ISSN 0016-8521, Geotectonics, 2024, Vol. 58, No. 6, pp. 611–638. © Pleiades Publishing, Ltd., 2024.

Tectono-Gravitational Detachments in the Alpine Cover of the Northern Slope of the Greater Caucasus and Western Pre-Caucasus Basin (Adygean Segment)

S. Yu. Kolodyazhny[a], N. B. Kuznetsov[a, *], E. I. Makhinya[a], E. A. Shalaeva[a], K. I. Dantsova[a, b], T. V. Romanyuk[c], M. P. Antipov[a], and G. E. Parfenov[a]

[a] *Geological Institute, Russian Academy of Sciences, Moscow, 119017 Russia*

[b] *Gubkin Russian State University of Oil and Gas (National Research University), Moscow, 119991 Russia*

[c] *Schmidt Institute of Physics of the Earth, Russian Academy of Sciences, Moscow, 123242 Russia*

**e-mail: kouznikbor@mail.ru*

Received April 23, 2024; revised September 5, 2024; accepted September 12, 2024

Abstract—The Adygean segment encompasses the transition zone between the Central and Western segments of the Greater Caucasus (GC). It is located within the western part of the Laba–Malka monoclinal zone (northern slope of the GC). North of this area, the Western Kuban and Eastern Kuban basins are situated. They are separated by the Adygean uplift and generally form the southern part of the Western Pre-Caucasus basins. We have carried out geological and structural studies of the lower part of the Alpine cover (Middle–Upper Jurassic) within the Adygean segment and interpreted seismic profiles of Mesozoic–Cenozoic strata in the Western Pre-Caucasian basins. It was revealed that tectonogravitational detachments are widespread within the Adygean segment on the northern slope of the GC and in the southern part of the Pre-Caucasus basins. They occurred as a result of slipping of sedimentary layers mainly in the north direction: down the slope of the GC orogen. Our tectonophysical studies have shown that the detachments took place in conditions of reverse and normal faulting due to vertical-oblique flattening and predominantly subhorizontal stretching. We have concluded that tectonogravitational detachments were formed by the interaction of two factors: vertical uplift of the GC orogen, caused by endogenic (tectonic) reasons, and gravitational slip of geomasses from the slopes of this mountain structure. Analysis of seismic sections crossing the Western Pre-Caucasian basins has shown the widespread development of clinoforms, which are paleodeltas of terrigenous material brought from the Scythian Plate and the East European Platform. The distribution of clinoforms in Cenozoic strata of the Pre-Caucasus basins allows us to suggest that southward-directed sedimentary flows existed from the Paleocene to the Late Pliocene. Based on this, we believe that the formation of the modern GC orogen and accompanying coarse molasse began no earlier than the end of the Pliocene, probably in the Eopleistocene. The formation of tectonogravitational detachments, which is one form of manifestation of the recent orogeny of the GC, led to the development of various structures: asymmetric folds, small thrusts, domino structures, faults, ramp folds, and thrust duplexes. Along the detachments there are ramp structures of local tension and compression, which form multisized cells of lateral rock-mass transport. Such cells facilitate activation of the migration, redistribution, and localization of hydrocarbons.

Keywords: geodynamics, tectonophysics, detachments, ramp structures, Greater Caucasus, Western Pre-Caucasus basin

DOI: 10.1134/S0016852124700407

INTRODUCTION

At present, there is no unified geodynamic model that explains the causes of and mechanisms underlying the formation of epiplatform orogenic belts. The modern mountain structure of the Greater Caucasus, which formed on the southern periphery of the epi-Hercynian Scythian Plate in the Late Alpine epoch, is an example of a typical epiplatform orogen [5, 25, 26]. Its structure includes rocks complexes that are indicators of subduction, accretion, and collision geodynamic settings of various stages of tectogenesis (Cadomian, Caledonian, Hercynian, Cimmerian, Alpine). At the same time, the processes of formation of the modern mountain structure of the Greater Caucasus reveal no clear spatiotemporal relationship with paleosubduction or paleocollision and are separated from them by a long era of platform (pericratonic) development [5].

There is no generally accepted opinion about the timing of the formation of the modern Greater Caucasus orogen. Different authors have and still place the boundary of the beginning of the Afro-Arabian-Eurasian collision and associated orogeny of the Greater Caucasus at different age levels of the Paleogene [16, 23, 41, 45, 46, 63], or even the Cretaceous [8, 10, 26], at the beginning or end of the Miocene [7, 15, 48], as well as the Pliocene [50].

According to the opinion of M.L. Kopp [5], the modern structure of the Caucasus region was formed in its different parts at different times: in the Lesser Caucasus, Late Cretaceous–Early Paleogene; in the central segments of the Greater Caucasus, in the Paleogene; and the subsidence of this orogen, in the middle of the Neogene–Quaternary.

It should be noted that there is no clear distinction between such phenomena as syncollisional folding and nappe formation, on the one hand, and orogenic uplift of elements of ancient accretionary–collisional structures recognizable in the internal structure of meganticlinoria, on the other. With respect to collision events, orogenic processes often manifest themselves somewhat later and can be characterized as superimposed [13].

Reliable indicators of orogeny include the processes of formation of marginal depressions filled with molasse—mountain erosion products [25, 29, 37]. In the Pre-Caucasus Basins, the onset of the formation of complexes traditionally associated with the lower (fine-grained) molasse is usually attributed to the Oligocene (Maykop Group, Upper Oligocene–Lower Miocene) [25, 26]. It is generally accepted that the accumulation of fine molasse indicates the occurrence in the Oligocene of low-mountainous island(s) in the area of the modern Greater Caucasus, from which fine-grained material could have been brought to the Pre-Caucasus Basins.

In the last decade, many high-resolution seismic profiles materials have become publicly available. These profiles allow to find out the internal structure of the western part of the Pre-Caucasus Basins. These materials suggest that the supply of fine detritus into these paleobasins occurred not from south to north (from the Greater Caucasus), but from north to south, from the vast areas of the East European continent, consisting of the East European Platform and the Scythian Plate [19, 35, 36].

This is evidenced, in particular, by the seismic facies complexes identified in seismic records: numerous buried scarps and clinoforms, which are paleodeltas—sedimentary bodies composed of detrital material brought to the wide shelf of the East European continent. The scarp walls and the inclination of the cross-bedded series within the clinoforms, indicating the directions of progradation of the paleodeltas, are oriented to the south.

Seismic complexes with similar configurations are noted not only in the layers of the Maykop Group, but also in the overlying strata of Western Pre-Caucasus up to the Neogene (Pliocene)–Quaternary (Gelasian) [19] stratigraphic boundary (2.6 Ma). It should be noted that these conclusions are based on the results from analyzing individual seismic profiles [19]. Further research is needed to fully substantiate them.

At the same time, taking into account these preliminary conclusions, it should be assumed that the sediments filling the depressions of Western Pre-Caucasus and covering the stratigraphic interval from the Oligocene to the Pliocene inclusive are not orogenic molasse (lower or upper), since they do not contain erosion products of the Caucasian orogen. Up to a certain point in time, the Cenozoic depressions of Western Pre-Caucasus were not marginal depressions dynamically associated with the mountain uplift of the Greater Caucasus, which probably did not exist at least until the end of the Pliocene. These continental marginal depressions were formed in the area of the wide shelf of the southern periphery of the East European continent and were filled with erosion products of the continent.

The depressions of Western Pre-Caucasus constituted part of the East Paratethys Sea basin, which included the area of the future Greater Caucasus orogen and areas farther to the south, including modern West and East Black Sea residual basins [19]. In [21], it was proposed to consider coarse Quaternary deposits containing isotope-geochronological (detrital zircon) and lithological (detrital material) markers of provenance areas in the Greater Caucasus region as orogenic molasse. These formations have a small thickness, and their age (2.6–2 Ma to the present) suggests a fast formation of the modern mountain structure of the Greater Caucasus.

The significant growth rates of the Greater Caucasus orogen and small volumes of its erosion products that accumulated in the depressions of Western Pre-Caucasus represent a contradictory phenomenon requiring explanation. In this regard, we have set a number of tasks.

— To identify additional seismostratigraphic features characterizing the direction of flows of detrital material, which filled the depressions of Western Pre-Caucasus.

— To identify and interpret the structures of the northern slope of the Greater Caucasus and the southern part of the Western Pre-Caucasus Basins, which together may have resulted from the recent orogeny.

— To identify mechanisms of rapid denudation of the plate cover and exposure on the surface in the area of erosion of the granite-metamorphic basement complexes of the Greater Caucasus.

To resolve these tasks, we conducted a geological and structural study of the sequence of the lower part of the plate cover within the Adygean segment of the Greater Caucasus, and also analyzed and interpreted seismic sections characterizing the structure of some troughs of Western Pre-Caucasus. During field studies and interpretation of seismic sections, methods of structural-kinematic, tectonophysical and paragenetic analysis were used [11, 14, 40, 55, 59].

We have studied the spatial orientation and morphological parameters of folds, various kinematic indicators, as well as faults and fractures with signs of displacement (marker displacements, slickenlines and scarps on slip planes, near-fault flexures, etc.). For statistical processing of kinematic data, we used FaultKin6 software [56].

GEOLOGY

Tectonics of the Western and Central Parts of the Greater Caucasus

The Greater Caucasus is one of the youngest orogens of the Black Sea sector of the Alpine–Himalayan mobile belt. Folds and fracture patterns of the modern Greater Caucasus formed on the Epi-Hercynian basement of the Scythian Plate during the Cimmerian and Alpine (Early and Late Alpine) epochs of tectogenesis [5, 25, 26, 30, 43, 49].

In its moderm structure, the Greater Caucasus is an asymmetrical meganticlinorium with a wide and gently sloping northern side and narrower and steep southern side (Fig. 1).

On the northern side of the meganticlorium, complexes of the Hercynian basement are exposed fragmentarily from under weakly dislocated and gently northward-dipping Mesozoic–Cenozoic strata of the cover of the Scythian Plate (the Paleozoic core of the Greater Caucasus). In the central segment of the meganticlinorium, the Greater Caucasus rock complexes are part of the Laba–Malka monoclinal zone [5]. A characteristic feature of this zone is a weakly deformed plate cover, represented by Middle Jurassic (Callovian)–Cenozoic strata. Conversely, the southern side of the meganticlinorium (southern slope of the Greater Caucasus) is formed by a unit of fold–thrust sheets, intensely compressed and overturned to the south. The unit is composed of Mesozoic (Lower Jurassic–Cretaceous) and Cenozoic sedimentary and volcanosedimentary rocks.

In the axial part of the Greater Caucasus, several along orogen segments are distinguished: western, central, and eastern [5, 26]. Within the axial portion of the central segment, in the core of the meganticlinorium, crystalline complexes of the basement of the Scythian Plate are exposed. The complexes were involved in the Cimmerian and Alpine deformations and were locally thrust onto Jurassic strata of the southern slope along the Main Caucasian Fault (Fig. 1) [5].

In the structure of the central and western parts it is customary to distinguish strata of the Hercinian, transitional (Indo-Sinian), Cimmerian, and Alpine tectonic stages [5, 17, 18, 43] (Fig. 2).

The strata of the Hercynian stage form the basement of the Scythian Plate and is represented by Paleozoic polyfolded metamorphic and igneous rock complexes, which host the Late Precambrian and Early Paleozoic blocks of Peri-Gondwanian origin (the Cadomides) [5, 22, 54, 60]. These formations are intruded by Late Paleozoic granitoids and covered by Late Paleozoic molasse.

The strata of the Transitional (Taphrogenic Indo-Sinian) stage, which occupies an intermediate position between the basement and the Mesozoic cover, is composed mainly of sedimentary and volcanosedimentary rocks of the Triassic and, partially, Upper Permian [9]. Within the Greater Caucasus these formations are spatially associated with outcrops of rocks of the Hercynian basement.

Sedimentary and volcanosedimentary rocks of the Lower–Middle Jurassic make up the strata of the Cimmerian stage. These formations, among others, are also developed on the northern slope of the Greater Caucasus. Here, they are usually considered as the lower strata of the Scythian Plate cover [9, 43]. The rock complexes of the Alpine stage are represented mainly by sedimentary formations of the Middle Jurassic (Callovian)–Cenozoic age. On the northern slope of the Greater Caucasus, they form a gently sloping cover (the middle and upper structural layers of the cover, according to [43]) and with a sharp unconformity overlie rocks of the Cimmerian and Hercynian structural levels. On the southern slope in the Greater Caucasus, such sharp unconformities at the base of the Alpine level are hardly manifested at all, due to the fact that the Alpine and Cimmerian rock complexes are involved in intensive fold–thrust deformations. In a number of cases, it has been noted that the rocks of the Alpine cover on the northern slope of the Greater Caucasus are tectonically detached from the underlying rock complexes [12, 17, 18]. However, the extent of this phenomenon has not been sufficiently studied.

Structure of the Troughs of Western Pre-Caucasus

North of the western part of the Greater Caucasus orogen, there is a system of depressions and buried uplifts, which together constitute the West Pre-Caucasus Basin. The depressions are filled with Mesozoic and Cenozoic sediments, which are the plate cover (Alpine stage) of the Scythian Platform.

From the Oligocene to the middle Miocene, in the western part of the Pre-Caucasus Basin, there was an accumulation of predominantly marine clayey (Maykop Group) and siltstone deposits, containing only rare and thin layers of sandy and gravel material, as well as pebbles, including fragments of sedimentary rocks of the underlying Alpine strata [2, 9] (Fig. 2).

Subcontinental (lagoon, lake, beach, and less often alluvial) deposits, indicating marine regressions in the Pre-Caucasus Basin (paleobasins) (Paratethys), are present in the sedimentary sequences since the Middle Miocene (Sarmatian) [2–4]. These formations are interlayered with marine sediments and contain very small volumes of coarse-grained material, represented

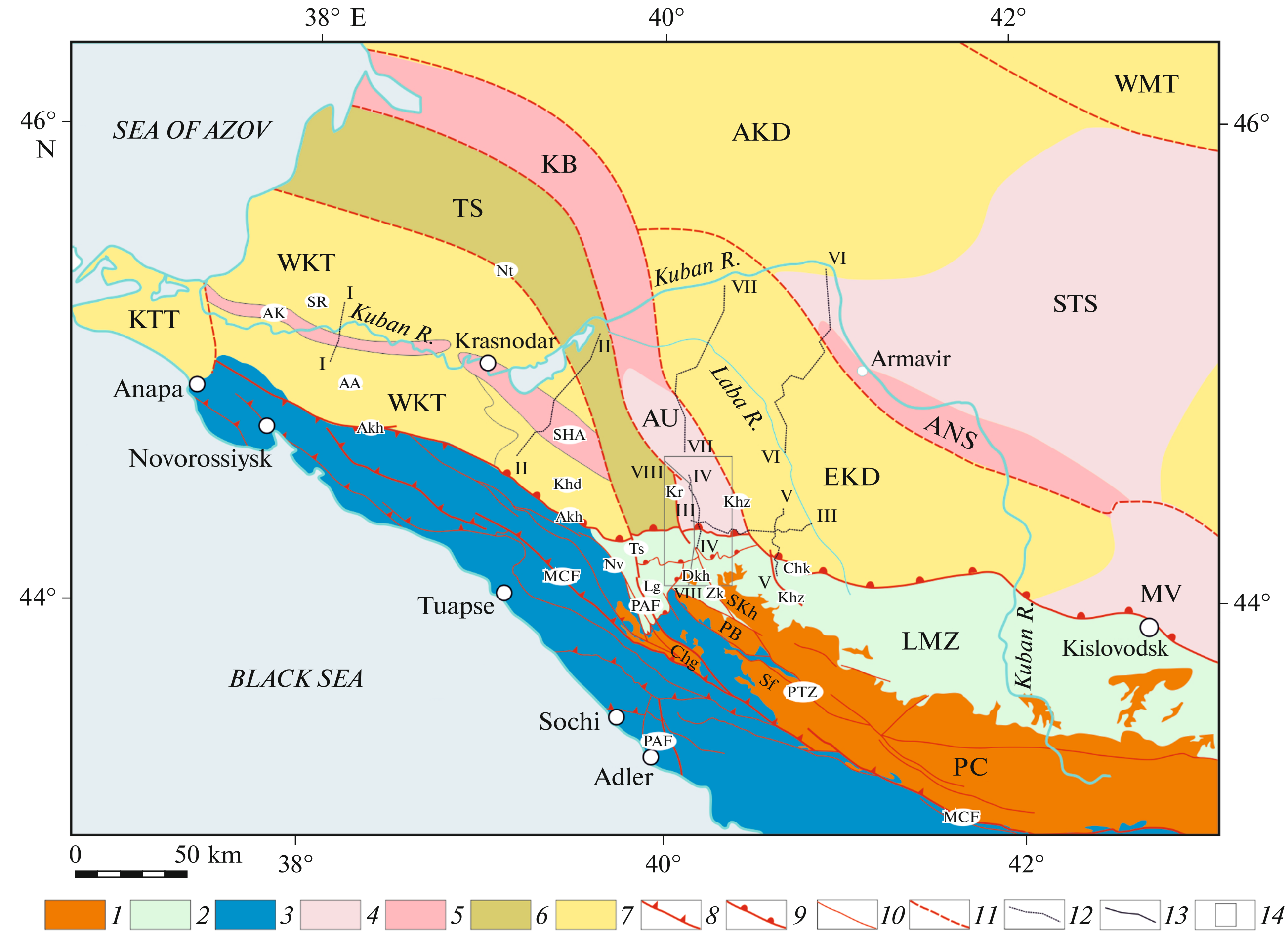


Fig. 1. Tectonic scheme of western part of Greater Caucasus and Western Pre-Caucasus trough (according to data from [5, 9, 26], modified). Legend (Roman numerals): seismic sections I–I'–VIII–VIII'. Notation: PC, Paleozoic core of Greater Caucasus, LMZ, Laba–Malka zone; *troughs and depressions*: KTT, Kerchinsky–Tamansky; WKT, West Kuban (SR, Slavyansky–Ryazansky depression; AA, Adagumo-Afip depression; KhD, Khadyzhensky monocline); EKD, East Kuban; AKD, Azov–Kuban; WMT, West Manych; *arches and highs*: STA, Stavropol; MV, Mineralovodsky; AU, Adygean uplift; *linear systems of uplifts*: AK, Anastasiev–Krasnodar anticline; SHA, Shapsug–Apsheron swell; KB, Kanevsky–Berezansky uplift system; ANS, Armavir–Nevinnomyssk swell; *fault zones*: MCF, Main Caucasian Fault, PTZ, Pshekish-Tyrnyauz; PAF, Pshekha–Adler; Akh, Akhtyr; Chk, Circassian; Nt, Novotitarovsky; *faults*: Nv, Navaginsky; Ts, Tsitsa; Kr, Kurdzhip; Zk, Zakan; Khz, Khodza; *pre-Jurassic basement highs*: SKh, Sakhray–Khodza; Dkh, Dakhovsky; PB, Pshekha–Bambak; Sf, Sofia; Chg, Chugushk; Lg, Lagonaki plateau. *Greater Caucasus fold belt (1–3)*: *1*, Hercynian basement; *2–3*, Cimmerian and Alpine complexes: *2*, northern slop monoclines; *3*, southern slope and Western Caucasus; *4–7*, structures of West Caucasian trough: *4*, projections; *5*, systems of swell-like uplifts; *6*, steps; *7*, depressions; *8–11*, faults: *8*, reverse faults and thrusts; *9*, detachments; *10*, unspecified kinematics; *11*, buried under sediment cover; *12–13*, section lines: *12*, seismic; *13*, geological; *14*, research area in Adygean segment.

