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Lithomineralogical and Geostructural Characteristics of the Lower–Middle Jurassic Terrigenous Complex of the Greater Caucasus (the Terek River Area)

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Abstract—Characteristics of Lower–Middle Jurassic sedimentary sequences of a folded structure, such as the dislocation degree, cleavage development, and mineral composition of rocks, as well as stress parameters, are considered using a geological profile across the central part of the Greater Caucasus as an example. It is shown that these characteristics regularly change along the profile. The results suggest a significant influence of stress on the intensity of secondary alterations in terrigenous sequences of the Greater Caucasus.

The Lower–Middle Jurassic terrigenous rocks in the Greater Caucasus reveal regular changes in essential characteristics of sedimentary deposits, such as the dislocation degree, intensity of development in secondary structures (cleavage), and rock composition. Each of these characteristics were discussed in sufficient detail between scientists from different points of view. Nonetheless, until now there has not been a prime example exhibiting these changes and thus, answering the question of which factor or factors caused postsedimentation structural-textural and lithological transformations.

Our studies were aimed at considering the above-mentioned characteristics together in a single structure while applying modern methods of investigation. The first stage of these works included comprehensive analyses along the profile across the entire area of development for Lower–Middle Jurassic rocks in the central part of the Greater Caucasus (the Terek River valley).

METHODS

Investigations of Lower–Middle Jurassic sequences along the Military Georgian Road included the study of lithology and stratigraphy, as well as the revision of available materials. These studies were accompanied by geological mapping. When geological observations in the Terek River valley were impossible for various reasons, the relevant intervals of the Jurassic sequence were studied in adjacent areas (tributary valleys or watersheds between them). As a result, the geological cross section passing along the Terek River valley and crossing the entire area of development of the Lower–Middle Jurassic rocks was compiled (Fig. 1). During these investigations, particular attention was paid to the study of cleavage developed in Jurassic sequences, and to its relationships with the folded structure. During laboratory processing of the material, the cleavage was studied in thin sections.

The most significant irregularity in cleavage development along the profile required its intensity to be estimated through quantitative characteristics. The first signs of cleavage, registered in thin sections by the appearance of isolated cleavage zones, became evident at grain deformation (shortening ε) equal to 5–7%. At a shortening of 10–12%, the rocks reveal a distinct structure. At a deformation of 20–25%, cleavage penetrates the rock. Such correlation between the cleavage development and the rock deformation degree at the level of mineral grains can be used as an approximate assessment of rock deformation. We used such a scale in this work.

An analysis of the folded structure and cleavage development was accompanied by the study of mineral composition of clayey rocks to establish a phase composition of clay minerals and to trace changes in their associations, as well as changes in polytype modifications of micaceous minerals. The crystallinity index of micas in samples from different parts of the profile was also estimated. The representative collection of samples was studied using different diffraction methods. The X-ray study of more than 80 oriented specimens and powders of the clay fraction <0.001 mm was carried out using the diffractometer DRON-2 (CuKα-emission). The electron diffraction study of 35 samples was performed using the method of oblique structures to establish polytype modifications of micas and to determine parameters of their elementary cells. These investigations were carried out using an electron diffraction device ER-100 with an accelerating voltage of 100 kV.

In addition, stress analysis was carried out to determine parameters of the rock stress under deformation.

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This method is used to define stress facies (σ-facies), in which a certain deformation structure of the mineral grain (in our case, an indicator mineral is represented by quartz) corresponds to a certain level of nominal and local stresses. These σ-facies, which are then divided into subfacies, are defined in rock complexes ranging from undeformed rocks (with quartz grains lacking stress signs) to strongly deformed rocks (with quartz granulation and abundant destruction zones in their peripheral areas). Data obtained in the course of investigations are shown in Fig. 2.

GEOLOGICAL STRUCTURE OF THE LOWER–MIDDLE JURASSIC TERRIGENOUS COMPLEX

Along the Military Georgian Road, Lower and Middle Jurassic rocks compose a band about 45 km wide and mark the axial part of the Greater Caucasus meganticlinorium. This band is divided by a major longitudinal thrust (Shaukhokh fault, according to N.A. Enna) into two zones (Fig. 2).

The northern part (Dzherakh–Dar'yal subzone, according to Rengarten, 1932) represents the western termination of the Dar'yal–Bogos anticlinorium (Gushchin and Panov, 1992), which, like in the Eastern Caucasus, is sharply asymmetric but characterized by its fan-shaped structure. Its axial part is marked by Paleozoic Dar'yal and Gvileti granite-gneiss massifs (Paleozoic granitoid massifs), which outcrop from beneath the Lower Jurassic sequences. In this area, the southern limb of anticlinorium, which is formed by a system of southward-overturned folds and is almost completely destroyed in the Eastern Caucasus, is cut by the Shaukhokh thrust and only preserved as fragments. The wider northern limb represents a very steep, sometimes overturned monocline with separate folds overturned northward and dissected by numerous fractures that are also steeply inclined northward. North of the Chmi Settlement, Lower–Middle Jurassic rocks are unconformably overlain by the Upper Jurassic strata of the northern limb of meganticlinorium. The southern part of the Lower–Middle Jurassic rock area is occupied by the accretionary thrust system of the southern slope. The latter is composed of a series of tectonic nappes divided by northward-dipping thrust planes. The age of nappe rocks becomes successively younger in the north–south direction, and deposits in every nappe are characterized by the normal stratigraphic succession with northward-dipping beds. The Tib thrust (Vardanyants, 1935) separates this system from the southern Chiauri–Dibrar synclinorium filled with Upper Jurassic and Cretaceous flysch deposits.

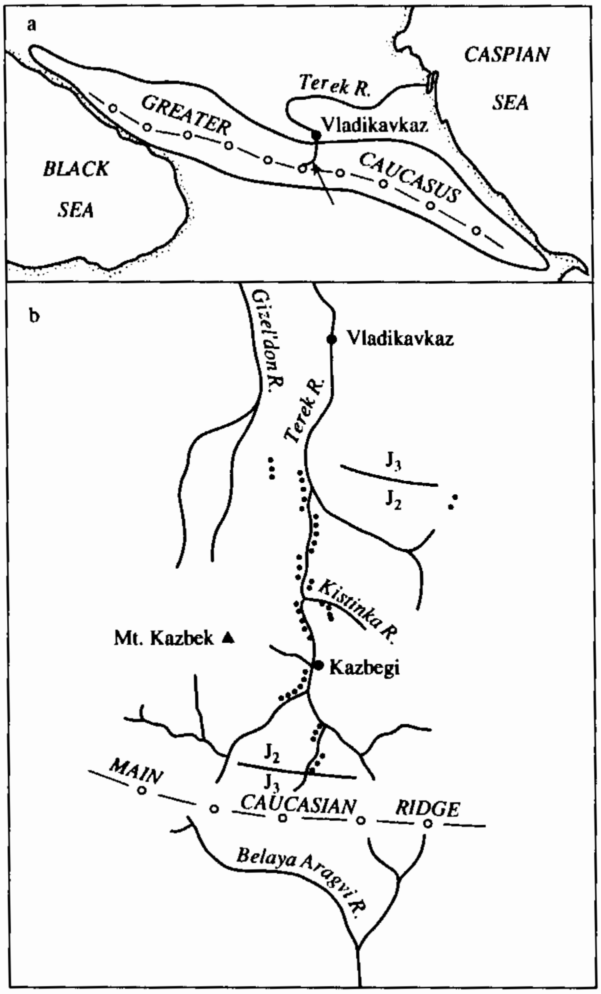

Fig. 1. Location of (a) study area and (b) sequences (designated by points) in the Terek River basin.

The structure of the northern area of the Lower–Middle Jurassic rocks. The northern limb of the Dar'yal–Bogos anticlinorium is subdivided by a system of steep southward-dipping fractures into a series of nappes (blocks), those rocks differ in age and composition, intensity of tectonic deformations, cleavage development, and so on. The northernmost nappe (Fig. 2, I) represents an almost vertical monocline composed of upper Aalenian rocks of the Koirakh and Fortaukh formations (Panov and Gushchin, 1993; *Yura Kavkaza*, 1992). It consists of distinctly bedded sequences of alternating argillites, siltstones, and sandstones outcropping in scarps between the Suargom River mouth and Chmi Settlement. Northward, in the Skalistyi Ridge, flat Bajocian–Bathonian layers occur locally under unconformably overlying Upper Jurassic limestones.

