The Post-Oceanic Geodynamics of the South Tien Shan Region

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[1] The subject of this paper is the structural and lithologic evolution of the earth crust in the South Tien Shan region, especially during the post-collision, platform and orogenic periods of its geologic history. The author describes the Mz-Kz intraplate and orogenic basins, including the Karakul, Zidda, Ravat, Zeravshan, and other basins, in terms of their tectonic structure, evolution, their positions in the general morphostructure of the recent orogenic belt, and its relationship with the crustal structure of the region. Particular attention is given to the problem whether the evolution of this territory in the Alpine time was or was not controlled by the previous stages of its evolution. It is concluded that the formation of the Alpine morphostructure of the region was controlled by the voluminous rock flows at different depth levels of the crust and lithosphere. The mechanisms of the material and structural reconstruction of the Paleozoic basement rocks are ranked as the processes of melange formation, voluminous cataclasis, brittle-plastic flow, and dynamic recrystallization, which had operated during the Alpine tectogenesis. The rock masses of the basement were proved to have experienced significant 3D movements during the platform and orogenic periods of the geologic history of the region discussed. *INDEX TERMS:* 1033 Geochemistry: Intra-plate processes; 1209 Geodesy and Gravity: Tectonic deformation; 3040 Marine Geology and Geophysics: Plate tectonics; *KEYWORDS:* geodynamics, orogeny, Tien Shan Region, tectonics.

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Introduction

[2] In the structural and tectonic sense, the South Tien Shan Belt (Figure 1) is a part of the Ural-Mongolian Belt (for the Paleozoic time) or a part of the Eurasian intracontinental orogenic belt (for the Alpine time). The region of recent tectonic reactivation embraces the significant crustal and lithospheric regions of Eurasia of different prehistory, such as fold and nappe belts, regions of young and old platforms, internal crystalline rock massifs, and others. The South Tien Shan Belt is located in the southwestern part of this orogenic belt. In the north and northeast it borders the structural features of the Middle Tien Shan Belt along the South Fergana and Inylchek faults, and in the south it borders the Tarim and South Gissar old crystalline massifs and the Pamir structural features located between them. At the present time the South Tien Shan Mountains are surrounded, almost on all sides, by young molasse basins.

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[3] The South Tien Shan Mountains experienced a complex and multistage geologic history which resulted in the formation of a fold-nappe structural feature, elevated by the present time to the heights of more than 5000 m above the sea level. The complexity and heterogeneity of the multistage formation of the South Tien Shan Mountains caused a great difference between the views offered for the normal mechanisms of the geotectonic evolution of these mountains and for the geodynamic processes that had operated in the region discussed. These mechanisms are the block (fold-block) mechanism of vertical movements [*Beloussov et al.,* 1984; *Rezvoi,* 1956; *Schultz,* 1979, to name but a few]; the mechanism of general horizontal compression, and the significant horizontal movements of crustal blocks [*Makarov,* 1990; *Nikonov,* 1990; *Zakharov,* 1970; *Zonenshain and Savostin,* 1979, to name but a few]. Other geologists admitted the joint effect of general compression, mantle processes, and the transformation of compression stress to a fold-block structural feature [*Chedia and Utkina,* 1990; *Kuchai,* 1981; *Makarov,* 1990]. *Artyushkov* [1978] and *Belousov et al.* [1984] suggested the leading role of mantle and asthenospheric diapirs. Substantial attention was given to the effect of the structural layering of the lithosphere in the region

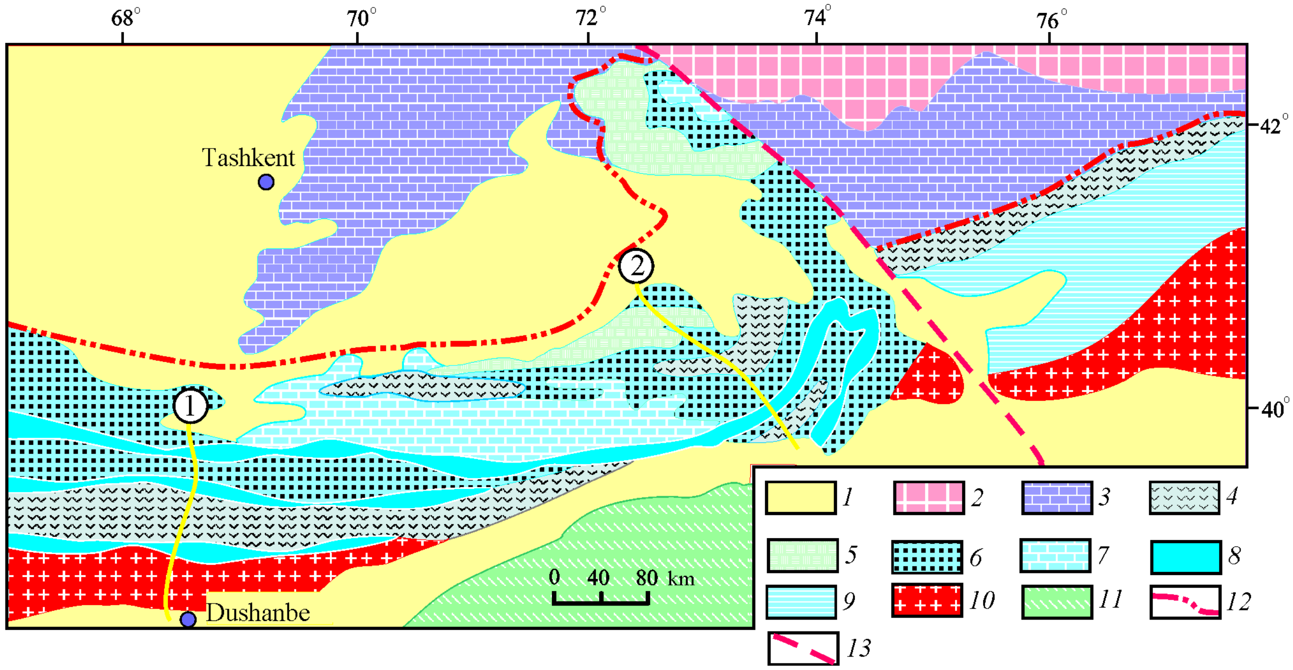

Figure 1. The schematic map of the main South Tien Shan structural elements: (1) the Mz-Kz sediments around the Tien Shan Mountains; (2) the Northern Tien Shan; (3) the Middle Tien Shan; (4–9) the Southern Tien Shan: (4) the outcropping rocks of the intrabasin volcanic (parautochthonous) ridges; (5) the outcropping allochthonous rocks of the intrabasin volcanic ridges; (6) the outcropping rocks of the relatively deep-sea troughs, volcanic high slopes, and shallow-sea bar slopes; (7) the outcrops of microcontinent sediments; (8) major collision sutures, listed in the northern direction: Karakul-Zidda, Zeravshan, Nuratau-Kurganak; (9) South Tien Shan structural features, undifferentiated; (10) Tarim, Suluterek, and South Gissar massifs; (11) Pamir structural features; (12) South Fergana ophiolite suture (Paleo-Turkestan ocean suture); (13) Talas-Fergana fault.

concerned [*Makarov,* 1990]. Views differ also concerning the inherited or independent geodynamics of the Alpine geologic history of the region [*Rezvoi,* 1956; *Yablonskaya,* 1989]. In my earlier paper [*Leonov,* 1996] I offered a view about the complex interference between different mechanisms and tectonic conditions during the formation of the modern structure of the South Tien Shan orogenic belt, although the main attention was given to the Paleozoic geologic history of the region, whereas in this paper the main attention is given to its Mz-Kz (Alpine) history, reflecting the geodynamics of the mobile belts during the post-collision, platform and orogenic, periods of their evolution.

The Modern Structure and Formation Types of the South Tien Shan Region

[4] The southern segment of the South Tien Shan orogenic belt is a nappe-type fold belt of a divergent structure, which includes structural elements of different tectonic styles and different rocks [*Leonov,* 1996] which can be classified into at least three major types of structural and formation assemblages (Figure 2). The first type is represented by the zones of antivergent structure showing the anticlinal style of folding, the axial planes of the folds flattening from the middle of the fold to its limbs, the gentle overthrusts in the marginal parts of the zone, and with well developed axial cleavage. Classified as this type are the Zeravshan-Turkestan and Turkestan-Alai zones composed of thick Lower Paleozoic and Silurian terrigenous-argillaceous-carbonate sedimentary rocks, often showing a typical flysch appearance. With the exception of orogenic granitoids, volcanic and metamorphic rocks are almost absent in these zones.

[5] The second type of the structural and formation assemblages, mapped in the Zeravshan-Gissar, Yagnob-Sugut, and Kan-Milisuy structural zones, is distinguished by the general synform structure, overthrusts and tectonic nappes, recumbent folds, flow texture and schistosity, as well as greenschist metamorphism. The synform zones are composed of Ordovician-Lower Carboniferous terrigenous, siliceous terrigenous, and siliceous-terrigenous-carbonate deposits. The siliceous terrigenous rocks often give way to the accumulation of basic volcanic rocks (tholeite basalts and the like). Also typical of these synform zones are the associations of basic volcanics and reef limestones. Associated with the synform zones are the outcrops of ultrabasic rocks and ophiolite melange. The synform zones are represented by two varieties: parauthochthonous and allochthonous ones repre-

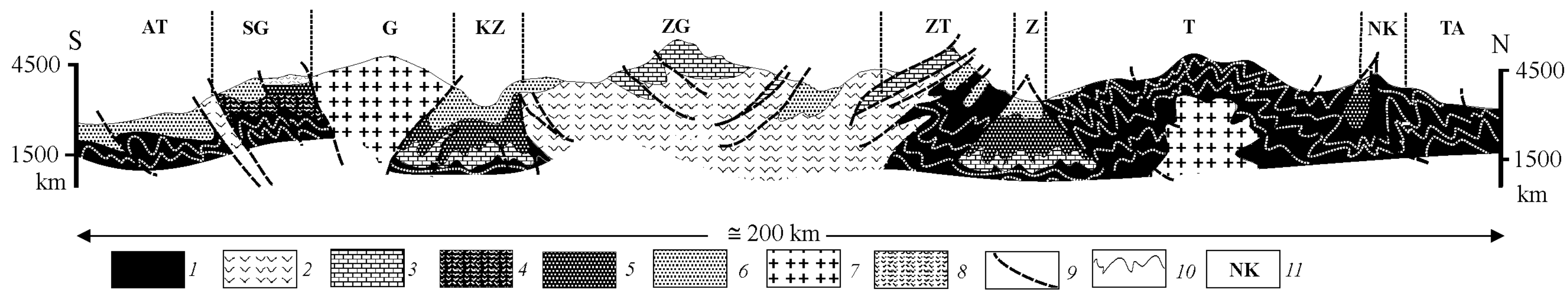

Figure 2. Schematic geological profile across the South Tien Shan along the meridian of 68°30′ (profile 1 in Figure 1): (1) the Low-Middle Paleozoic rocks, undifferentiated; (2) the Ordovician-Lower Silurian rocks, partially metamorphozed to green schists (Yagnob Schist); (3) the Devonian-Middle Carboniferous carbonate and siliceous-carbonate sediments; (4) the Lower-Middle Carboniferous volcanic rocks; (5) the Late Paleozoic, mostly flysch and molasse deposits; (6) the Mz-Kz platform-type terrigenous-carbonate and coarse-clastic molasse deposits; (7) granitoids of the Gissar batholith and of the axial zone of the Turkestan Ridge; (8) Permian acid volcanic rocks; (9) faults; (10) lithologic boundary; (11) names of the lithostructural zones: (AT) Afgan-Tajik, (SG) South-Gissar (including the Osmantala Zone), (CG) Central-Gissar, (KZ) Karakul-Zidda, (ZG) Zeravshan-Gissar (Yagnob), (ZT) Zeravshan-Turkestan, (Z) Zeravshan, (T) Turkestan, (NK) Nuratau-Kurganak, and (TA) Turkestan-Alai.

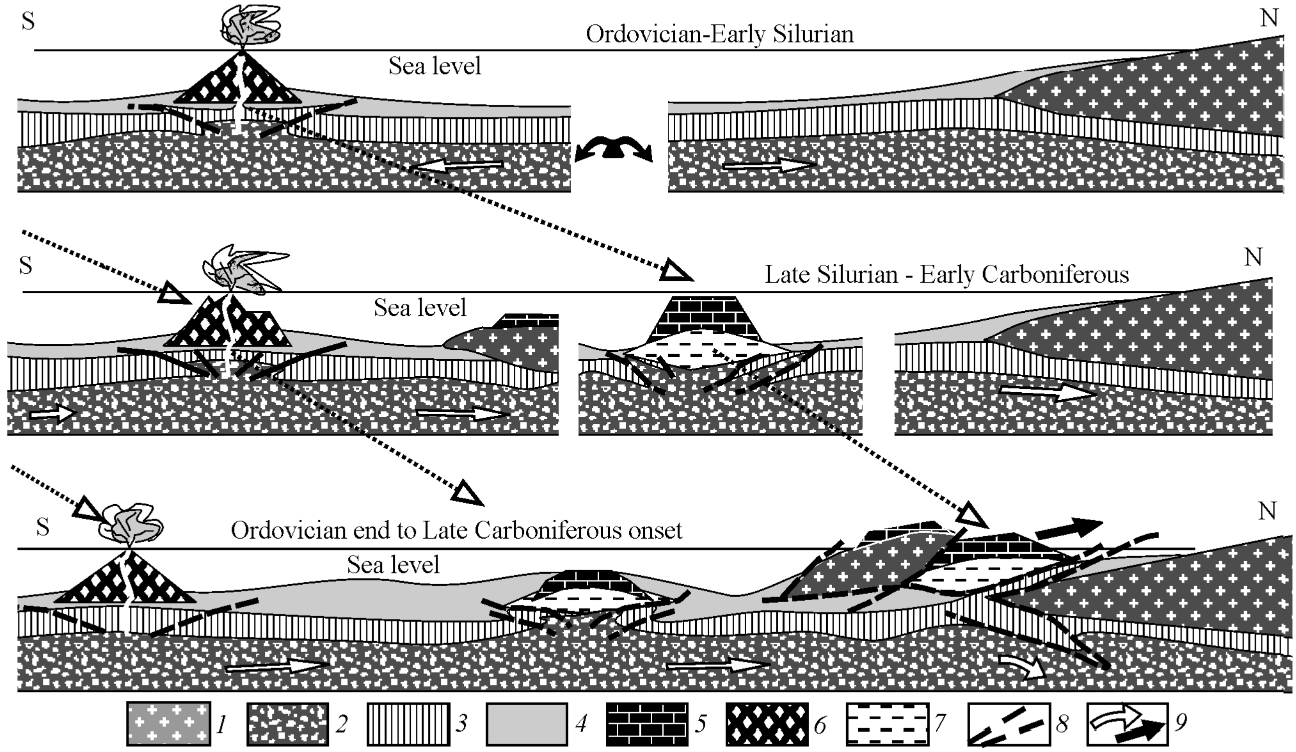

Figure 3. The geodynamic evolution of the South Tien Shan in Ordovician-Late Carboniferous time (Alai segment, see Profile 2 in Figure 1): (1) Kazakhstan-Kirgizian continental block; Turkestan paleooceanic crust: (2) ultramafic layer, (3) gabbro-basalt layer, (4) sedimentary layer, (5) carbonate deposits (reefs), (6) intrabasin volcanic underwater and island-arc ranges; (7) metamorphic volcanic rocks of intrabasin ridges; (8) faults; (9) trend of tectonic rock flow.

sented by similar rock sequences, the allochthonous rocks being the tectonically detached rock masses of the para-autochthone type.

[6] The third type is represented by narrow suture zones (Karakul-Zidda, Zeravshan, Naratau-Kurganak, Kulgedzha, to name but a few) with subvertical or steep fan-like arrangement of rock layers and structural elements. These zones show concentric dislocations and local dynamic metamorphism. No volcanic activity is recorded. Two types of suture zones have been identified. One of them is composed of relatively deep-sea carbonate and siliceous rocks (Devonian-Carboniferous), replaced upward by Late Paleozoic flysch and tectonic-gravitational mixtite bodies. The second type of the suture zones is distinguished by their terrigenous and terrigenous-carbonate deposits (Ordovician-Devonian), Carboniferous carbonate reef deposits, and Late Paleozoic molassoid block-conglomerate deposits. Associated with the suture zones are Mz-Kz basins filled with Jurassic-Eocene platform deposits and Neogene-Quaternary molasse rocks.

[7] The cross sections of the nappe-fold structural feature show the alternation of the zones of these different types. The zones of the two former (synclinorium and anticlinorium) types contact one another along the systems of head-on overthrusts, or are separated by narrow subvertical zones of the third type. Both the interior parts of the zones and their contacts show nappes and overthrusts, though no huge surface overthrusts, accompanied by the large-magnitude overlaps of the rocks of some zones over the other, are characteristic of the South Tien Shan Belt. The exception is the northern margin of the region (the piedmont of the Alai Range and the Nuratau Mts. yet even in these regions the nappe magnitudes are not greater that 10–20 km, the overthrusting showing a head-on direction.