by thin layers and lenses of conglomerates. The presence of large boulders and blocks in these conglomerates, poor sorting and rounding of the fragments allow us to assume that these deposits were formed as a result of local erosion of the underlying layers and were transported to a short distance. Sedimentary molasse-like rocks make up only the uppermost part of the Western Pre-Caucasus sequence. They have an insignificant thickness, but they host horizons of polymictic gravelstones and conglomerates with fragments of granitoids and metamorphic rocks lithologically similar to those in the Paleozoic core of the Caucasus orogen. Being based only on rare faunal finds, the age of these formations has not been estimated accurately enough, it varies in a wide range from the Pliocene to the Quaternary [2]. The authors of [21] provide information that allows us to suggest that these molasse deposits accumulated no earlier than the Pleistocene. In the West Pre-Caucasus Basin along the base of the Alpine rock complex (Middle Jurassic, Callovian), a number of large structural elements are distinguished [9] (Fig. 1). In the northern part of the territory, there is a zone of Manych troughs and the Azov–Kuban depression, which narrows towards the south and turns into the East Kuban depression. From the west, these depressions are bounded by basement ruptures that border the Kanev–

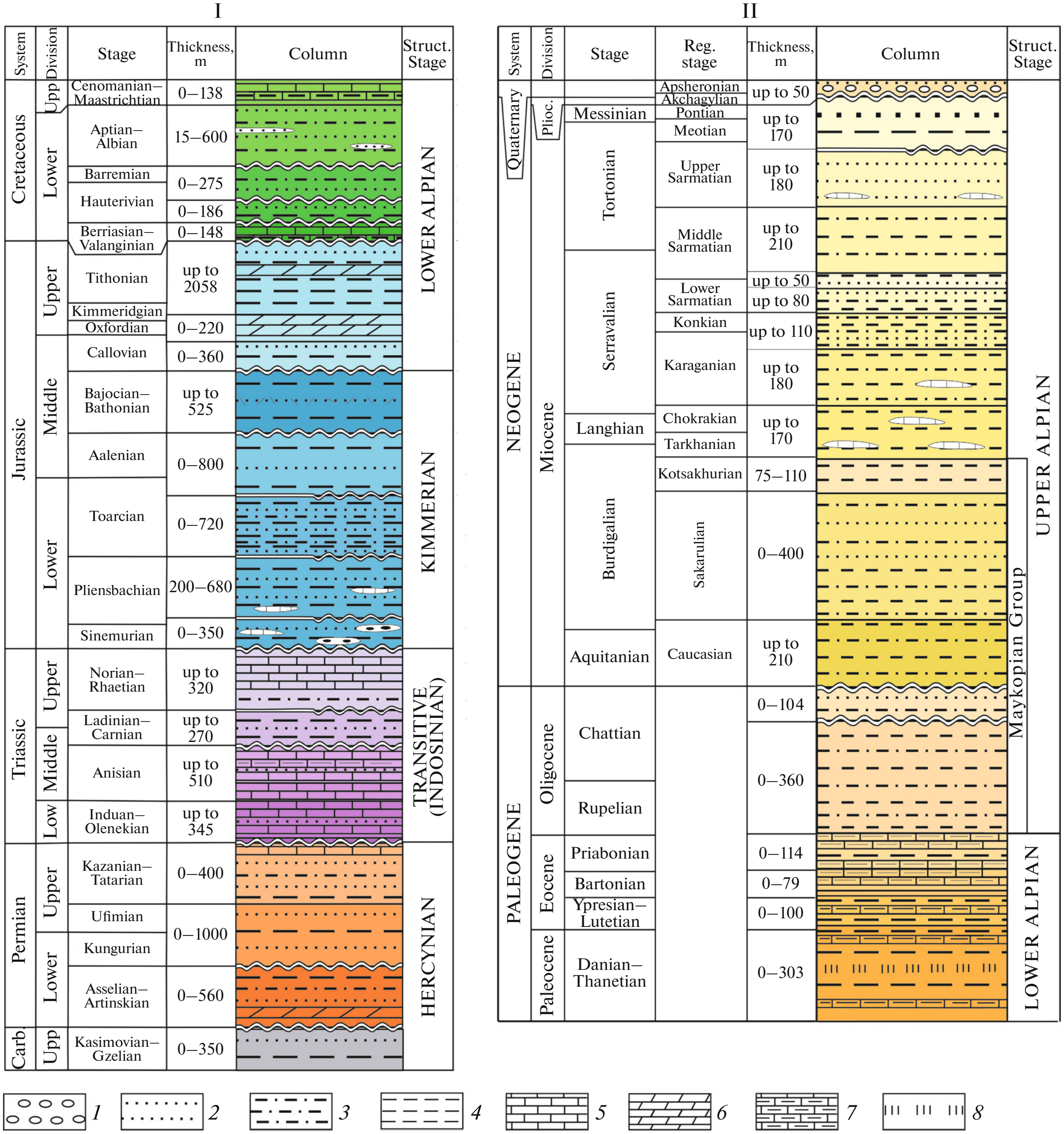


Fig. 2. Combined stratigraphic column of northern slope of Greater Caucasus and Western Pre-Caucasus (Belaya River basin) in area of Adygean sector (according to data [17]). Parts of section are marked as follows: I, Paleozoic–Mesozoic; II, Cenozoic. *1*, Gravelstones and conglomerates; *2*, sandstones; *3*, siltstones; *4*, argillites; *5*, limestones; *6*, dolomites; *7*, marls and clayey limestones; *8*, silicites.

Berezan system of uplifts and the Adygean uplift located on their southern continuation. To the southwest of these uplifts is the Timashevsky monoclinal step, which is separated from the West Kuban trough by the large Novotitarovsky fault. The system of depressions of the Western Pre-Caucasus is bounded from the east by the Stavropol and Mineralovodsky uplifts.

The thickness of the Mesozoic–Cenozoic complex in the axial part of the Pre-Caucasus Basins reaches 10–15 km [28]. In areas of uplifts, there is a significant reduction in thickness of the sedimentary cover. The uplifts, steps, and depressions of the Western Pre-Caucasus are usually bounded by flexural fault zones, developed mainly in the lower part of the plate cover and in basement rocks. Upwards, in the cover sequence, these faults gradually vanish. They are replaced by zones of sharp changes in thiknesses of layers and disappearance of individual horizons from the section. The southern slope of the Western Pre-Caucasus Basin is separated from the area of syn-Alpine folding of the Greater Caucasus by the Akhtyrsky (in the west) and Cherkessky (in the east) fault zones, which are longitudinally connected in the southern part of the Adygean uplift (Fig. 1).

DETACHMENT SYSTEMS AT THE BASE OF THE PLATE COVER IN THE ADYGEAN SECTOR OF THE GREATER CAUCASUS

Geological Structure of the Adygean Sector

The Adygean sector encompasses the transitional region between the Central and Western segments of the Greater Caucasus within the Belaya River basin in the western part of the monocline of the northern slope of this orogen, known as the Laba–Malka zone (Fig. 1). In this area, periclinal dipping of the crystalline core complexes of the Greater Caucasus occurs under the folded Mesozoic and Cenozoic strata of the Western Pre-Caucasus segment. According to [25], this dipping is caused by the submeridional (transcaucasian) Pshekha–Adler flexural-fault zone, diagonal to the strike of the structures of the Caucasian (NW) direction, along which the stepped uplift of the Central Caucasus takes place with respect to its Western segment.

According to other interpretations, the Pshekha–Adler (Adygean–Laba) zone is a wide trans-Caucasian uplift extending NNW (submeridionally), complicated by longitudinal faults, depressions, and uplifts in the pre-Jurassic basement of the Greater Caucasus [17]. As a result of tectonic events of the Cimmerian and Alpine stages of the region's evolution, along with rejuvenation of submeridional structures, structural ensembles of the Caucasian direction were formed—ruptures, folds and protrusions/horsts. The largest Cimmerian–Alpine faults are the Pshekha–Tyrnyauz suture zone and the Main Caucasian Fault. The faults that make up the Pshekha–Adler zone are diagonally connected and usually merge gradually with the structures of the Caucasian direction (Fig. 1). The combination of such differently oriented structures determines the style of tectonics of the sector under consideration.

In the northern part of the Adygean sector, pre-Jurassic rocks constitute three massifs: the Rufabgo, Sakhrai–Khodza and Dakhovsky (Figs. 3, 4). Within the Rufabgo and Sakhrai-Khodza massifs, carbonate, terrigenous–carbonate, and terrigenous rocks of the Triassic transitional stage outcrop at the surface. These sedimentary groups often form gently sloping monoclines, broken by sublayer faults and thrusts, which are dynamically related to shear fault zones of the trans-Caucasian direction [17].

The Dakhovsky massif includes fragments of Cadomian metamorphic and Hercynian igneous rocks. Along the periphery, the massif comprises amphibolites and amphibole gneisses of the Neoproterozoic Balkan Group (Cadomides). The amphibolite–gneiss complex consists of fragments of deformed Precambrian tectonic covers, along the boundaries of which serpentinite plates and protrusions and zones of polymictic (serpentinite–amphibolite-gneiss) mélange are widespread.

In the central part of the Dakhovsky massif, gneisses are intruded by the Late Paleozoic (Hercynian) granitoids (plagiogranite–diorite Malka complex [17]). The massif underwent significant deformations during the Cimmerian and Alpine stages of tectogenesis; it is dissected with numerous faults, tectonically stratified, and is bounded by normal and strike-slip faults along the edges. Judging from drilling results, the Dakhovsky massif is thrusted onto Triassic rocks [17] (Fig. 4).

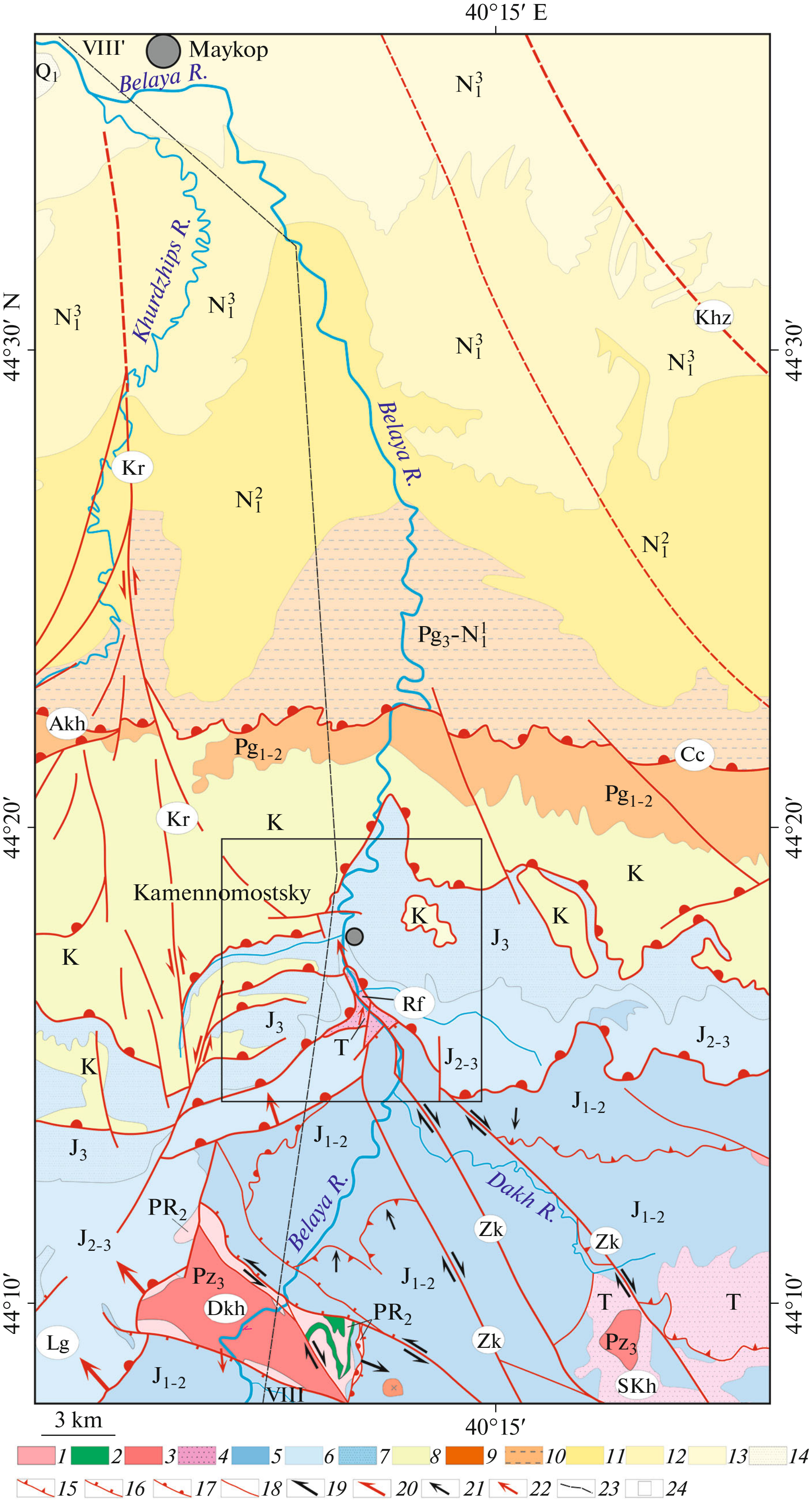
The sandy–shale coal-bearing and terrigenous–volcanic rocks of the Lower and Middle Jurassic make up strata of the Cimmerian stage (Fig. 2).

These rocks are relatively weakly deformed, forming open folds and gently sloping monoclines, but along tectonic faults, the intensity of their deformations increases significantly. At the same time, compressed folds, cleavage, and faults with transpressional and transtensional kinematics are manifested here. The Jurassic strata include thrusts and nappes, as well as associated systems of small asymmetric folds (Fig. 3).

There is a point of view that fold–thrust structures are dynamically related to the faults of the trans-Caucasian strike, which have a shear displacement component [17].

The strata of the Alpine stage is represented by carbonate, carbonate–evaporite–terrigenous, terrigenous and, less commonly, volcanic rocks of the Middle Jurassic (Callovian)–Cenozoic age (Fig. 2).

Fig. 3. Geological map of northern part of Adygean segment of Greater Caucasus (according to [17], modified). Notation: *pre-Jurassic basement highs*: Rf, Rufabgo; SKh, Sakhray–Khodza; Dkh, Dakhovsky; *fault zones*: Akh, Akhtyr; Cc, Circassian; Zk, Zakan; Kr, Kurdzhip; Lg, Lagonaki plateau. *1–3*, Hercynian rock complexes: *1*, Late Proterozoic gneisses; *2*, serpentinites (age not determined); *3,* Late Paleozoic granitoids; *4*, Indo-Sinian sedimentary complexes of Triassic; *5,* Cimmerian sedimentary complexes of Lower–Middle Jurassic; *6–14*, Alpine sedimentary complexes: *6*, Middle–Upper Jurassic; *7*, Upper Jurassic; *8*, Cretaceous; *9*, Paleocene–Eocene (undifferentiated); *10*, Oligocene–Lower Miocene; *11*, Middle Miocene; *12*, lower part of Upper Miocene; *13*, upper part of Upper Miocene; *14*, Pliocene–Quaternary; *15–18*, discontinuous faults: *15*, reverse faults and thrusts; *16*, normal faults; *17*, detachments; *18*, other; *19–20*, directions of shear displacements (stages): *19*, Cimmerian; *20*, Alpine; *22–23*, directions of horizontal (tangential) movements (stages): *21,* Cimmerian; *22,* Alpine; *23*, geological section along line VIII–VIII'; *24*, research area in Rufabgo scarp.


