]. In the course of intense uplift, the early landforms can be destroyed and the retained landforms will not reflect its true vertical amplitude corresponding to the recent structure of the mountain range.

Thus, origination and growth of the mountain edifice is recorded in a complex of sedimentary, geomorphic, and structural–geological attributes. The set of such attributes, while not allowing the identification of all features of regional mountain building, nevertheless gives an idea of the general tendencies of uplifting.

Mountainous Upl.fts As a Result cf Collisional Compression

As was shown above (Figs. 1, 3–5), the Alpine-Himalayan Belt underwent recent transverse shortening under the effect of collisional compression. The process was accompanied by rotation of separate microplates [23]. The orientation of the compression axis changed with time. At the first stage (Oligocene–early Miocene), the NNW orientation was predominant; during the second stage (early–middle Miocene) it was oriented in the NE direction; at the third stage (late Miocene–early Pliocene), the orientation again was NNE or N–S; meridional compression dominated during the fourth stage (middle Pliocene–Quaternary).

The geodynamic correlation is outlined between the tectonic events at the northern flanks of the Arabian Plate and the evolution of the Aden–Red Sea Rift System [20, 87, 93]. At the first stage, the rift system propagated westward and the Aden Rift pulled apart more intensely than the Red Sea Rift. The Arabian Plate correspondingly moved to the north-northwest. At the second stage, the Red Sea Rift pulled apart more intense than the Aden Rift and the Arabian Plate moved to the northeast. During the third stage, the intensity of pulling apart increased because of the breakup of the continental crust and the onset of spreading [20, 72]. Inasmuch as the breakup of the crust and spreading developed in the Aden Rift earlier than in the Red Sea Rift, the plate moved to the north-northwest. Finally, at the fourth stage, the Red Sea Rift was involved in spreading as well, and the plate began to move northward. As was shown above, similar variations of orientation of the compression axis also took place in other segments of the orogenic belt, which were not related to the drift of the Arabian Plate. It is obvious that they were controlled by more general geodynamic factors that determined the drift of Arabia among other phenomena.

In contrast to the above variations of the stress-and-strain state, the progradation of the Zagroz Foldbelt was characterized by the stress state that remained unchanged over all stages of its evolution. This is indicated by the parallel orientation of folds differing in age and the conjugated NW-trending faults. In contrast to the folds, the structural framework of the Zagros as the contemporary right-lateral Main Strike-Slip Fault is inscribed into the setting of near-meridional compression established in Pliocene–Quaternary. At that time, the Main Fault was separated from the zone of active folding by previously formed fold zones, where folding ceased. The stress fields different in rank are probably combined here: the transregional field controlled by common movement and interaction of lithospheric plates and microplates and the regional field involving only the Zagros. These may be caused by the wedge shape of the Arabian Plate, which creates compression of its northeastern margin in the process of the northward drift. A similar progradation with the same geodynamic consequences took place in the Himalayans, where the front of the maximal displacements and deformations migrated in the post-middle Miocene time from the Central Thrust Fault to the Advanced and then Frontal faults and now is shifting to the Sub-Himalayas.

From the late Eocene to the early Pliocene, the uplifts expressed in the topography arose and developed in the same tectonic zones of the belt, which underwent the strongest compression and shortening.

Such uplifts can be regarded as a result of isostatic compensation of thickening of the crust due to its compression. Differentiation of the peneplain with the formation of uplifts and intermontane basins took place also beyond the regions of collisional diastrophism expressed in folding and thrusting, e.g., in the Tien Shan and Altai. Their compression could have been induced by the microplates migrating in the course of collision. In the Central Tien Shan, this was the pressure of the Tarim microplate. Because the direction of principal compression varied during the recent stage, the uplifts differing in strike originated at different times. The growth of mountains could lag behind the onset of diastrophism, probably, owing to inertia of isostatic compensation.

The height of the mountain ranges was estimated in two ways: first, using the geologic–geomorphic method discussed in the preceding article, and second, by calculation based on correlation between rise of the crust and its deformational shortening. Artyushkov [47] proposed the following formula for this purpose:

$$\begin{aligned} \Delta\zeta_{up} &= [(\rho_m - \rho_c)/\rho_m](h_1 - h_0) \\ &= [(\rho_m - \rho_c)/\rho_m][(L_0 - L_1)/L_1]h_0, \end{aligned} \qquad (1)$$

where $\Delta\zeta_{up}$ is the value of isostatic uplift under conditions of local isostasy, m; L_0, L_1, h_0, h_1 are the initial and final widths of compressed region (L) and thickness of the crust (h) therein; ρ_m and ρ_c are the densities of the upper mantle and average density of the crust, respectively, kg/m^3. The value and rate of shortening and corresponding initial width of the region were estimated from structural–geological data, whereas the initial thickness of the crust was estimated from its thickness in the adjacent undeformed regions with similar initial characteristics of the crust (Tien Shan, Himalayas) or on the basis of geological features if the region under consideration initially differed from the adjacent territories (Pamir, Zagros).