Next, the southern nappe II (Fig. 2) located between the Suargom (Suargom River mouth) and Pui (northern margin of the Nizhnii Lars Settlement) faults is composed of upper Toarcian, and lower–upper Aalenian rocks: alternating argillites, siltstones, and sandstones of the Dzhimarin and Koirakh formations, and argillites and siltstones of the Salgin and Fortuakh formations (Panov and Gushchin, 1993; *Yura Kavkaza*, 1992). These strata form a large northward-overturned and strongly compressed synclinal fold whose northern limb is almost completely cut by the Suargom fault,

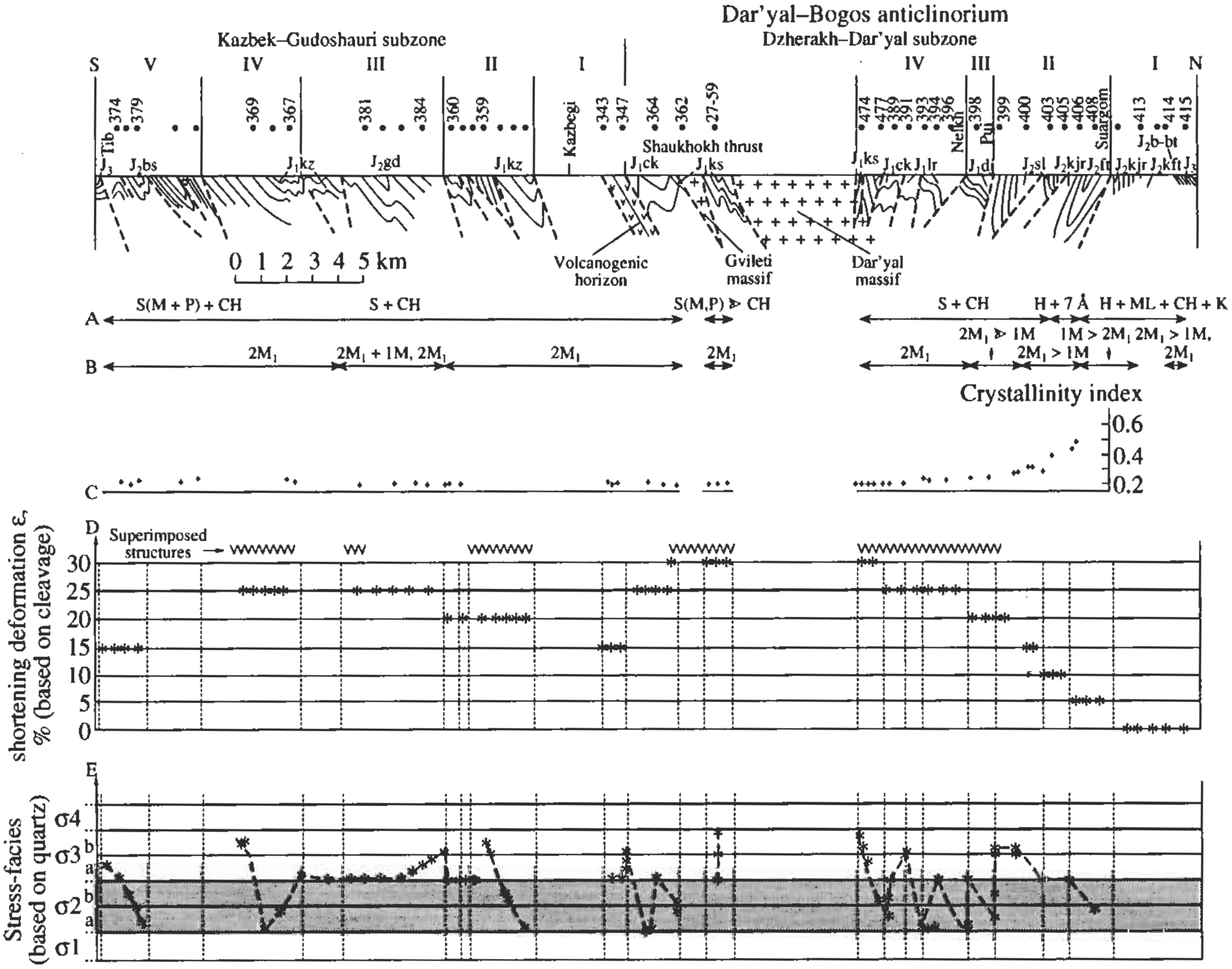

Fig. 2. Geological section across the distribution area of Lower–Middle Jurassic rocks in the Central Caucasus (Terek River) with typical mineralogical and geostructural characteristics. (A) Distribution of associations of clay minerals: (H + ML + CH + K) hydromica–mixed-layer illite/smectite mineral–chlorite–kaolinite association; (S + CH) sericite + chlorite association; (S(M,P)) sericite of muscovite affinity with an admixture of paragonite; (H + 7 Å) hydromica–berthierine association. (B) Distribution and ratios of different clay mineral polytypes (1*M* and $2M_1$). (C) Changes in the crystallinity index of micas. (D) Changes in the intensity of the cleavage development. (E) Parameters of stress-facies based on quartz.

whereas, the wide southern limb representing a steep (locally overturned) monocline is complicated by several southward-dipping faults. The latter faults are reflected, similarly to the Pui fault bounding the nappe in the south, in a wide (up to several dozens of meters) fracture and boudinage zone with abundant quartz veins. Rocks in the southern part of the nappe also reveal a vague southward-dipping cleavage.

The nappe III (Fig. 2) is located southerly between the Pui and Nelkh (or Buron–Lars) faults (Gushchin and Panov, 1992) and is entirely composed of upper Toarcian argillites and clayey–silty rocks of the Dzhimarin Formation creating a northward-flattening monocline with a normal dip of 70° to 25°. These rocks reveal no macroscopic cleavage. The Nelkh fault, which dips southward (maximum 50°) and bounds the nappe in the south, is also marked by a wide fracture and boudinage zone with slickensides and abundant quartz veins.

The southernmost nappe (Fig. 2, IV) of the northern limb (between core and the Nelkh fault) is mainly represented by clayey and banded clayey–silty shales of the Tsiklauri (upper Pliensbachian–lower Toarcian) and Lars (Toarcian) formations (Panov, 1976; Panov and Gushchin, 1987), as well as by a thin fragment of the Kistinka Formation (Sinemurian–lower Pliensbachian) composed of quartzite and siltstone. This nappe hosts a system of folds dipping northward, which is complicated by a series of fractures. Folds are size-variable (from several dozens of meters to 300–500 m), open, symmetrical, steep (dip angle up to 60°–70°) or slightly overturned northward, sometimes acquiring a "chair-shaped" form. Rocks are marked by perfect cleavage with almost vertical planes, dipping steeply southward. Faults, mainly reversed, steep, and dipping southward, are marked by fracture zones, compression, and quartz development. In the southern part of the nappe, the Tsiklauri Formation rocks are saturated with abundant and distinctly transverse diabase dikes, which, in turn, are characterized by the boudinage texture.

The axial part of the Dar'yal–Bogos anticlinorium is bound by a system of subvertical faults in the north (near the northern entry into the Dar'yal canyon), and by the Gvileti reversed fault in the south (near the village of Gvileti). The pre-Jurassic basement represented by Upper Paleozoic granite-gneisses, which is locally fringed by thin variegated quartzites and phyllites of the Upper Carboniferous Baddon Formation, outcrops in two massifs (Dar'yal in the north and Gvileti in the south, Fig. 2). Both massifs are enveloped by quartzitic sandstones, gritstones, and shales of the Sinemuran–lower Pliensbachian Kistinka Formation (Panov, 1976) and, probably, represent rootless diapir-shaped anticlinal folds that are cut in the south by faults. These structures were formed during the Alpine deformation of the pre-Jurassic basement and sedimentary cover. The contact of the Kistinka Formation with granite-gneisses is stratigraphic. This is evident from their composition and enclosed pebbles of Dar'yal granite-gneiss (Kipiani, 1985), but it is often detached in the course of tectonic deformations. Granite-gneiss rocks show the development of hornfels and hydrothermal alteration in the contact zone. The area near the Gurgala Mountain at the eastern periphery of the Dar'yal massif is covered by a tectonic nappe: granite-gneisses and the lower Kistinka Formation rest upon the upper layers of the Kistinka Formation with a horizontal displacement magnitude of at least 1 km.

In the north, directly in the Terek River canyon, the Dar'yal massif is bound by subvertical faults, between which wedges (up to several dozens of meters thick) of the Kistinka Formation rocks are squeezed. In the south, the massif is bound by the Amali reversed fault whose plane dips at an angle of 65°–75° N (Kipiani, 1985). Along the latter fault, the massif is thrust onto the northern flank of the Gvileti massif, which is composed of the Kistinka Formation rocks intensely deformed into strongly compressed folds. The southern part of the Gvileti massif is also composed of the Kistinka Formation rocks that form a vertical or steep, southward, overturned, monocline that is complicated by abundant fractures; they are preserved only fragmentarily, and are almost completely cut by the Gvileti reversed fault whose plane dips at an angle of 60°–70° N. Rocks of the Kistinka Formation show intense cleavage and are densely saturated with transverse diabase dikes, which, in turn, enclose abundant boudines.

The southern limb of the Dar'yal anticlinorium in the Terek River canyon represents a narrow (1.5 km) strip of very steeply dipping, banded, clayey–silty shales of the Tsiklauri Formation, which are overlain by banded clayey–silty and sandy–silty shales of the Lars Formation on the right-hand slope of the valley in the Terek–Kistinka watershed area. Rocks are marked by intense cleavage and form conjugated, strongly compressed synclinal and southward overturned anticlinal folds with dip angles of 75°–90° in limbs. The clayey shale sequence hosts abundant diabase dikes with boudines and obliquely crossing bedding planes; locally, dikes prevail over enclosing rocks. The Shaukhokh fault, bounding the Dar'yal–Bogos anticlinorium in the south, represents a thrust with the plane dipping at an angle of about 50° and is marked by a wide boudinage and compression zone with abundant quartz veins. The compression zone, which is traced along the right-hand side of the Terek River, was first mapped by Rengarten (1932).