40°15′ E
VIII'
Maykop
Belaya R.
Khurdzhips R.
44°30′ N
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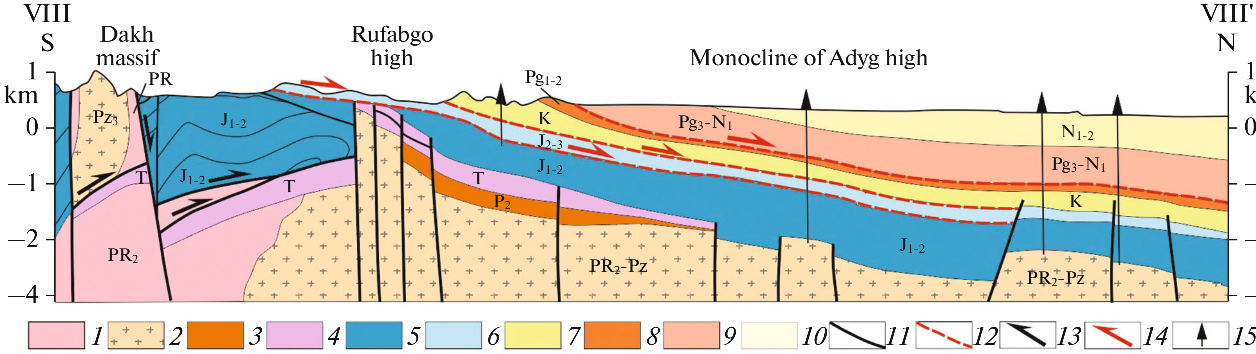


Fig. 4. Geological section along line VIII–VIII' (according to data from [17], modified). Position of section VIII–VIII', Fig. 3. *1–3*, Hercynian rock complexes: *1*, Late Proterozoic gneisses; *2*, Late Paleozoic granitoids; *3*, Late Permian sediments; *4*, Indo-Sinian sedimentary complexes of Triassic; *5*, Cimmerian sedimentary complexes of Lower–Middle Jurassic; *6–10*, Alpine sedimentary complexes: *6*, Middle–Upper Jurassic; *7*, Cretaceous; *8*, Paleocene and Eocene; *9*, Oligocene–Lower Miocene; *10*, Lower–Middle Miocene; *11–12*, stages of fault formation: *11*, Cimmerian; *12*, Alpine; *13–14*, stages of manifestation of movements: *13*, Cimmerian; *14*, Alpine; *15*, wells.

In the territory under consideration, these rocks form a weakly deformed cover, gently dipping (3°–10°) in the north. The Alpine cover complexes with a sharp structural unconformity overlie formations of the Cadomian–Hercynian, transitional, and Cimmerian stages (Figs. 3, 4).

Within the Main Caucasus Range the Alpine rock complexes consitute the vast Lagonaki plateau, which is considered a relict of the Miocene peneplain [25] (Fig. 1).

At the base of the Alpine cover, on the complexes of the pre-Alpine base, transgressively lies a horizon of terrigenous rocks of the Lower–Middle Callovian (Kamennomostsky Unit [17]). At its base are gravelstones (0.7 m), overlain by sandstones, siltstones, and argillites. These rocks mark a major stratigraphic nconformity between the Cimmerian and Alpine stages, have a variable thickness (0–65 m), form lenticular-shaped bodies, and are probably partially detached from their structural base.

The overlying strata of dolomites and limestones of the Middle–Upper Jurassic are significantly deformed along zones of sublayer detachments developed at the base and within its section. The monoclinal occurrence of the Alpine cover is deformed by gently sloping and ridge-shaped folds, locally developed along submeridional faults.

The outer boundary of the Alpine cover complexes is represented by steep scarps comprising carbonate rocks of the Middle–Upper Jurassic. At the base of the scarps, boulder–rubble giant bands are widespread. Their width (2–5 km) is several orders of magnitude greater than the height (50–150 m) of the rocky scarps, which suggests the seismic nature of the latter. The wave-shaped, relief-subordinated outer boundary of the cover complexes excludes a possible relationship between the scarps and the steeply dipping faults. The most likely reason for their formation is seismic movements along sublayer detachments, which contributed to the origin of significant volumes of disintegrated material.

The Pshekha–Adler zone is represented by a fault system of a NNW strike, developed in a wide (up to 60 km) zone. The longest structures are represented by the Tsitsa, Kurdzhips, and Khodza faults, as well as the Zakan tectonic zone [17, 25, 26]. According to geophysical data and drilling, some of these faults have been traced to the West Pre-Caucasus Basin, within which the Kurdzhips and Khodza faults limit the Adygean uplift [2, 17] (Fig. 1).

The faults under consideration are characterized by significant vertical and horizontal displacements. The vertical offset along these steeply dipping faults decreases from the upper part of the basement upwards to the Upper Miocene strata from 400–500 m to tens of meters respectively and almost completely vanishes in the Pliocene–Quaternary deposits [2]. In the area of the Alpine cover distribution, the shear offset along faults is as follows [2]:

— the Kurdzhips fault (sinistral), 3.5–4.0 km;

— Khodza fault (dextral), 10.0–12.0 km.

In the Late Miocene and Pliocene–Quaternary formations, shear displacements are weakly manifested. The Zakan zone probably represents the southeastern continuation of the Kurdzhips fault at a greater depths, in the rocks of the pre-Alpine basement. For this zone, as well as for the Khodza fault, significant right-lateral displacements have been established, which at the Cimmerian stage of development led to the formation of strike-slip fold–thrust and nappe structures [17].

At the same time, on the opposite sides of the Zakan zone, the vergence of the fold–thrust structures changes to the opposite, depending on the directions of the relative horizontal movements of the adjacent blocks.

On the southwestern side of this zone, nappe structures with northern vergence have been established, and on the northeastern side—with southern vergence (Fig. 3).

Presumably, the thrust movements on the southwestern side of the Zakan zone leaded the Dakhovsky crystalline massif to thrusting onto Triassic rocks [17] (Fig. 4).

RESULTS

Data from Geological-Structural and Tectonophysical Studies in the Area of the Rufabgo Massif

Near the village of Kamennomostsky, the Belaya River and its tributaries cut through a system of deep canyons known as Rufabgo, in which the lower part of the Alpine cover section and underlying complexes of the pre-Alpine basement outcrop (Fig. 5).

The walls of the canyons are a stepped system of subvertical and overhanging scarps 20–60 m high, forming an escarpment of the outer contour of the distribution area of the Alpine cover. Walls of the canyons are composed of limestones and dolomites (Middle–Upper Jurassic, Gerpegem Formation), covered by argillites and siltstones with interlayers/lenses of sandstones, dolomites and gypsum (Upper Jurassic, Mezmai Formation). These strata, which form the lower part of the Alpine cover, gently (3°–10°) dip to the north.

At the bottom of the canyons, two Triassic strata of the transitional structural level were exposed, forming the Rufabgo uplift: thin-layered pelitomorphic limestones with thin interlayers of marl (Lower Triassic, Yatyrgvarta Formation) and predominantly terrigenous (siltstones, argillites, sandstones) strata with a horizon of polymictic conglomerates (Middle–Upper Triassic, Dakhovsky sequence) [17]. Along the contact of these strata, a steeply dipping fault has developed, which can be traced NNW along the bed of the Belaya River and is one of the elements of the Zakan shear-thrust fault zone (Fig. 5, Fig. 3).

Geological, structural, and tectonophysical studies were conducted in the area of the Rufabgo canyons along the bed of the Belaya River and its left tributary, Rufabgo Creek. The studied section is approximately 6 km from bottom to top.The results for key sites (1–5) within the section are given below.The position of sites is shown in Figs. 5a, 5b.

Site no. 1. This Site is located at coordinates 44.30139500° N/40.17650500° E–44.29678400° N/ 40.17478600° E. The beginning of the Rufabgo canyon in the Belaya River valley is located within the Kamennomostsky village. Here, the canyon wall is up to 20 m high and is composed of the Upper Jurassic red argillites,siltstones, and sandstones (Mezmai Formation). The layers dip to the north (3°–8°). Along individual horizons of argillites and along their contacts with sandstones, sublayer faults—detachments—are often observed. They are expressed in thin zones of shearing and boudinage of rocks. Along the breaks, feathered synthetic Riedel shears (R) are present. Some of these faults are gently sloping listric faults that merge smoothly with the fault surfaces and have displacement amplitudes of up to 1 m (Fig. 6a).

When synthetic Riedel shears intersect layers of sandstones, structures of pinch and swelling arise in the latter, associated with the initial stages of the formation of asymmetric boudins (Fig. 6a).

The results of tectonophysical studies in the area of site no. 1 showed three groups of structures, the kinematic parameters of which are presented on the stereographic projection (Fig. 7, graphic table).

The first group of structures unites the entire set of studied shear cracks and small ruptures that have kinematic signs of displacement. The group presents various paragenetic associations of disorders, which reflects some average characteristics of heterogeneous dynamic parameters.

The second group of structures contains a selection of kinematically (paragenetically) interconnected structures: sublayer detachments, their feathering synthetic Riedel shears and antithetic steeply dipping normal faults (Fig. 7, Fig. 6a).

The displacement vectors along gently sloping faults indicate predominant horizontal displacements of layer packages to the north. Statistical analysis of the kinematic parameters of the structures of this group on the stereographic projection allows us to reconstruct the predominant normal-reverse fault (transitional from nappe to subduction) deformation regime with an inclined position of the main deformation axes lying in the submeridional plane (Fig. 7, mini-stereogram).

In this case, the axis of maximum elongation (stretching) has a position close to horizontal, and the axis of compression is steeply inclined, which indicates the predominance of horizontal stretching conditions during the formation of detachments.

The third group includes two fault systems of submeridional and northwestern strike, which were formed under conditions of a normal fault deformation regime with a sublatitudinal (WSW–ENE) orientation of the extension axis. Judging by the observations in the outcrops, the relationships between the structures of the second and third groups are ambiguous, mutual-intersection.

Site no. 2. This site is located at coordinates 44.28378300° N/40.18028300° E. In the quarry, located on the southern outskirts of the village of Kamennomostsky, limestones and dolomites of the Middle–Upper Jurassic (Gerpegem Formation) were exposed (Fig. 5). In the southwestern wall of the quarry, two extended detachments are established, obliquely cutting the layers of the carbonate strata and having a gently dipping fault planes to the west-northwest (Fig. 8a).

The system of listric faults feathers the detachments. The detachment located in the lower part of the section is accompanied by a thick (more than 10 m) zone of fine lenticularization, crushing and mylonitization of limestones. The surfaces of the detachments are complicated by ramp scarps—accompanying (decompression) and preventing (compression) slipping. Above the decompression ramps of the lower structural level detachment, gentle subsidence synclines are developed in the limestones, cut off in the wings by normal faults. The compression ramp of the upper detachment controls the asymmetrical flat anticline (Fig. 8a).

Tectonophysical research in the area of site no. 2 showed the following results. Measurement group no. 1 characterizes the entire set of studied structures. The sample of structures associated with the formation of detachments is represented by the second group of structures. The displacement vectors along these gently sloping faults are directed NNW (Fig. 7).

The deformation mode is normal-reverse fault regime, the main axes of deformation are inclined and lie in a plane parallel to the direction of tectonic transport. The close position to horizontal plane of the axis of maximum elongation, as in the previous case (site no. 1), indicates the predominance of stretching conditions during the formation of detachments.

The third group of structures is characterized predominantly by a strike-slip deformation fault regime and is represented by a combination of right-hand strike-slip faults of north-west orientation, left-hand strike-slip faults of north-east orientation, and submeridional normal faults. The formation of faults occurred under conditions of horizontal meridional compression and latitudinal extension.

The detachment zones identified in the quarry located near the village of Kamennomostsky (site no. 2) have been traced far in the south-eastern and southwestern directions.

They are marked on the right rocky side of the Belaya River in the area of site no. 5 (Figs. 5, 8b).

Within the northern part of the Lagonaki plateau, in outcrops that expose the middle level of the Gerpegem Formation section, a powerful zone of tectonic boudinage is exposed, in the structure of which domino-type structures participate, indicating the slipping of layers in the northwest direction (Figs. 6b, 6c).

Site no. 3. This site is located at coordinates 44.28425900° N/ 40.17533000° E. 300 m to the west of site no. 2, at the bottom of the Rufabgo canyon, the Belaya River cuts through the lower part of the section of carbonate rocks of the Middle–Upper Jurassic (Gerpegem Formation) (Fig. 5).

Here, at the base of the layer of grey massive limestones and dolomites, there is a wide (more than 10 m) zone of tectonically reworked rocks thinly layered (lenticular-shaped) carbonate rocks and mylonites (calcmylonites), lenticular-banded marbled and brecciated limestones (Fig. 6d).

The zone is conformal to the layering, dips to the north at angles of 7°–10° and represents a major detachment. Numerous structural features indicate that the carbonate rock mass has experienced downward slipping along a detachment.

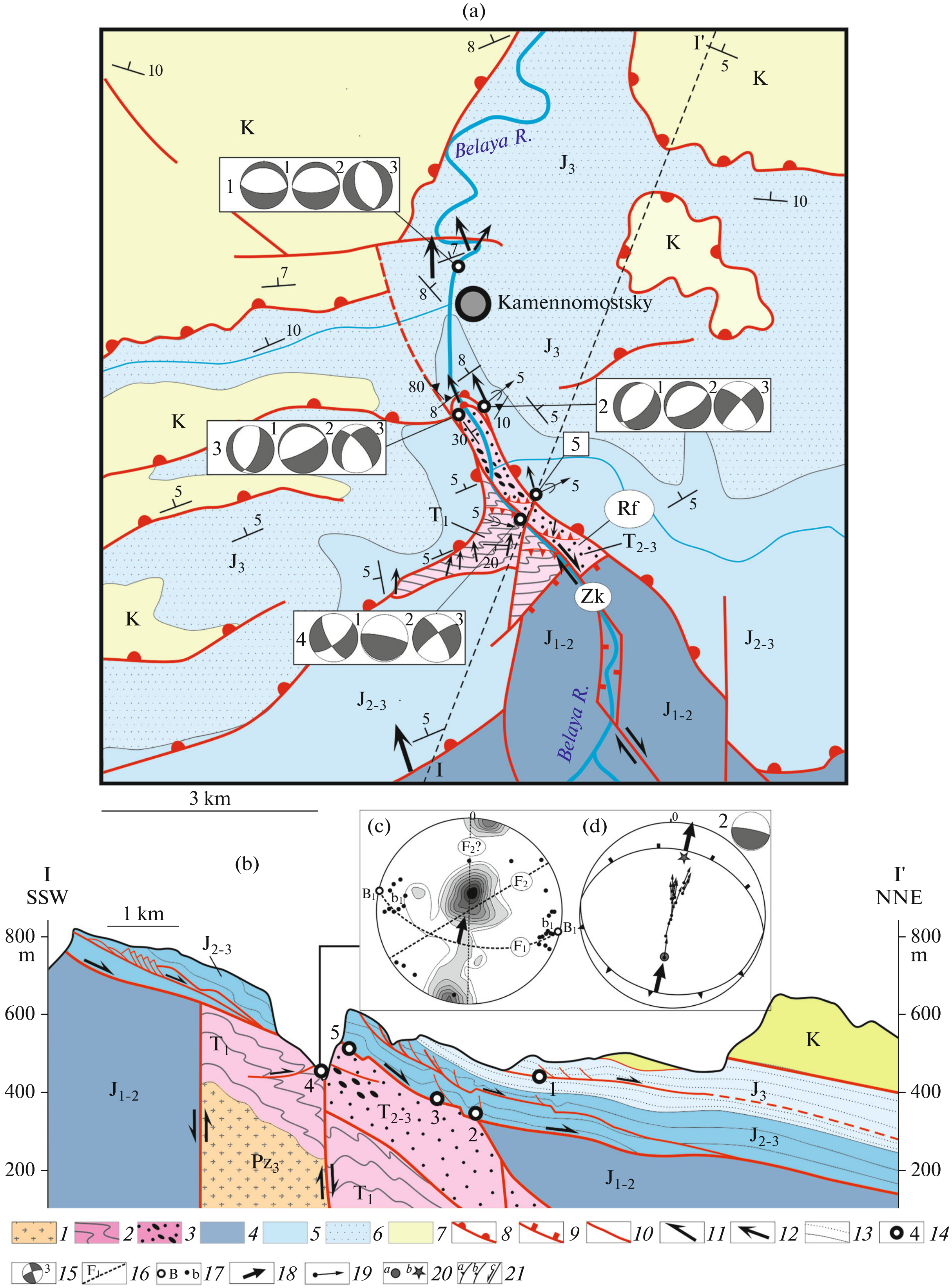
The structure of the tectonite zone involves conjugate systems of Riedel shears—synthetic (R) and antithetic (R') shears, which have normal and reverse kinematics, respectively (Figs. 6d, 6e).