The Prealpine History of the South Tien Shan Orogenic Belt

[8] The pre-Alpine geological history of the region was discussed in many papers and monographs (see the references in [*Leonov,* 1996]), its Paleozoic evolution being discussed briefly in this paper.

[9] Almost throughout the whole of the Paleozoic, the South Tien Shan territory was an oceanic space with a complex, dissected morphostructure (Figures 3 and 4). The early (Riphean-Early Silurian) periods of its geological history were dominated by general extension which resulted in the opening of the Turkestan paleoocean and was associated with the lateral movements of the oceanic lithosphere and with the formation of diffused spreading zones. This period

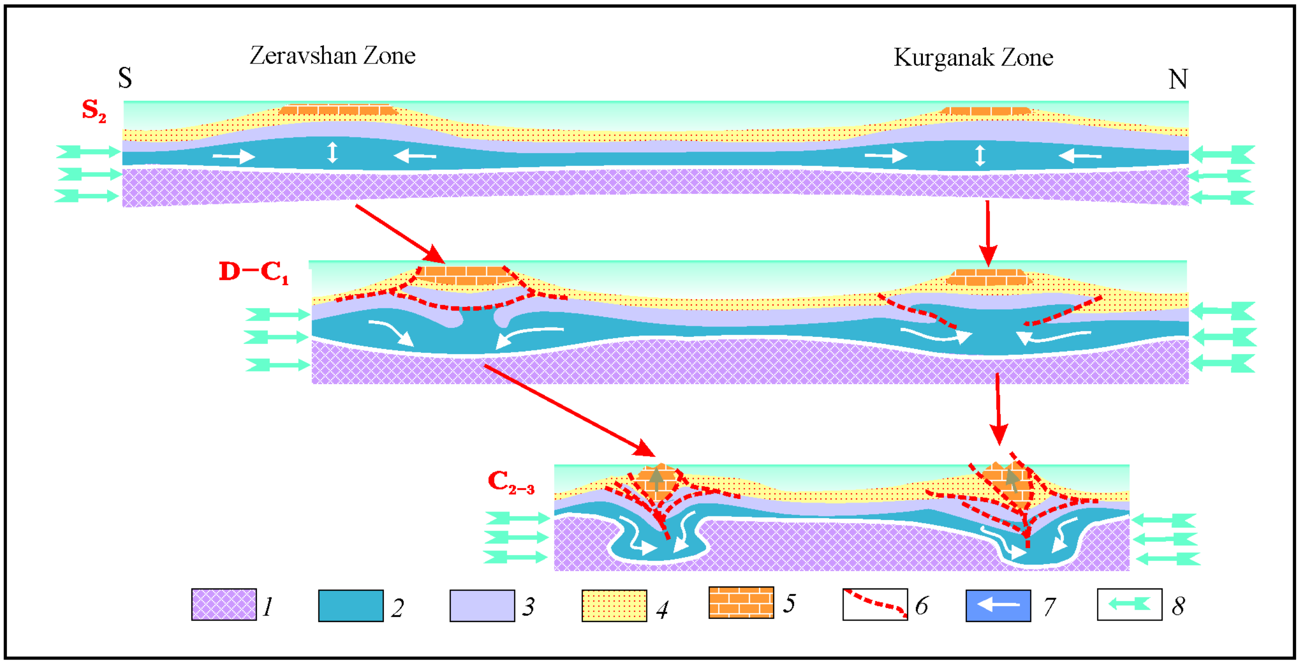

Figure 4. The geodynamic evolution of the Central Tajik segment of the South Tien Shan in Late Silurian-Carboniferous time (Profile 1 in Figure 1). Shown in this figure is the morphostructural breaking of the sea basin floor and the formation of the lock-joint subduction and ophiolite-devoid collision sutures. (1) Mantle; (2) serpentinized ultrabasic rocks; (3) gabbro-basalt; (4) sedimentary rocks; (5) deposits of intra-basin carbonate banks and platforms; (6) faults and tectonic break zones; (7) tectonic flow and rock mass movement trends; (8) general compression of the region.

of time also witnessed the formation of positive morphostructures, such as volcanic sea-floor and island-arc ridges. These morphostructures experienced the operation of the mechanism of gravitational instability, which caused the formation of specific structural-metamorphic rock assemblages with the formation, accompanied by the origin of synform structural features, plastic flow zones, and other specific features. This means that the early stage was marked by two independent geodynamic processes: the general motion of the oceanic lithosphere and the local process of the structural and lithologic reworking of the volcanic geomorphic structures under the conditions of gravitational instability.

[10] The Middle Silurian time was marked by the beginning of the morphostructural differentiation of the oceanic basin with the formation of linear uplifts (carbonate platforms and island chains with reef buildings). The origin of the morphostructural differentiation seems to have been associated with changes in the state of the oceanic lithosphere, namely, with the replacement of extension by compression. This involved the mechanism of bending instability [*Lobkovskii,* 1988], which stimulated the lateral subplastic redistribution of rock masses in the crustal layer, the formation of outflow zones (lows) and of discharge zones (highs), as well as the shaping of the respective topographic features and structural parageneses (see Figure 4).

[11] The mid-Carboniferous time witnessed the general space contraction and the successive accretion of morphostructural elements to the northern side of the paleoocean (see Figure 3). Under the compression conditions its embryonic forms grew more complex and were transformed, together with the basin deposits, to the complex nappe-fold structural feature of the region as whole. Some morphostructural elements were replaced by intrabasin collision sutures and by the zones of shallow self-closing subduction (see Figure 4). By the end of the Carboniferous to the beginning of the Permian a complex nappe-fold area was formed in the oceanic space. The main structure formation mechanism operating during that period of time was submeridional compression and space shrinking. A nappe-fold area with dissected topography was formed in the area of the former ocean. The newly formed erosion areas supplied the terrigenous material of the Late Paleozoic molasse to the remaining basins of the region. This period of time was marked by some granitoid magmatism.

[12] It follows that the Paleozoic structure of the Tien Shan Mountains was formed mainly at the expense of the transverse shortening of the space and horizontal submeridional compression. However, at the background of this geodynamic activity, associated with the paleooceanic lithosphere evolution (initially during its extension and later during its compression), the paleoocean experienced the effects of independent geodynamic effects, such as the gravitational instability of the rock masses in the intraoceanic positive morphostructures, and also the mechanism of the bending instability of the rheologically stratified lithosphere, which resulted in the initial warping of the ocean floor, and in the

formation of new intrabasin topography. These mechanisms produced specific structural assemblages which affected substantially the morphostructural evolution of the region during the Mesozoic and Cenozoic periods.

[13] To sum up, a complex heterogeneous nappe-fold structural feature was formed in the region of the former Turkestan paleoocean. Taking into account the specific structure of the paleoocean, its evolution, and present-day structure, namely, its fold-nappe structure, the structural combination of different morphostructural elements, rocks, and crust types, the presence of lenses and bands of metamorphic rocks, the significant volumes of orogenic volcanic rocks, the tectonic layering of the crust, the partial removal of the sedimentary cover, and other factors, it can be assumed that the final result of the Paleozoic evolution of the region was the formation of a thick crustal layer. However, this layer differed radically from its present-day analog, because it had not experienced any separation into a "granite-metamorphic" and a "basic" rock layer. By the end of the Paleozoic, this crustal layer seems to have been represented by some intricately bedded, heterogeneous structural melange of igneous and metamorphic rocks reworked during the long-lasting evolution of the fragments of the basic crust, the granite metamorphic rock layer of the microcontinents (such as the well known Alai microcontinent), the rocks of the Paleozoic volcanogenic sedimentary cover, and early metamorphic rocks. It is obvious that the heterogeneous combination of these rock masses must have existed in a thermodynamically and isostatically unstable state with a variable with a variable and complex internal stress field.

The Alpine History of the South Tien Shan Region

General Characteristics

[14] The main structural and lithologic manifestations of geologic events and the indicators of the geodynamic mechanisms that had operated during the plate stage of the evolution of the mobile belts are the intraplate and orogenic structural basins (discrete sedimentation basins). It is the tectonic structure of the intraplate and inter-mountain basins, their relations with the basement rocks, the specific manifestations of igneous activity and sedimentation, and the character of the secondary structural and diagenetic transformation of the volcanic-sedimentary rock cover provide information for the consolidated crust geodynamics and specific evolution during the platform and orogenic stages in the Earth mobile zones. Some information is provided by the studies of the topography and its association with the internal structure of the rock masses, and also by the data available for the crustal structure of the study region and for the thermal state of its interior.

[15] In the South Tien Shan territory the platform sediments and orogenic rocks are preserved mainly on the southern and northern slopes of these mountains and in the numerous intermontane basins arranged as individual beaded structural features, usually restricted to the suture zones, described briefly above. In this chapter we will discuss the structure of the Mz-Kz intermontane basins at the intersection of the Gissar and Alai ridges (see Figure 1 and Figure 2). Arranged in the northward direction are the following structural elements composed of Mz-Kz rocks: the Afgan-Tajik basin, the zone of the South slope of the Gissar Ridge, the Karakul-Zidda basin system, the Ravat Basin, the Zeravshan Basin, the Nuratau-Kurganak zone on the northern slope of the Turkestan Range, and the Fergana Basin.

[16] The schematic correlation of the stratigraphic rock sequences, compiled by the author of this paper, using the numerous publications available and personal observations, reflects the main characteristics of the structure, composition and age of the deposits (Figure 5). The Mz-Kz rocks of the intermountain basins are fairly monotonous, except for the substantial differences in the thicknesses and facies of the rocks, some missing layers and rock sequences, and the like.

[17] The Triassic rocks were mapped in the margins of the Afgan-Tajik Basin, where they are represented by red conglomerates, and also in the Zidda and Fan-Yagnob basins, where they are composed of the products of weathering. The Jurassic rocks are present in many basins where they consist of gray and black sandstone, clay with coal lenses and seams, conglomerates, and gravelite of continental facies (lake, swarm, and alluvial deposits). The maximum thickness of the Jurassic deposits was found in the Fan-Yagnob River area, ranging from 100 m to 300 m elsewhere.

[18] In contrast to the Jurassic rocks, the Cretaceous deposits showed a wider areal extent, varying greatly in lithology. The Lower Cretaceous rocks are represented by red continental and continental-margin deposits. They overlie transgressively all of the older sediments. One type of the continental deposits is represented mainly by pebble beds, sandstones, and alluvial-proluvial and alluvial-deltaic facies. Those of the second type are terrigenous rocks with interbeds and members of thin layers of dolomite limestone, as well as of their detrital varieties, enriched more or less with a sand material. Also present their are gray clay and siltstone interbeds. All of these rocks characterize the marginal environment of a large epicontinental sea basin. In the internal basins marine deposits may be absent, the Early Cretaceous ones being represented by red sandstones and conglomerates. The thickness of the Lower Cretaceous rocks varies from basin to basin from 0 m to 300–400 m (see Figure 5). The products of the Late Cretaceous sedimentation differ substantially from those described above. They do not contain any coarse terrigenous material, the number of red beds diminishes, whereas the volume of carbonate deposits grows notably. On the whole, the Late Cretaceous rocks are typical of a shallow epicontinental sea basin of intricate configuration, and also of lagoons, sometimes separated from the sea basin. A series of transgression and regression cycles can be traced, yet, later the sea transgression embraced almost the entire Gissar-Alai territory. The thickness of the Late Cretaceous deposits varies greatly amounting to 100–500 m in some basins.

[19] The Paleogene deposits cover all of the older rocks with a distinct erosion. They are missing in some basins.

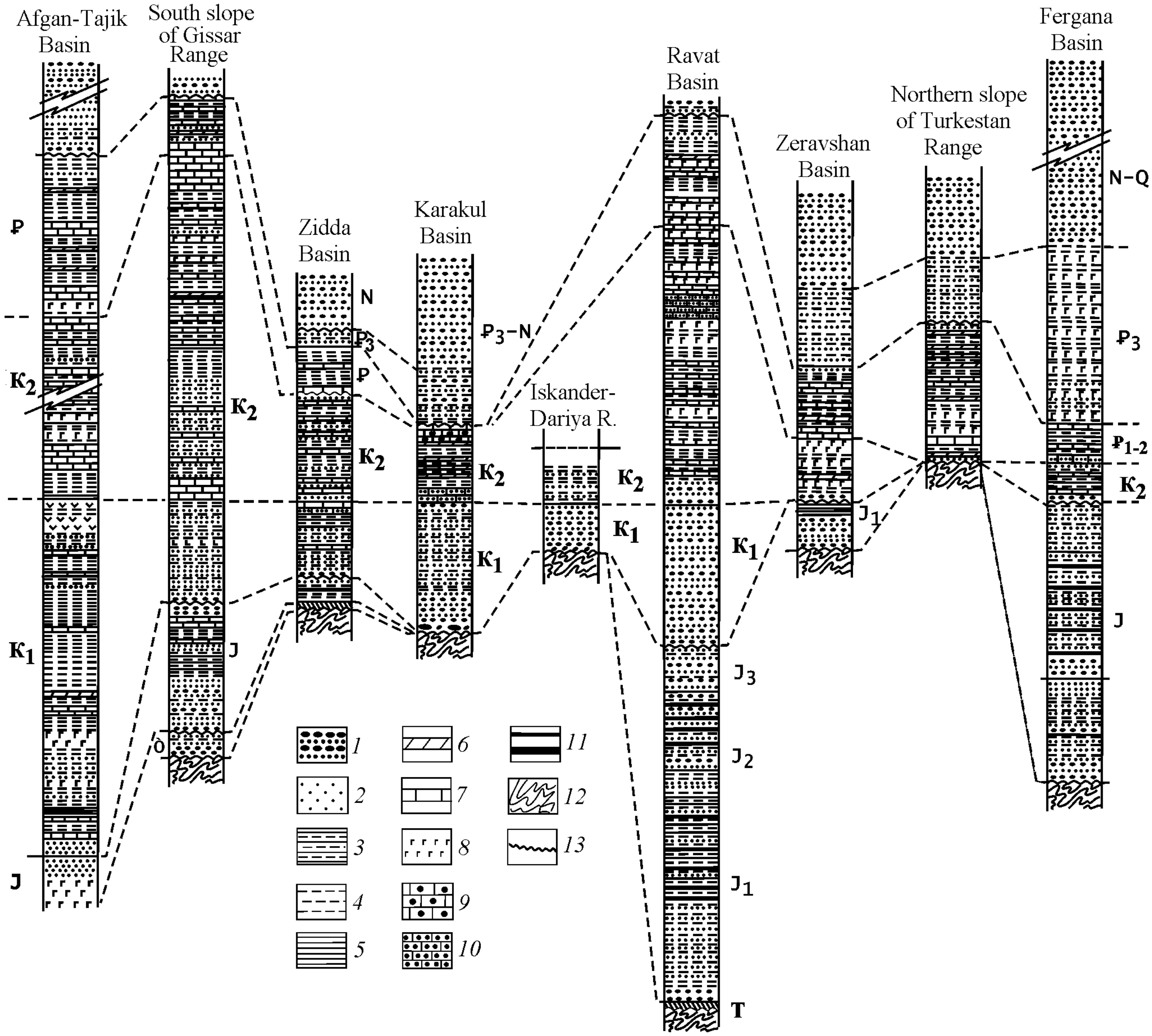

Figure 5. The schematic correlation of the stratigraphic sections across the Mz-Kz basins in the Central Tajik segment of the Southern Tien Shan region, based on the data reported in the following papers: [*Bebeshev,* 1988; *Bosov,* 1972; *Davidzon et al.,* 1982; *Dzhalilov et al.,* 1971; *Kazakov et al.,* 1985; *Luchnikov,* 1979; *Polyanskii,* 1989; *Tadzhibekov,* 1986; *Timofeev et al.,* 1985] and (Biostratigraphic Mapping of South Tajikistan, 1973): (1) gravelite and conglomerates; (2) sandstone; (3) siltstone and claystone; (4) siltstone; (5) claystone; (6) marl; (7) limestone; (8) gypsum and salt; (9) oyster limestone; (10) sandy limestone; (11) coal; (12) the rocks of the Paleozoic folded metamorphic basement; (13) the surfaces of stratigraphic and angular unconformities.

The Paleogene rock sequence is composed of dolomite, limestone (including its algal and oolitic varieties), claystone, some sand and silt material, and bituminous shale with phosphorite and gypsum intercalations. The sedimentation conditions show the oscillations of the basin floor with intermittent, short epochs of transgressions and regressions which developed under the conditions of a shallow, epiplatform sea basin. The thickness of the Paleogene deposits from basin to basin from 0 m to 200–300 m.