As was noted, the nappe–thrust system of the Southern slope (the Kazbek–Gudoshauri subzone, according to Rengarten, 1932) is subdivided by longitudinal faults into several nappes, five of which are outstanding in dimensions (Fig. 2). The first (northernmost) nappe is composed of shales of the Tsiklauri Formation, which encloses abundant stratiform diabase bodies probably representing pre-fold sills concordant with host rocks and likewise characterized by cleavage development, as well as several amygdaloid pillow-lava flows. At all accessible sites, they have normal bedding and dip (60°–80° N), i.e., they form a steep monocline. Magmatic rocks prevail in the northern (upper) part of the section, which is separated from the southern one by a major reversed fault; the southern (lower) part is strongly dominated by shales. In the south, the first nappe is bound by the major Adaikom–Kazbek reversed fault–thrust (Rengarten, 1932; Vardanyants, 1935), whose plane dips at angles of 40°–60° N.

The second nappe exposed in the Terek River valley between Gergeti and Pansheti settlements is composed of alternating shales, banded sandy–silty shales and sandstone beds of the Kazbek (Toarcian) Formation (Panov, 1976; Panov and Gushchin, 1987). The Kazbek Formation beds form a steep monocline characterized by the normal dip (60°–85° N) and complicated by a series of reversed faults and thrusts, whose planes also dip in the same direction at angles of 60°–70°. Owing to this, there is repetition in a partial section. Nonetheless, normal layer succession is not disturbed: the lower part of the Kazbek Formation, approximately half of which is represented by sandstones, is exposed in the southern part of the monocline; the upper one, where banded shales prevail, is exposed in its northern areas. All rocks of the Kazbek Formation are marked by well-developed cleavage whose planes dip southward or very steeply northward (85°–90°). Diabase dikes are absent here and in the successive nappes. In the south, the nappe is bound by a reversed fault, whose plane, according to Kipiani, dips northward at angles 55°–60°. Only in the area directly adjacent to the latter, the Kazbek Formation beds are deformed into a series of folds 100–300 m across and locally overturned southward.

The third nappe, which is composed of sandstones and clayey–silty shales of the Kazbek Formation in the southern part, and of the Gudoshauri Formation (Aalenian) shales in its northern areas, forms a monocline with a normal dip at angles of 60°–80° N. In the south, it is limited by the Arsha reversed fault–thrust (dip angle about 50° N), which is marked by wide foliation and a fracture zone with quartz veins. Near the latter, the Kazbek Formation beds, separated by another reversed fault from the Gudoshauri Formation in the north, form a sufficiently large, box-shaped, anticlinal structure complicated by small overturned folds near the Arsha fault to the south. Rather large (500–600 m) symmetrical folds with dip angles of 60°–70° at their limbs also occur within Gudoshauri shales, but they only complicate the general monoclinal bedding and are mainly confined to the northern margin of the nappe, under the Arkhoti reversed fault.

The fourth nappe in the Terek River valley is composed of Kazbek Formation rocks forming a monocline with a normal dip at angles of 45°–65°. Only its northernmost part, located under the Arsha reversed fault, shows the presence of folds slightly overturned southward. In the south, the nappe is limited by a major fault of the reversed fault–thrust type with the northward plane dip at an angle of 50°.

The fifth nappe composed of shales and sandstones of the Bajocian or Bathonian Busarchil Formation consists, at least, of four sheets subdivided by several reversed fault–thrusts dipping at angles of 50°–70°. Each of these sheets represents a steep monocline with the northward normal dip at angles 40°–75° and is usually complicated near faults by small folds, commonly overturned northward. Similar to the previous nappes, rocks here are characterized by well-developed cleavage with a northward dip at angles of 60°–90°. In the south, the entire Kazbek–Gudoshauri subzone is bound by the major Tib thrust (Rengarten 1932; Vardanyants, 1935), along which, the fifth nappe is thrust over the Upper Jurassic and Cretaceous flysch sequences.

CHARACTERISTICS OF CLEAVAGE

All Jurassic rocks in the study area show the presence of cleavage developed to different extents. It should be noted that, contrary to the definition of cleavage as *the capability of rocks to be split into thin plates and small lenses*, accepted in Russian literature, we follow the participants of the Penrose conference, 1976 (*Atlas…*, 1982; Powell, 1979) and designate the term "cleavage" as *certain microstructures* favoring the appearance of anisotropy of mechanical properties in rocks (Talitskii and Galkin, 1988). Two types of such structures can be defined: *disjunctive cleavage* and *crenulation cleavage.*

Disjunctive cleavage occurs solely in rocks with irregular granular structure, e.g., siltstones, sandstones, tuffstones, some types of limestone, and others. Thin sections oriented perpendicular to the structure strike reveal the system of thin subparallel threadlike zones dividing the rock into thin lamellas, or small lenses (Fig. 3). These threadlike zones (cleavage zones) and small lenses (microlithons) distinctly differ in mineral composition and structure (Fig. 4). They represent aggregates of very small flakes of clay (usually micaceous) minerals, fine grains of ore minerals, and finely dispersed organic matter. The mineral composition of these zones is always constant regardless of the rock composition itself. The main peculiarity of the mineral composition in cleavage zones is the virtually complete absence of some minerals, such as quartz, calcite, and feldspar. As a rule, grains in microlithons located between cleavage zones and irregularly distributed, but are sometimes oriented along the strike of these zones. In other cases, ends of more or less isometric grains are marked by a fringe that is composed of small, newly formed grains of quartz, calcite, and some other minerals. In this case, the entire aggregate consisting of grain and fringe is elongated parallel to cleavage zones (Fig. 4).

Crenulation cleavage is developed in rocks, which previously had an anisotropic structure, e.g., rocks with a schistose structure or perfect disjunctive cleavage. When studying such structures under the microscope, one can see that planes of the previous structure are deformed into microscopic folds (microcrenulation) or microflexures. Limbs of microfolds enclosed upon each other, or common limbs of microflexures located one above another usually form straight zones; the steeper the microfold limbs or common limbs of microflexures, the better these zones are expressed. Such zones (cleavage zones) cut the rock into bands and small lenses (microlithons), which enclose microfold hinges or low-angle limbs of microflexures. Limb zones are almost completely composed of mica flakes, fine grains of ore minerals, and finely dispersed organic matter. Small lenses bear relics of the earlier structure, which forms either microfold hinges or low-angle limbs of microflexures. Relative to limb zones, lenses are always enriched in quartz, calcite, and feldspar.

The understanding of cleavage as a structure organized in a certain way is important because this suggests an opportunity to study the genesis of cleavage of different types based on their morphology. Cleavage in studied Jurassic terrigenous complexes of the Greater Caucasus is mainly represented by the intergranular type, therefore, it seems reasonable to first consider the genesis of this structure. The formation of this cleavage is related to nonuniform rock texture at the grain level. Compression of rocks results in elevated pressure at grain contact lying perpendicular to the compression axis; under a certain value of compression force, the least amount in the contact area, the higher the pressure (pressure is equal to the force/contact area ratio, i.e., $P = F/S$). In areas of elevated pressure, the crystal texture is "looser" (caused by the elevated concentration of lattice vacancies) and favors the intense solution of substance in the fluid. Such processes of oriented dissolution result in shortening deformation along the compression axis, whereas dissolution zones concentrate insoluble substances and form cleavage zones (dissolution results in transformation of point contacts into planar ones). A dissolved substance is carried by fluid and redeposited in sites with lowered pressure, e.g., at grain sides parallel to the compression axis, in microhollows, microfissures, and so on. In more detail, the mechanism of the disjunctive cleavage formation is described in (Nicolas, 1984; Talitskii, 1989a). The mechanism of crenulation cleavage formation is only distinguished by the following feature: the pressure difference during its formation is generated between limbs and hinges of microfolds, or between short and long limbs of microflexures (Cosgrove, 1976; Gray, 1979; Talitskii, 1983). Dissolution under pressure in the course of cleavage formation was confirmed by many finds of partly dissolved organic remains, oolites, and other bodies of known morphology in rocks with cleavage. Moreover, such observations suggest that visually distinct cleavage appears in rocks under deformation shortening by approximately 20–25%. Under lower values, the structure is seen only under the microscope.