Along the main slipping planes of the detachment zone and the faults that surround them, there are fragments of slipping mirrors. The foliation of the mylonites and the secondary lamination of the limestones are subparallel to the main faults. In calcmylonites and marbleized limestones, secondary banding and foliation are developed, as well as numerous meso- and microstructural kinematic indicators: domino structures and sigmoidal nodules, small symmetrical and asymmetrical boudins, as well as isolated fragments of asymmetrical folds of sublayer slip and C–S structures (Fig. 6f).

The linearity of the extension of lenticular separations (linearity of transport) is conformal to the surface of the detachment zone and is directed downwards along its dip.

Tectonophysical studies near site no. 3 were carried out in the zone of detachment development and in the massif of carbonate rocks of the Middle–Upper Jurassic lying above the section. The combined stereogram, which combines all measurements in the first group of structures, shows dynamically incompatible structures that cannot be statistically interpreted.

Fig. 5. Geological structure of Rufabgo canyon and adjacent territories along section I–I'. Position of section I–I', Fig. 1. Notation: RF, Rufabgo canyon; Zk, Zakan fault zone. (a) Geological and structural diagram of work site; (b) section along line I–I'; (c) equiareal stereographic projection of orientation of layers poles and fold hinges in carbonate strata of Lower Triassic (lower projection, 155 measurements, isolines 1.5, 3, 4.5, 6, 7.5, 9, 10.5, 12, 13.5%); (d) equiareal stereographic projection (lower hemisphere) of orientation of poles of faults with displacement vectors of hanging wall. *1*, Late Paleozoic granitoids (on section); *2*, Lower Triassic carbonate rocks; *3*, terrigenous rocks of Middle–Upper Triassic; *4*, sandy-shale strata of Lower–Middle Jurassic; *5*, mainly carbonate rocks of Middle–Upper Jurassic; *6*, terrigenous rocks of Upper Jurassic; *7*, carbonate and terrigenous rocks of Cretaceous; *8–10*, discontinuous faults: *8*, detachments; *9*, normal faults; *10*, other; *13*, layering (on sections); *14*, stakes of tectonophysical observations and their numbers; *15*, ministereograms of paleostress orientation of structural groups 1–3 (extension segment (gray), compression segment (white)); *16–18*, structural elements on stereogram (c): *16*, axial planes of first- (F_1) and second- (F_2) generation folds; *17*, fold hinges found geometrically (c) and by measurements in outcrops b_1; *18*, direction of vergence of structures; *19–21*, structural elements on stereogram (d): *19*, poles of faults with displacement vectors of hanging wall; *20*, axes: (*a*) tension; (*b*) compression; 21, planes of reverse faults and thrusts (*a*), normal faults (*b*), strike-slips (*c*).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Belaya R.
Kamennomostsky
K
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J2-3
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(d)
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1 km
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1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20 21

The second group of structures presents faults associated with the formation of the detachment zone (Fig. 7).

The displacement vectors along the main slipping surfaces and the synthetic Riedel shears (R) feathering them are directed to the NNW. Antithetic shears have reverse fault kinematics and reverse vergence relative to the direction of slipping along the detachment. The normal–reverse deformation fault regime has been reconstructed, and the angular relationships of the main deformation axes with the horizontal plane indicate the predominance of tension conditions during the formation of a large detachment zone developed at the base of the Alpine cover.

The third group of structures combines systems of strike-slip, normal-strike-slip, reverse and normal faults (Fig. 7).

This paragenesis of structures was formed under conditions of a strike-slip deformation fault regime with a submeridional position of the compression axis and a sublatitudinal orientation of the extension axis.

Site no. 4. This site is located at coordinates 44.27053300° N/40.18598400° E. On the left side of the Belaya River and its left tributary, Rufabgo Creek, thin-layered limestones of the Lower Triassic (Yatyrgvartinskaya Formation) outcrop, bounded from the northeast by one of the faults of the Zakans fault zone, which extends northwest (Fig. 5).

The thickness of limestones is complicated by a system of asymmetric folds associated with numerous sublayer faults and thrusts, which limit the units of layers with a separate fold structure (Figs. 6g, 6h).

The folds are disharmonious: their morphology and outlines of hinges change from sinusoidal to chevron even within one plicative structure (Fig. 6h).

In the transverse direction to the axial planes of the folds, one can observe alternating compressed, open, and flexurelike structures. Inclined and overturned folds, as well as small recumbent folds—thrusts—are common. Frequent, gently sloping undulation of the fold hinges is noted (Fig. 6h).

On the stereographic projection, the orientation of the layering poles of the considered Lower Triassic limestone formations form two dispersion belts (Fig. 5c).

The most representative belt in terms of the number of measurements is located along the arc of a large circle and characterizes the main system of cylindrical folds. Their axial planes have a WNW orientation and are inclined in southern directions. The second poorly expressed belt corresponds to the arc of a small circle and characterizes gently sloping conical folds with ENE extension of the axial planes. These structures are newer, overprinted on the folds of the main (first) generation. Their development caused the undulation of early folds, which can be seen in the nature of the distribution of their hinges on the stereogram (Fig. 5c).

In general, the entire system of early folds and associated thrusts has a general northern vergence (Figs. 5c, 5d).

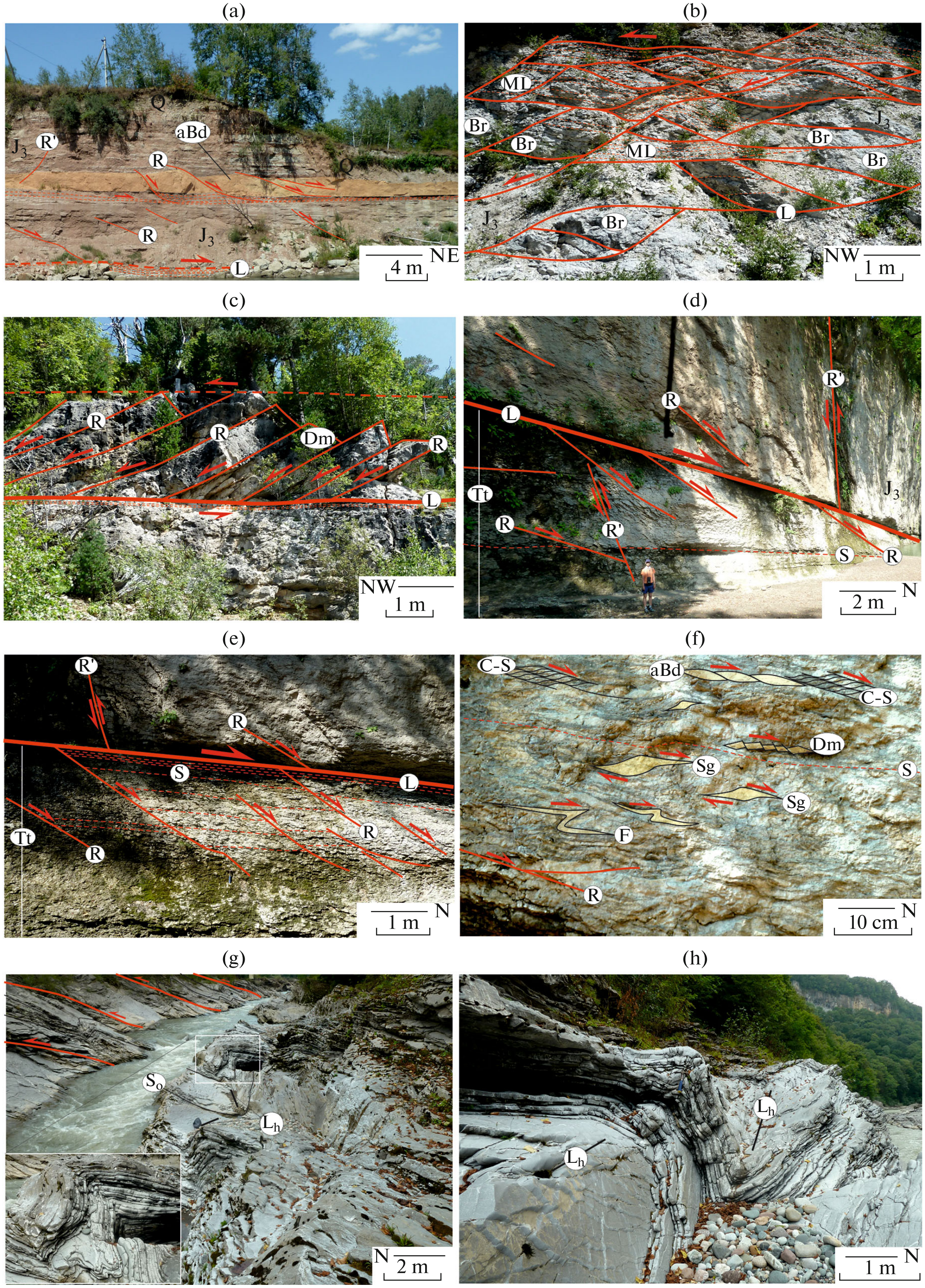
In this case, the fold–thrust system in the the Lower Triassic carbonate layer is oriented diagonally to the fault of the Zakan deformation zone, which developed along the Belaya River bed. On the northeastern wall of this fault, similar folds and thrusts are developed in the layers of a Middle–Upper Triassic terrigenous sequence, but they have an inverse southern vergence [17].

Hence it follows that these deformation structures are associated with right-lateral displacements along the faults of the Zakan zone. This is consistent with the fact that farther away from the strike-slip fault that developed in the Belaya River valley, in a southwesterly direction up the stream. Rufabgo, the amplitude of the folds gradually decreases. At a distance of 400–500 m from the fault, the folds are replaced by infrequent flexures and faults, which developed against the background of a monoclinal bedding.

Thus, the considered first-generation fold–thrust structures that developed in the Triassic strata are associated with right-lateral displacements along the Zakan deformation zone. As established in [17], this zone was formed during the Cimmerian stage of tectogenesis. The gently sloping second-generation plicative structures that developed in the Triassic limestones are probably associated with the Alpine stage of development.

This is confirmed by the fact that 500 m northeast of site no. 4, gently sloping folds with a similar ENE orientation of the axial planes developed in the carbonate layer of the Alpine cover (Fig. 5, site no. 5).

Fig. 6. Field photographs of structures associated with development of gently sloping tectonic faults. Notation: L, main detachment fault; R, R', synthetic and antithetic Riedel shears, respectively; S_o, layering; S, foliation; C-S, structures; F, folds; L_h, fold hinges; Tt, tectonite zones (mylonites, breccias, etc.); Sg, sigmoidal structures (nodules); aBd, asymmetric boudins; Dm, domino structures. (a) Sublayer detachments and their en echelon arrangement of synthetic Riedel shears (R) in variegated layer of Upper Jurassic (Mezmai Formation, northern part of Rufabgo canyon); (b, c), zones of boudinage and development of domino structures along detachments in middle part of section of Middle–Upper Jurassic carbonate strata (Gerpegem Formation, northeastern part of Lagonaki plateau); (d, e) large detachment (L) and associated synthetic (R) and antithetic (R') Riedel shears at base of section of carbonate strata of Middle–Upper Jurassic (Gerpegem Formation, middle part of Rufabgo canyon); (f) mesostructural kinematic indicators in zone of tectonite development (banded marbleized limestones and calc-mylonites) along detachment at base of section of carbonate strata of Middle, Upper Jurassic (Gerpegem Formation, middle part of Rufabgo canyon); (g, h) asymmetric folds and thrusts in sequence of thin-layered limestones of Lower Triassic (Yatyrgvarta Formation, southern part of Rufabgo canyon), rectangular contours with white line.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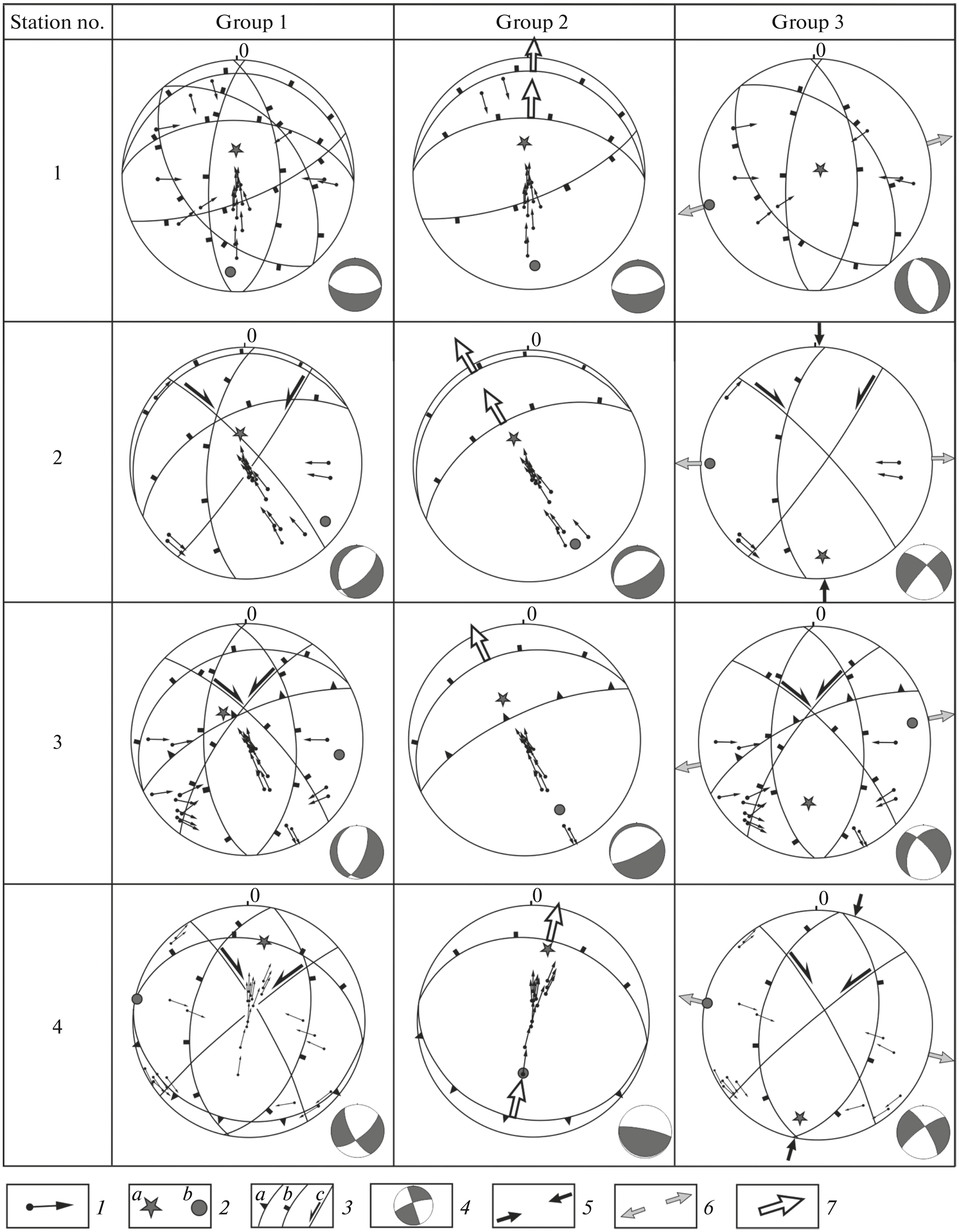


Fig. 7. Equiareal stereographic projections (lower hemisphere) of fault poles with hanging limb displacement vectors for structures of groups 1–3. Shown: group 1, all measurements; group 2, paragenesis of structures associated with detachments; group 3, paragenesis of faults mainly with strike-slip, normal, and combined kinematics. *1*, Poles of fault planes and displacement vectors of hanging limb; *2*, principal axes of paleostresses: (*a*) compression; (*b*) tension; *3*, average positions of fault planes: (*a*) reverse faults and thrusts; (*b*) normal faults; (*c*) strike-slips; *4*, ministereograms of paleostress orientation (segments: (gray) tension, (white) compression); *5–6*, projections of principal axes of deformation onto horizontal plane: *5*, compression axes; *6*, tension axes; *7*, directions of horizontal (tangential) movements.