[20] All rocks, mentioned above, are overlain, with angular unconformity and erosion, by Oligocene-Neogene red siltstones, sandstones, gritstones, conglomerates of continental origin. The thickness of these rock sequences vary from 500 m to 800 m, amounting to 1500 m in some basins (for example, in the Magian Basin). Quaternary rocks are also widespread in many basins. These are usually sand, sandy loam, and pebble beds, being as thick as 300–500 m.

[21] Apart from the typical deposits of intermontane troughs,

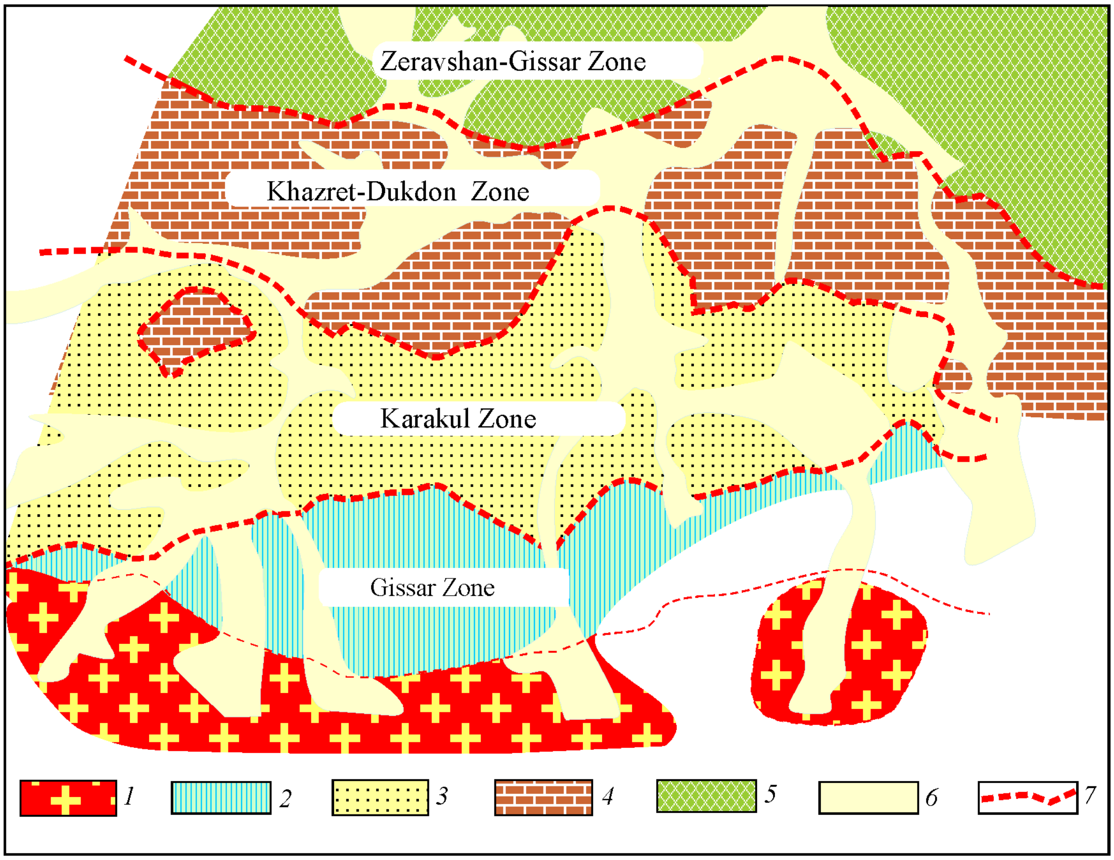

Figure 6. The structural position of the Karakul Zone (Mz-Kz depression) in the Central Tajik segment of the South Tien Shan: (1) the granitoids of the Central Gissar batholith; (2) the Middle-Late Paleozoic terrigenous and carbonate rocks of the Gissar Zone; (3) the Late Paleozoic (flysch) and Mz-Kz (terrigenous and carbonate) rocks of the Karakul Zone; (4) the Middle Paleozoic carbonate rocks of the Dukdon Zone; (5) the Lower and Middle Paleozoic rocks (metamorphic schist, shale, chert, and limestone) of the Zeravshan-Gissar Zone; (6) the lines of the overthrust boundaries of the lithostructural zones; (7) Boundary of structure formation zones.

the Quaternary sediments include trains of peculiar coarse-clastic carbonate breccias, which were classified in this study as tectonic-gravitational mixtite. These rocks will be discussed here in detail later. The Neogene-Quaternary rocks prove intensive mountain-formation activity in the Gissar-Alai region.

Tectonic Patterns of the Mz-Kz Basins

[22] **Karakul Basin.** This basin is restricted to the Karakul-Zidda tectonic zone which extends as a narrow (0–5 km) belt along the general Tien Shan trend over a distance of more than 300 km. The description of this structural feature, as well as of the Zidda basin which will be discussed here later, is based on my own data and on the results reported by many other geologists, such as, [*Kazakov et al.,* 1985; *Nesmeyanov and Barkhatov,* 1978; *Tadzhibekov,* 1986, to name but a few].

[23] The Karakul Basin is situated at the contact of two large structure-formation zones of the southern Tien Shan region (see Figures 1 and 2), namely, the Zeravshan-Gissar zone in the north and the South Gissar zone (together with the Osmantala zone) in the south. This basin is bounded by large thrust faults, being sort of pressed between them. The northern and southern thrust faults are spaced some distance apart in some areas and merge together in the others producing a system of scalloped features composed of Mesozoic and Cenozoic rocks. The areas of the convergence show the structures of "tectonic linkage". The merging of the thrust faults and the "collapse" of the zone are restricted to high hypsometric levels and are generally found in pass zones at the heights of 3500–4000 m. Along the strike and toward the valleys the faults diverge slowly allowing one to observe the internal parts of the zone. The scalloped outcrops of the Mesozoic and Cenozoic rocks are usually found in relative (yet significant) topographic lows (depressions), which separate the mountain ridges accompanied by longitudinal river valleys filled with alluvial-proluvial, mainly

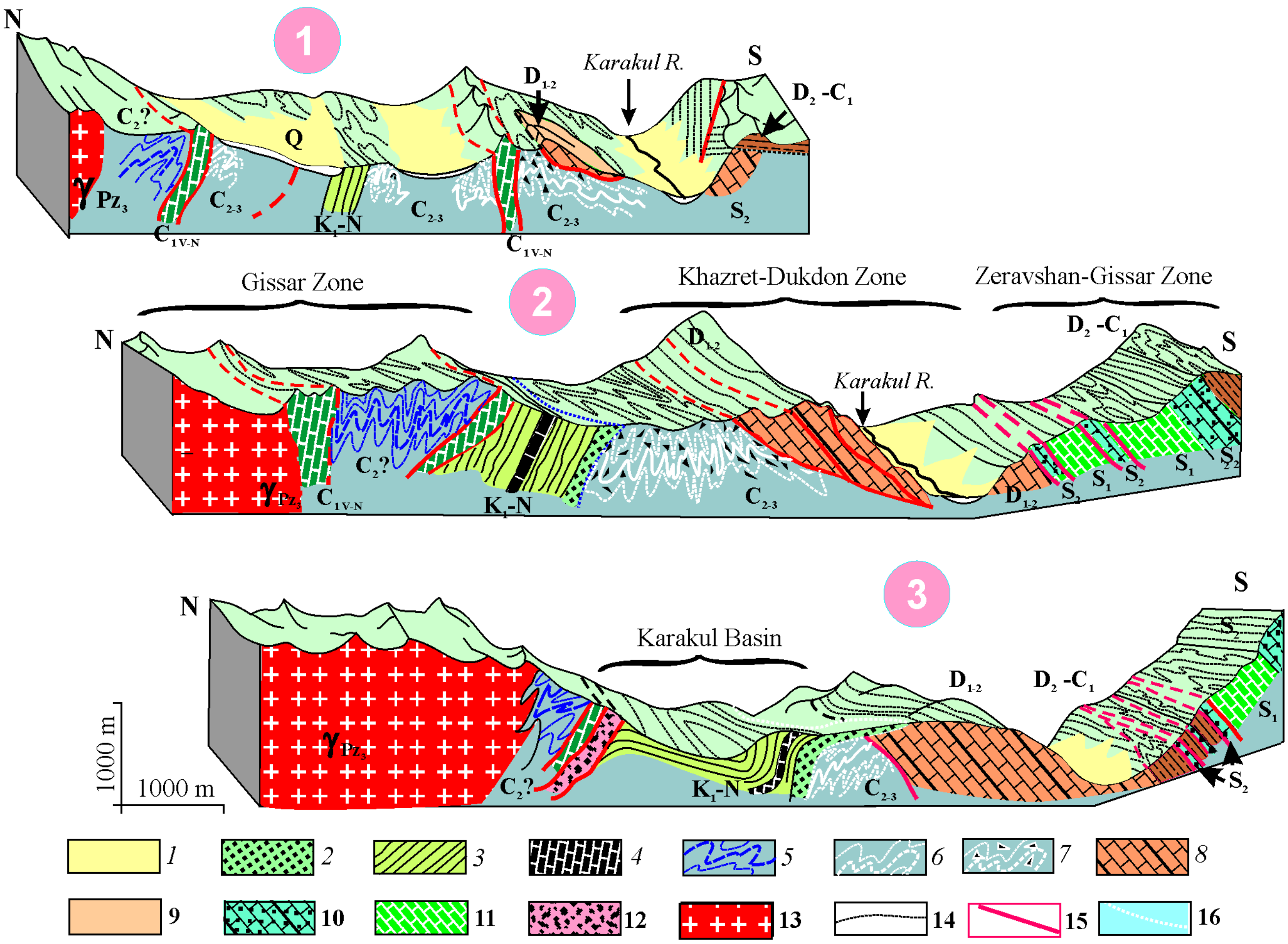

Figure 7. The tectonic structure of the Karakul Basin: (1) Quaternary deposits; (2–4) Cretaceous-Neogene deposits: (2) conglomerates and coarse sandstones, (3) claystone, siltstone, sandstone, and gypsum, (4) limestone; (5) Middle-Late Carboniferous sandstone and shale; (6) Middle-Late Carboniferous flysch; (7) Middle-Late Carboniferous flysch transformed to "sedimentary" melange; (8) Lower-Middle Devonian limestone; (9) outlier of the Lower-Middle Devonian limestone nappe; (10) Upper Silurian limestone and dolomite; (11) Lower Silurian limestone, dolomite and shale; (12) tectonic breccia; (13) Gissar Batholith granite; (14) structural line; (15) faults; (16) transgression contact. The figures in circles are the numbers of the sections.

coarse-clastic deposits. The segment discussed includes two Karakul and Zidda basins.

[24] The area studied in the Karakul River basin (Figures 6 and 7) includes four tectonic elements, differing in their internal structural patterns and in the rocks composing them: the Gissar zone, which is underlain by the granitoids of the Gissar batholith and by the Lower-Middle Paleozoic rocks surrounding it; the Karakul-Zidda Zone proper, composed of Late Paleozoic flysch and Mz-Kz rocks; the Khazret-Dukdon Zone of Late Silurian and Carboniferous carbonate sediments; and the Zeravshan-Gissar (Yagnob) Zone of polyfacies Ordovician-Carboniferous rocks and metamorphic greenschists developed after them. The general tectonic style of the area described is controlled by a series of tectonic nappes and thrusts, inclined in the opposite directions, by the complex structure of flysch deposits in the central part of the zone, and by the presence of two structural stages in the Karakul Zone: the Paleozoic and Mz-Kz ones, separated by an erosion and angular unconformity. In the modern topography of this segment of the South Tien Shan region the Karakul Zone is marked by a relative depression (Karakul Basin). This basin is pressed between the neighboring structure-formation zones which are thrust over it from the north and south and rise high in the topography.

[25] The area south of the Karakul Basin is occupied by the Gissar Zone of Lower Carboniferous crystalline limestone, the monotonous argillaceous and carbonate rocks of the Mura Formation, supposedly dated Middle Carboniferous, and by the granites of the Gissar Batholith. All of these rock units are separated from one another by faults, their contacts with the granitoids being often intrusive, and are thrust northward as a system of tectonic slabs over the de-

posits of the Karakul Basin. The surface of the overthrust is marked by a band, up to 100 m wide, of tectonic breccias and mylonite. The fault plane dip varies from 20° to 80°. The rocks of the hanging wall bordering the fault plane vary greatly up to the Gissar batholith granitoids. In some areas the Paleozoic rocks of the Gissar zone are thrust over not only the Paleozoic rocks, but also over the Mesozoic and Cenozoic deposits.

[26] Similar to the situation south of the Karakul Basin, the carbonate rocks of the Khazret-Dukdon zone were thrust over the rocks of the Karakul Zone, the former corresponding structurally to the tectonic nappes of the Devonian-Carboniferous limestone, resting as allochthon outliers on the Upper Paleozoic flysch-molasse rocks in the area of the Zidda Basin described above, the fault plane dipping northward at angles varying from 20° to 40°, being even steeper in some areas. The carbonate rock massif is broken by a series of faults into individual slabs and tectonic wedges with insignificant displacements. The western edge of the outcrop of the Devonian limestone looks (in the map) as a tectonic nappe outlier resting on the flysch. This overthrust has been dated pre-Early Cretaceous, because it was sealed by Lower Cretaceous rocks. The rocks of the Khazret-Dukdon Zone (and eastward of the Karakul Zone) are, in turn, covered tectonically by the polyfacies rocks of the Zeravshan-Gissar Zone, and by those of the Yagnob Zone in more eastern areas. The rocks of this zone compose a series of tectonic slabs and wedges of different thickness and length, which had experienced dislocation-type greenschist and epidote-amphibolite facies metamorphism. The plastic rocks are deformed to folds with their axial planes inclined in the north direction.

[27] The central part of the region is occupied by the Karakul Zone proper. Its axial segment is composed of Upper Paleozoic flysch deposits with interlayers and lenses of conglomerates, breccias, and tectonic-gravitation mixtite. The flysch sequence includes individual long and thick limestone slabs dated supposedly Carboniferous (mesoliths and synsedimentation nappes). The flysch of these zones is crumpled to small folds opened fan-like downward. In the southern and northern segments of the region the flysch rocks are broken and rolled out, the sandstone beds being broken and boudinaged. The flysch of these zones is transformed to a tectonic mixture, which is a kind of a sedimentary melange. The northern segment of the melanged flysch is restricted to an overthrust where the flysch is overthrust by the limestone of the Khazret-Dukdon type. On the whole, the flysch rocks are deformed to an anticlinal fold with the rock layers in its limbs dipping in opposite directions. However, proceding from the discovery of rock layers with reverse gradational bedding (which is masked by fine bedding) and from its analogy with the Zidda Basin (see below), it can be inferred that here we deal with a fanlike syncline whose limbs are overturned opposite to one another.

[28] The intricately dislocated and melanged flysch deposits are overlain, with an erosion and with an angular (up to 90°) and stratigraphic unconformity, by Mesozoic and Cenozoic deposits (see Figure 5). The eastern part of the basin includes Jurassic rock fragments which are overlain with a stratigraphic unconformity by the Lower Cretaceous red beds and Upper Cretaceous red beds and Upper Cretaceous terrigenous argillaceous and carbonate rocks. The thickness of these rocks is not higher than 200 m. The Cretaceous rocks are overlain with a stratigraphic unconformity by the Oligocene-Neogene rocks represented by red sandstones and conglomerates, measuring about 600 m in thickness. The Quaternary rocks are represented here by glacial, slope-type deposits of Middle Pleistocene-Holocene age.

[29] Occurring as one rock sequence, the Mesozoic and Cenozoic deposits form an asymmetric syncline with a steeply dipping, fault-cut southern limb. The syncline is complicated by second-order bends which complicate its structure. The bends of this kind are observed along its southern boundary, where small faults of the overthrust type are observed in the Neogene rocks. A pronounced bend was fond also in the northern limb of the syncline, where the Mz-Kz rocks grade to an almost horizontal position with the dip angles lower than 10°. The interbed tectonic deformations of the Mesozoic and Cenozoic rocks are almost absent, though some beds show their lateral thickness variation over insignificant distances. This is especially obvious in the carbonate and gypsum-bearing rock sequences, which can be an indirect indication of the tectonic flow of these rocks, which has been reported for the rocks of similar age and composition [*Davidzon et al.,* 1982].

[30] Worthy of mention is the position of the pre-Mesozoic unconformity plane, which corresponds to the surface of the Paleozoic basement and, at the same time, to the surface of the pre-Mesozoic (pre-Cretaceous in our case) peneplain. The block diagram shows that this plane experienced intensive deformation, which is proved by its incline, amounting locally to 70–80°, and also flexure-type bending without any breaks of its surface. This proves that this plane experienced plication, which can be observed now thanks to the deep erosion of the rocks. It should be noted that the most intensive deformation of the flysch sequences, located below the surface of the unconformity, and their transformation to the melange, are restricted to the region of the maximum deformation of the Paleozoic basement surface.