Calculation of isostatic uplift since the Oligocene as a result of thickening of the crust by compression made by Artyushkov for the Central Tien Shan has shown that by the end of the Pliocene (onset of intense rising), the uplift reached 0.6–0.9 km [47]. This estimate is consistent with geologic–geomorphic estimates according to which the height of the uplift by the end of the Pliocene did not exceed 1.5 km; the difference in the heights of the uplifts and the surface of the basins was 1 km and the average height of the Central Tien Shan was close to 1 km, i.e., ~0.7 km higher than the height of the initial preorogenic peneplain. In other words, before intensification of mountain building, the growth of uplifts could be entirely determined by regional compression.

The characteristics of the molasse complexes and rare estimates of correlative downcuttings in the planation surfaces and pediplanes in other mountain systems of the belt lead to similar conclusions. The uplifts which arose from the Oligocene to the early Pliocene, aside from occasional local deviations, also towered above the preorogenic peneplain by not more than ~1.5 km. It is quite possible that they were created by thickening of the crust owing to compression. In the Eastern Carpathians and the Greater Caucasus, the deformational thickening of the crust in the middle–late Miocene did not result in the corresponding rise of the territory. In the Carpathians, the cause could have been related to compaction of the lower crust [60]. A similar situation probably took place in the Caucasus. The formation of the Pannonian Basin also could be caused by an increase in the lower crust density.

The transverse segmentation of the belt is clearly expressed in its contemporary structure. The eastern segment separated from the central segment by the Darwaza–Chaman Fault System is characterized by a maximum of rising that involves not only mountain systems but also most conjugated basins and plains (microplates). This feature of the eastern segment was retained through the entire recent stage, increasing with time. Cenozoic granitic magmatism is widespread here, whereas in other segments granites are not so abundant in contrast to the intense recent volcanic activity. The central segment, separated from the western one by the Dead Sea Transform and East Anatolian Fault Zone, is characterized by more differentiated topography. The mountain systems here are lower than in the east, and the Caspian Basin and Persian Bay are situated at the periphery of the belt. The belt of intense recent volcanism extends along the western margin of the segment. In the western segment, mountainous domains are combined with basins. Both rising of mountains (the Alps, the Carpathians, Middle East) and subsiding of basins were intensified here.

It would be suggested that the hypsometric features of the segments are related to different rates of their compression, or transverse shortening. It actually reaches a maximum in the eastern segment owing to substantial displacement of the Indian Plate; however, in the same segment, the highest tectonic uplift took place in the Himalayas rather than in the Punjab–Pamir arc, where shortening was highest. It is evident that differences in the intensity of vertical movements were related not only to intensity of compression but also to the features of the lithosphere determined by the tectonic history and deep geodynamics.

Pliocene–Quaternary Mountain Building

The Pliocene–Quaternary mountain building fundamentally differs from the preceding stages of the evolution of the orogenic belt not only in the higher intensity of uplifting but also in the extensiveness of the involved territories irrespective of their tectonic history. The uplifts embraced the whole of Central Asia and developed in other regions of the belt.

The intensification of Pliocene–Quaternary rising is not related to acceleration of plate movements and increase in collisional compression. On the contrary, the intensity of compression locally decreased. For example, in the Alps and the Western Carpathians, collision was completed as early as the middle Miocene, whereas the mountains began to grow in the Pliocene against the background of diminished compression. In the Greater Caucasus, the growth of uplifts accelerated in the Pliocene–Quaternary against the background of decreasing compression rate recorded in the GPS data [54] and total displacement along active faults [48]. Even in the regions where compression increased (Himalayas, Pamir, Central Tien Shan), the amplitudes of uplifting related to the thickening of the crust by compression are only a part of the total amplitude of the uplift over this time (Fig. 6). If the compression rate in the Central Tien Shan estimated from the data on Late Quaternary displacements along faults and the results of GPS measurements (5–10 mm/yr) is extrapolated over the entire phase of intense mountain building (late Pliocene–Quaternary), then it will be higher than the average compression rate in the preceding epochs (2.5–3.0 mm/yr). The isostatic uplift at this rate of crust compression estimated from formula (1) is 180–360 m, i.e., 9–18% of the increment of average height of mountain edifice of ~2 km [47]. A similar calculation of the height increment of the Himalayas and Pamir in the Pliocene and Quaternary yielded no more than 40–50%. Most intermontane basins rose also, though not so intensely, and this hardly can be a manifestation of compression. Thus, regardless of either increase or decrease of regional compression in the Pliocene and Quaternary, this factor may explain only a part of the rate of uplifting, and not everywhere. The remainder must be explained by the contribution of other factors.