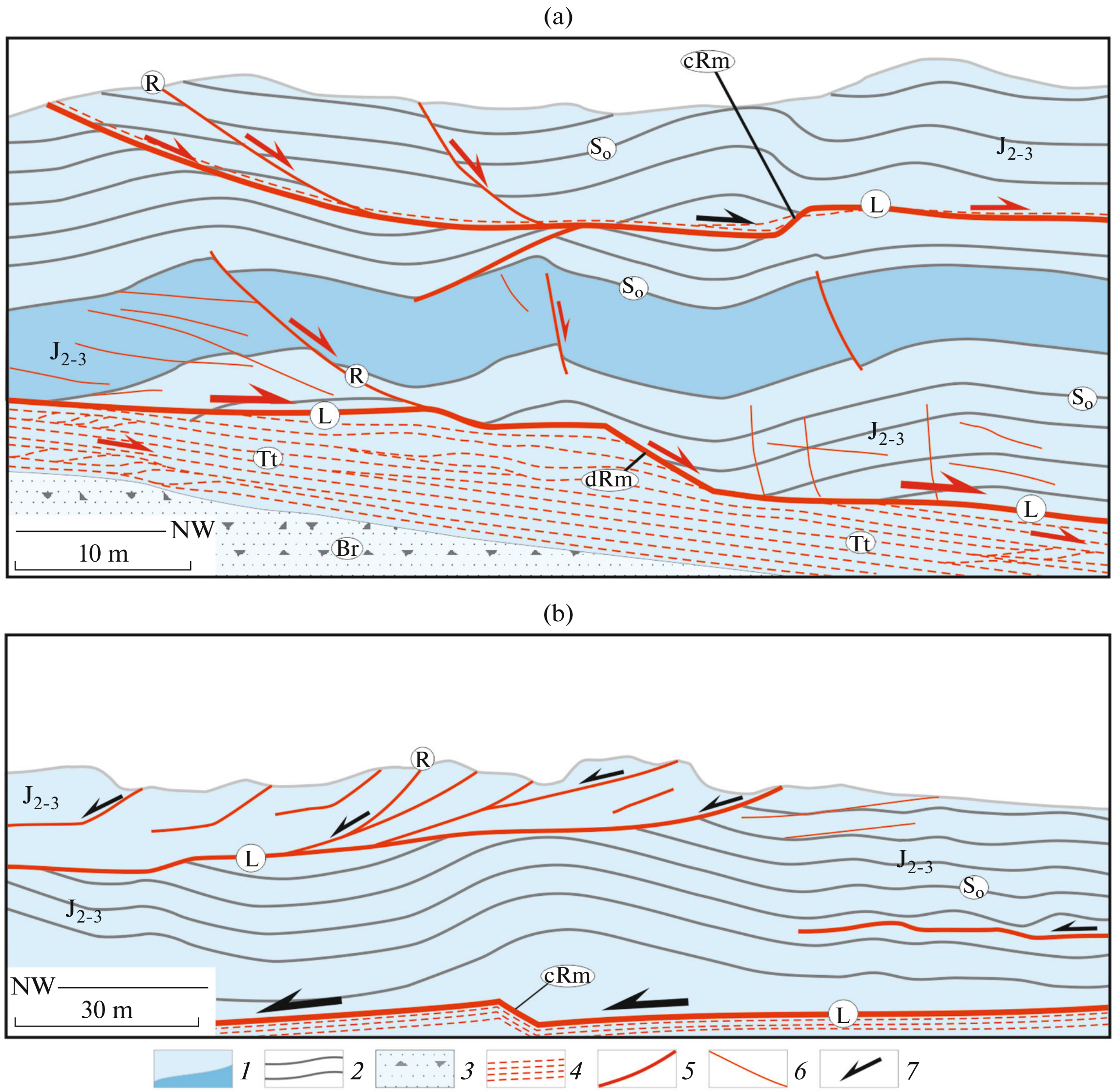


Fig. 8. Sublayer detachments and associated structures in carbonate strata of Middle–Upper Jurassic (Gerpegem Formation). Notation: L, main detachment faults; R, synthetic normal faults; cRm, compression ramps; dRm, decompression ramps; So, layering; Tt, tectonite development zones; Br, breccias; So, layering. (a) Wall of a quarry on southern outskirts of village of Kamennomostsky (site no. 2); (b) rocky scarp on right side of Belaya River in southern part of Rufabgo canyon (site no. 5). *1*, Limestones and dolomites of Middle–Upper Jurassic; *2*, layering; *3*, breccias; *4*, zones of tectonite development (schistosity, boudinage, mylonitization, brecciation of rocks); *5*, faults; *6*, fractures; *7*, directions of displacements along faults.

These structures reveal a clear paragenetic relationship with the sublayer detachments of the cover, or more precisely, with ramp scarps that prevent the slippage of layers in the NNW direction (Fig. 8b).

The gently sloping second-generation folds that developed in the Triassic rocks were probably formed as a result of the dynamic influence of Alpine cover detachments.

The results of the tectonophysical studies in the vicinity of site no. 4 are presented in three groups (Fig. 7).

The first group of structures combines the entire set. Most of them are dynamically coordinated and can be considered a single paragenesis that formed under conditions close to a strike-slip deformation fault regime.

The second group includes thrusts and sublayer faults that formed under conditions of a thrust (normal–reverse) deformation regime as a result of subhorizontal movements in the northern direction.

The third group combines systems of dextral and sinistral strike-slip faults with a northwestern and northeastern strike, respectively, as well as a system of submeridional normal faults. The normal-shear deformation regime was reconstructed under conditions of sublatitudinal (WNW–ESE) extension and submeridional (NNE–SSW) compression.

SEISMOSTRATIGRAPHIC ANALYSIS OF THE STRUCTURE OF THE SOUTHERN PART OF THE WEST PRE-CAUCASUS BASIN

The Greater Caucasus orogen is framed in the north by large and deep depressions, which most researchers consider marginal depressions associated with the formation of this orogen [1, 9, 26, 30].

West Kuban Trough

The West Kuban trough is the eastern part of the Indolo-Kuban trough and can be traced from the waters of the Sea of Azov ESE to the Adygean uplift (Fig. 1).

The West Kuban trough is more than 250 km long and 90 km wide. In the north, the trough is separated from the Timashevskaya step by the Novotitarovskaya flexural-fault zone, which is arcuately curved and in the south-eastern direction passes along strike into the Tsitsa fault zone, traced within the northern slope of the Greater Caucasus [2, 17]. In the south, the West Kuban trough along the Akhtyr fault zone is connected with the Soberbash–Gunai synclinorium of the Northwest Caucasus.

In cross section, the West Kuban trough has a sharply asymmetric structure (Fig. 9a).

Its northern board is gently sloping and weakly dislocated, while the southern board is steep and has a complex structure. A particularly sharp asymmetry is noted in the west, where the axial part of the trough at the level of the Maykop Group is closely situated to its southern board, complicated by the fault system of the Akhtyr zone.

Some researchers consider the Akhtyr zone as a complex imbricated fold–thrust structure that formed due to thrusting of the strata of the marginal part of the Western Caucasus onto the southern board of the West Kuban trough [27, 28, 34]. Others note that the Greater Caucasus is hardly anywhere thrust onto Pre-Caucasus troughs and, in particular, the upper part of the Akhtyr fault plunges steeply (75°–80°) to the south, and at greater depths subvertically passes into the mantle [47].

In the axial part of the West Kuban trough, there is the longitudinal Anastasiev–Krasnodar rootless anticline zone, which divides the western part of the trough into two basins: the sharply asymmetrical and wide Slavyansky–Ryazansky in the north and the narrow Adagumo-Afipsky in the south. On the continuation of the Anastasiev–Krasnodar anticlinal zone to the east is the Shapsug–Apsheron paleouplift (Fig. 1).

Within the southern board of the West Kuban trough (Adagumo-Afipsky depression), five anticlinal zones are distinguished, they have subsided stepwise to the north. Some of them are overlain by thrusts that developed within the Akhtyr fault zone, along which the marginal part of the Western Caucasus has been thrust onto the West Kuban trough [24, 31]. To the east of the anticlinal zones the Khadyzhensky block with a monoclinal structure is situated (Fig. 1).

In this area, fold–thrust deformations associated with the Akhtyr zone have not been established.

In almost all Cenozoic strata of the West Kuban trough, numerous buried paleoscarps and obliquely layered seismic complexes, which represent clinoforms, are noted (Figs. 9a, 9b).

The inclination of the oblique series within the clinoforms, corresponding to the direction of their progradation, as well as scarps, are directed to the south. These formations can be considered as paleodeltas (alluvial fans) composed of sediments transported from the north into the shallow shelf of the southern margin of the East European continent. In the seismic record, clinoforms can be fairly confidently interpreted in the Paleogene, Miocene, and Pliocene strata (Fig. 9a, 9b).

The seismic section also shows that, in addition to the lateral progradation of paleodeltas within individual horizons, their southward-directed migration is also observed in the vertical section of Cenozoic strata at different stratigraphic levels (Fig. 9a), probably due to the periodic southward retreating of the coastline (Eastern European continent), beginning in the Oligocene to the Pliocene.

In the strata of the Upper Miocene (Pontian regional stage) and Lower Pliocene, deep (up to 100 m) paleoincisions are noted on seismic sections, within which chaotic seismic complexes are present (Fig. 9a).

We consider these formations as paleovalleys filled with alluvial sediments, chaotic slope-landslide (olistostrome) complexes and their rewashing boulder–pebble products. The deep incision of the valleys occurred as a result of a sharp drop in the erosion base during periods of regression of the Paratethys marine basin [57, 62].

This corresponds to the idea that, beginning in the Middle Miocene (Sarmatian), the marine basins of Ciscaucasia experienced periodic regressions, during which subcontinental (subaerial) sedimentation conditions were established in some parts of the basins.

The rootless Anastasiev–Krasnodar anticlinal zone is complicated by longitudinal faults and manifestations of diapirism and intense mud volcanism [28]. In the base of the Maykop Group and in deeper horizons, this anticlinal zone is not expressed and corresponds to a depression. In its western part, large, high-amplitude, rootless diapir folds (Kurchan, West-Anastasiev, and Anastasiev–Troitsk) are distinguished, their cores consist of Maykop clays.

Some of these structures are expressed in the section from the Oligocene (lower part of the Maykop Group) to the Anthropogene. In the eastern direction, the Anastasiev–Krasnodar anticlinal zone vanishes. Only small folded forms can be found here, the amplitude of which in the Miocene layers does not exceed 25 m [28].

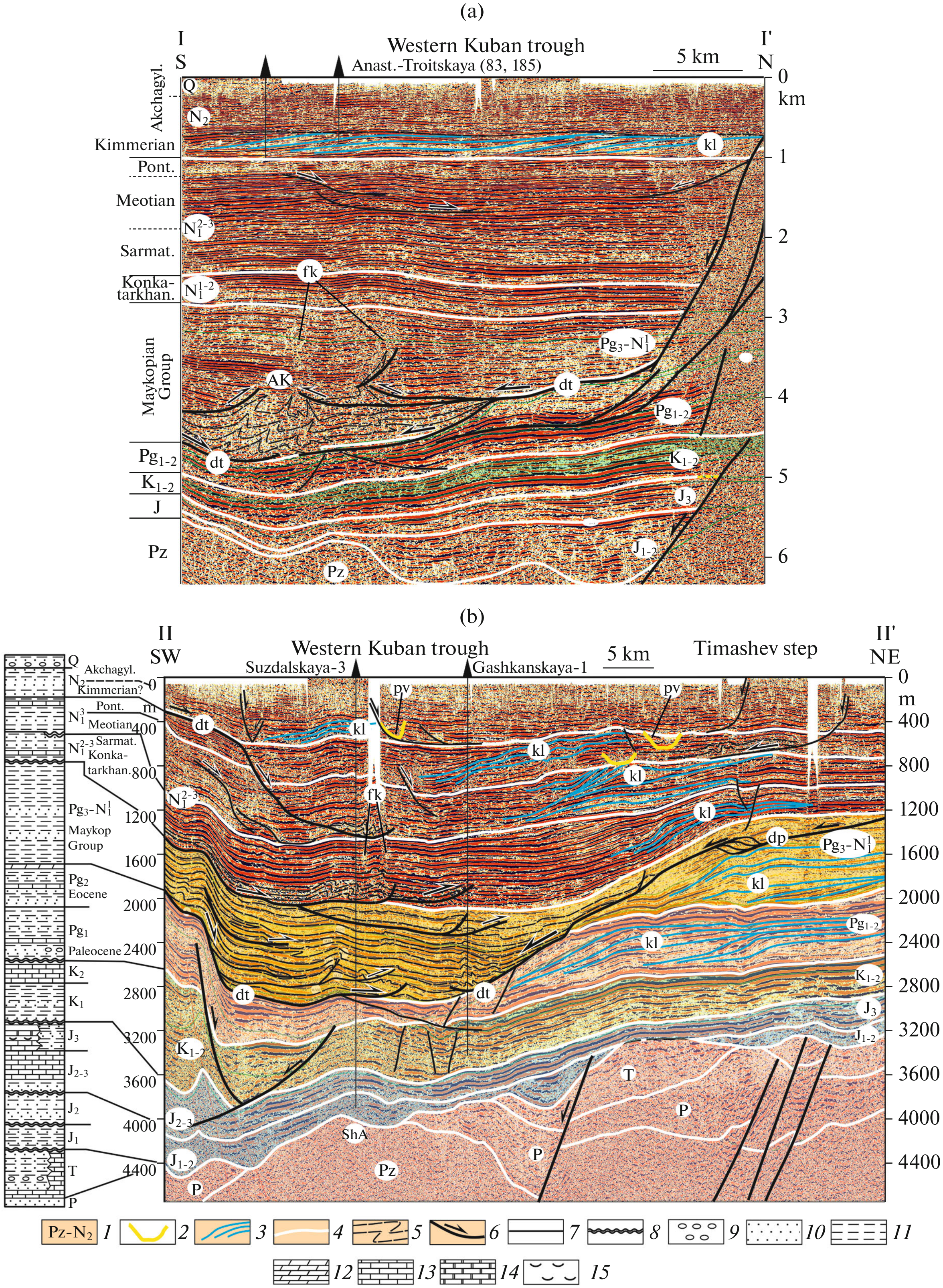


Fig. 9. Seismostratigraphic sections of West Kuban trough along lines of sections I–I' and II–II'. (a) Section I–I' (section position, Fig. 1); (b) section II–II' (position, Fig. 1). Notation: dt, detachments; dp, duplexes; fk, flame-shaped structures; kl, clinoforms; pv, paleo-incisions. *1–6*, Designations on seismic sections: *1*, age indices; *2*, paleosections; *3*, clinoforms; *4*, stratigraphic boundaries; *5*, layers; *6*, faults; *7–15*, designations on stratigraphic column: *7*, concordant stratigraphic contacts; *8*, unconformable bedding; *9*, gravelstones and conglomerates; *10*, sandstones; *11*, aleuropelites; *12*, dolomites; *13*, limestones; *14*, marls and clayey limestones; *15*, evaporites.

The Shapsug–Apsheron uplift is located on the continuation of the Anastasiev–Krasnodar anticlinal zone. To the east it is a paleouplift buried under Cenozoic strata, and is clearly identified on a seismic profile (Fig. 9a).

In the apical part of the uplift, a threefold decrease in the thickness of the Paleocene and Eocene strata was noted, indicating their local erosion. Conversely, in the overlying, predominantly clayey part of the Maykop Group, we can see a significant increase in its thickness with respect to the boards of the West Kuban trough. At the base of the Maykop Group, above the uplift, there is a depression.

In the northern board of the depression, a gently sloping fault dipping to the south and obliquely intersecting the section of the Maykop Group is clearly visible in the seismic record. This fault down dip turns into a sublayer detachment, which bounds the Maykop Group from below (Fig. 9a). Immediately below and above the detachment surface, the seismic record has a corrugated appearance, probably indicating the presence of small folds along the detachment in the over- and underlying layers.

On the northern continuation of the gently sloping fault in the area of the ramp that complicates it and prevents slippage, a system of thrust compression duplexes has been identified: kinematic signs of southward-directed slipping of strata downslope. The significant decrease in the thickness of the seismic complex, represented here by the Maykop Group, is spatially related to the zone of development of a gently sloping fault. We explain this situation by lateral extension and tectonic shortening of the Maykop Group section along the fault in question in accordance with the mechanism of formation of asymmetric boudins.

In the southern board of the West Kuban trough, a steeply dipping flexural fault zone is developed, along which the layers of the Maykop Group also thin significantly. To the north of this zone, numerous faults of the synphase axes are noted in the seismic image of the Maykop Group, manifesting branching, doubling, and attenuation, as well as in division of seismic reflections into short dotted segments, outlining small folded forms. The asymmetry of some folds allows us to associate them with sublayer faults and thrust displacements in the north direction (Fig. 9a).

We interpret this system of structures (gently sloping southern dipping normal fault and steeply northward dipping flexural normal fault zone) as the result of slipping of plastic clay rocks of the Maykop Group from the slopes of the West Kuban trough along systems of detachments, gently sloping normal faults and thrusts developed at the base and within the section of the Maykop Group (Fig. 9a).

As a result, tectonic crowding of material occurred in the axial part of the depression, which caused a significant increase in the thickness of the seismic complex, represented by the Maykop Group.

In our opinion, it is possible in a similar way to explain the formation of the Anastasiev–Krasnodar anticlinal zone, formed by rootless folds—clay diapirs that developed in the axial part of the West Kuban trough. On the seismic profile, such folds are visible in the lower part of the section of the seismic complex represented by the Maykop Group, in the part of the profile where detachments of the northern and southern slopes of the trough converge.

The counterslipping of clay masses along detachments caused the development of rootless plicative structures in the area of crowding and suturing of faults with opposite signs of displacements (Fig. 9b).

Higher upsection, the folds decay, but are replaced by flame-shaped structures, probably corresponding to zones of increased permeability of fluid-gas flows, or water-saturated and plasticized Maykop clays that form that diapirs (Figs. 9a, 9b).

In the Miocene and Pliocene strata overlying the Maykop Group, it is also possible in seismic profiles to recognize systems of step and listric faults, which turn into local sublayer detachments (Fig. 9a).

The offsets along these faults are significantly smaller than in the thickness of the Maykop clays, but they have the same dynamic trend associated with slipping of geomasses from the sides of the West Kuban trough into its axial part. This allows us to suggest that the entire fault system in the Cenozoic sediments formed no earlier than the Pliocene. The intensity of these dislocations is probably due to the rheological properties of the rocks subjected to these dislocations.