[31] To sum up, the upper and lower structural stages differ in the deformation intensity and style. The rocks of the lower stage are highly dislocated, folded, and melanged. The rocks of the upper stage are deformed to a relatively simple syncline, the limb of which is complicated by flexure-type folds, any internal deformations being absent. Yet, the plane separating the structural stages, suffered fairly intensive plicate deformation without any breaks. The bending of the pre-Mesozoic peneplane surface without any breaks was possible thanks to the fact the flysch rocks were transformed to melange, lost their coherence, and, hence, were able to experience volumetric brittle-plastic flow.

[32] **Zidda Basin.** Similar to the Karakul structural feature, this basin belongs to the Karakul-Zidda tectonic zone (see Figures 1 and 2, and also Figures 8 and 9) and is similar in many respects to the former [*Leonov,* 1995]. The Zidda Basin is situated between the water-shed areas of the Gissar Range (in the north) and the Osmantala and Sangi-Navishta ridges (in the south). This Mz-Kz basin is composed of epiplatform and orogenic rocks, ranging from

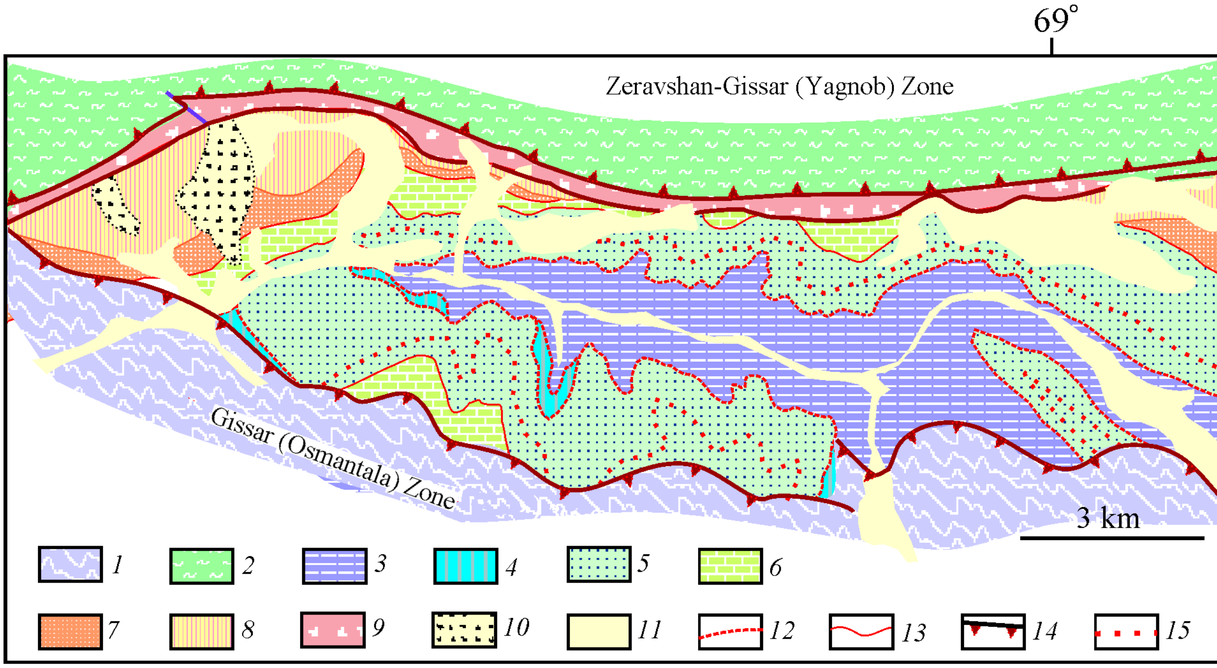

Figure 8. Schematic geological map of the Zidda Basin area: (1) Gissar (Osmantala) Zone; (2) Zeravshan-Gissar (Yagnob) Zone; (3) the Paleozoic folded metamorphic rock basement of the Karakul-Zidda structural feature; (4–8) the Mz-Kz deposits of the Zidda Basin: (4) Jurassic, (5) Lower Cretaceous, (6) Upper Cretaceous, (7) Paleocene-Eocene, (8) Oligocene-Miocene; (9) tectonic breccia; (10) trails of recent tectonic-gravitational mixtite of carbonate composition; (11) Quaternary deposits; (12–15) boundaries: (12) stratigraphic transgressive boundaries and those marking angular unconformities, (13) normal stratigraphic contacts, (14) boundaries between structural formation zones, including overthrusts; (15) intraformation markers.

the Triassic to Quaternary ones, which rest with erosion on the underlying Paleozoic rocks of the basement. Below follows their brief description [*Tadzhibekov,* 1986]. The Triassic rocks, 10–20 m thick, are represented by the products of the weathering of the Paleozoic rocks. The Jurassic rocks, up to 100 m thick, consist of sandstone and claystone with lenses and layers of coal, gravelite, and conglomerate. The Lower Cretaceous interval (250 m) is composed mainly of continental red sandstone and conglomerates. The Upper Cretaceous interval (300 m) consists of marine limestone, claystone, and sandstone. The Paleogene deposits (200 m thick) are also of shallow-sea origin, being represented by siltstone, claystone, limestone, and dolomite, the Oligocene interval being dominated by red rocks, such as sandstone, siltstone, and claystone. The Neogene rock sequence (350 m) is composed of sandstones intercalated by siltstone and sandy claystone layers at the bottom, which are overlain by a sequence of pinkish-brown sandstone, gravelite, and conglomerates of the orogenic rock complex at the top.

[33] The general regularities of the Zidda tectonic structure can be seen in Figures 8 and 9. One can see that they are similar to those of the Karakul Basin and do not call for any detailed comments. Yet, the main specific features that controlled the tectonic style of the zone in this region are offered below.

[34] The structure of the Zidda Basin in the general plan is a megasyncline filled with the complex rocks of the Paleozoic folded basement and the overlying deposits of the Mz-Kz sedimentary cover. This syncline is pressed between two large extensive faults, namely between the Major Gissar fault in the south and the Anzob reverse fault in the north. The syncline has an asymmetric structure and a different expression in the basement and in the sedimentary cover. The rocks of the basement and those of the sedimentary cover occur as two distinct structural stages of different dislocations.

[35] The structure of the Paleozoic rocks is represented in the general plan by a complex megasyncline with the limbs overturned opposite to each other. The southern limb is more overturned and, hence, has an asymmetric structure. In the south the granites of the Central Gissar batholith and the Paleozoic rocks of the Osmantala Zone are thrust over the Paleozoic rocks. In the north the Zidda basin deposits are restricted by the overthrust surface of the Zeravshan-Gissar (Yagnob) zone. The southern overturned limb of the megasyncline, composed of the flysch deposits of the Middle-Late Carboniferous Maikhura Formation, developed as an upthrust, where the Lower Paleozoic rocks are thrust over the Mesozoic and Cenozoic rocks of the sedimentary cover. The overthrust attenuates in the dip and strike directions

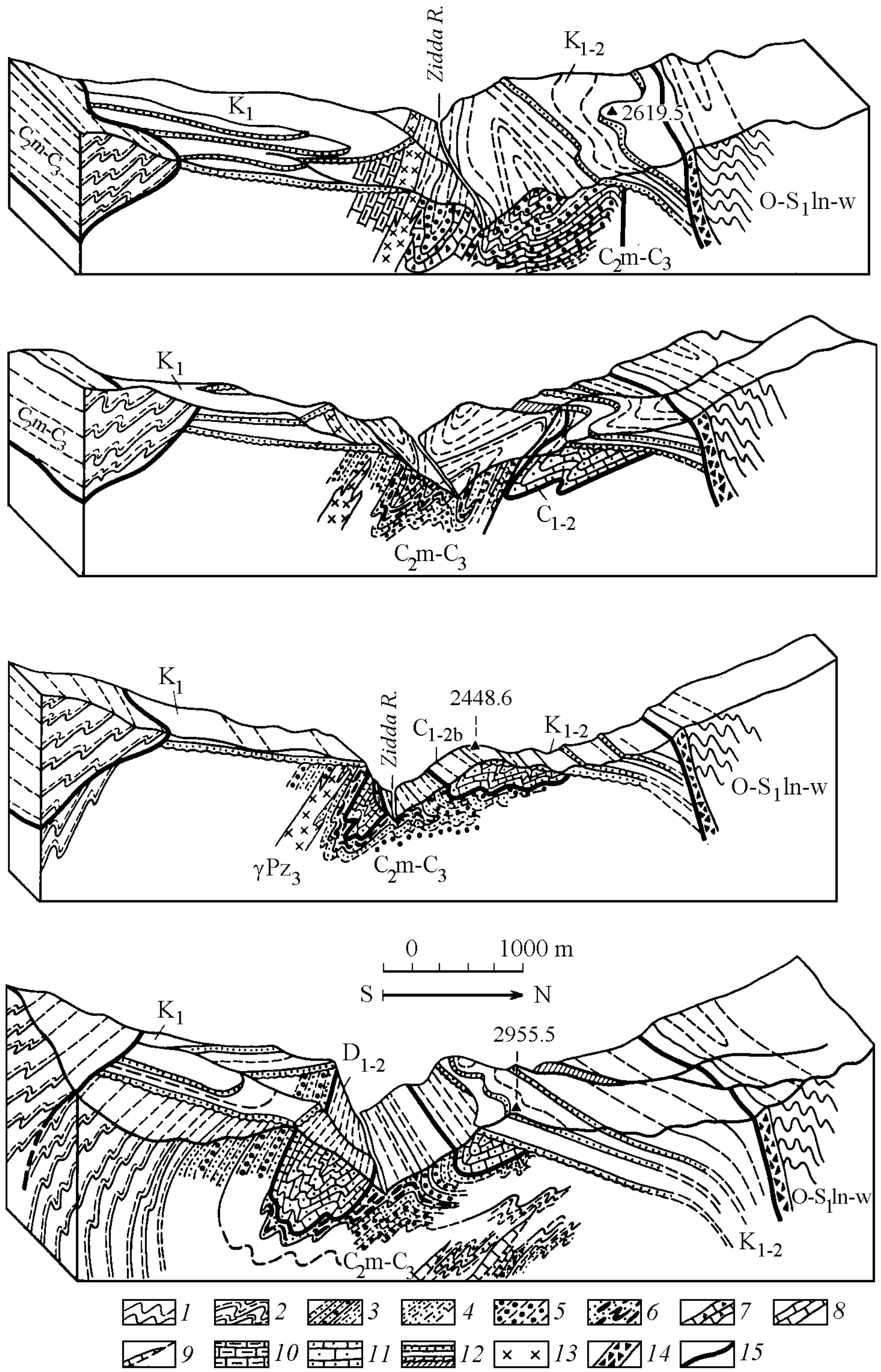

Figure 9. The tectonic structure of the Zidda Basin (its central and eastern parts): (1) the dolomite, limestone, and greenschist of the Yagnob tectonic zone (Ordovician-Silurian); (2) the shale and sandstone (flysch) of the Maikhura Formation (Middle-Late Carboniferous); (3–10) the rocks of the Zidda Formation (Moscow Stage of the Middle Carboniferous to the Late Carboniferous, possibly, the Early Permian, as suggested in [*Leonov,* 1995]: (3) shale and sandstone with landslide breccia interbeds, (4) shale, siltstone, and sandstone, (5) sandstone and conglomerates, (6) sandy-clayey rock sequences of chaotic structure (gravitational mixtite); (7) carbonate breccia; (8) interbeds of pelitomorphic ooze and algal limestone; (9) limestone slabs in sandstone and shale sequences; (10) marble; (11) Devomian and Lower-Middle Carboniferous limestone (synsedimentation tectonic nappes); (12) transgressively overlapping Mesozoic rocks and markers; (13) granodiorite and diorite; (14) tectonic crush zone; (15) overthrust and other faults.

in the sedimentary rocks of the Maikhura Formation, yet, it was mapped in the sides of the intermontane basins in the west and east.

[36] The megasyncline discussed is complicated by two asymmetric second-order synclines which are connected with each other by a compressed ridge-like anticline. The axes of these folds are oriented WNW-ESE, being somewhat oblique to the general strike of this structural feature, suggesting the presence of a shear component. The axial planes dip SSW at 60–70°. The large folds are complicated by minor folds, restricted either to the curves of the large folds or complicating the bedding of the thin-bedded arenaceous-argillaceous and carbonate rocks, this producing some intrabed disharmonic folding pattern. The cores of the synclinal folds shape the bodies of the synsedimentation tectonic nappes composed of Lower-Middle Devonian limestone.

[37] The Mz-Kz deposits, resting in the transgressive and discordant manner on the Paleozoic rocks of the folded basement, also form a syncline, though a simple one, with a gently dipping, almost horizontal, floor which shows a gentle anticlinal bend in the middle of the basin. The sides of the syncline are deformed to single tense large-magnitude folds. The central section of the basin shows that in the north the sedimentary rocks are gently dipping under the Anzob reverse fault, and in the south they dip slightly to the north, having been cut off by the Gissar fault. However, westward these two reverse faults converge in the form of a tectonic suture. In these areas, like in the Karakul basin, the sediments occur as a fan-shaped syncline with its limbs overturned opposite to each other. It should be noted, however, that in the eastern direction the Gissar fault deviates from the Mz-Kz boundary into the field of the flysch rocks of the Paleozoic basement and dies out. Yet, eastward it manifests itself again at the boundary between these two rock complexes. The deformation of the sedimentary rocks agrees with that of the pre-Mesozoic peneplane surface. It is important that the floor of the basin is deformed to a lesser extent, compared to its margins.

[38] It should be noted that, like the Karakul Basin, the Mz-Kz and recent basin coincides in space with the region underlain by the Late Paleozoic flysch, this suggesting that the formation of this structural feature had been inherited form the Paleozoic period of its evolution.

[39] **Zeravshan Basin.** This structural feature is located in the Zeravshan tectonic zone (see Figures 1 and 2, and also Figures 10 and 11). The Zeravshan Zone is located at the contact between two largest lithostructural zones of the South Tien Shan region, between the Zeravshan-Gissar Zone (in the south) and the Zeravshan-Turkestan Zone (in the north) [*Kukhtikov,* 1968; *Leonov,* 1989; *Nesmeyanov and Barkhatov,* 1978]. The Zeravshan Zone extends for many hundreds of kilometers in the latitudinal direction and has a width of less than a few kilometers. Restricted to the Zeravshan Zone are the outcrops of the Mesozoic and Cenozoic rocks. This zone is marked in the modern topography by a relative depression, where the Zeravshan R. Valley is situated, and is accompanied by a number of small basins filled with Mesozoic and Cenozoic rocks and elongated in the general Tien Shan direction.

[40] The Zeravshan-Turkestan Zone is represented in this region by a terrigenous flysch complex (up to 4000 m thick) of Llandoverian-Wenkockian age. The tectonic style of this zone is controlled by a series of size-varying asymmetric folds, overturned to the south, and overthrusts [*Rogozhin,* 1977]. The morphology of these structural features, as well as their orientation, suggest that these rock masses had been overthrust in the southern direction.

[41] The Zeravshan-Gissar Zone is a highly complicated tectonic unit. Structurally, this is a series of tectonic sheets separated by steep overthrust faults which grow flatter with depth. As follows from the general S-N orientation of the structural elements, such as, the overthrust vergence, the axial surfaces of the folds, and the positions of the corrugation and kink zones, which are opposite to those observed in the Zeravshan-Turkestan Zone, the general trend of the rock mass movement was from the south to the north, that is, opposite to the trend observed in the Zeravshan-Turkestan Zone. The Zeravshan-Turkestan and Zeravshan-Gissar zones contact the Zeravshan Zone, located between them, along large, extensive overthrust faults. These faults dip in the opposite directions, being steep in the upper segments and flattening with depth. The area where the northern and southern faults join together shows a subvertical suture where the rocks of the Zeravshan-Turkestan and Zeravshan-Gissar zones contact each other. Therefore, the Zeravshan Zone is a wedge compressed between two faults, which grows wider downward and eastward. As mentioned above, this zone corresponds to a basin filled with the Alpine deposits.

[42] Mapped in the Zeravshan Zone are two structural stages separated by an erosion surface and an abrupt (up to 90°) angular unconformity. The lower stage corresponds to the Paleozoic basement of folded metamorphic rocks, the lower, to a platform and molasse rock cover. The lower stage is composed of the Carboniferous rocks of the Vasha (C_{1-2}) and Darakhtisurkh (C_{2-3}) formations [*Cherenkov,* 1973]. *Saltovskaya* [1964] dated the rock of the Vasha Formation Namurian and those of the Darakhtisurkh Formation, Early Moscovian, which is not important in the context of this paper. The Vasha Formation is represented by a thick (up to 500 m) sequence of thin-bedded pelitomorphic limestone, polymictic sandstone, and chert, the Darakhtisurkh Formation being represented by terrigenous flysch, more than 500-meter thick, with interlayers of block breccias and less common conglomerates and gravelstones, including individual blocks and slabs (up to more than 1 km long) of limestone older than the enclosing rocks [*Rogozhin,* 1977], including some detached masses of the Vasha rocks. The deposits of the Darakhtisurkh Formation were identified as flysch including the bodies of olistostromes [*Cherenkov,* 1973] or of tectonic-gravitational mixtite [*Leonov,* 1981]. The rocks of the condensed Kshtut-Urmeta sequence, earlier ranked as a bedrock sequence [*Saltovskaya,* 1964; *Torshin,* 1970], is also a large mesolith or a detached mass of the synsedimentation nappe imbedded in the flysch deposits. This proves that we deal with a detachment of some zone, unknown in the bedrock and hidden from the observation in some suture zone.