The study of disjunctive cleavage intensity under the microscope allows the rock deformation degree to be roughly estimated at the grain level. Figure 5 demonstrates the dependence of spherical grain contact areas (S), with the diameter taken to be unity ($d = 1$), on the pressure (P) over these areas under the constant force (F) equal to unity. The plot also shows deformation shortening in percentages. It is evident that the increase, due to dissolution of the contact area, results in a gradual pressure fall. Under a shortening of 5 to 50%, pressure at contacts decreases almost by an order, with the most significant falling under a shortening value of 5 to 15%. Under deformation shortening of 15 to 25%, the dependence between the pressure fall and contact areas is almost proportional, whereas under a greater deformation, the pressure fall sharply decelerates. Such a curve pattern suggests that deformations resulting from dissolution under pressure represent typical deformations accompanied by rock stabilization. Most readily, they take place under the limited area of grain contacts (5–15%). Under deformations of 15–25%, distinct planar structures appear (the area of grain contacts substantially increases) and as a result, further deformation sharply decelerates.

Earlier, we performed a genetic analysis of the relationships between cleavage and folds of longitudinal winding (Talitskii, 1989b), which showed that universally developed transverse cleavage is formed either at the early stage of folding and longitudinal bed shortening, or at the later stage of general flattening, i.e., either prior to folding or after termination of the latter. The feature indicating that it was formed at an early stage is the radiating distribution of cleavage zones and their breakage at layer boundaries in limbs, whereas its formation at the later stage is indicated by parallel axial-plane cleavage penetrating all layers regardless of their lithology. In addition to arguments of the genetic relation between cleavage and folds discussed in the above-mentioned paper, the S–P plot can be used as another. It is apparent that under joint deformation by longitudinal shortening of fine-grained and coarse-grained rock layers, disjunctive cleavage develops more readily in fine-grained rocks, in which cleavage is always expressed better (Fig. 3), i.e., such rocks are stabilized faster. This results in the growth of the viscosity contrast, loss of rock rigidity in the bedded sequence, and a folding of the latter. Further cleavage perfection in such folds occurs later, at the stage of general flattening. At this stage, cleavage can also develop in coarse-grained rocks. Thus, the intensity of folding deformation can be judged from the intensity of cleavage development in different rocks.

Cleavage developed in rocks from the considered Jurassic section of the Caucasus was studied both visually and by using a microscope. The degree of rock fissility and its orientation were registered visually, whereas the morphological type of cleavage and degree of rock deformation in the course of this structure's formation were estimated under the microscope. In total, approximately 300 thin sections were studied. As mentioned above, the disjunctive cleavage represents a principal structure developed in the rocks in question. Some blocks revealed superimposition of crenulation cleavage on the intergranular one. In addition, sometimes, rocks bear structures that differ from cleavage; the latter are discussed below when describing these rocks. All thin sections with traces of bedding reveal breaking of cleavage planes at the layer boundaries (Fig. 3), allowing the considered cleavage to be referred to as "prefolding"-type cleavage formed at the stage of longitudinal layer shortening. Later, through a course of folding deformation, the cleavage underwent reorganization and thus, perfection.

In the intensity of cleavage development, rocks from the northern and southern parts of the profile differ from each other: in the southern part, they reveal intense cleavage, whereas in northern areas, it is less developed or, sometimes, missing. The rock deformation intensity estimated, based on the cleavage development, is shown in Fig. 2d. Figure 3 demonstrates a series of photomicrographs showing a change in cleavage intensity for different blocks (nappes) of the northern part of the profile.

In rocks from block I located in the northern limb of the Dar'yal anticlinorium, cleavage is absent (Fig. 3, sample 413). Visually, the cleavage is first registered only in the southern part of block II, where cleavage planes have a steep southward dip. Microscopically, cleavage is first registered in the northern part of block II. To the south, the cleavage intensity gradually increases (Fig. 3, samples 408–399). Rocks from the northern

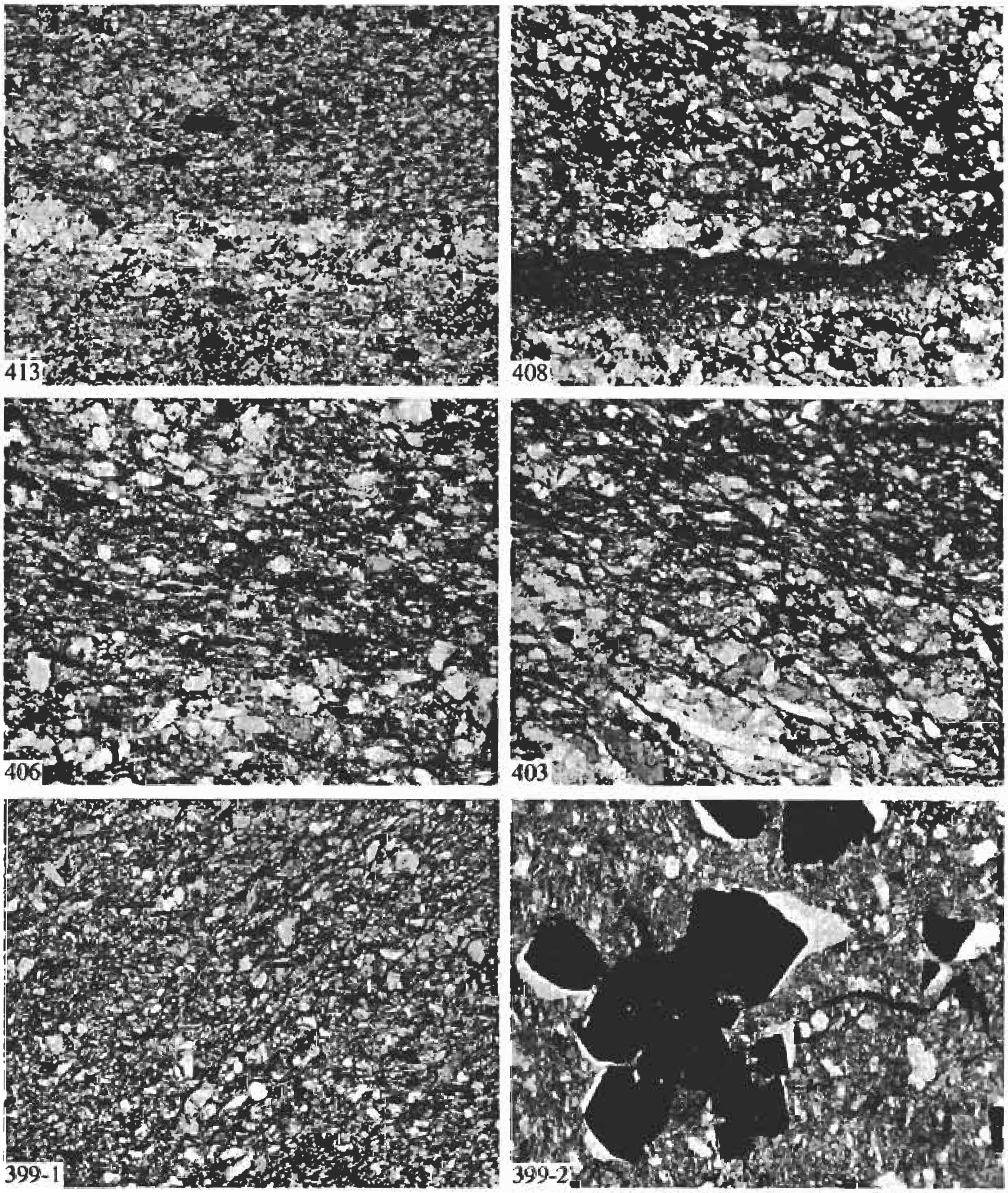

Fig. 3. Photomicrographs of rock samples from different blocks from the northern part of the profile ranging from the cleavage-free argillite zone (sample 413) to the shale zone with perfect cleavage (samples 391–396). Numbers correspond to sample numbers in Fig. 2. The width of the horizontal side is 1 mm.

part of this block reveal only short isolated cleavage zones, whereas some thin sections of fine-grained siltstones from its southern part show quite perfect cleavage. Areas near pyrite crystal inclusions in cherty concretions reveal the presence of "pressure shadows" resulting from compression strain and filled with authigenic quartz (Fig. 3, sample 399-2).

Blocks III and IV, located further to the south, expose different cleavage intensity. Visually, rocks from block III virtually lack cleavage, whereas, under the microscope it is quite distinct (Fig. 3, sample 398). Rocks from block IV reveal well-pronounced, subvertical, or steep, southward dipping structures; an intense cleavage is also registered under the microscope in clayey siltstone (Fig. 3, samples 391–396), which allows the deformation shortening to be estimated at 25% (Fig. 2d).