Adygean Uplift

The Adygean uplift is a submeridional tectonic structure located between the West Kuban trough and the East Kuban depression [33]. The uplift is bounded by faults: from the west by the Kurdzhips fault, and from the east by the Khodzin fault (Fig. 1).

The Adygean uplift is characterized by a sharp reduction in thickness of the Mesozoic, Paleocene, and Eocene sequences. However, in the overlying horizons, including the Maykop Group, the uplift is not expressed (Fig. 10a), since in the Oligocene, the region within which the Adygean uplift is located generally subsided [2, 25].

In cross section, the Adygean uplift is a slightly asymmetric positive structure with a steeper western and gently sloping eastern slope (Fig. 10a).

In addition to the large flank faults that limit the uplift, it is complicated by numerous smaller ruptures that infrequently penetrate the Cenozoic strata, but control gently sloping plicative flexures in them. As a result, the seismic image of the cover in the area of the Adygean uplift has a parallel-wavy internal structure associated with the development of gently sloping submeridional folds. Within various horizons of the cover, these folds are harmonious and cover the entire column of rocks visible in the section from the Upper Jurassic to the Cenozoic. No signs of structural unconformities were noted (Fig. 10a).

On the seismic sections longitudinal to the strike of the Adygean uplift, a gently sloping monocline is visible, formed by rocks of the cover, including complexes of its syn-Alpine part (Figs. 10b, 10c).

The difference in absolute elevations of the base of the cover from south to north on profile sections 30–35 km long is 2400–2500 m. Transverse (sublatitudinal) faults in the Adygean uplift have been traced only in basement complexes and the structural stage of the Cimmerian.

In the strata of the syn-Alpine cover, systems of parallel and gently undulating reflections. Along the boundaries of lithostratigraphic units, with sharply different rheological properties, finely corrugated, often asymmetrically oblique, seismic complexes are expressed (Figs. 10b, 10c).

Based on the bends and displacements of reflections, small asymmetric folds and small thrusts can be identified among them. These structures gradually decay upwards along the section. They are replaced by structures with wavy and parallel reflections. Dislocations of this type are manifested at several stratigraphic levels: in the Upper Jurassic, in the Maykop Group, and in the Middle–Upper Miocene. The manifestation of these sublayer faults determines the general structural disharmony of the cover complexes, probably due to the development of detachments along the boundaries of rheologically contrasting strata (Figs. 10b, 10c).

In the northern and southwestern parts of the Adygean uplift, in the strata of the Early Alpine structural stage (Callovian–Eocene), sublatitudinal linear folds are developed with the limbs dipping up to 20°. In this case, the axes of the anticlines are displaced to the north by 2–3 km relative to structures with similar geometric parameters, expressed in the strata of the Cimmerian stage. Similar cases of horizontal displacement of fold axes were noted for two anticlines developed in the lower parts of the seismic complex section represented by the Maykop Group (upper Alpine level) in the north of the scarp. These structures, while maintaining a parallel orientation with the anticlines of the lower Alpine stage, are shifted relative to them by 1–1.5 km [17].

Such examples of lateral dissociation of plicative structures probably indicate horizontal displacements along detachments in the north direction. The total amplitude of the relative displacement of the upper elements of the section of the Alpine cover, taking into account only these structural benchmarks (folds), can be estimated at 3–4.5 km. This does not take into account possible displacements along the detachments of other structural levels of the Alpine cover, as well as lateral movements associated with volumetric flow and redistribution plastic rocks, e.g., clays of the Maykop Group.

Other signs of detachment manifestation are zones of intermittent (dotted), lenticular, and tiled reflections. The last can be compared with extension duplexes or domino-type structures formed in detachment zones at the base of Cretaceous and Paleogene strata (Fig. 10c, right).

Detachments and associated fold–thrust faults in the section are manifested unevenly and not always along uniform rheological boundaries. The attenuation of detachments or their transition to another structural level of the section usually occurs along ramp scarps. Above the scarps that prevent slipping (compression ramps), ramp folds are developed, which compensate for displacements. Examples of such structures are noted at the base of seismic complexes represented by the Upper Jurassic and middle of the Maykop Group section (Figs. 10b, 10c).

Ramp scarps of detachments, accompanying slipping (decompression unloading ramps), cause the formation of accompanying local extension structures: faults and subsidence synclines. A large unloading ramp, represented by a well-defined system of step faults in the seismic record, connects sections of detachments at the base of seismic complexes represented by the Cretaceous and lower part of the Upper Jurassic section (Fig. 10c).

In the upthrown wall, a laterally extended (7 km) massif of Upper Jurassic rocks is reflected in the seismic record as a complex boudinage system of reflections. We identify such a seismic image as a thick (400 m) zone of sublayer boudinage, development of boudins, and longitudinal extension, located above a flat detachment surface. Farther along the dip of the detachment at the base of the boudinage zone, its surface forms a compression scarp, above which the Jurassic layers form ramp asymmetric folds complicated by thrusting (Fig. 10c).

The ensemble of these structures represents a single compensatory paragenesis. This paragenesis can be characterized as a dynamically coupled triad: (i) zone of normal fault detachment (unloading ramp)–zone of longitudinal extension and (ii) slipping (transport) along inclined detachment–squeezing, as well as (iii) the development of ramp fold–thrust structures (Fig. 10c).

Similar types of triads (normal fault/rupture–transport–squeezing) have been established in other parts of the Cenozoic section within the Adygean uplift, in particular, in the lower stratigraphic levels of the Maykop Group. Here, a combination of listric faults is expressed, passing down dip into detachments, on the lateral continuation of which ramp squeezed structures are developed (Fig. 10c).

East Kuban Depression

The East Kuban depression is bounded in the west by the Kanevsky–Berezan system of uplifts and the Adygean uplift, and in the east by the Mineralovodsky scarp and the Stavropol uplift, located on the continuation of the Transcaucasian transverse uplift. The depression stretches from northwest to southeast for approximately 250 km, its width reaches 70 km. The depth of the East Kuban depression along the base of the cover exceeds 6500 m. The structure of the depression includes deposits from the Lower Jurassic to the Quaternary.

The Upper Jurassic deposits of the central East Kuban depression host evaporite strata with a considerable thickness, as well as reef structures and their collapse trails (clastic limestones), controlled by buried fault scarps (Fig. 11).

Farther from the slopes of the Greater Caucasus, the thick strata of the Middle and Upper Jurassic of the central East Kuban depression are cut off by the pre-Cretaceous denudation surface on the sides of this depression (Fig. 11).

At the same time, on the slopes and in the apical parts of the uplifts framing the East Kuban depression, formations of the Cimmerian and, sometimes, Hercynian structural levels were subjected to erosion. Favorable combinations of various factors make it possible to predict the possibility of localizing hydrocarbons in the walls of the East Kuban depression in traps of structural (anticlines, flexures), lithological (reef and clastic limestones covered by screening horizons), and stratigraphic (disconformities) types (Fig. 11).

On seismic sections crossing the East Kuban depression, detachments were identified at the base of seismic complexes represented by Paleocene and Oligocene–Miocene strata (Maykop Group). The formation of these detachments is obviously caused by the slipping of rock masses in the northern directions from the slopes of the Adygean uplift and the Caucasian orogen. In this case, indicators of detachments are large structurally isolated seismic complexes, expressed in the multiple doubling and crowding of rock volumes mass with sigmoidal outlines and bounded by inclined reflecting surfaces (Fig. 11).

We consider such structures as thrust duplexes, the formation of which is caused by sublayer failures and multiple accumulation of layer packages. We consider the characteristic properties of these structures to be the cross-cutting of layers (reflections) by gently sloping discontinuities in the lower and upper parts of the duplexes, as well as the development of antiform structures (squeezed structures) in the overlying strata. The latter indicates that the thrust duplexes were formed after the accumulation of overlying sediments. As a result of the crowding of tectonic plates within the duplexes, local uplifts were formed above them, squeezed antiforms (Fig. 11a, fragment).

The considered of structures (duplexes, antiforms) are confined to high-gradient areas of decreasing intensity of inclination of rock stratification, as well as their slip surfaces—detachments.

DISCUSSION

It is known that classic marginal (piedmont) depressions (Pre-Ural, Pre-Appalachian, Pre-Alpine, Pre-Carpathian, etc.) are located in between mountain ranges and adjacent platforms [29, 37, 44]. However, the Greater Caucasus, the foreland of which is the southern margin of the Eastern European continent, is characterized mainly by an inverse southern vergence [5].

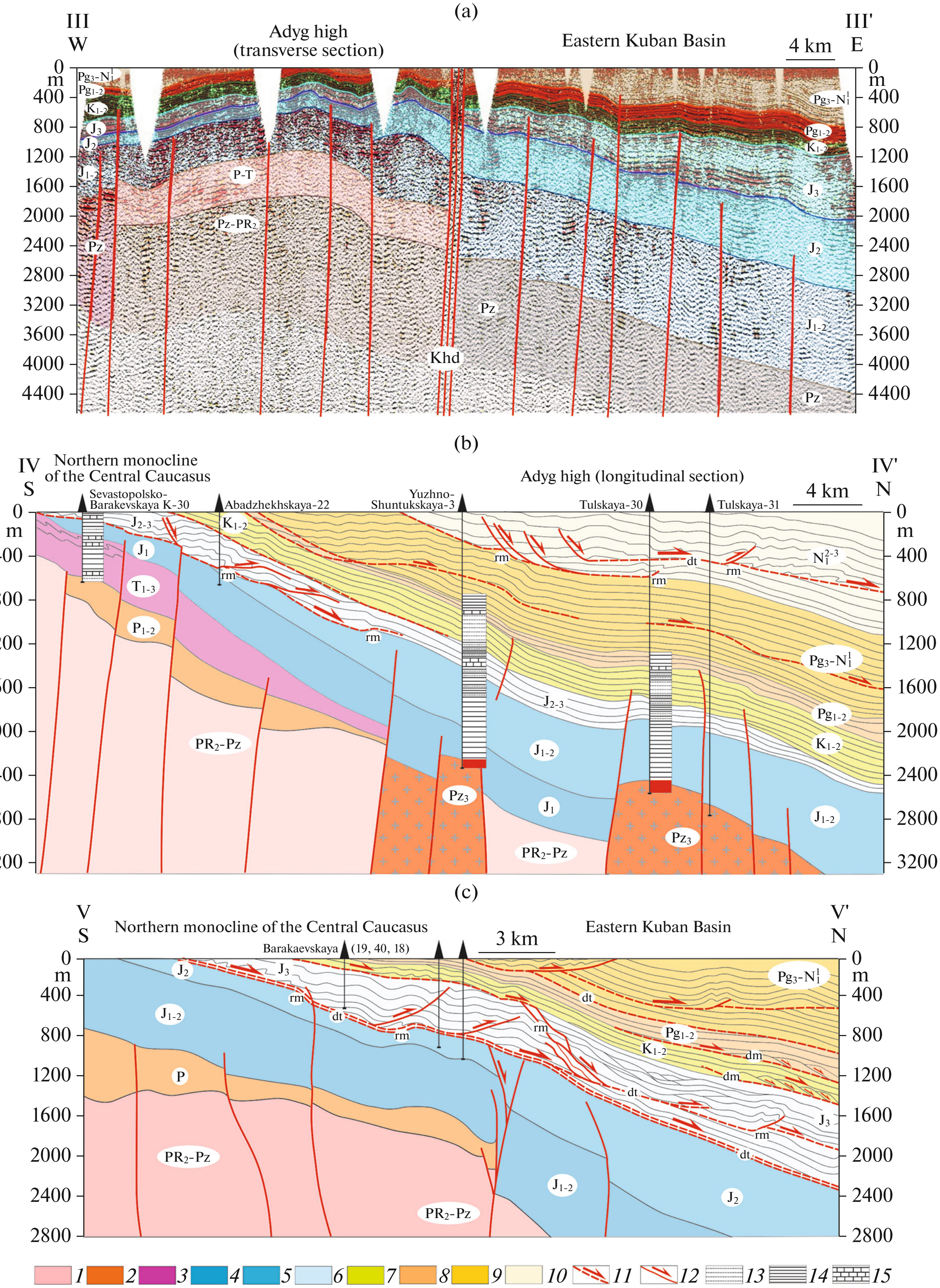
The Pre-Caucasus Basin are not typical in that they are located not in the front of but behind the main system of thrusts of the orogen. In addition, the depth of the troughs is inversely related to the mountain altitudes. Maximum depths are along low altitudes of the Greater Caucasus while minimum depths are adjacent to the high altitudes of the central part of the Greater Caucasus [9].

The considered materials allow us to suggest that the formation of the troughs of Western Pre-Caucasus was not associated with orogeny, at least up to the Pliocene, inclusive. This conclusion agrees with authors' pilot studies [19, 20]. These, to a significant extent, novel ideas touch upon questions about the time of onset and conditions of the Greater Caucasus orogeny.

The numerous buried scarps and clinoforms that we have identified on seismic sections of the eastern part of the West Kuban trough can be considered as relicts of the paleodeltas of ancient river systems, which transported detrital material from north to south—from the vast area of the East European continent to its southern margin (shelf area) (Fig. 9).

The distribution of clinoforms in the Cenozoic filling of the West Kuban trough allows us to assert that sedimentary flows were directed from the East European continent, at least from the Paleogene to the Early Pliocene, inclusive. At the same time, in the eastern part of the West Kuban trough, the migration of paleodeltas in the southern direction to the foothills of the modern Greater Caucasus (which did not exist

Fig. 10. Interpretation of seismostratigraphic sections crossing Adygean uplift and its eastern limb along lines of sections: (a) Section III–III'; (b) section IV–IV'; (c) section V–V'. Position of sections, Fig. 1. Notation: dt, detachments; rm, ramps; dm, domino structures. *Rock Complex (1–9)*: *1–2*, Hercynian (Pz–PR_2, R); *3*, transitional (T_{1-3}); *4–5*, Cimmerian (J_{1-2}); *Alpine Cover* *(J_3–N_1^{2-3}) (6–10)*: *6*, Middle–Upper Jurassic (J_{2-3}) and Upper Jurassic (J_3) thickness; *7*, Cretaceous strata (K_{1-2}); *8*, Lower–Middle Paleogene strata (Pg_{1-2}); *9*, Oligocene–Lower Miocene Maykop Group (Pg_3–N_1^1); *10*, Middle–Upper Miocene strata (N_1^{2-3}); *11*, detachments; *12*, faults; *13–15*, composition of rocks in well section: *13*, sandstones; *14*, aleuropelites; *15*, limestones.


(a)
III
W
Adyg high
(transverse section)
Eastern Kuban Basin
4 km
III'
E
0
m
400
800
1200
1600
2000
2400
2800
3200
3600
4000
4400
P-T
Pz-PR2
Pz
Khd
J3
J2
J1-2
(b)
IV
S
Northern monocline
of the Central Caucasus
Adyg high (longitudinal section)
IV'
N
Sevastopolsko-
Barakevskaya K-30
Abadzhekhskaya-22
Yuzhno-
Shuntukskaya-3
Tulskaya-30
Tulskaya-31
J2-3
K1-2
J1
T1-3
P1-2
PR2-Pz
Pz3
rm
dt
(c)
V
S
Northern monocline of the Central Caucasus
Barakaevskaya
(19, 40, 18)
3 km
Eastern Kuban Basin
V'
N
J2
J3
J1-2
P
PR2-Pz
Pg1-2
K1-2
dm
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15

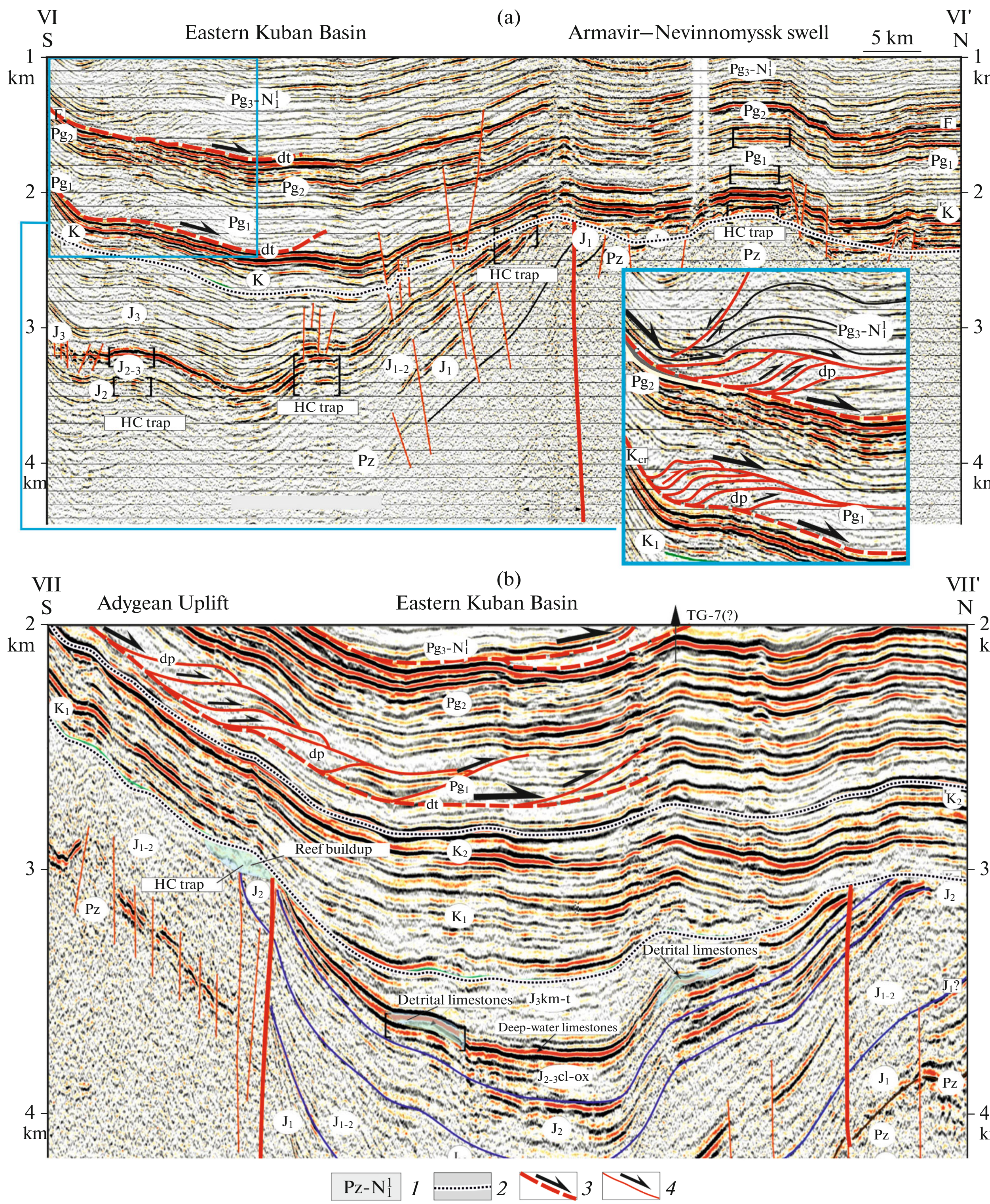


Fig. 11. Seismostratigraphic sections of East Kuban depression along lines of sections VI–VI' and VII–VII'. (a) Section VI–VI'; (b) section VII–VII'. Position of sections—see Fig. 1. Notation: dt, detachments; dp, duplexes. *1*, Age indices; *2*, boundaries of unconformity; *3*, detachments; *4*, faults.

up to the Pliocene) has been established (Fig. 9a). This confirms the previously obtained results from analyzing seismic profiles of the western part of the West Kuban trough [19, 36].