[43] The Paleozoic rocks of the lower structural stage are deformed to steep folds compressed to isoclinal folds with

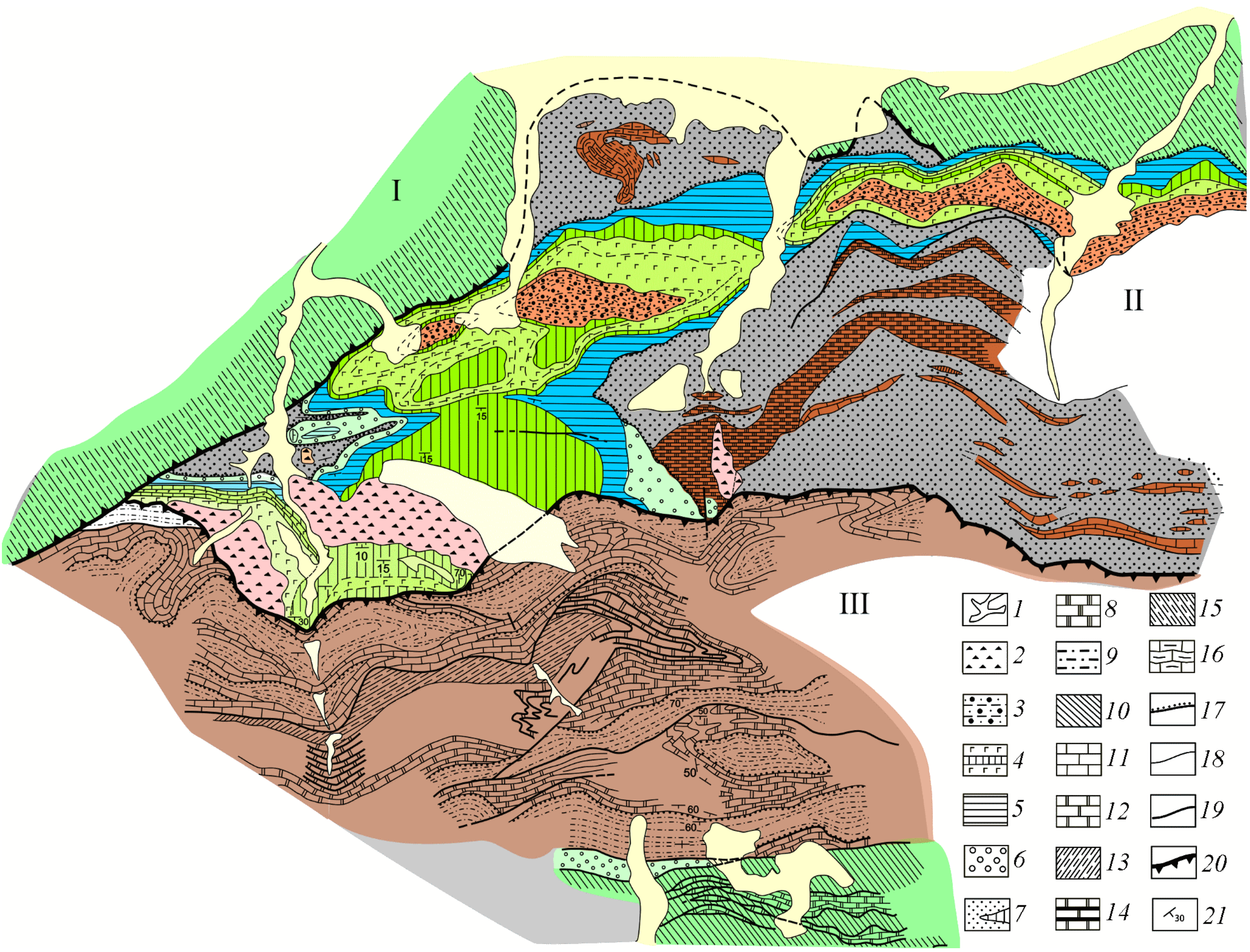

Figure 10. The geological map of the Zeravshan basin area (the basins of the Rivat, Vashan, and Madm Creeks): (1) Quaternary deposits; (2) Quaternary tectonic and gravitational mixtites; (3) Oligocene-Neogene conglomerates, sandstone, and claystone; (4) Late Cretaceous-Paleogene marl, claystone, sandstone, and gypsum (the vertical hatching shows the limestone and carbonate sandstone markers); (5) Late Cretaceous conglomerates, sandstones, and claystones; (6) Liassic conglomerates and sandstones; (7) Late Paleozoic flysch with limestone blocks and mesoliths (C_{1-2}); (8) Middle (Lower?) Carboniferous limestone and chert; (9) unmetamorphozed rocks; (10) Middle Devonian-Lower Carboniferous chert and terrigenous rocks in greenschist facies; (11) Upper Ludlovian limestone; (12) Lower Ludlovian limestone and dolomite; (13–14) Llandoverian-Wenlockian rocks: (13) mainly limestone, (14) mainly terrigenous rocks and green schists; (15) Lower Silurian sandy and shaly rocks of the Zeravshan-Turkestan Zone; (16) olistolith composed of the rocks of the Kshtut-Urmeta Type; (17) the transgressive boundary between the Mz-Kz rocks and the Paleozoic basement (pre-Mesozoic peneplane surface); (18) stratigraphic boundary; (19) faults; (20) boundaries between the structure-formation zones; (21) dip and strike.

subvertical axial planes and acute hinges. The beds stand on their heads or dip to the south (in the southern side) or to the north (in the northern side), producing a fan-shaped structure, slightly open downward. In some areas the tectonic reworking was intensive enough so that the rocks lost their bedding and stratification. The sandstone beds are boudinaged, broken, and rolled out, their argillaceous varieties being transformed to a structureless mass. This produces a chaotic structure with intricate broken antiform folds, the morphology of which suggests the tectonic flow of the rock material and the forced flow of the rock material to the cores of the anticlinal folds produced by the unconformity plane (the basement surface or the pre-Mesozoic peneplain).

[44] The upper structural stage, corresponding to the sedimentary cover, is composed of Liassic, Upper Cretaceous, Paleogene, and Miocene sediments [*Davidzon et al.,* 1982]. Like the unconformity plane, the sediments of this complex are deformed to large, conjugated synclinal and anticlinal folds, slightly overturned to the north. The thick competent conglomerate deposits of Liassic and Miocene age are

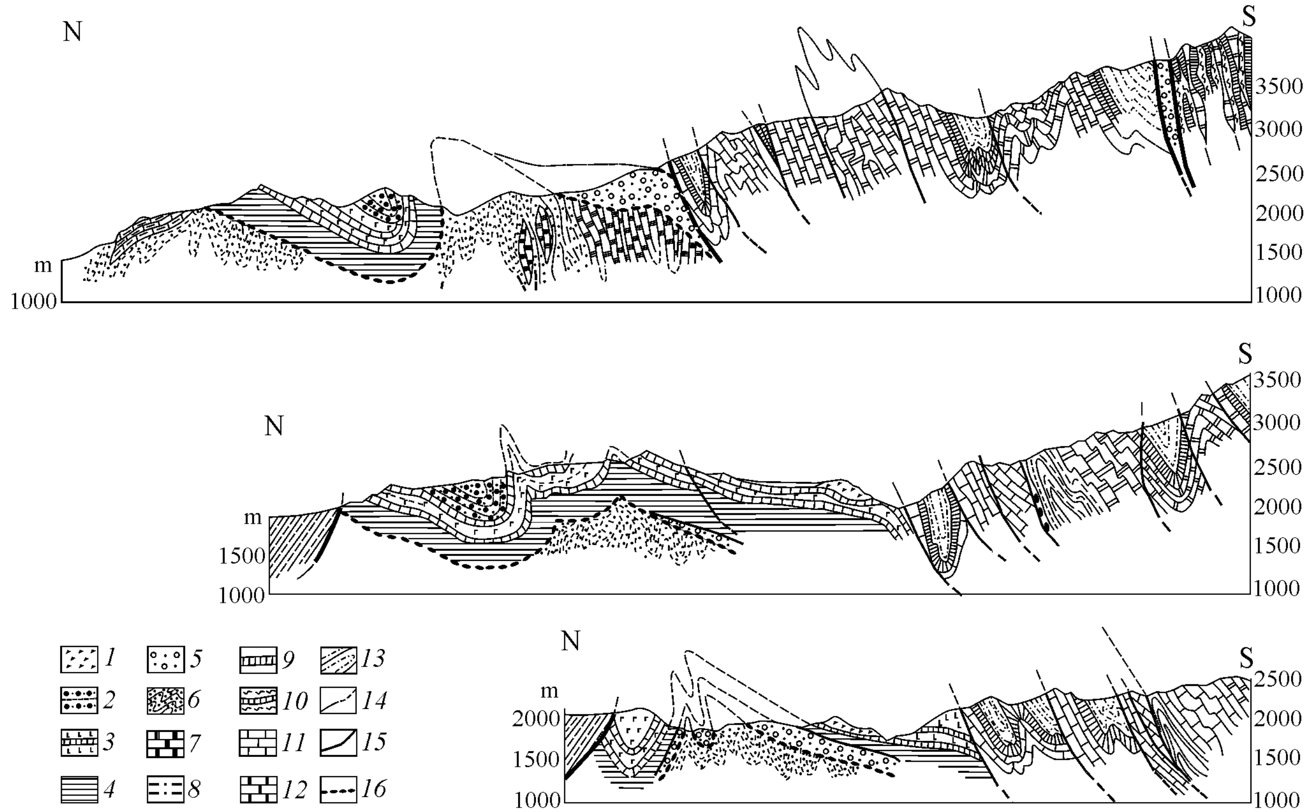

Figure 11. Geological profiles for the map shown in Figure 10: (1) modern tectono-gravitational mixtite; (2) Oligocene-Neogene red sandstone, gravelstone, and conglomerate; (3) Upper Cretaceous-Paleogene marl, claystone, sandstone, gypsum, and limestone; (4) Upper Cretaceous sandstone and claystone; (5) Jurassic conglomerate and sandstone; (6) Late Paleozoic flysch; (7) Lower-Middle Carboniferous chert and limestone; (8) Middle Devonian-Early Carboniferous sandstone and shale; (9) Middle Devonian-Early Carboniferous chert and limestone; (10) same as (9) in greenschist facies; (11) Late Ludlovian limestone; (12) Early Ludlovian limestone and dolomite; (13) Lower Silurian terrigenous rocks and greenschists; (14) faults; (15) boundaries of lithostructural zones; (16) pre-Mesozoic peneplane surface.

deformed to simple structural forms. The core of the anticlinal fold, composed of Cretaceous-Paleogene plastic claystone, limestone, and gypsum, shows a series of second-order strained carinate folds. The gypsum beds show changes in their thicknesses, associated with the sublayered tectonic flow of the plastic rocks.

[45] The Mesozoic and Cenozoic deposits are bounded in the south by an overthrust fault, yet, in the western area thrust over the younger rocks are the Paleozoic rocks of the Zeravshan-Gissar Zone, whereas in the more eastern part of the region the thrust was formed from the limb of a north-overturned fold composed of the Late Paleozoic flysch of the Zeravshan Zone itself. The tectonic pattern here is similar to the pattern observed in the Zidda Basin. The Zeravshan Zone is also bounded by an overthrust in the north, yet, in contrast to the southern area, it was not mapped everywhere, because eastward, in the Aini Village area, it extends under the transgressive Upper Cretaceous deposits.

[46] The morphology of the structural features of the Zeravshan Zone and its relationships with the morphotectonic elements of the South Tien Shan show that the tectonic pattern of this zone during the Alpine (Neogene-Quaternary) period of time was formed under the conditions of its bilateral compression between the opposite overthrusts under the conditions of the active subplastic redistribution of the rock masses. The overthrusting of the rocks masses of the neighboring structure-formation zones resulted not only in the lateral flow of the rock masses, but also in their vertical redistribution, namely, in the pressing them downward with the formation of molasse basins and in the pressing them upward with the formation of pseudodiapirs or, more likely, of protrusions. The southern overthrust was more active than the northern one. The latter was not rejuvenated in many areas during the recent time. The southern fault is bordered by almost recent and recent tectonic-gravitation mixtites which occur as extensive and thick (up to 100 m) fields bordering the front of this overthrust.

[47] To sum up, the transverse spatial reduction of the Zeravshan Zone was compensated, in addition to folding and thrusting, by the squeezing out of the rocks of the Zeravshan-Gissar and Zeravshan-Turkestan zones and by the subsidence (apparently, with lateral spreading) of the

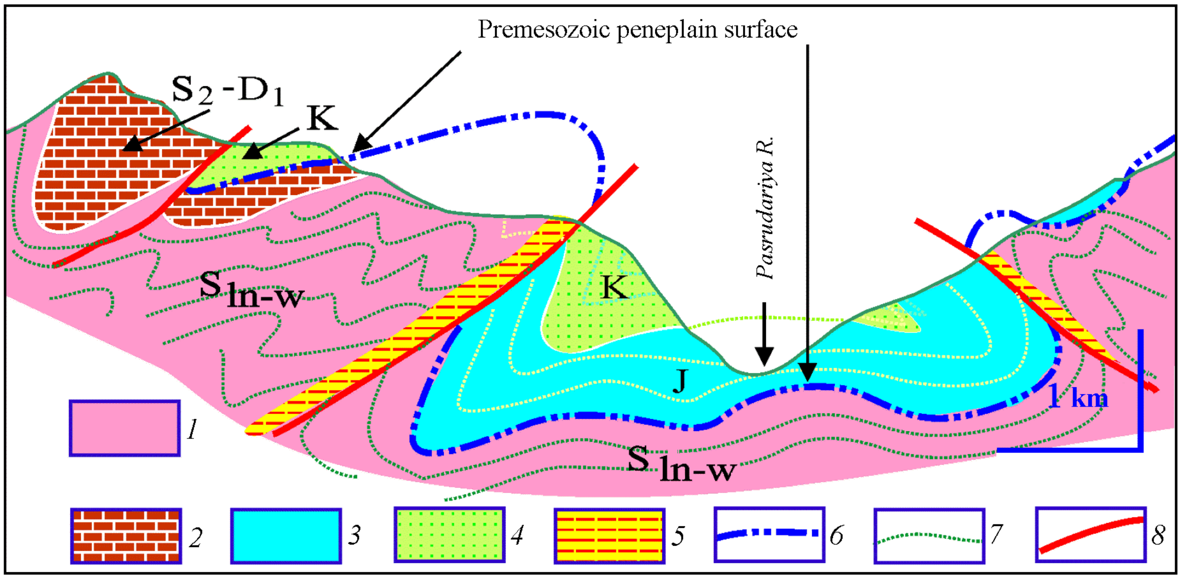

Figure 12. The profile across the Fan-Yagnob Basin (Pasrud-Dariya R. basin): (1) Ordovician-Lower Silurian metamorphic greenschist; (2) Upper Silurian-Lower Devonian limestone and marble; (3) Jurassic sandstone, conglomerate, claystone, and coal; (4) Lower Cretaceous red conglomerate, sandstone, siltstone, and claystone; (5) zones of mylonite developed after greenschist; (6) pre-Mesozoic peneplain surface sealed by Mesozoic deposits; (7) structural lines; (8) faults.

Zeravshan rocks proper. In the upper horizons the overthrusts grow steeper and converge to produce a subvertical suture. Broadening downward in a fan geometry, the deposits of the Zeravshan Zone dip under the allochthon masses of the neighboring structure-formation zones.

[48] The movements responsible for the structural reconstruction of the sedimentary cover modified the configuration of the basement surface the bends of which became conformable with the folds in the sedimentary cover. Moreover, the intricately folded rocks of the sedimentary cover were reworked repeatedly, lost their coherence, and were transformed to a structurally complex tectonic mixture, known as tectonic mixtite or as sedimentary melange. The loss of coherence as a result of melange formation provided for the 3D mobility of the basement rocks and for the curvature of the basement surface without any breaks. The melange formation takes place in this case in areas of most intensive form changes. The tectonomixtites compose the cores of the anticlines, as well as diapir-like and protrusive bodies. Similar relationships are characteristic of the other Southern Tien Shan zones (see below).