Both siltstone and sandstones of the Jurassic Kistinka Formation reveal intense cleavage in the axial part of the Dar'yal–Bogos anticlinorium. Some mem-

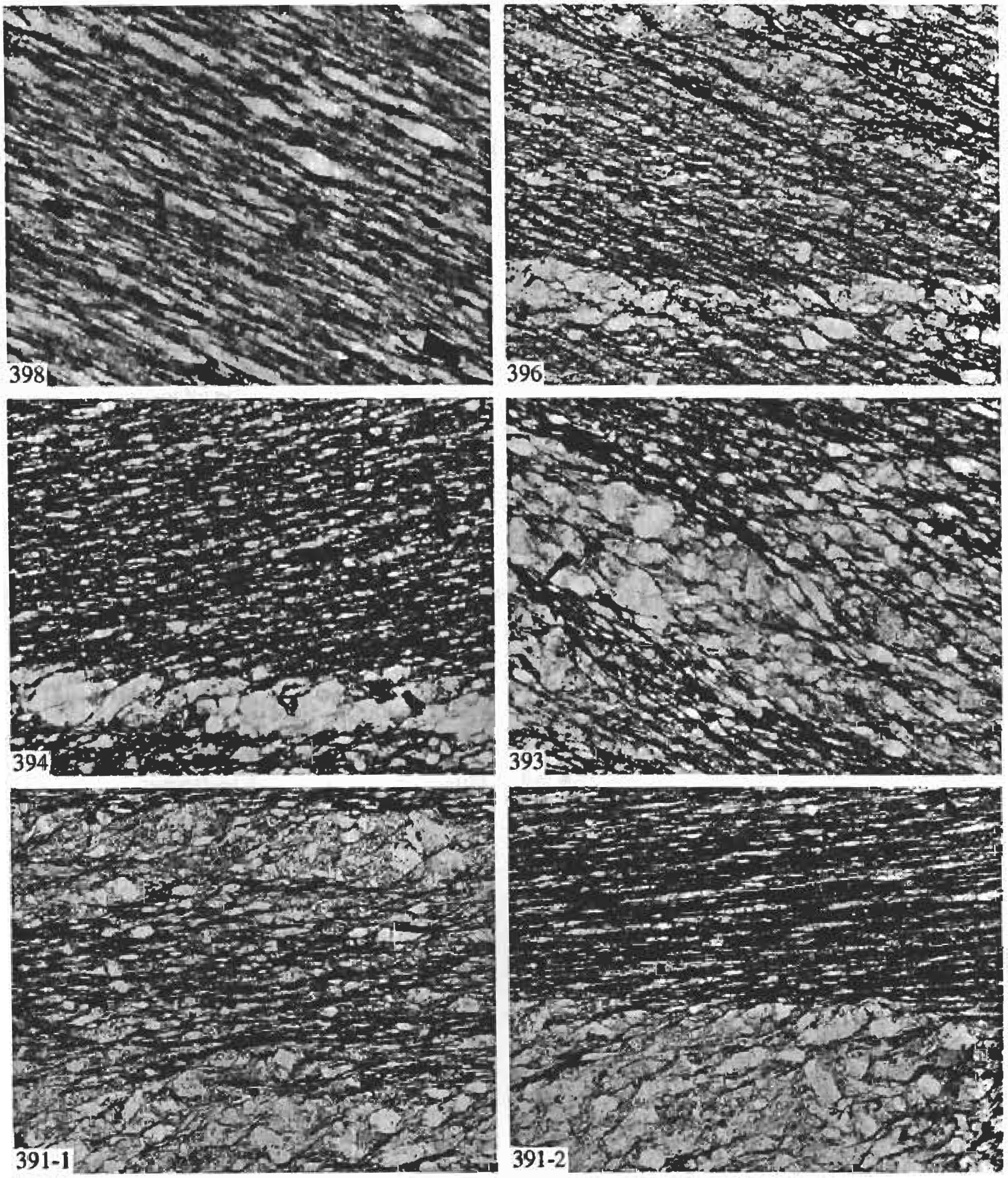

Fig. 3. (Contd.)

bers of clayey siltstone contain large andalusite porphyroblasts. Pre-cleavage porphyroblasts are usually rounded, or "rough-hewn." In the course of deformation and cleavage formation, these porphyroblasts experienced rotation owing to which cleavage planes around porphyroblasts are often twisted and deformed into microfolds.

Quartzitic sandstones from the Kistinka Formation show the development of a peculiar texture, which is sometimes termed as a mylonite or cataclasite texture. This structure is represented by large quartz grains with serrate boundaries and fringed by small, yet abundant quartz grains. Often, large grains of quartz are uniformly elongated. A structure of this kind is formed under oriented compression of rocks in the process of polygonization and rotational crystallization (Poirier, 1985). Rotational crystallization affects marginal parts of large quartz grains, whereas their cores show undulatory and cloudy extinction indicating polygonization but not a recrystallization process.

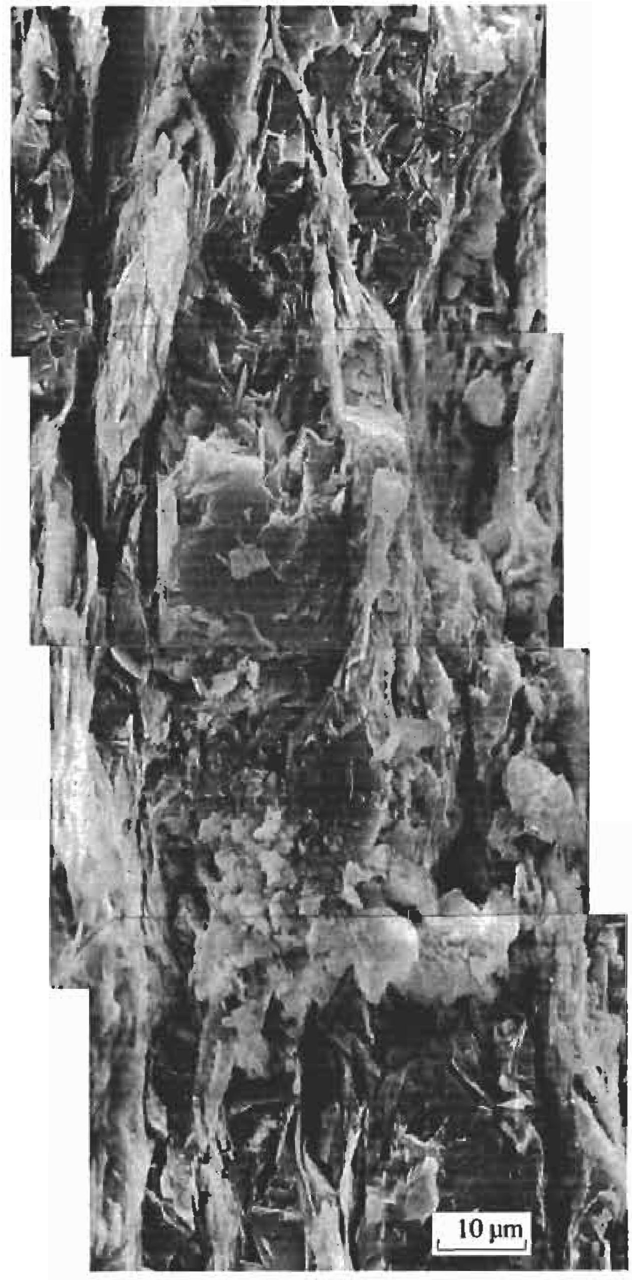

Fig. 4. SEM image of clayey–silty shales from the zone of the intense cleavage development in block IV of the northern part of the profile (sample 394; Figs. 2, 3). Zones with cleavage are marked by the prevalence of authigenic mica (sericite). Quartz grains are observed in microlithons in spaces between cleavage zones.

Rocks of the Tsiklauri Formation in the southern limb of the Dar'yal–Bogos anticlinorium reveal a well-developed cleavage. The microscopic study of these rocks showed that intense cleavage developes in both siltstones and sandstones. The estimated shortening, based on cleavage in sandstone, makes up approximately 25%.

The block located directly south of the Shaukhokh thrust is composed of volcanogenic and sedimentary rocks of the Tsiklauri Formation. Microscopically, volcanogenic rocks are marked by relatively weak cleavage. Based on this cleavage, the shortening is estimated to be 15%. Sedimentary rocks from this block, as well as those from block II, reveal at the macroscopic level, well-developed cleavage with steep subvertical planes. Studying these rocks under the microscope shows that cleavage is better developed in siltstones and less intense in sandstones. The deformation in the course of formation of steep cleavage is estimated to be about 20%.

The third and fourth nappes are composed of Middle and Lower Jurassic rocks, respectively. Nonetheless, these rocks show similar cleavage intensity both at macroscopic and microscopic levels. Perfect cleavage in these nappes is developed both in siltstones and sandstones, which indicates significant rock deformation.

The fifth block of the southern part of the section is marked by perfect cleavage, which is clearly visible; under the microscope, however, well-developed cleavage is detected only in siltstones whereas, in sandstones, it is either absent or forms primordial, poorly pronounced cleavage zones. Based on this kind of cleavage, deformation shortening is estimated to be about 15–20%.

Beneath the microscope, it was evident that Jurassic sequences along the considered profile across the Greater Caucasus are characterized by uneven cleavage development. It is less intense in the northern part of the northern limb of the Dar'yal–Bogos anticlinorium and best-developed in the core of the structure. Cleavage-related deformations in the southern part of the section were significant but also uneven. The greatest intensity is seen in rocks of the Middle Jurassic Gudoshauri Formation and the Lower Jurassic Kazbek Formation that compose blocks III and IV, where the well-developed cleavage is characteristic of both fine-grained and coarse-grained varieties. The latter fact probably suggests strong folding deformations.