Role of the Asthenosphere in Mountain Building

The transfer of lithospheric plates is caused by lateral flow of rock masses in the deeper layers of the upper mantle, primarily, in the asthenosphere, as a manifestation of multilevel mantle convection [34, 45, 50]. The Tethys originated in the Late Paleozoic to the southwest of the closed Paleozoic ocean, owing to the rifting that resulted in the breakdown of Pangea. The high activity of the Tethyan asthenosphere caused by two factors was important for the neotectonics of the Alpine–Himalayan Belt.

The first factor is related to the geodynamic impact of the Ethiopian–Afar superplume ascending from the lower mantle [65]. When estimating its role, it should be kept in mind that Eocene and Miocene volcanism is noted in all segments of the belt [25, 26, 39, 75, 83, 94]. The volcanic activity of that time was not restricted to the subduction zones at the margins of the closed basins and often bore attributes of ensialic island-arc magmatism. Grandiose multiphase granite formation continued in the eastern segment until the Miocene [12, 53, 89]. The decrease in the density of crustal material caused by generation of granites could be an additional driving force for uplifting. The lateral asthenospheric flow from the Ethiopian–Afar superplume was manifested in vigorous volcanism of the Arabian–Caucasus segment of the belt [13, 93]. The earliest occurrences of volcanic activity in Ethiopia are Eocene in age. By the end of the Miocene, the asthenospheric flow spread up to the Greater Caucasus, reflected in the progressive northward rejuvenation of volcanism. In our opinion, the implications of the superplume are not limited to this migration.

In the contemporary structure, the superplume is localized at the boundary between the African and the Somali plates and at the triple junction of both with the Arabian Plate. If a long-term stationary position of the plume is assumed, then before Eocene it was located to the east of the minor Somali Plate that occupied its present-day position, owing to the eastward drift of Africa in the process of opening of the Atlantic. The seismic tomographic data show that the superplume is a near-meridional elongated zone spreading southward almost up to Malawi [44]. It cannot be ruled out that in the past it was longer. When drifting above this zone, the Gondwana asthenosphere was enriched in hot components and the velocity of hot asthenospheric flows increased. More and more fragments had been detached from Gondwana and moved together with the Tethyan lithosphere to the northeast, where it was subducted beneath Eurasia.

The second factor related to the activity of the Tethyan asthenosphere was its enrichment in fluids. The seismic tomographic data on the subduction zones and its continental framework obtained for northeastern Asia [14] and subsequently confirmed for the Indonesian segment of the Alpine–Himalayan Belt [44] have shown that the subducted slabs only occasionally subside deeper than ~700 km. More often, at a depth of 400–700 km they pass completely or partly into the horizontal high-velocity bodies plunging beneath the continent. The origin and crystal chemical features of these bodies indicate that they contain chemically bound aqueous components [42]. Because volcanism related to the Ethiopian–Afar superplume has reached the Greater Caucasus, the subducted cold slabs were fragmented after their plunging, so that asthenospheric flows migrated through them. The destruction and dehydration of high-velocity bodies at a depth of 400–700 km led to enrichment of the asthenosphere in fluids. After closure of the Tethys, collision of the Gondwana and Eurasian plates slowed down their convergence, but the hot hydrous asthenosphere of the former Tethys penetrated beneath the entire Alpine–Himalayan Belt in Cenozoic.