[49] It should be noted that the mantle-type accumulations of tectonic-gravitational carbonate mixtite were mapped not only in the Zeravshan Zone but also in the Zidda Zone along the overthrusts limiting this zone in the south. They occur as a coarse chaotic breccia up to 100 m thick. The area covered by these "mantles" is about 10 km^2. The rock fragments are poorly rounded or not rounded at all. There is no indications of the rock layering. Both the fragments and the cement of these rocks are composed exclusively of the limestone and dolomite of the Silurian rocks composing the hanging wall of the overthrust. Earlier, proceeding from the lithology of their rock fragments, containing Silurian fauna remains, and from their resting on the Mesozoic deposits, these rocks were interpreted as tectonic nappes. Along with some other indications, the formation of such breccias suggests the modern activity of the nappe-thrust structural features located along the sides of the intermontane basins.

[50] **Some additional evidence on the structure of the intermontane basins.** Earlier, using the example of the fairly detailed description of the Karakul, Zidda, and Zeravshan basins, I described the main structural patterns and some features of the paleotectonic evolution of the negative Mz-Kz structural features of the intermountain basins in the Zeravshan-Gissar mountainous region. Here I describe briefly some of my observations and facts that are important for the further discussion of the general geodynamic evolution of the region.

[51] As mentioned above, many intermontane Alpine basins coincide in area with the Late Paleozoic flysch troughs or, to be more exact, with the modern areas of flysch development. Such patterns are observed in the Karakul-Zidda, Zeravshan, and Nuratau-Kurganak zones. These regular patterns can be interpreted as the evidence of their evolution inherited from their Paleozoic history, which seems to be a real fact. At the same time some of the young basins originated outside of the flysch areas.

[52] In particular, the system of the Fan-Yagnob basins originated and evolved in the area underlain by the Ordovician-Silurian metamorphic schists. This system is similar in many respects to those described above, yet it provides some additional information (Figure 12). In this area the Mesozoic and Cenozoic rocks compose a very thick sequence and show

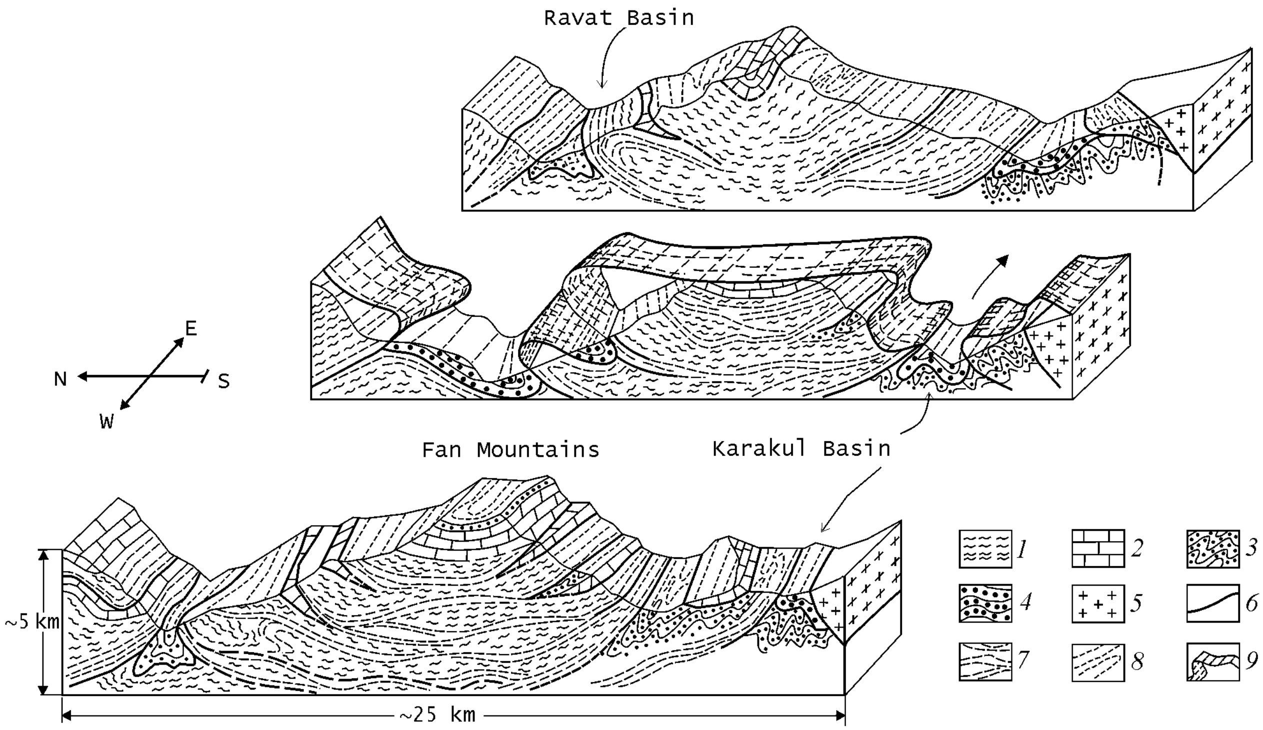

Figure 13. Deformation of the pre-Mesozoic peneplain surface in the area of the Fan Mountains: (1) Ordovician-Silurian metamorphic schists; (2) Late Silurian-Early Devonian limestone and marble; (3) Middle Devonian-Early Carboniferous terrigenous siliceous-carbonate and flyschoid deposits and Late Paleozoic flysch; (4) Mz-Kz conglomerate, sandstone, gypsum, claystone, limestone, and coal; (5) Central Gissar Batholith granitoids; (6) faults; (7) the zone of high schistosity and mylonitization; (8) structural and bedding lines; (9) pre-Mesozoic peneplain.

high similarity to those described above (see Figure 5), except that they are much thicker here. One can clearly see the deformation of the pre-Mesozoic peneplain, the surface of which shows its large plication forms. The floors of the basins (pre-Mesozoic peneplane surface) are deformed to synclinal folds with the slightly wavy gently dipping floors and the limbs overturned opposite to one another. The limbs of the folds are cut off by overthrust faults, along which the Mz-Kz deposits are overlain tectonically by Paleozoic metamorphic schists. The overthrusts are inclined in opposite directions with the fault planes dipping at the angles of 30°–60°. The overlapping magnitude is not more than 1–2 km. Outside of the area of sedimentary rocks the thrust faults can not be traced, obviously because they die out. This is especially well seen in the northern side of the basin, where the young sedimentary rocks rest on the Paleozoic ones.

[53] The morphology of the general structural pattern of the Fan Mountains (Figure 13) and the specific deformation of the zone bordering the overthrusts (both in the Paleozoic and Mz-Kz rocks) suggest that the formation of the overturned asymmetric (and symmetric) synclines had been associated with differential movements along the old plastic flow zone regenerated in Recent time. As a result of the plastic redistribution of the Paleozoic rock masses, the basin margins were pressed closer to one another from one or both sides with the formation at the boundaries of thrust-type or nappe structural features of small amplitude. The rock material flowed in the lateral direction away from the Fan Mountain massif to its periphery. The orientation of the general rock mass movement, recorded in the structural pattern of the Paleozoic rocks and in the deformation pattern of the sediments, shows that the lateral redistribution of the rock material was associated with the vertical unilateral pressure, which in this case might have been caused only by the weight of the overlying rocks. The real possibility of this mechanism was produced by the presence of the intricately dissected high-mountain topography causing the gravitational instability of the rock masses.

[54] In other words, the intensive plication folding of the pre-Mesozoic peneplain without its breaks (see Figure 13) operated there at the expense of the volumetric mobility of the rocks of the metamorphic basement, provided by the differential brittle-plastic flow of the Paleozoic rocks [*Leonov,* 1991, 1993].

[55] Another system of young intermountain basins is restricted to the Nuratau-Kurganak suture zone between the Zeravshan-Turkestan and Turkestan-Alai structural zones (see Figures 1 and 2). The shapes of the basins and the types of their sedimentary rocks are similar to those described above. Also recorded here is the plication-type deformation of the pre-Mesozoic peneplain, its asymmetric development, and the presence of the "sedimentary melange"

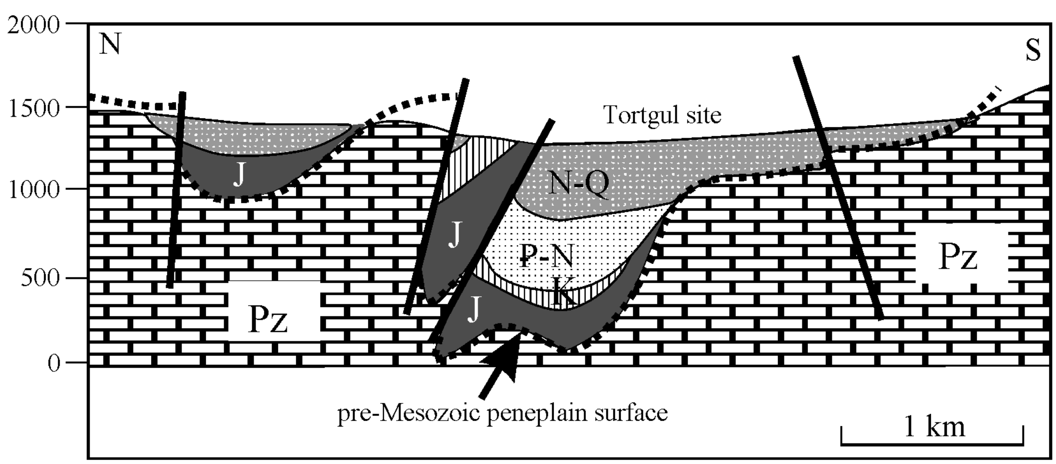

Figure 14. A profile across one of the basins restricted to the Kurganak suture zone, after [*Nesmeyanov and Barkhatov,* 1978]. See the text for explanations.

developed form the Paleozoic rocks and forming pseudoanticlines or protrusions (Figures 14 and 15), cutting the sedimentary rocks. The Nuratau-Kurganak Zone differs from the other zones by its shear movements and lateral plastic flow expressed both in the Paleozoic and Mz-Kz rocks.

General Patterns of the Formation of the Alpine South Tien Shan Region

The Triassic-Early Eocene Platform-Type History of the Region

[56] In spite of the very scarce information on the lithologic and petrologic history of the rock masses for the Early Triassic to the Early Eocene, some assumptions can be proposed here. During the Late Permian to the Early-Middle Triassic the territory of the South Tien Shan region existed as a platform with the formation of a peneplain and areal weathering crust (see Figures 6 and 7). The time period, when the platform was highly elevated, witnessed the formation of numerous explosion pipes which brought the rocks of deep crustal origin to the ground surface, indicative of the significant heterogeneity of the deep-seated rocks of the region (The Earth Crust of Uzbekistan, 1974). The formation of explosion pipes and the high elevation of the territory might have been associated with the overstrain of the lower layers of the heterogeneous crustal layer, and also with the processes of gravitational, lithologic, and dynamic equilibration, which resulted in the redistribution and transformation of the rock material with the formation of a "granite-metamorphic" and a "basalt" layer.

[57] During the Late Triassic and Early and Middle Jurassic, the platform conditions continued to exist, this time period being marked by some tectonic reactivation, by the differentiation of the topography, by the folding of the pre-Mesozoic peneplain, and by the formation of a system of troughs and low highs. The troughs were filled mainly with alluvial and swamp-lake deposits. Some basins experienced local crumpling. This reactivation was partially associated with the tectonic events in the neighboring Pamir-Hindu Kush region [*Nikonov,* 1990; *Shcherba,* 1990]. The fact that the crust of this region was highly responsive to the tectonic events that had taken place in the Tethys region suggests its relative mobility and incomplete consolidation. However, these processes were not associated with the general compression between the "rigid" blocks, because the formation of some troughs was accompanied in some cases by some extension, which is proved by the presence of quartz porphyry flows and extrusions in the mountains surrounding the Fergana region.

[58] During the Late Jurassic and the Early and Late Cretaceous, the South Tien Shan platform continued to react to the phases of the Alpine tectogenesis operating in the more southern region [*Chedia and Utkina,* 1990], yet the effect of the latter diminished with time. During the second half of the Cretaceous to the end of the Eocene, the subsidence became insignificant, and the whole territory concerned experienced the accumulation of thin continental carbonate, argilaceous and gypsic rocks, as well as of the lagoonal and shallow-sea sediments of insignificant thickness. Some of the erosion zones remained, yet, judging by the lithology and grain size of the resoling rocks, the topography was relatively low, except for some benches and scarps [*Shcherba,* 1990]. The platform conditions with chemogenous carbonate clay sedimentation continued to exist to the end of the Eocene. During that time (Late Jurassic to Eocene) the territory concerned experienced the alternation of the epochs of relative isostatic and geodynamic equilibrium (for example, the epoch of pre-late Cretaceous peneplain development) and the epochs of insignificant tectonic reactivation. These periods were marked by the tectonic deformation of the pre-Mesozoic peneplain surface and by the formation of extensive narrow lows and highs which might have been associated with the lateral redistribution of the rock material in the basement, namely, with its flow from

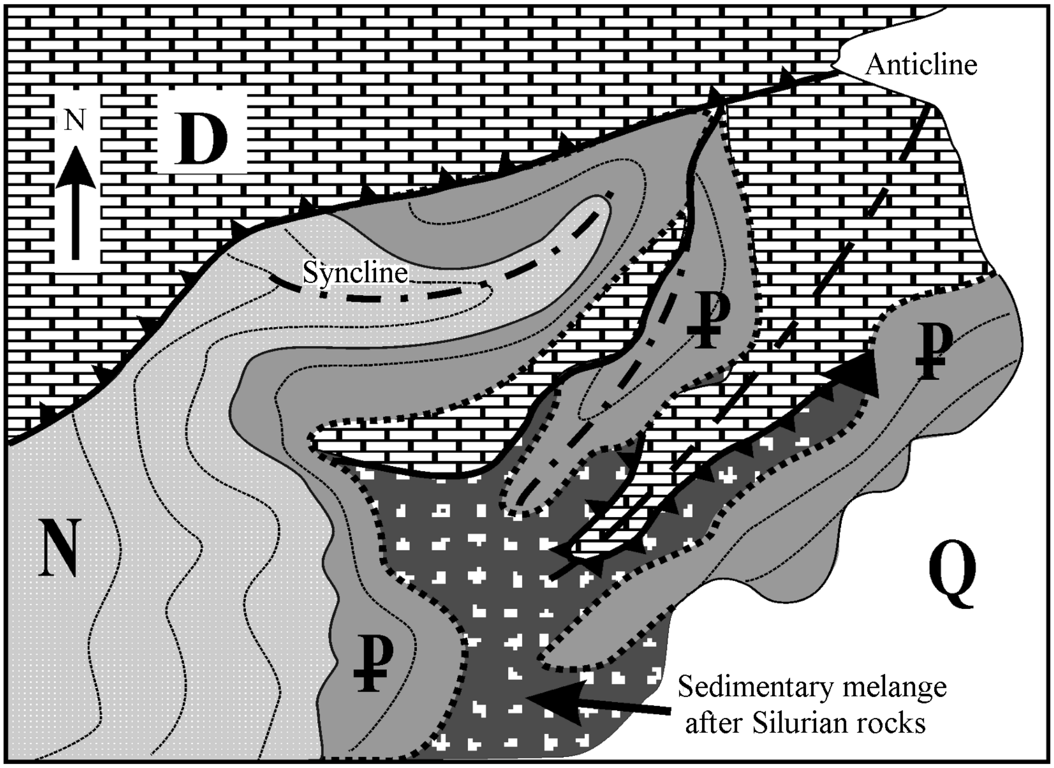

Figure 15. The "sedimentary" melange (tectonic mixtite) after the Silurian arenaceous-argillaceous rocks in the Shakhristan-Leilyak Basin restricted to the Kurganak suture zone. See the text.

the descending areas to the elevated zones of the tectonic topography following the mechanism of bending instability [*Lobkovskii,* 1988]. The voluminous rock mass flow during that period of time is proved by the origin and development of a system of lows and highs, at least up to the middle of the Miocene, as plicated structural features deformed by faults only in some scarce areas complicated by faults. By that time the magnitude of the tectonic topography, measured using the surface of the pre-Mesozoic peneplain was as high as 3–6 km.

The End of the Eocene to the Present Time: Intracontinental Orogenic Belt

[59] A new stage, namely, the stage of the tectonic reactivation and the transformation of some segment of the epi-Paleozoic platform to an intracontinental orogenic belt began, according to some authors [*Mossakovskii et al.,* 1993], at the end of the Eocene, according to other writers [*Nikonov,* 1990], during the Late Oligocene. The geodynamic interpretations of this region for its neotectonic evolution vary greatly, reflecting, as has been mentioned at the beginning of this paper, the total spectrum of potential views. However, by the present time many new data have been accumulated, which allow one to improve substantially the model available for the evolution of the South Tien Shan Belt in the course of its orogenic history (Figure 16).

[60] During the Oligocene-Early Miocene the conditions of this region were close to the platform ones, with the formation of some regional unconformities, but never intermitted by any folding events. This period was distinguished by the general low-magnitude uplifts and by the formation of the gentle folds of the basement with a large curvature radius.