Thin sections of rocks from different blocks along the profile reveal textures superimposed on disjunctive cleavage. Textures of such kind are manifested as rare kink zones, as well as in variably developed crenulation cleavage. Throughout the course of visual studies, these textures were not registered nor measured and therefore, they are not described in detail, although their presence in the cross section is marked by a special symbol (Fig. 2).

CHARACTERISTICS OF STRESS-FACIES PARAMETERS

In addition to cleavage, deformation microtextures developed in quartz grains from terrigenous rocks were also studied. The presence of such textures allows for qualitative estimation in the level of strains affecting rocks in the course of their deformation. The method of

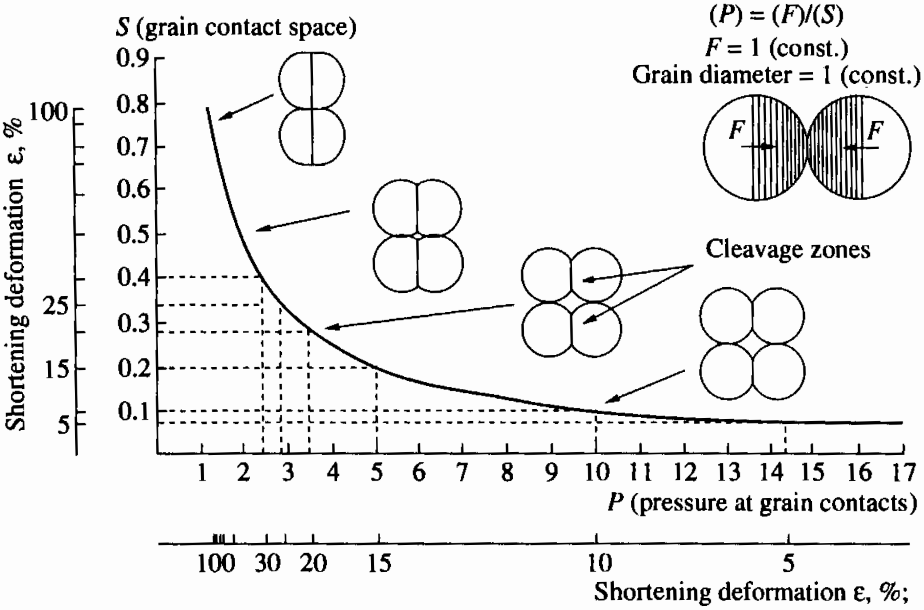

Fig. 5. The relation between contact area of grains (*S*) and pressure (*P*) under constant force (*F*).

such assessment is described in detail in (Galkin, 1992).

Examining deformation textures registered in quartz grain from different types of deformed rocks under the microscope allows us to establish a general succession in the development of such textures (stress-facies):

(σ1) Rocks are undeformed; signs of planar, linear, and other oriented structures are absent; quartz grains lack signs of deformation in marginal zones and in the core, or these deformations reveal no spatial regularities (buried textures) at all;

(σ2) rocks acquire elements of oriented structures of different intensity;

(σ2a) quartz grains can be corroded along their periphery (in low-temperature facies), and other deformation textures are absent;

(σ2b) quartz grains in peripheral zones are granulated, and deformation textures in the central part are missing;

(σ3a) quartz grains reveal undulatory extinction and Boehm lamellae in their central part;

(σ3b) intense undulatory extinction and distortion of orientation in some sites, accompanied by the appearance of distinct grain boundaries and by partition of grain into 3–4 fragments;

(σ4) granulation and abundant signs of destruction in the peripheral and central parts.

According to the considered method, succession of these facies reflects a gradual increase in external (nominal) stress.

Rocks of block II in the northern part of the profile show a regular growth of stress influence. This correlates well with the increase in cleavage intensity observed in the same direction and, as shown below, with changes in the mineral composition of clayey rocks. In all the other southerly located blocks of both northern and southern parts of the profile, stress-facies parameters show significant variations, even within the same block.

In the stress-facies plot (Fig. 2), facies σ2 corresponds to the shaded area. Cleavage was formed under stress corresponding to this facies. Figure 2 demonstrates that many data points fall into the field of the facies σ2 development although there are points located beyond the limit of the latter. As a rule, such data points are confined to the zones adjacent to faults, suggesting that after cleavage and fold formation, the further increase of load on rocks resulted in their uneven stabilization and formation of faults along the areas maximum in values. Further deformation was related to block movements along fractures.

Unfortunately, the available material is insufficient for compiling a well-substantiated curve of paleostress along the entire profile, therefore, the conclusion drawn is preliminary.

CHARACTERISTICS OF CLAY MINERALS

Micaceous minerals and chlorite are the most abundant clay minerals in the Lower–Middle Jurassic rocks exposed along the considered profile; less developed are kaolinite, mixed-layer mica–smectite minerals. Other phyllosilicates occur rarely. Despite the phase-composition, similarity of clayey rock varieties, rocks from different parts of the profile reveal peculiar mineralogical features characterizing specific conditions of their formation. Figure 2 demonstrates some mineral characteristics of rocks developed in different parts of the profile.

Micaceous minerals.[1] All samples studied reveal a variable amount of dioctahedral micaceous mineral mainly of the muscovite type and virtually lacking expanding interlayers.

The northern part of the profile, where the cleavage is absent, is occupied by rocks, which are noticeably different from those developed in southern areas. Here, clayey rocks of block I and in the northern part of block II are marked by the presence of mixed-layer mica–smectite minerals, which contain up to 20–25% of expanding interlayers (Gavrilov *et al.*, 1992). The analysis of diffraction patterns of samples indicates that the alteration of micaceous and smectite layers in the mixed-layer mineral show the tendency toward regular interlayering. Similar diffraction patterns were obtained for all the samples of this part of the section. In some samples from block I, the micaceous mineral representing the main mineral phase contains 10–15% of expanding interlayers. It should also be noted that value d of the third order in diffractograms of many ignited samples is very low (d_{003} = 3.31–3.32 Å at d_{001} = 9.98–10 Å), this is apparently related to high dispersion degree of micaceous minerals.

The analysis of electron diffraction patterns of oblique structures allowed us to establish that micaceous minerals are represented by polytypes $1M$ and $2M_1$, whose distribution is quite regular (Fig. 2). Rocks of the Kistinka Formation contain mica of the $2M_1$ polytype. North of the Dar'yal granite-gneiss massif, similar $2M_1$ micas (block IV, the area of the intense cleavage development) occur in rocks occupying a relatively wide zone. Rocks from block III show the presence of rare micas of the $1M$ polytype. Further to the north, the content of the latter in rocks of block II gradually increases to become prevalent in Aalenian deposits. As seen in Fig. 2, the cleavage intensity regularly decreases in the same direction.

In the Bajocian and Bathonian deposits, the leading role again belongs to the $2M_1$ polytype; moreover, some samples lack the $1M$ polytype.

Deposits with well-developed cleavage from the southern part of the profile are mainly characterized by a $2M_1$ polytype. However, some samples from the Gudoshauri Formation show a noticeable admixture of mica belonging to the $1M$ polytype.

In micas from clayey shales of the Kistinka Formation, basal reflection of the first order is often characterized by an unusually low d value (d = 9.82 Å), whereas those of lower orders show a distinct splitting: 4.97 and 4.83; 3.32 and 3.21; 1.994 and 1.928 Å. Special studies showed that the appearance of an additional series of reflections indicates the presence of paragonite in rocks. In addition, the presence of low-intensity reflection with interplanar distances of 4.90, 3.26, and 1.960 Å allows the occurrence of a certain amount of mixed-layer paragonite–muscovite phase to be suggested (Frey, 1970, 1987).

The study of samples with a maximum paragonite content, using the electron diffraction method of oblique structures, unambiguously indicates the presence of the phase mixture of paragonite and muscovite; moreover, both dioctahedral micas are characterized by the $2M_1$ polytype.

Not all samples studied show the presence of paragonite. Its maximum content (30–40%) is registered in basal layers of the section near the contact with granite-gneiss of the Dar'yal massif (Gurgala Mountain area). Samples from the upper part of the formation (the southern slope of the Molchechkort Mountain) reveal the lower paragonite content (≈10%); sometimes, it is completely absent.

No paragonite phase is reliably established in deposits of other stratigraphic intervals. Nonetheless, diffraction patterns of some samples from the uppermost layers of the southern limb (Busarchil Formation) show weak reflections with $d \cong 4.87$ and 3.25 Å, probably indicating the presence of some amount of paragonite in micas.