Let us emphasize here once again that in the Pliocene, the border area of the shallow and deep-water shelf, where the paleodeltas formed, was as close as possible to the area of the northern foothills of the modern Greater Caucasus. Farther south, this border area was probably represented by the deep-water shelf and continental slope of the East European continent, the sedimentary complexes of which covered the base of the future Greater Caucasus orogen, but were subsequently removed.

Sedimentary coarse-grained polymictic formations, which can be compared with typical orogenic molasse compose only the uppermost part of Western Ciscaucasia sedimentary sequence. They have insignificant thickness and contain lithological and isotope-geochronological (U-Pb dating of detrital zircon) signs of the flow of detrital material from the Greater Caucasus orogen [2, 20, 21]. Study [21] provides data that allows us to consider that the accumulation of these coarse-grained polymictic formations began no earlier than the Eopleistocene.

From this preliminary information, it is possible to formulate a number of questions that require further analysis.

— The areas within which the western segment of the Greater Caucasus orogen and the troughs of Western Pre-Caucasus are now located, in the Mesozoic and Cenozoic up to the end of the Pliocene, represented a marginal continental sedimentary basin (the western part of the Crimean–Caucasian basin, according to [19]), which was part of the Paratethys megabasin. In the basin terrigenous material brought from the East-European continent accumulated. By the end of the Neogene the Upper Mesozoic–Cenozoic sedimentary sequence (cover of the Scythian Platform) was several kilometers in thickness.

— The Greater Caucasus orogen began to rise no earlier than the Pliocene, and possibly even later, only at the beginning of the Quaternary (2.6–2 Ma (?)). The orogen experienced rapid uplift over a short period of time (2.6–2 Ma). The overlying layers of the plate cover experienced rapid denudation (hypergene and tectonic (?) erosion). As a result the granite and metamorphic complexes of the axial zone ot the Greater Caucases were exposed to the surface and the products of their erosion started to accumulate in Quaternary molasses.

— The erosion products of the Greater Caucasus constitute thin strata of orogenic (coarse) molasse of Quaternary age in Western Ciscaucasia depressions. These formations have extremely small volumes, which are incomparable to the rate of the uplift of the Greater Caucasus and the inferred thickness (many kilometers (?)) of the strata that overlie Paleozoic complexes.

— The high growth rates of the western segment of the Greater Caucasus orogen and the small volumes of its erosion products accumulated in the West Pre-Caucasus Basin, taken together, represent a contradictory phenomenon that cannot be explained solely by hypergene erosion of the Greater Caucasus.

Our research has yielded additional information partly confirming and expanding the content of the above provisions. In particular, there are grounds to consider that the plate cover complexes, which presumably overlay the Greater Caucasus before the beginning of the Quaternary, experienced not only hypergene denudation, but also tectonic-gravitational slipping of rock masses from the slopes of the growing orogen. Thus, a wide distribution of detachments has been established at the base and within the Alpine cover of the Scythian Plate on the northern slope of the Greater Caucasus and in the southern part of the West Pre-Caucasus Basin.

The results of field geological and structural studies in the northern part of the Adygean segment of the Greater Caucasus (the mountainous part of the Belaya River basin) show that the formation of structural parageneses associated with the development of detachments was caused by the slipping of units of cover layers mainly in the northern direction, down the slope of the northern monoclinal side of the Greater Caucasus (Fig. 7, group 2).

Detachments and associated structures were revealed by field observations in outcrops, as well as in seismic profiles. It is possible to see a high degree of similarity of structures with respect to their morphology and principles of spatial organization observed:

— in outcrops (Figs. 5b, 6, 8);

— on seismic profiles (Figs. 9–11).

In both cases, the structural disharmony of the layer units separated by detachments is visible. Numerous en echelon structures of different ranks accompany sublayer detachment zones. These are synthetic, Riedel shears, and, less commonly, antithetical faults with reverse- and normal-fault kinematics. Compressional and decompressional ramp scarps, their accompanying compressional structures (ramp folds, thrust duplexes) and extensions, have been established by direct observations and interpretaiton of seismic profiles. Field observations along the detachment zones have recorded asymmetric folds and small thrusts, domino structures, and boudinage zones, which have also been identified in seismic images.

Thus, a wide distribution of detachments has been established at the base and within the Alpine cover of the Scythian Plate on the northern slope of the Greater Caucasus and in the southern part of the West Pre-Caucasus Basin.

Our analysis of seismostratigraphic sections of the southern part of Western Pre-Caucasus showed the widespread development of detachments in depressions framing the modern Greater Caucasus orogen. As a rule, detachments are confined to the boundaries of strata with different rheological properties and areas of significant inclination of the layering of sedimentary complexes on the northern side of the modern Greater Caucasus orogen and, less often, buried uplifts. Various structures associated with slip processes along detachments have been established: asymmetric folds

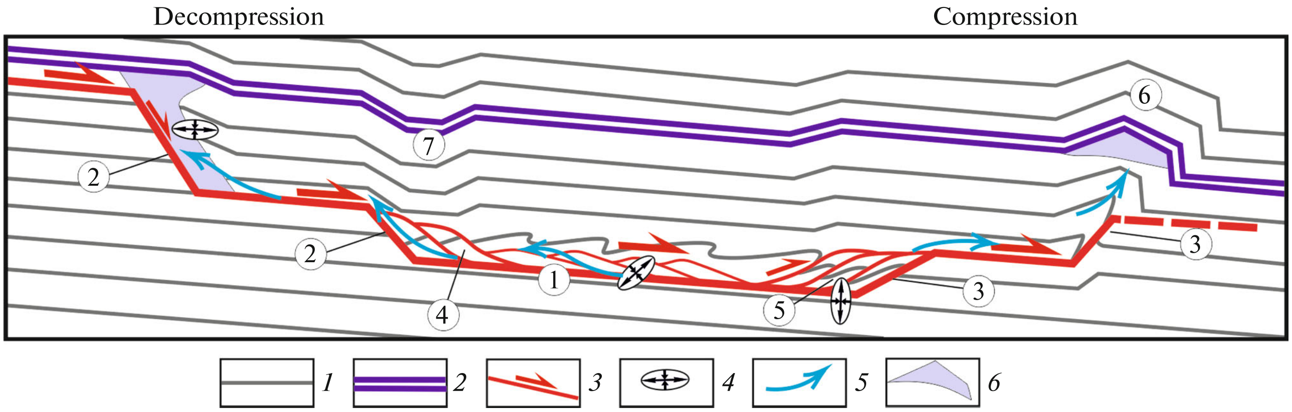


Fig. 12. Model of structure of structural triad “decompression ramp–transport zone–compression ramp” and possible methods of redistribution and localization of hydrocarbons. Indicated (numerals in circles): 1, detachment; 2, decompression ramp; 3, compression ramp; 4, extension duplexes; 5, compression duplexes; 6, ramp folds; 7, supra-ramp subsidence syncline. *1*, Layers; *2*, screening horizon; *3*, faults; *4*, orientation of compression and tension axes; *5*, expected migration routes of hydrocarbons; *6*, potential structural traps for hydrocarbons.

and small thrusts, domino structures, and boudinage zones (Figs. 9–11).

In the ramp scarps of detachments accompanying slippage, local decompression structures (faults, subsidence synclines) have developed. Along ramps preventing displacements, fold–thrust compression structures (ramp folds, thrust duplexes) are localized (Figs. 10b, 10c, 11).

The combined development of both structures in a number of cases leads to the development of unique cells of lateral rock mass transport. The structure of such cells involves a dynamically coupled triad of structural ensembles (Fig. 12):

— zone of dynamic unloading and development of fault ruptures (decompression ramp);

— zone of boudinage and slip (transport) along the inclined detachment;

— zone of squeezing and formation of ramp folds, thrusts, and thrust duplexes (compression ramp) compensating for movements along the detachment.

The aggregate structures formed within such cells represent a compensatory structural paragenesis that reflects individual deformation components: displacement/slip and the compensating components of tensile and compressive deformation.

It is possible to observe the multirank nature of how lateral geomass transport cells operate along detachments. They have lengths from several tens of meters to several (10–15) kilometers (Figs. 8–11).

In general, lateral series of such small cells constitute a single system (a first-order cell) associated with the slipping of the cover layers from the slopes of the Greater Caucasus orogen and their compensatory crowding/squeezing in the central parts of depressions framing this modern mountain structure.

Dynamic lateral transport cells are also interesting from the practical perspective. Detachments within rocks are subjected to intense deformation and disintegration may represent zones favorable for the activation of hydrocarbon migration processes.

The existence of relatively high compression and decompression segments within the considered cells is a factor governing the creation of secondary porosity in tectonic rock disintegration zones, as well as the redistribution (migration) of hydrocarbons into the decompression sector. The migration of hydrocarbons is also hydrodynamically favored by the direction of ascent of detachment and disintegration zones (Fig. 12).

When there is an overlying cap horizon (fluid seal) in the zone of the decompression ramp, formation of hydrocarbon traps can be expected. The formation of structural traps is also possible in the locks of ramp folds above compression duplexes and ramps, in the area of which, as a result of dynamic loading, hydrocarbons are squeezed out (Fig. 12).

A unique ensemble of structures was established in connection with the development of detachments in the walls of the West Kuban trough. In this case, the counterslipping of plastic, predominantly clayey rocks of the Maykop Group along detachments led to crowding and a multiple increase in their volume in the axial part of the depression (Fig. 9).

In the suture zones of oppositely directed detachments, as a result of countermovements of geomasses, squeezed folds were identified (Fig. 9b).

Flame-shaped structures are developed above such squeeze zones, which may represent either water-saturated clay diapirs or high-permeability and hydrocarbon-migration zones. In any of these cases, it can be suggest that the formation of both diapirs and hydrocarbon flows could have been caused by processes of masses squeezed out of a high dynamic load zone in which squeezed folds were formed.

In the examined sections of the West Kuban trough, no clearly expressed signs of clay diapirism were noted in the strata overlying the Maykop Group or on the surface. It can be suggested that diapirism in the eastern part of the West Kuban trough is in its initial stages of development. Quite possibly, the clay diapirs of the West Kuban trough, which developed along the western part of the Anastasiev–Krasnodar anticlinal zone, are also dynamically associated with the phenomena of squeezing out of Maykop clays as a result of their counterslip from the opposite walls of the trough.

There are ideas that the Anastasiev–Krasnodar anticlinal zone began its formation in the Sarmatian and its synsedimentary development continued until the Pleistocene [28]. In accordance with our ideas, the signs of consedimentary development of this structure (local unconformities, wedging out of individual horizons) cannot currently be considered strictly proven. Such relationships of sedimentary strata can arise as a result of the hydrodynamic breakthrough of water-saturated plastic clays to the surface, which leads to the formation of detached contacts, ruptures, and crush zones, as well as sediment slip structures on the sides of diapir uplifts.

Considering the fact that most of the diapiric anticlines of the Anastasiev–Krasnodar anticlinal zone are well expressed in the modern relief and disrupt the normal occurrence of Quaternary deposits, we believe that the active development of this structure started only in the Quaternary. Accordingly,indirectly, it can be assumed that the development of detachments, which caused the squeezing of masses at the base of this rootless anticlinal zone did not occur earlier than the Quaternary.

Based on the results of our tectonophysical studies, two groups of structural parageneses that developed in the strata of the Alpine cover were established. One of the groups has isolated dynamic parameters with respect to the structures associated with detachment formation. The tectonic faults of this group formed under conditions of normal-fault, transtensional, and strike-slip deformation regimes (Fig. 7, group 3).

The orientation of the compression axis varies from subvertical to subhorizontal position of the submeridional direction. In this case, the subhorizontal axis of extension has a sublatitudinal (W–E, WSW–ENE, WNW–ESE) orientation.

Currently, based on seismological and tectonophysical data, it has been established that the principal Late Alpine and modern stress fields of the Central and Western Caucasus are characterized predominantly by conditions of meridional compression and latitudinal extension [39, 42]. The mechanism of the 2004 Pshekha earthquake with a magnitude of 4.5, which occurred in the Adygean segment of the Greater Caucasus, corresponds to horizontal extension with a strike-slip component [6].

This allows us to suggest that the structural paragenesis we identified (group 3) characterizes manifestations of the neotectonic deformations associated with minor transverse compression (contraction) and longitudinal extension of the considered marginal part of the Greater Caucasus orogen. In this case, movements along strike-slip faults of the trans-Caucasian (NNW, NE) direction were probably of major importance.

Detachments in the northern part of the Adygean segment of the Greater Caucasus developed under conditions of a reverse-normal fault deformation regime with an inclined position of the compression and tension axes approximately in the same plane as the displacement vectors. In this case, the compression axes are oriented at large angles (60°–70°), while the tension axes form small angles (20°–30°) with a horizontal plane (Fig. 7, group 2).

Such kinematic and dynamic parameters indicate conditions of vertical-oblique flattening and predominant subhorizontal extension. The most probable cause of these dynamic conditions is the interaction of two interrelated factors: vertical uplift of the Greater Caucasus orogen, caused by endogenic (tectonic) factors, and gravitational sliding of geomasses from the slopes of this mountain structure.

The most probable mechanism for the formation of detachments in Western Pre-Caucasus is the mechanism of tectonic-gravitational sliding of sedimentary cover layers from the slopes of the growing Greater Caucasus orogen. Such phenomena are widely developed in many orogens. They are commonly considered in so-called "thin-skinned tectonics"—tectonics of a cover detached from its structural base [44, 58]. The mechanisms at work in this case are endogenic (tectonic [51, 61, 64]) and gravitational factors, which together cause postcollisional extension processes and collapse of the orogen [52, 53].

There are also grounds to consider that plate cover complexes that until the end of the Neogene covered the area corresponding to the modern Greater Caucasus in the Quaternary underwent not only denudation, but also tectonic erosion due to gravitational sliding from the slopes of the growing Greater Caucasus orogen. These processes are a form of the recent orogeny in of the Greater Caucasus and, possibly, together with hypergene factors, led to erosional–tectonic exhumation of the granite–metamorphic basement of the Caucasus orogen.

CONCLUSIONS

1. Within the West Kuban trough, clinoforms are widespread, representing paleodeltas composed of terrigenous material brought from the Scythian Plate and East European Platform into the wide shelf zone of the Eastern Paratethys. The distribution of clinoforms allows us to suggest that southward-directed sedimentary flows existed until the Late Pliocene, inclusive.

2. Formation of the modern Greater Caucasus orogen and coarse molasse deposits associated with the erosion of the orogen began no earlier than the end of the Pliocene, probably in the Eopleistocene.

3. In the structure of the Adygean sector of the northern slope of the Greater Caucasus and the southern part of the West Kuban trough, tectonic-gravitational detachments are widely developed. They are one of manifistations of recent orogeny in the Greater Caucasus.

4. Detachments in the northern part of the Adygean segment of the Greater Caucasus developed under conditions of a reverse-deformation fault regime in a setting of vertical-oblique flattening and predominant subhorizontal extension.

5. Tectonic-gravitational detachments are formed by the interaction of two factors: vertical uplift of the Greater Caucasus orogen, caused by endogenic (tectonic) factors, and gravitational sliding of geomasses from the slopes of this mountain structure.