The first three stages of neotectonic evolution of the belt are distinguished by (1) intense deformation; (2) intense tectonic delamination and related large-scale lateral displacements with participation of basement blocks and sheets; and (3) collisional volcanism and granite formation. These features are caused, to a great extent, by the effect of the asthenosphere. The impact of asthenospheric fluids brought about softening of the lithosphere. According to [2], this made possible intense recent deformations resulting in the arising of uplifts expressed in the topography. The fluid facilitated the tectonic delamination of the lithosphere along the zones of high gradients of physical properties and large lateral offsets. Magmatic activity was excited by the active asthenosphere, which deformed the lithosphere in the process of movement, providing local decompression, penetration of fluids and other components of the asthenosphere, forming intralithospheric, including crustal, magma sources [33]. In the opinion of Koronovsky and Demina [26], the magma sources in the lower crust and the uppermost mantle of the Caucasus region arose under the effect of heat and mass transfer and oxidation of reduced fluids supplied from deep mantle levels.

The large-scale deformation of the crustal masses from the Oligocene to early Pliocene accompanied by metamorphism and crustal magmatism gave rise to homogenization and consolidation of the Earth's crust in those domains of the belt, where these processes had not proceeded earlier, and thus prepared the next, fourth Pliocene–Quaternary stage of the neotectonic evolution. Consolidation of the crust was expressed at this stage by cessation of large-scale granite formation and localization of volcanic activity within limited zones often related to strike-slip faults [25, 73, 93, 96]. The latter became the leading form of transverse shortening of the belt, whereas the fold–nappe deformations concentrated within the basins with thick sedimentary cover (the Sub-Himalayans, Afghan–Tajik Depression, foothills of the Taurus, the Lower Zagros, periclines of the Caucasus).

The tectonically delaminated lithospheric mantle overlain by the consolidated crust was partly replaced by low-density asthenosphere [1, 2, 47], and this abruptly intensified the growth of mountain ranges. This is indicated by the lowered seismic wave velocities beneath the highest mountain systems of Central Asia (Himalayas, Tibet, Kunlun, Pamir–Hindu Kush–Karakoram region, Central and Eastern Tien Shan) [8, 9, 35, 37, 78], as direct evidence for the low-density upper mantle suggested from gravity measurements [59, 71]. Kaban [18] noted the same features in the gravity field of the Lesser Caucasus. The lowered seismic wave velocity related to the ascent of the asthenosphere was revealed beneath the Eastern Carpathians [60].

The decompaction of the lower crustal masses as a result of retrograde metamorphism under the effect of cooled Pliocene asthenospheric fluids may be the second factor of intensification of mountain growth. This factor probably became dominant in the uplift of the Greater Caucasus and the Western Tien Shan, where low-density mantle domains have not been detected except for the Elbrus magma chamber. A flow of active asthenosphere penetrated beneath both mountain systems only in the late Miocene–early Pliocene after closure of the troughs framing them in the south. Therefore, its impact remained insufficient for partial replacement of the lithospheric mantle but was able to facilitate metamorphic decompaction of lower crustal high-grade metamorphic rocks.

CONCLUSIONS

In the Cenozoic, especially since the late Eocene, different zones of the Alpine–Himalayan Belt have undergone collisional compression caused by convergence of the Gondwana plates with the Eurasian Plate. This compression was expressed in folding, thrusting of the sheets of continental crust over one another, and closure of the Neotethyan basins and related backarc seas and eventually led to the local thickening of the crust and its isostatic uplifting. The uplifts that grew in such a way, as a rule, did not exceed the hypsometric level of low and moderately high mountains. The process continued like this up to the early Pliocene and the areas occupied by uplifts became more extensive with each new tectonic phase. In other words, before the early Pliocene, the growth of mountain edifices was almost completely caused by collisional compression of the belt, although local deviations from the isostatic compensation of compression arose, being directed toward the greater amplitudes of uplift, e.g., in the eastern segment where low-density granites were formed, and to the lower amplitudes in the Eastern Carpathians and the Greater Caucasus, probably due to metamorphic compaction of the lower crust.

The isostatic uplifting of the crust thickened by compression developed further in the Pliocene–Quaternary, locally even more intensely than before, but general rising of most part of the orogenic belt was added to this process. The general rise was greater in amplitude than the contribution of the uplift caused by local thickening of the crust by compression; it did not depend on the Cenozoic history of either territory, involved not only mountain edifices but also the majority of adjacent basins, and eventually led to the formation of the contemporary mountain topography of the belt. A possible cause of this rise, unrelated to collisional compression, is the effect of the active asthenosphere of the closed Tethys, which spread beneath the orogenic belt. The isostatic reaction to the decompaction of the upper mantle was a result of partial replacement of the lithospheric mantle with the asthenospheric material, whereas the decompaction of the lower crust was caused by retrograde metamorphism under the effect of cooled asthenospheric fluids. The deep transformations also caused deepening of some basins and increased contrast of transverse segmentation of the belt.

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