A Model of the Alpine South Tien Shan Geodynamics

[61] The generalization of the data reported above and some new data allow one to discuss a model for the Alpine geodynamics of the South Tien Shan region. This model was based on the analysis of the data available in the literature: see a list of more than 100 references in [*Leonov,* 1993], and on the data obtained by the author of this paper. This model (Figures 17 and 18) was based on four groups of data: conceptual, model, geophysical, and geological data.

[62] **Conceptual data.** The model proposed in this paper is based on the recognition of the rheologic and structural layering of the lithosphere and on the internal 3D mobility of the basement, the theoretical possibility and factologic substantiation of which were discussed in my numerous publications and were discussed, in part, in the previous sections of this paper. As applied to the South Tien Shan region, this approach allows one to find the satisfactory so-

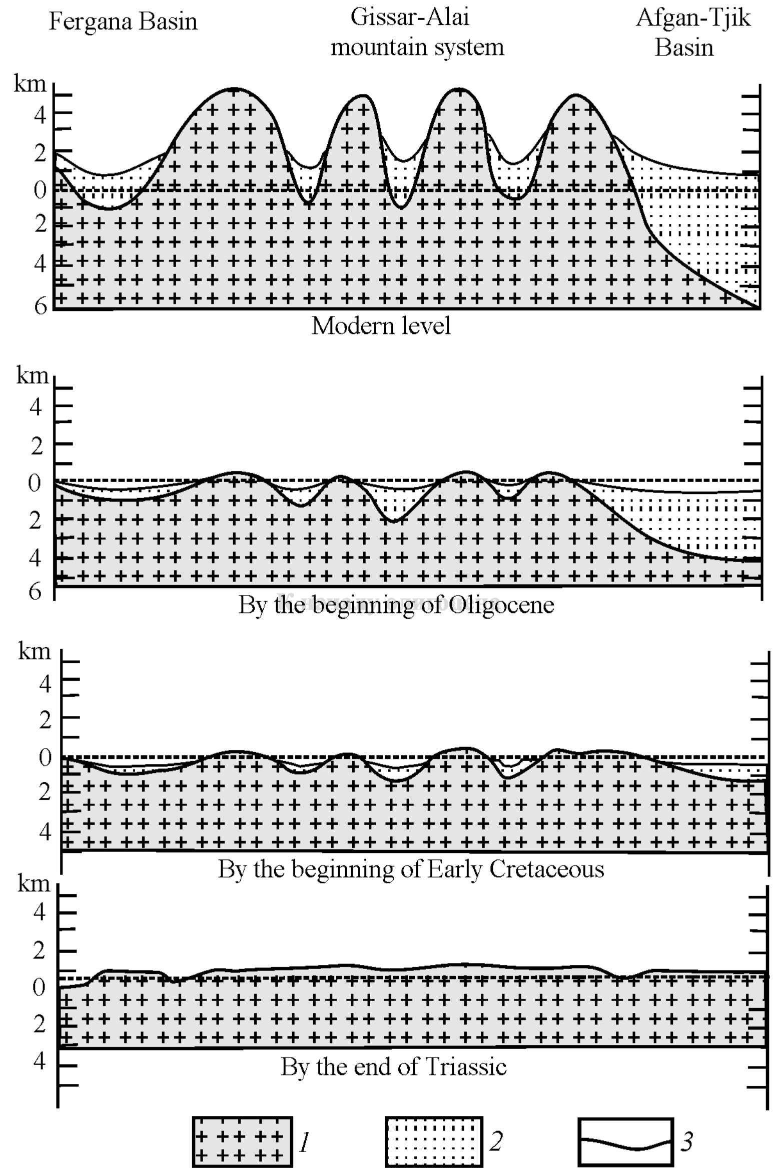

Figure 16. Schematic map of the evolution of the Mz-Kz basins and the deformation of the pre-Mesozoic South Tien Shan peneplain, based on the same data as Figure 5, using the data reported in [*Bogdanova,* 1972; *Chedia and Utkina,* 1990; *Kazakov et al.,* 1985; *Kostenko,* 1964; *Krestnikov,* 1962; *Lukina,* 1977; *Makarov,* 1990; *Nesmeyanov and Barkhatov,* 1978; *Nikonov,* 1990; *Sadybakasov,* 1990; *Shultz,* 1979; and others]. (1) Paleozoic folded-metomophic basement; (2) Deposits of Mz-Kz basins; (3) pre-Mesozoic peneplain surface.

lution of some problems of the modern structural pattern of the region and of its origin and evolution [*Makarov,* 1990].

[63] **The data provided by numerical and physical models.** The physical and mathematical modeling confirmed the existence of the layered lithosphere and the rock material mobility in the basements of the platforms, see, for example, [*Leonov,* 1972, 1976; *Lobkovskii,* 1988]. In this study I used the model of the curving-type instability of the rheologically stratified lithosphere [*Lobkovskii,* 1988], which reflects the mechanism of the formation of relatively plastic rock layers in the asthenosphere, and the ability of the rock material to flow in its solid state with the formation of the regions of the outflow, laminar flow, and pumping of the rock material. This process has a wave character. The length of the wave propagating from one pumping area to another has a value controlled by the thickness of the deformed layer. Note that the wave pattern of the pumping zone and outflow zone distribution is illustrated well by the location pattern of the Mz-Kz basins with the distances between their axes being 20–25 km (Figure 17). This process develops under the conditions of unstationary lithostatic pressure pro-

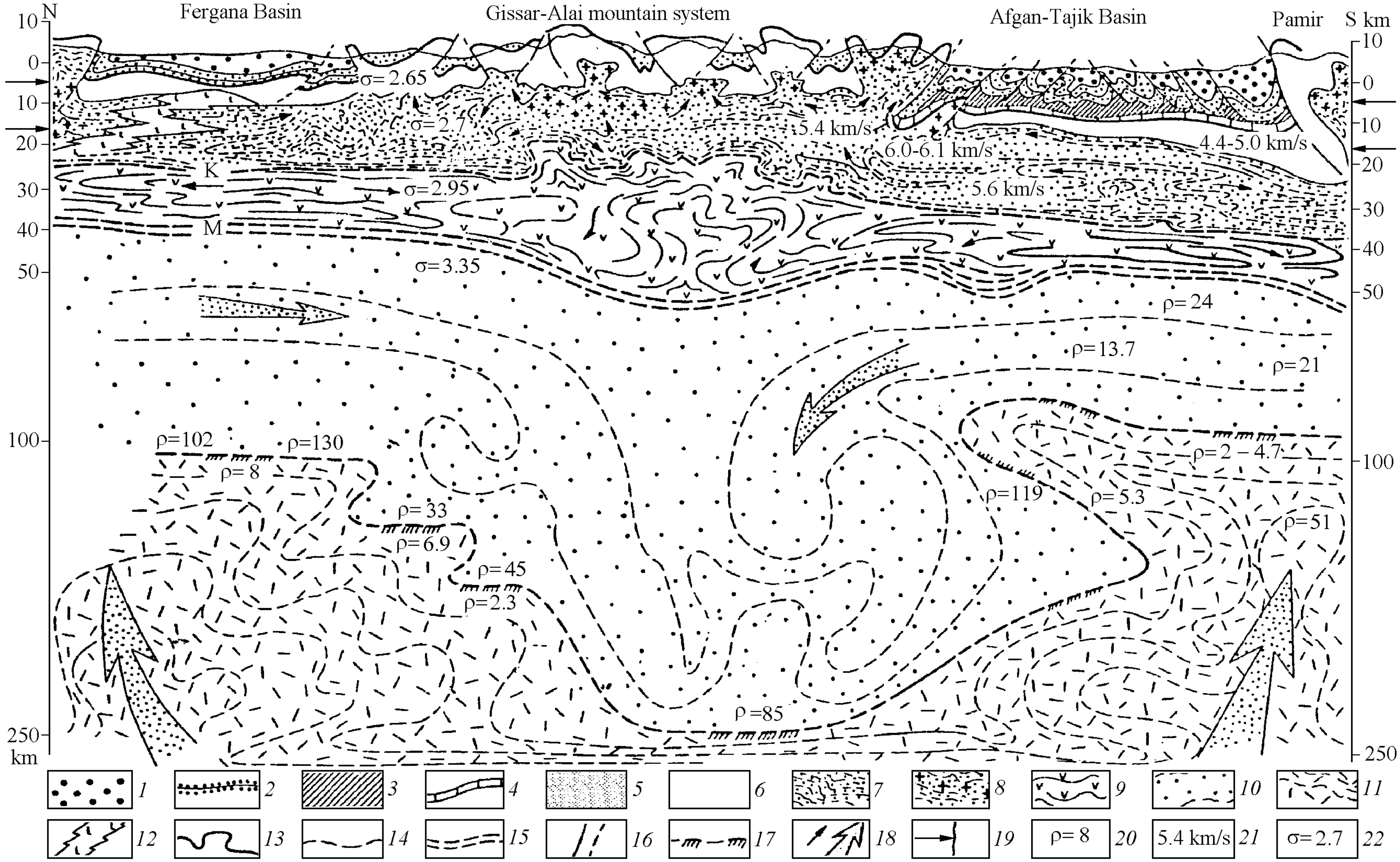

Figure 17. The geological and geophysical profile reflecting the modern structure, crustal structure, and Alpine geodynamics of the South Tien Shan region. This model was derived using the data published in the literature (more than 40 references given in [*Leonov*, 1993]), and my own data. (1) Neogene-Quaternary molasse; (2) the Mesozoic-Paleogene deposits of the Fergana and Afgan-Tajik basins; (3) salt-bearing rocks; (4) subsalt rocks; (5) Mz-Kz deposits in the internal basins of the Gissar-Alai mountain system; (6) Paleozoic deposits of the sedimentary-metamorphic rock complex that had accumulated by the end of the Paleozoic; (7) "granite-metamorphic" (upper crust) layer; (8) potential regions of melt origin, inferred from geothermal and gravity data (with Paleozoic granites found at shallow depths); (9) lower-crust "basic" rock layer; (10) relatively cold upper mantle; (11) relatively hot low-density mantle (asthenospheric layer); (12) potential trace of the Paleozoic marginal subduction zone; (13) the surface of the pre-Mesozoic basement (pre-Mesozoic peneplane); (14) Conrad discontinuity; (15) Moho discontinuity; (16) overthrusts and faults in the top of the crustal layer; (17) the conventional boundary between the low-conductivity (cold) and high-conductivity (hot and low-density) mantle (the cross-hatched areas denote the boundary zones recorded by magnetotelluric sounding); (18) tectonic flow trends; (19) general subhorizontal compression stress; (20–22) the values of resistivity (20), seismic velocity (21), and density (22).

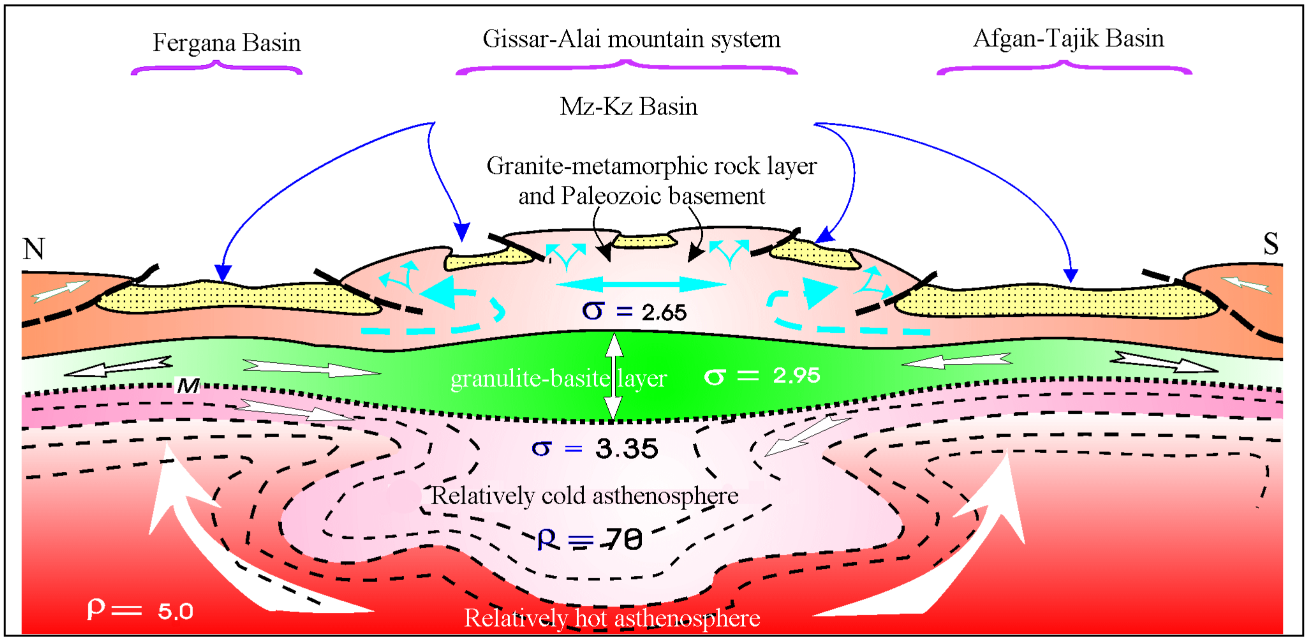

Figure 18. The schematic map of the modern geodynamics of the Gissar-Alai mountain system (the schematic generalized version of Figure 17). See the text for the explanations.

duced by the shear stress. This pattern was proved by the experimental and field data available for the origin of tectonic flows [*Luk'yanov,* 1991; *Miller,* 1982; *Patalakha et al.,* 1995]. Of great importance if the model of cascade convection [*Pushcharovskii et al.,* 1989] which allows one, as mentioned by *Chekunov* [1991], to reconstruct the kinematics of deep-seated processes, primarily, the processes of the distribution and transformation of tectonic movements at different depth levels.

[64] **Geophysical data.** There are numerous data for the structure of the region, and many structural features of the Earth crust in the South Tien Shan region can be interpreted with high confidence. Presented below are some data which are important in terms of the problem discussed.

[65] The crustal layer and the upper mantle were found to contain heterogeneities of different sizes, which are reflected in gravity anomalies and also in the complex combinations of rock volumes with different seismic wave velocities and varying electric conductivity. A great number of wave guides have been recorded in the upper and lower crustal layers. In the upper crust they occur as broken, not extensive rock volumes. Also recorded were the mushroom-shaped rolls of high-velocity rocks over the low-velocity ones. An extensive thick waveguide was recorded between the upper "seismic" layer of the crust and the lower aseismic one in a depth interval of about 24–38 km. The crustal layer has a lower thickness under the Fergana and Afgan-Tajik basins, being thicker under the South Tien Shan Rise at the expense of the bulging of the lower crustal ("basite") layer and, to a lesser extent, of the upper ("granite") layer.

[66] The thinning of the crust at the transition from the mountains to the intermontane basins owed its origin to the rising of the subcrustal material and to some descending of all interface surfaces. The Conrad boundary is not recorded under the folded mountains. The upper boundary of the basement in the Fergana and Afgan-Tajik basins shows a gentle, subhorizontal position. Some segments of the Moho boundary show some regions of the "crust-mantle mixture" with seismic velocities of 7.4 km s^{-1} to 7.7 km s^{-1}. The Moho surface shows narrow (15–20 km) and extensive flexure-type synform flexures with relatively steep angles. They are marked by negative gravity anomalies and by the suture zones of concentrated deformation and Alpine-type depressions, such as the Zeravshan, Karakul-Zidda, and other depressions. It is likely that these synforms are the traces of the Late Paleozoic zones of lock-joint subduction.

[67] The South Tien Shan area shows a regional gravity low situated inside of an extensive Central Asia low, the latter being associated with the low-density rocks of the mantle at depths of about 200 km. The varying density of the upper mantle is also marked by its varying electric conductivity. Magnetotelluric data suggest some volumes of some hot, low-density mantle material under the Fergana Basin (in its northern part) and under the Afgan-Tajik Basin, which has been proved by geothermal data. In the territory discussed in this paper, namely, under the large Afgan-Tajik and Fergana basins, the asthenosphere resides at the depths lower than under the Gissar-Alai Mountains. Seismic activity is restricted mainly to the upper crustal layers, being restricted to the depth interval of 10, 10–20, and 30–35 km.