The mica crystallinity degree. Comparison of diffraction pattern of samples from different parts of the profile shows that minerals differ in the crystallinity degree. To assess this parameter of phyllosilicates, different crystallinity indices (Weaver, Kübler, Weber, and Flemming) are used. As was noted in (Frey, 1970; Drits *et al.*, 1997; and others), the crystallinity index (CI) has certain shortcomings because it partly depends on factors directly unrelated to the state of minerals. Nonetheless, with the availability of many determinations, the crystallinity index provides statistic data reflecting the true tendencies on the transformation of phyllosilicates. Crystallinity indices of mica from rocks of the sequence in question were determined using the method described in (Kübler, 1964; Kisch, 1991a). The Kübler crystallinity index is better applied to rocks that experienced significant postdiagenetic alterations, which are characteristic of rocks distributed along the considered profile. Figure 2 demonstrates the CI variations in different parts of the profile. In the Dzherakh–Dar'yal subzone, minimal CI values (0.21° Δ 2θ) are peculiar of deposits of the Kistinka, Tsiklauri, and lower Lars formations, which are characterized by the best-developed cleavage. Northward, the CI values increase to 0.5° Δ 2θ); simultaneously, the cleavage intensity decreases. In some samples from the Aalenian deposits of block I, where cleavage is completely missing, reflections in diffractograms become wider at the expense of reflections of mixed-layer illite–smectite varieties superimposed on those of mica proper. It is interesting that in the Bajocian and Bathonian deposits, the width of reflections slightly decreases.

The CI values in rocks with intense cleavage, which developed in the Kazbek–Gudoshauri subzone, vary to

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[1] Terms "hydromica" and "sericite" are used in this work in line with classification discussed in (Omel'yanenko *et al.*, 1982): sericite is taken to mean finely dispersed micaceous (muscovite type) minerals with expanding layers making up less than 5%.

a lesser extent (Fig. 2c). The Tsiklauri, Kazbek, and Gudoshauri formations are characterized by close CI values (0.20°–0.22° Δ 2θ); younger deposits of the Busarchil Formation show slightly higher CI values (up to 0.25° Δ 2θ).

Chlorite is second in abundance of clay minerals. Its content in rocks varies along the profile rather regularly. In the northern part of the profile, chlorite is almost universally subordinate except for samples from the basal layers of the Lars Formation, where it is a preponderant clay mineral. In the northernmost part of the profile, which is marked by the appearance of kaolinite in the Aalenian deposits, the chlorite content decreases, and in some samples, it is replaced by kaolinite.

Deposits of the Kistinka Formation are peculiar, because chlorite is either completely missing or is sharply subordinate.

A different situation arises with chlorite in deposits of the southern part of the profile: in the Tsiklauri Formation rocks, it is universally preponderant, moreover, in clayey shale interbeds of the volcanogenic–sedimentary sequence where its amount significantly increases and reaches maximum contents. In the central part of the southern section (the Kazbek and Gudoshauri formations), the mica/chlorite ratio is variable with the preponderance of either mineral. In the upper part of the section, mica prevails over chlorite.

As in the case of micas, the crystallinity degree of chlorite along the profile significantly varies, particularly, in its northern part: crystallinity index variation is noticeably lower than that in mica (0.2°–0.25° Δ 2θ). Changes in the crystallinity degree are generally unidirectional.

The establishment of chlorite polytypes represents a rather difficult task because of the permanent presence of clay minerals whose reflections are superimposed on reflections characteristic of different chlorite polytypes. Polytypism can be reliably established only in those cases, when chlorite strongly prevails over mica. For instance, the chlorite polytype II*b* is almost unambiguously established in the Tsiklauri Formation rocks. The Alpine-type quartz veinlets (Tsiklauri and Kazbek formations) sufficiently reveal abundant segregations of chlorite, which is readily identified as the polytype II*b*. Taking into consideration that this chlorite variety represents the less stable mineral, in rocks formed under the elevated pressure and temperature, one can assume, with sufficient confidence, that there is a wide distribution of the chlorite polytype II*b* in the cleavage development area.

Even greater difficulties arise when defining chlorite polytypes in the least altered argillites from the northern part of the profile, where hydromica prevails among clay minerals. Nonetheless, there are data available that suggest the presence of chlorite polytype 1*b* in these rocks (Gavrilov *et al.*, 1992).

The transition from argillites to clayey shales in the northern part of the profile is of a particular interest because this very zone is characterized by significant mineral transformations. In particular, there is a narrow (a few kilometers) zone within the boundaries of the block II, where micaceous minerals show, first a significant decrease in the development, and then, disappearance of mixed-layer patterns when moving in a north–south direction. This zone is marked, first of all, by the development of mineral phase in clayey rocks, which is missing in other areas of the profile. Diffractograms reveal reflections with interplanar distances of 7.1 (most intense reflections) and 3.53 Å. By this and other features, the phase can be considered as berthierine.

Kaolinite occurs only in Middle Jurassic argillites distributed in the northern margins of the profile (in the northern part of block II and in block I), and is subordinate to hydromica; cleavage in these rocks is missing.

In addition to clay minerals, shales locally contain other silicate minerals. The peculiar feature of some horizons in the Kistinka Formation is the presence of andalusite (chiastolite) and, sometimes, Mg-cordierite. The clayey member from the lower Kistinka formation, located close to the contact with granite-gneiss, is marked by segregations of chloritoid crystals particularly abundant in thin (1–2 mm) fissures crossing clayey shales. Diffraction patterns of the <0.001 mm fraction sometimes reveal reflections indicating the feldspar admixture in the latter. These reflections become more distinct with an increase in the degree of secondary rock alterations, i.e., southward. Maximum feldspar contents are depicted in clayey shale layers of the volcanogenic–sedimentary sequence; as evident from the chemical analysis, they are represented by sodium varieties. Authigenic quartz shows similar patterns: its reflections become more distinct in samples from the zone of the best-developed cleavage in rocks.

DESCRIPTION OF CONCRETIONS

Clayey–silty deposits exposed along the profile enclose concretions of a different mineral composition (calcareous, sulfide, and siliceous). They are most abundant in the upper Toarcian–Aalenian and Bajocian–Bathonian sequences; in the Kistinka, Tsiklauri, Lars, and Kazbek formations, concretions are rare. The composition of concretions changes along the profile depending on the intensity of secondary alterations. Rocks developed in the northernmost part of the profile (block I and adjacent areas of block II) are characterized by the development of diagenetic, calcareous, and mainly siderite concretions; however, beginning at the southern part of block II and moving southward, siderite concretions give way to concretions similar to the latter in morphology (lenticular and ellipsoidal) and dimensions (10–20 cm across) but different in mineral composition. As a rule, they are composed of a fine-grained quartz–chlorite mass, which sometimes encloses abundant sulfide mineralization in the form of well-edged pyrite crystals or their clusters (Fig. 3, sample 399-2). It should be emphasized that both siderite and quartz–chlorite concretions occur in deposits formed in similar facies environments, where processes of diagenetic mineral formation have also developed in a similar way, i.e., the early diagenetic stage was marked by the formation of mainly siderite and, sometimes, calcite–siderite concretions in the considered terrigenous deposits.

As is seen, changes in mineral composition of concretions occur in areas with a maximum gradient of changes in other mineral and structural characteristics of Jurassic rocks. A similar situation was earlier described in Lower–Middle Jurassic sections of western Dagestan (Gavrilov, 1982), in the transition from slightly altered Liassic rocks of the Laba–Malka structural–facial zone to strongly altered rocks of the Mukulan Formation of the Pshekish–Tyrnauz suture zone (Gavrilov and Tsipurskii, 1987). This tendency, in general, is characteristic of Jurassic rocks in those areas of the Greater Caucasus, where argillites grade into rocks marked by cleavage of variable intensity.

DISCUSSION

The material considered above shows that rocks along the studied profile are characterized by regular and concordant changes in some characteristics of sedimentary rock sequences, such as the deformation degree, cleavage development, stress parameters, mineral composition of clayey rocks (mineral associations, mineral crystallinity degree, mica and chlorite polytypes), and mineral composition of concretions. Similar relationships between cleavage development and some mineral parameters are also recorded in other folded areas (Aller *et al.*, 1987; Garcia-Lopez *et al.*, 1997; Kisch, 1991; Weaver *et al.*, 1984; and others).

It is clear that the modern mineral composition of terrigenous deposits and some of their structural–textural peculiarities have been formed mainly as a result of postdiagenetic processes. To understand the trend of these alterations, it is important to know the initial mineral composition of rocks. Earlier, this question was discussed in (Gavrilov *et al.*, 1992) and it was shown that the mineral composition of deposits is reconstructed with a high degree of confidence, at least for rocks from the northern part of the profile (accordingly, for the northern part of the Early–Middle Jurassic basin of the Central Caucasus).

Argillites from the northernmost part of the profile experienced less intense secondary alterations. Inasmuch as there are no principal differences in the terrigenous material provenance during the Domerian–Aalenian time, one can state with sufficient confidence that the mineral composition of Aalenian argillites from the northern part of the profile also reflects the initial lithology of the Domerian–Toarcian sequence. As was shown above, Aalenian rocks contain hydromica, a mixed-layer hydromica–smectite mineral, kaolinite, and chlorite. Deposits of this stratigraphic interval were subsided to depths of 3.5–4 km. In this connection, it can be assumed that mixed-layer minerals were initially characterized by a slightly higher content of smectite layers. Chlorite and kaolinite ratios varied in different facies. However, our data on the distribution of clay minerals show that Toarcian rocks in northerly sections are characterized by the prevalence of chlorite; the leading role among micaceous minerals belonged to hydromica polytype 1*M* (Gavrilov and Tsipurskii, 1987).