ACKNOWLEDGMENTS

The authors express their deep gratitude to reviewer Yu.L. Rebetsky (Schmidt Institute of Physics of the Earth, Russian Academy of Sciences, Moscow, Russia) and anonymous reviewers and members of the editorial board of the journal *Geotectonics* for valuable critical comments and advice that improved the article. The authors are thankful to editor M.N. Shupletsova (Geological Institute, Russian Academy of Sciences, Moscow, Russia) for careful editing.

FUNDING

The study was supported by the Russian Science Foundation (grant no. 24-27-00252, https://rscf.ru/project/24-27-00252/).

CONFLICT OF INTEREST

The authors of the work declare that they have no conflicts of interest.

REFERENCES

1. A. D. Arkhangel'skii, *Conditions of Oil Formation in the Northern Caucasus* (Nauch.-Izdat. Byuro SNP, Moscow, 1927) [in Russian].
2. E. V. Beluzhenko, I. G. Volkodav, M. G. Derkacheva, S. G. Korsakov, V. V. Sokolov, and V. I. Chernykh, *Oligocene and Neogene deposits of the Belaya River Valley (Adygea)* (AdygGU, Maykop, 2007).
3. E. V. Beluzhenko, "Middle–Upper Miocene and Pliocene stratigraphy between Psekups and Belaya rivers (Northwestern Caucasus). 1. Middle Miocene," Byull. Mosk. O–va Ispyt. Prir., Otd. Geol. **77** (1), 47–59 (2002a).
4. E. V. Beluzhenko, "Middle–Upper Miocene and Pliocene stratigraphy between Psekups and Belaya rivers (Northwestern Caucasus). 2. Upper Miocene and Pliocene," Byull. Mosk. O–va Ispyt. Prir., Otd. Geol. **78** (2), 51–61 (2002b).
5. *Greater Caucasus during the Alpine Orogenesis*, Ed. by Yu. G. Leonov (GEOS, Mosow, 2007) [in Russian].
6. I. P. Gabsatarova, "The Pshekh earthquake on November 15, 2004 goda s M_w = 4.5, 0I P = 5–6 (Krasnodar krai)," in *Earthquakes in Northern Eurasia, 2004* (GS RAN, Obninsk, 2010), pp. 386–394.
7. B. P. Zhizhchenko, *Methods of Stratigraphic Research in Oil and Gas Regions* (Nedra, Moscow, 1969) [in Russian].
8. V. M. Isaev, T. I. Gadzhiev, S. A. Ali-Zade, and T. N. Kengerli, "Tectonic nappes and olistostrom complexes of the South-Eastern Caucasus," Geotektonika, No. 1, 70–84 (1981).
9. N. V. Klavdieva, Candidate's Dissertation in Geology and Mineralogy (Mosk. Gos. Univ., Moscow, 2007).
10. A. L. Knipper, "Movements of the Austrian phase in the Tethys oceanic crust: Character of manifestation, consequences, and possible causes," Geotektonika, No. 2, 3–15 (1985).
11. S. Yu. Kolodyazhnyi, *Paleoproterozoic Structural-Kinematic Evolution of the Southeastern Part of the Baltic Shield* (GEOS, Moscow, 2006).
12. S. Yu. Kolodyazhnyi, E. I. Makhinya, E. A. Shalaeva, and K. I. Dantsova, "Features of the Late Alpine tectonics in the Adyghe sector of the Greater Caucasus," in *Proceedings of LV Tectonic Conference "Tectonics and Geodynamics of the Earth's Crust and Mantle: Fundamental problems–2024," Jan. 29–Febr. 3, 2024* (GEOS, Moscow, 2024), vol. 1, pp. 202–206.
13. M. L. Kopp, "Structures of lateral extrusion in the Alpine–Himalayan Collision Belt," in *Transactions of Geological Institute of Russian Academy of Sciences*, Vol. 506, Ed. by Yu. G. Leonov (Nauchn. Mir, Moscow, 1997) [in Russian].
14. M. L. Kopp, *Mobilistic Neotectonics of Cratons in the Southeastern Europe*, Ed. by Yu. G. Leonova (Nauka, Moscow, 2005) [in Russian].
15. M. L. Kopp and I. G. Shcherba, "The Late Alpine development of the Eastern Caucasus," Geotektonika, No. 6, 94–108 (1985).
16. M. L. Kopp and I. G. Shcherba, "The Caucasian Basin in Paleogene," Geotektonika, No. 2, 29–50 (1998).
17. S. G. Korsakov, I. N. Semenukha, E. V. Beluzhenko, V. I. Chernykh, G. R. Tuzikov, I. I. Grekov, V. N. Tokarev, M. G. Derkacheva, and V. V. Sokolov, *The 1 : 200000 State Geological Map of the Russian Federation, 2nd ed., Ser. Caucasian. Sheet L-37-XXXV. Maykop. Explanatory Note* (Vseross. Nauchno-Islsed. Geol. Inst., St. Petersburg, 2004) [in Russian].
18. S. G. Korsakov, I. N. Semenukha, S. M. Gorbova, et al., *The 1 : 200000 State Geological Map of the Russian Federation, 2nd ed., Ser. Caucasian. Sheet L-37-XXXIV.* Tuapse. *Explanatory Note* (Vseross. Nauchno-Islsed. Geol. Inst., St. Petersburg, 2002) [in Russian].

19. N. B. Kuznetsov, T. V. Romanyuk, K. I. Dantsova, I. V. Fedyukin, I. V. Latysheva, A. V. Shatsillo, O. A. Maslova, and S. D. Polina, "On the tectonic nature of the Western Kuban trough," Oil Industry, No. 9, 78–84 (2023).
https://doi.org/10.24887/0028-2448-2023-9-78-84

20. N. B. Kuznetsov, T. V. Romanyuk, K. I. Dantsova, I. V. Fedyukin, I. V. Latysheva, A. V. Shatsillo, O. A. Maslova, and S. D. Polina, "Characteristics of sedimentary strata of the Indolo-Kuban trough as indicated by the results of U–Pb isotopic dating of detrital zircons," Nedra Povolzh'ya Prikaspiya, No. 1, 4–15 (2024a).
https://doi.org/10.24412/1997-8316-2024-113-4-15

21. N. B. Kuznetsov, T. V. Romanyuk, A. V. Shatsillo, I. V. Latysheva, I. V. Fedyukin, A. S. Novikova, O. A. Maslova, K. I. Dantsova, A. V. Drazdova, A. S. Dubenskii, K. G. Erofeeva, and V. S. Sheshukov, "The ages of detrital zircon from the sandstones of the Belorechensk Formation (Western Cis-Caucasus): Preliminary conclusions about its age and the time of formation of a young orogen of the Greater Caucasus," in *Proceedings of LV Tectonic Conference "Tectonics and Geodynamics of the Earth's Crust and Mantle: Fundamental problems–2024," Jan. 29–Febr. 3, 2024* (GEOS, Moscow, 2024b), vol. 1, pp. 244–249.

22. N. B. Kuznetsov and T. V. Romanyuk, "Peri-Gondwanan blocks in the structure of the southern and southeastern framing of the East European Platform," Geotectonics **55** (4), 439–472 (2021).
https://doi.org/10.1134/S0016852121040105

23. M. G. Leonov, "Wild flysch of the Alpine Region," in *Transactions of Geological Institute of USSR Academy of Sciences*. Vol. 199, Ed. by A. V. Peive (Nauka, Moscow, 1975) [in Russian].

24. A. I. Letavin, V. E. Orel, S. M. Chernyshev, et al., *Tectonics and Petroleum Potential of Northern Caucasus*, Ed. by N. A. Krylov (Nedra, Moscow, 1987) [in Russian].

25. E. E. Milanovskii, *The Neotectonics of the Caucasus* (Nedra, Moscow, 1968), [in Russian].

26. E. E. Milanovskii and V. E. Khain, *Geological Structure of the Caucasus*, Ed. by A. A. Bogdanov, M. V. Muratov, and N. S. Shatsky (Mosk. Gos. Univ., Moscow, 1963) [in Russian].

27. Z. Kh. Mollaev, V. V. Dotsenko, and T. Kh. Bachaeva, "The concepts of forming the West Kuban Foredeep," in *Modern Problems of Geology, Geophysics, and Geoecology in the North Caucasus*, Ed. by A. O. Gliko and I. A. Kerimov (IIET RAN, Moscow, 2020), vol. X, Pt. 1, pp. 179–186 [in Russian].

28. Z. Kh. Mollaev, V. V. Dotsenko, and T. Kh. Bachaeva, "Structure and the formation of the West Kuban Foredeep," in *Modern Problems of Geology, Geophysics, and Geoecology in the North Caucasus*, Ed. by A. O. Gliko and I. A. Kerimov (IIET RAN, Moscow, 2020), vol. X, Pt. 1, pp. 187–199 [in Russian].

29. M. V. Muratov, "Types of depressions in the sedimentary cover of ancient platforms," Byull. Mosk. O–va Ispyt. Prir., Otd. Geol. **47** (5), 61–71 (1972).

30. A. M. Nikishin, A. V. Ershov, and V. A. Nikishin, "Geological history of Western Caucasus and adjacent foredeeps based on analysis of the regional balanced section," Dokl. Earth Sci. **430** (4), 155–157 (2010).

31. V. E. Orel, Yu. V. Raspopov, A. P. Skripkin, et al., *Geology and Petroleum Potential of the Cis-Caucasus Predkavkaz'ya*, Ed. by V. E. Orel (GEOS, Moscow, 2010) [in Russian].

32. I. S. Patina, Yu. G. Lenov, Yu. A. Volozh, M. L. Kopp, and M. P. Antipov, "Crimea–Kopet Dagh zone of concentrated orogenic deformations as a transregional late collisional right-lateral strike-slip fault," Geotectonics **51**, 353–365 (2017).

33. S. D. Polina, K. I. Dantsova, and G. I. Faizullin, "Current problems and issues of tectonics, geology, and petroleum potential of the northwestern Caucasus," Izv. OshTGU, No. 2-1, 269–276 (2023).

34. V. I. Popkov and N. A. Bondarenko, "Tectonics of orogenic structures of the Northwestern Caucasus," in *Proceedings of XLI Tectonic Conference "General and Regional Problems of Tectonics and Geodynamics," Jan. 29–Febr 1, 2008* (GEOS, Moscow, 2008), vol. 2, pp. 125–130.

35. S. V. Popov, M. P. Antipov, A. S. Zastrozhnov, E. E. Kurina, and T. N. Pinchuk, "Sea-level fluctuations on the northern shelf of the eastern Paratethys in the Oligocene–Neogene," Stratigr. Geol. Correl. **18** (2), 200–224 (2010).

36. I. S. Postnikova, I. S. Patina, and G. M. Gorkin, "Geological setting and formation of the erosional structure of Upper Miocene deposits in Western Ciscaucasia," Lithol. Miner. Resour. **59**, 517–525 (2024).
https://doi.org/10.1134/S0024490224700676

37. Yu. M. Pushcharovskii, "Foredeeps, their tectonic structure and development," in *Transactions of Geological Institute of Russian Academy of Sciences*. Vol. 28, Ed. by N. P. Kheraskov (Nauka, Moscow, 1959) [in Russian].

38. Yu. L. Rebetsky, "The current state of crustal stresses in the Caucasus according to the United Catalogue of Earthquake Focal Mechanisms," Geodynam. Tectonophys. **11**, 17–29 (2020).

39. Yu. L. Rebetsky, L. A. Sim, and A. V. Marinin, *From Slickensides to Tectonic Stresses. Methods and Algorithms*, Ed. by Yu. G. Leonov (GEOS, Moscow, 2017) [in Russian].

40. L. A. Sim and A. V. Marinin, "Methods of field tectonophysics for reconstruction of paleostresses," in *Modern Tectonophysics. Methods and Results*, Ed. by Yu. L. Rebetskii (IFZ, Moscow, 2015), vol. 2, pp. 47–76 [in Russian].

41. A. S. Stolyarov, "Paleogeography of the Cis-Caucasus, Volga–Don, and Southern Mangyshlak Region in the Late Eocene–Early Oligocene," Byull. Mosk. O–va Ispyt. Prir., Otd. Geol. **6** (4), 64–80 (1991).

42. T. Yu. Tveritinova, A. V. Marinin, and I. V. Bondar', "Structural features of the Pshekha–Adler fault zone of the Greater Caucasus according to structural studies," Moscow Univ. Geol. Bull. **79**, 375–387 (2024).

43. *Tectonics of Europe and Adjacent Regions: Variscides, Epi-Paleozoic Platforms, Alpides. Explanatory Note to the 1 : 2500000 International Tectonic Map of Europe and Adjacent Regions*, 2nd ed. (Nedra, Moscow, 1978).
44. V. E. Khain, *Tectonics of Continents and Oceans* (Nauka, Moscow, 2001).
45. V. N. Kholodov and R. I. Nedumov, "On the problem of the existence of the Caucasian paleoland during the Oligocene–Miocene," Stratigr. Geol. Korrel. **4** (2), 80–91 (1996).
46. V. F. Sharafutdinov, Doctoral Dissertation in Geology and Mineralogy (Mosk. Gos. Univ., Moscow, 2003) [in Russian].
47. A. G. Shempelev, N. I. Prutskii, I. S. Fel'dman, and S. U. Kukhmazov, "Geological–geophysical model along the Tuapse–Armavir profile," in *Tectonics of the Neogean: General and Regional Aspects. Proceedings of the XXXIV Tectonic Meeting, Jan. 30–Febr. 3, 2001*, (GEOS, Moscow, 2001), vol. 2, pp. 316–320.
48. I. G. Shcherba, *Stages and Phases of Development of Alpine Area in Cenozoic*, Ed. by Yu. G. Leonov (Nauka, Moscow, 1993) [in Russian].
49. S. Adamia, G. Zakariadze, T. Chkhotua, N. Sadradze, N. Tsereteli, A. Chabukiani, and A. Gventsadze, "Geology of the Caucasus: A review," Turkish J. Earth Sci. **20**, 489–544 (2011).
50. B. Avdeev and N. A. Niemi, "Rapid Pliocene exhumation of the central Greater Caucasus constrained by low-temperature thermochronometry," Tectonics **30**, 1–16 (2011).
51. J. H. Davies and F. Blanckenburg, "Slab breakoff: A model for syncollisional magmatism and tectonics in the Alps," Tectonics **14** (1), 120–131 (1995).
52. J. F. Dewey, "Extensional collaps of orogenes," Tectonics **7**, 1123–1139 (1988).
53. W. Frisch, I. Dunkl, and J. Kuhlemann, "Post-collisional orogen-parallel large-scale extension in the Eastern Alps," Tectonophysics **327**, 239–265 (2000). https://doi.org/10.1016/S0040-1951(00)00204-3
54. I. Gamkrelidze, D. Shengelia, G. Chichinadze, Y.-H. Lee, A. Okrostsvaridze, G. Beridze, and K. Vardanashvili, "U–Pb LA-ICP-MS dating of zoned zircons from the Greater Caucasus pre-Alpine crystalline basement: Evidence for Cadomian to Late Variscan evolution," Geol. Carpathica **71** (3), 249–263 (2020). https://doi.org/10.31577/GeolCarp.71.3.4
55. S. Hanmer and C. Passchier, "Shear-sense indicators: A review," Geol. Surv. Canada, Paper 90–17 (1991). https://doi.org/10.4095/132454
56. R. A. Marrett and R. W. Allmendinger, "Kinematic analysis of fault-slip data," J. Struct. Geol. **12** (8), 973–986 (1990).
57. D. V. Palcu, S. Lazarev, W. Krijgsman, I. S. Patina, I. Ţ Şandric, I. Vasiliev, and M. Stoica, "Late Mocene Megalake regressions in Eurasia," Sci. Rep. **11** (1), Art. 11471 (2021). https://doi.org/10.1038/s41598-021-91001-z
58. O. A. Pfiffner, "Thick-skinned and thin-skinned tectonics: A global perspective," Geosciences **7** (3), 71 (2017). https://doi.org/10.3390/geosciences7030071
59. J. G. Ramsay and M. I. Huber, "The techniques of modern structural geology," in *Folds and Fractures.* Vol. 2 (Acad. Press, London, 1987).
60. M. Somin, "Pre-Jurassic basement of the Greater Caucasus: Brief overview," Turkish J. Earth Sci. **20**, 545–610 (2011). https://doi.org/10.3906/yer-1008-6
61. V. Trifonov and S. Sokolov, "Late Cenozoic tectonic uplift producing mountain building in comparison with mantle structure in the Alpine-Himalayan belt," Int. J. Geosci. **5** (5), 497–518 (2014).
62. C. G. C. van Baak, W. Krijgsman, A. Grothe, T. M. Hoyle, I. Magyar, O. Sztanó, L. A. Golovina, I. S. Patina, E. P. Radionova, O. Mandic, S. V. Popov, M. Stoica, and I. Vasiliev, "Paratethys response to the Messinian salinity crisis," Earth-Sci. Rev. **172**, 193–223 (2017).
63. S. J. Vincent, A. C. Morton, A. Carter, S. Gibbs, and T. G. Barabadze, "Oligocene uplift of the Western Greater Caucasus; an effect of initial Arabia–Eurasia collision," Terra Nova **19**, 160–166 (2007).
64. E. Zor, "Tomographic evidence of slab detachment beneath eastern Turkey and the Caucasus," Geophys. J. Int. **175**, 1273–1282 (2008). https://doi.org/10.1111/j.1365-246X.2008.03946.x

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