[68] To sum up, the geophysical data correlate with the conceptual and model ones and confirm the layering of the crust and its lateral heterogeneity, suggesting some internal 3D mobility of its deep rock masses. The seismic and seismological data prove that the earthquake sources are con-

centrated along some critical depth levels, namely, along the tops and bottoms of the seismic wave guides, in the vicinity of the granite layer surface. In the Afgan-Tajik Basin, the earthquake foci coincide with the folded surface of the subsalt rock complex. Such levels are interpreted as the surface of lateral gliding and disharmonic breaking [*Bekker et al.,* 1988]. The mobile state of the crustal material seems to be proved also by the mapping of "scintillation"-type boundaries. The analysis of the time sections showed that the recording and not recording of these boundaries in time and space are associated with changes in the state of the rock materials. Also established was the heterogeneity of the upper mantle and a change in the position of the asthenospheric layer in space. The uplifted volumes of the high-conductivity and low-velocity mantle are interpreted as the diapirs of the relatively heated, low-density mantle material. Changes in the mantle structure, as we pass from one large outcropping structural feature to another, prove the mantle control of the morphostructural change of the region. At the same time, the structural features of hercynian age are often traced using gravity data only to the depths of 10–15 km and are not directly reflected in the morphostructures of the surface of the underlying granite-gneiss basement and "basalt" layer, this reflecting the disharmony between the structural styles of the different crustal layers, on the one hand, and the "consumption" of the folded metamorphic rocks of the Paleozoic basement by the processes of volumetric metamorphism and granitization.

[69] **Geological data.** By the present time a huge volume of data was collected for the Alpine geodynamics of the Southern Tien Shan and its surroundings (see the references in [*Leonov,* 1993, 1994], which will be used, along with own data, below. Here, I will mainly deal with two groups of geological facts which will allow me to discuss the mechanisms of the formation of the modern structure of the region, which have been described above, namely: (1) the dislocations of the upper surface of the Paleozoic basement and (2) the specific structural and material reworking of the folded and metamorphic rocks of the Paleozoic basement during the Alpine tectonic activity.

[70] The rocks of this region can be subdivided into two structural levels, the lower Paleozoic stage and the upper Mz-Kz stage, which are separated by the surface of angular (up to 90°) and stratigraphic unconformity. The consolidation of the Paleozoic rocks was associated with the hercynian tectogenesis. By the beginning of the Mesozoic Era, the South Tien Shan region was a slightly hilly continental plain. The Mesozoic-Paleogene time was marked by a qualitatively new, platform-type evolution period which was marked by the formation of the sedimentary cover composed of terrigenous-carbonate-gypsum shallow-sea marine, lagoonal, and continental deposits. From the end of the Eocene, or from the end of the Oligocene, as suggested by other geologists, the South Tien Shan region experienced its orogenic evolution with formation of mountains in the Neogene-Quaternary period of time. The surface of the pre-Mesozoic peneplane was deformed during its platform history and especially during the period of its recent reactivation. It should be noted that the surfaces of the pre-Mesozoic and younger (pre-Late Cretaceous and pre-Neogene) rocks are deformed almost conformably, therefore, below we will discuss merely the total deformation of the pre-Mesozoic peneplane.

[71] The South Tien Shan Mountains have the form of a megadome bordering the large negative structural features, such as, the Fergana and Afgan-Tajik basins. These structural features are treated here as the first-order basement folds or as megasynclines and megaanticlines. The basement surface is deformed differently in these large basins and in the Gissar-Alai Mountains. As follows from the results of the geophysical measurements, the basement surface producing the floors of the basins is fairly flat, being deformed merely by gentle bends with dip angles of 15° to 20°. Closer to the mountains, the pre-Mesozoic surface is more deformed, showing folds with the steep, overturned internal (mountain facing) limbs which are often cut by overthrusts. The vergence of the fore-mountain folds and overthrusts faces north in the southern side of the Fergana Basin, and south, in the northern side of the Afgan-Tajik Basin, the total structural pattern being divergent. The surface of the pre-Mesozoic peneplain shows a more complex deformation in these mountains. The background of this barrier-shaped (anticlinal) uplift shows a series of secondary-order bends of synclinal and anticlinal forms.

[72] The synclinal features are represented in the modern topography by basins with the preserved deposits of the Mz-Kz sedimentary cover. The floors of the synclines can be locally dislocated. The surface of the basement (the pre-Mesozoic peneplain), forming the floors of the basins, is usually poorly dislocated or undislocated at all. The marginal parts of the basins are more intensely deformed. The dip angles of the pre-Mesozoic peneplain are as high as 90°, the peneplain being often overturned or cut off by overthrusts with magnitudes of a few hundred meters to 2–3 km and larger in some areas.

[73] The basins have been classified as monovergent, divergent or convergent [*Sadybakasov,* 1990]. The monovergent basins are characteristic of the uplift limbs or of the zones of the contacts with the basins. The convergent forms are characteristic of the flanks and internal areas of the mountains. The divergent forms are restricted to the central part of the Gissar-Alai mountains, where they are uplifted highly and are almost undeformed (see Figure 13). To sum up, the form of the basins and their structural symmetry or asymmetry reflect the centrifugal movements of the rock masses from the axis of the mountains toward the depressions. In some of their segments the convergent basins are pressed between the opposite overthrusts with the formation of tectonic-suture structural features which were classified and mapped by *Luk'yanov* [1991]. However, as have been demonstrated above, these overthrusts are distinguished by the low magnitudes and decay rapidly with depth and along the strike, often grading to shear deformations or to the zones of longitudinal plastic flow.

[74] The modern mountain ridges are typical anticlinal structural features. The remnants of the pre-Mesozoic peneplanation surface reflect its significant plicated deformation. They show bends, including those of small curvature radii and dip angles, as high as 90°, and also the overturned posi-

tions of the peneplain surface. The forms of the pre-Mesozoic peneplain suggest the more intensive deformation in the uplifted areas compared to the basins. Along with the general "centrifugal" structural pattern, this suggests the relatively more "active behavior" of the rock masses in the anticlinal highs and bends in the course of the Alpine tectonic activity. Taking into account the rise of the rock masses to the elevations of more than 5000 m and the divergence structure of the region, it can be inferred that in the upper rocks of the elevation zones the vertical movements were transformed to horizontal ones, and the rock masses were "spread" from the middle of the fold belt toward the basins surrounding it, which was confirmed by the analysis of the stress state of the rocks [*Nikolaev,* 1992].

[75] The reconstructions of the pre-Mesozoic peneplain surface, based on the data reported by *Lukina* [1977] and *Makarov* [1990] and on my own observations, showed that the surface of the pre-Mesozoic peneplane was faulted and plicated in the course of the Alpine tectonic activity without any breaks. The curvatures of this surface are as high as many degrees and vary from subhorizontal to vertical and overturned positions. The example of these deformations is shown in Figure 14 for the area of the Fan Mountains.

[76] The style of the structural and lithological reworking of the rocks of the South Tien Shan folded metamorphic-rock basement, which provided their volumetric mobility, was highly variable. It is known that these variabilities of the mechanisms are associated, first of all, with the rheologic properties of the rocks and with their ability of tectonic volumetric flow or of rheid deformation [*Beroush,* 1991; *King,* 1967; *Leonov,* 1991, 1993, 1996, 1997; *Patalakha,* 1966, 1971]. In the Karakul, Zeravshan, and Kurganak zones, the central parts of which are composed of flysch deposits, the loss of coherence was caused by the process of melange formation; in the Ravat area, by the development of plastic deformation; in the area of the Gissar batholith granites, by volumetric cataclasis; in the carbonate massifs of the Fan mountains, by dynamic recrystallization. The loss of coherence resulted in rheid tectonics which, in its turn, permitted the bending of the basement surface (the top of the pre-Mesozoic peneplane) without its breaking.

[77] To sum up, the significant internal 3D mobility of the pre-Mesozoic folded metamorphic basement was recorded in the complex plicative form of its surface. The overthrusts complicating the sides of the basin are secondary and older than the pre-Mesozoic folds of the peneplane, which is confirmed by the attenuation of the faults along their strikes and dips, by their transformation to the zones of brittle-plastic flow, by the relatively poor deformation of the valley floors, and by the historical analysis of the development of negative structural features and of the uplifts surrounding them (for instance, V. D. Bosov, I. V. Koreshkov, V. N. Krestnikov, I. Sadybakasov, S. S. Shultz, to name but a few).

[78] The internal deformation of the rocks of the basement also confirms the conclusion of its 3D mobility. As follows from the data discussed above, three mechanisms might have been responsible for its mobility. These are the plastic deformation of the metamorphic schists of the Fan Mountains and Kurganak Zone; the melanging activity and the formation of the sedimentary melange protrusions in the Paleozoic rocks of the Zeravshan and Kurganak zones; the volumetric cataclasis of the significant volumes of the Central Gissar Batholith; the dynamic recrystallization of the carbonate massifs, and the formation and rejuvenation of the zones of vertical and horizontal plastic flow. In other words we have the real confirmation of the large-scale plastic, brittle-plastic, and cataclastic flow of the rocks in the area discussed. The internal deformation of the rocks of the pre-Mesozoic folded metamorphic basement agrees well with the folded surface of the pre-Mesozoic peneplain.

[79] In choosing a model for the structural shaping of some or other region, a question arises concerning the predominance of the vertical or horizontal crustal movements, and also concerning the stresses and forces that caused these movements. Many geoscientists (for instance, *Makarov* [1990], *Zonenshain and Savostin* [1979], to name but a few) supported the mechanism of general horizontal compression during the Alpine stage, which had been caused by the convergence of the Eurasian and Hindustan lithospheric plates. This view was based on the general warping of the Earth surface, the formation of folds, nappes, and tectonic sheets, as well as, on the existence of submeridional compression in the region.

[80] However, the space shrinking in the South Tien Shan region for the recent time was estimated by *Chedia and Utkina* [1990] to be merely 6 km, 12 km, and 14 km for the width of the zone measuring 80 km, 240 km, and 350 km, respectively. The tangential compression factor was found vary from 0.01 to 0.3, its average value being 0.04–0.05. Assuming that the South Tien Shan structure was shaped at the expense of the pressure from Hindustan and Pamir, the highest compression must have been concentrated at the meridian of the Pamir Arc curvature trend. However, this has not been proved, although the compression in the Alai Valley was found to be as high as 0.3, this suggesting that the Pamir pressure operated and simultaneously relaxed in the course of the accumulation of the rocks composing the Alai segment of the Afgan-Tajik depression. At the same time in the central and western parts of the Afgan-Tajik depression the surfaces of the "Paleozoic basement-sedimentary cover" boundaries and of the "Paleozoic basement-granite and metamorphic layer" boundaries are fairly flat and do not show any substantial bends or folds, suggesting that the tangential compression stress acting from the Pamir arcs and from the Hindustan Plate were not recorded structurally in the deformation of the boundaries mentioned above. If this assumption is correct, it is not clear how this pressure could cause the highly complex transformations of the rock masses and the deformation of the basement surface of the mountains located north of the region discussed.

[81] Some information of the space reduction might have been provided by the overthrusts and folds of Alpine age, yet, here, too, we deal with many problems. Many overthrusts are not of compression origin but reflect the extension conditions, including the formation of the subhorizontal tectonic flow zones. The potential operation of such processes has been demonstrated using models and geological examples [*Luk'yanov,* 1991; *Ramberg,* 1986]. As can be seen in Figures 9, 12 and 13 these overthrusts have insignificant magnitudes and attenuate along their strikes and dips or are

transformed to the zones of brittle-plastic shear flow (in mechanical sense) of the overthrust kinematics. The deformation of the general compression associated with some large-scale reduction of the space is discarded by the relatively simple shapes of the internal syncline floors, compared to their sides. Combined with the overthrusting attenuation, this suggests that the synclines were compressed merely by the stresses arising in the basin sides. The formation of the structural pattern of the sedimentary cover in the Afgan-Tajik Basin, which is disharmonic relative to the basement surface, was controlled by another mechanism, namely, by the mechanism of lateral extrusion between the relatively rigid blocks of the Pamir and Tien Shan mountains [*Kopp,* 1997; *Zakharov,* 1970].

[82] As follows from the data reported by *Nikolaev* [1992], the crustal stress field, reconstructed for the lower layers of the crust in the Gissar-Alai mountains, was marked by submeridional compression and subvertical extension. The lower crust became shorter as it was forced toward the center of the mountains, and the rock masses were squeezed up. On the contrary, the regional surface field suggests the lateral subhorizontal extension and subvertical compression (flattening), which are reflected in the model offered in this paper. This view does not contradict the results of the light-type range finder measurements, which show that the Tajik Depression grows wider in spite its subhorizontal compression. This geodynamic interpretation agrees well with our interpretation [*Guseva et al.,* 1993].

[83] The complex pattern of the pre-Mesozoic peneplain surface and the internal structural and material transformation of the basement rocks suggest that the foundation of the epi-Paleozoic platform, reactivated in recent time, behaved as a quasi-plastic body. The structural and lithologic transformations reflect the plastic, brittle-plastic, and cataclastic flow of huge rock masses. The degree of the Alpine structural and material transformations of the basement rocks was found to be higher than that of the sedimentary rocks, the structure of the basement rocks, associated with the Alpine tectonics, was found to be more stressed than that of the sedimentary rocks. Similar conditions have been discovered recently in the Central Tien Shan region, and an attempt was made to substantiate this phenomenon in physical and mathematical terms [*Mikolaichuk et al.,* 2003].

[84] The studies of the general structural style, particular structural features, geophysical data, as well as of the physico-mathematical models available, suggest that the basic formation mechanism of the structural and morphologic styles of the South Tien Shan region is controlled at the present time by the volumetric redistribution of the crustal material, namely, by the flow of the rock masses from the basin regions and their displacement to the mountain region. This is confirmed by the geological and geophysical data available, by the deformation pattern of the pre-Mesozoic adjustment surface, by the operation of the mechanisms responsible for the flow and accumulation of rocks in elevated regions, by the geometry of earthquake sources, and other factors.

[85] The lateral tectonic flow operates at different levels of the upper and lower crustal layers, resulting in the origin of subhorizontal tectonic layering and in the formation of a disharmonic structure inside each layer. The origin of lateral flows, which are transformed to the zones of vertical rising and pumping, with subsequent spreading, seems to be associated with the development of heterogeneities (asthenospheric diapirs, convective and advective flows) in the reactivated low-density mantle, the existence of which is inferred from the results of geophysical measurements. The total subhorizontal compression stress, typical of the modern stress state of the region, were expressed in the mechanism of curvature instability, in the wave pattern of the distribution of the domes and basins of various sizes, and in the redistribution of the rock material in space. The compression stress controlled the spatial arrangement of the structural style and the formation (or regeneration) of some structural elements, such as diagonal strike-slip faults and longitudinal flattening zones (viscous faults and shear zones). It should be added that the recent time witnessed the operation of the mechanism of rock mass gravitational instability which was especially pronounced in the region of the Fan mountains. The gravitational instability of the highly elevated Fan rock masses (with the previous formation of their subhorizontal layering and the presence of low-viscosity rocks) resulted in the divergent subhorizontal "spreading" of the massif, the formation of overthrusts and sublayer breaks, the tectonic crushing of the young basins surrounding this rock massif, and the formation of the structural features of the "tectonic sewing" type.

[86] It follows that the formation of the Alpine structure of the South Tien Shan region was controlled by, at least, three different, relatively independent mechanisms, although associated with one another, namely, by the main compression-ejection mechanism, associated with the 3D lateral rock-mass redistribution [*Leonov,* 1993], and the accompanying mechanisms of bending and gravitational instability. The factors responsible for the mantle reactivation are not discussed here. This seems to have been a planetary event, unassociated with any regional mechanisms and environments, yet this event can be treated as the mantle response to the underplating of the Hindustan plate and its movement in the northern direction.

Conclusions

[87] The results of this study demonstrate that the modern South Tien Shan structure was formed as a result of the multistage evolution (paleoocean $\Longrightarrow$ fold-nappe region $\Longrightarrow$ platform $\Longrightarrow$ intracontinental orogenic belt) and of the multifactor tectonic activity, reflecting the complex interference between different geodynamic processes and structure-formation mechanisms.

[88] Each of the major stages of this region evolution was characterized by its own geodynamic behavior, charateristics only of that particular stage. At the same time, each of these stages might have been accompanied by some particular geodynamic activity associated with or independent of the major one. These particular geodynamic processes, usually resulted in the formation of some specific structural-tectonic assemblages, and operated in some particular zones

(environments) or embraced vast spaces, overlapping the effects of the main geodynamic process.

[89] The origin of some mechanisms and conditions, both particular and general ones, was not controlled by the geological history of the previous geological epochs. As proved by this study, the geodynamic conditions of the neotectonic stage were not controlled by the geodynamic conditions which produced the nappe-fold region at the end of the Paleozoic, being of absolutely different origin. This conclusion confirms the thesis advanced earlier [*Leonov,* 1972, 1976] concerning the independence of the orogenic events of the geological prehistory of some or other territory.

[90] Some geodynamic mechanisms (such as, bending or gravitational instability), operating at different stages of the geologic evolution of the region, suggest the operation of some processes which had a more general significance, compared to the common mechanisms of structure formation, associated with the effects of regional compression, extension, or pressure development. The processes of this kind are associated with the fundamental properties of the Earth and its lithosphere, such as, its gravitational instability, the 3D mobility of rock masses, and the varying stress state of the lithosphere. These processes develop during different periods of the lithosphere evolution, and in different geotectonic environments.

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