Deposition of the Kistinka Formation corresponding to the base of the section occurred in slightly different environs, owing to which it differs in some mineralogical and geochemical characteristics from deposits of the remaining Jurassic sequence. As shown above, clayey rocks of the Kistinka Formation are characterized by the preponderance of micaceous minerals and by a distinct subordinate content, or complete absence, of chlorite. Shales also contain andalusite (chiastolite) and products of its replacement as well as an often significant admixture of paragonte, in addition to a muscovite-type mica. The formation of the modern association of silicate minerals in shales was related not only to intense postdiagenetic alterations but, to a great extent, to the initial mineral and chemical composition of Liassic sediments. When reconstructing the initial lithology of rocks, their deposition environments should be taken into consideration.

The results of the facies analysis, both previous (Chikhradze, 1979; and others) and in this work, suggest that the Kistinka Formation deposits were accumulated in the shallow basin. The latter was apparently rich in islands; for some, its bays were pinched and overgrown, which promoted the accumulation of organic matter in sediments. The terrigenous material transported into the basin from land was represented by erosion products: of Paleozoic crystalline schist and granite-gneiss. It is quite natural that sediments contained abundant micaceous minerals. Simultaneously, the land likely provided kaolinite as well, which was formed in weathering crust and lacustrine–boggy systems. This assumption is supported by data on mineral composition of continental deposits developed in northerly areas. For instance, in clayey rocks of the lower Pliensbachian coal-bearing Khumarin Formation (Laba–Malka structural–facies zone), the dominant role belongs to kaolinite, whereas hydromica is subordinate (Gavrilov and Tsipurskii, 1987). Coeval marine deposits (Kyrtyk depression) are characterized by a hydromica–kaolinite association; a similar composition of clayey rocks is also registered in other areas.

Thus, in the early Pliensbachian time, when the central part of the Greater Caucasus trough represented an area of accumulation of the Kistinka Formation deposits, in adjacent northerly areas, kaolinite and hydromica were the preponderant constituent of coeval clayey sediments. Basins that accumulated the Kistinka Formation sediments were characterized by a peculiar environs, which favored, owing to diagenetic processes, the formation of authigenic kaolinite in sediments highly enriched in plant organic matter. Therefore, one can suggest that mineral composition of the Kistinka Formation sediments was determined by the presence of mica, hydromica, and kaolinite (allothigenic and authigenic); chlorite was present in a subordinate amount.

The structural reorganization of the area, that occurred in the Bajocian–Bathonian time and was reflected in the appearance of a system of rises in the central part of the basin, resulted in slight changes of the terrigenous material provenance and in the development of a new association of accessory minerals in sandy–silty rocks (Bezborodov, 1961). Rises were composed of pre-Bajocian terrigenous deposits (in addition to crystalline rocks of the basement), which experienced substantial postdiagenetic changes by that time. As a result of their erosion and transport of the sedimentary material into basin, sediments became enriched in the hydromica polytype $2M_1$ to its absolute preponderance. However, Bajocian hydromicas contain some amount of expanding interlayers and are less crystallized as compared with the cleavage development zone, where micas are also represented by the polytype $2M_1$.

As for the initial lithology of clayey rocks from the southern part of the profile, they are characterized by some peculiar features against a background of general similarity between modern mineral parageneses from the northern and southern areas. First, it should be noted that parameter b of sericites is higher than that in micaceous minerals from the northern part of the profile. Maximum values of b are characteristic of clayey shales from the volcanogenic–sedimentary horizon of the Tsiklauri Formation. Noteworthy is the coincidence of parameter b growth and the chlorite content increase in rocks. Second, the entire section of the southern part of the profile is characterized by sericite phengitization. One of the probable causes of this event is the primary composition of the terrigenous material transported from southern land. As Chikhradze (1979) notes, biotite and chlorite are persistently present among minerals of the heavy fraction in the southern Dzirul and Svanetiya terrigenous–mineralogical provinces (TMP) that provided the sedimentary material, whereas in deposits of the Malka–Argun and Urukh–Ardon TMPs (in the northern areas), biotite is missing. The appearance of biotite in the initial material at the stage of strong secondary alterations could result in sericite phengitization (Gavrilov *et al.,* 1992). It means that the initial mineral composition of clayey sediments in the southern part of the basin was similar to that in the northern part, only differing in its presence of biotite.

Thus, the initial composition of clayey sediments from the central part of the Early–Middle Jurassic basin of the Greater Caucasus can be characterized, with great confidence, as follows. In the Sinemurian–early Pliensbachian (Kistinka) time, sediments were marked by the mica–hydromica–kaolinite association, whereas in the late Pliensbachian–Bathonian time, preponderant was the hydromica–mixed-layer hydromica/smectite mineral–kaolinite–chlorite association, which was probably supplemented by biotite in the axial and southern parts of the basin. Indeed, proportions of minerals in this paragenesis could sometimes vary significantly during sedimentation.

It should be noted that there are no grounds to believe that the Jurassic basin sediments contained any appreciable amount of montmorillonite, whose presence in the terrigenous sequence was assumed by some researchers to substantiate conclusions on the mechanism of deformations.

As shown above, the clay minerals in modern clayey rocks that developed along the major part of the profile are represented by the sericite–chlorite association formed in the course of intense secondary alterations. The question on the nature of these secondary alterations is of principal importance for reconstruction of evolution of the Lower–Middle Jurassic sedimentary complex. The presented factual material shows that textural, structural, and mineral characteristics of sedimentary sequences regularly change along the profile. In this respect, of peculiar interest is the northern limb of the Dar'yal–Bogos anticlinorium because this area is marked by a transition from rocks least affected by secondary alterations to those with a maximum alteration degree. This area is also characterized by complication of structure from simple monocline in the northern block (I) to sufficiently deformed deposits in the southern block (IV) adjacent to the Dar'yal massif and marked by distinct folding (Fig. 2). Intermediate blocks represent a compressed syncline (the northern part of block II) and monocline complicated by small folds in fracture zones.

The cleavage intensity drastically changes in the same (north–south) direction: from its complete absence in block I to the maximum development in block IV (Figs. 2, 3). Blocks III and IV reveal superimposed structures.

It is interesting to note that the most drastic mineral and structural changes are characteristic of block II. This is reflected in the replacement of clay mineral associations (hydromica–mixed-layer hydromica/smectite mineral–kaolinite–chlorite by sericite–chlorite), in the substantial change in the mica polytype ratio (from $1M > 2M_1$ to $2M_1 \gg 1M$), in the increase of the mica crystallinity degree, of changes in the mineral composition of concretions from carbonate (mainly, siderite) to quartz–chlorite–sulfide composition, and in the fast growth of cleavage intensity (seen under the microscope). In other words, this zone located between the Suargom fault in the north and Pui fault in the south is characterized by a maximum gradient of changes in mineral composition of rocks, structural patterns (cleavage), and stress parameters of sedimentary rocks. The fact that different parameters reveal similar trends in changes apparently indicates their relation to the same cause, i.e., to stress. It was noted above that stress-induced cleavage is accompanied by dissolution of many primary components of sediments. Owing to diffusion and migration, the dissolved substance can be redeposited at other sites. This process was likely responsible for the appearance of abundant quartz veins in the Pui fault zone and near fractures in the southern part of block II; in the northern areas of the profile, their amount is substantially less. Under growing stress and cleavage development, siderite in diagenetic concretions also became unstable, which resulted in its replacement by silicate (quartz and chlorite) and sulfide (mainly pyrite) minerals. The latter were partly formed at the expense of Fe, which was formerly a constituent of siderite.

At the same time, it should be taken into consideration that lower layers of the Jurassic sequence experienced, in addition to stress, the influence of significant geostatic load. When estimating values of this load, one should take the total thickness of the Aalenian sequence to be about 5.5–6 km.

The thickness assessment of the Aalenian sequence, which could build on the sedimentary section in the central part of the trough, is a difficult task. In the Bajocian–Bathonian time, the formerly common basin was divided, due to structural reorganization of the region, into autonomous northern and southern basins, whose depocenters were located substantially to the north and south, respectively. In addition, this time was marked by erosion of central uplifts and transport of their materials into the Middle Jurassic post-Aalenian basins. Thus, the thickness of the Bajocian–Bathonian deposits can be neglected in calculations (at least for the axial part of the Liassic–Aalenian basin). The thickness of the Upper Jurassic–Cretaceous sequence is also unclear, nonetheless, the analysis of northern and southern sections suggests that it hardly exceeded 2–2.5 km. Thus, when estimating the geostatic load, the total thickness of Mesozoic deposits of the Greater Caucasus paleobasin should be taken to be 7.5–8.5 km.

The relationship of catagenetic and metagenetic rock transformations in the course of their subsidence and alteration, due to development of stress (so-called stress-metamorphism or dynamic metamorphism), represents an urgent scientific problem. Different aspects of the latter were considered by many researchers (Luk'yanov, 1991; Luk'yanova, 1995; Marakushev, 1988; Yapaskurt, 1992; and others). The considered material also allows some aspects of this problem to be discussed.

In this connection, it is important to compare some coeval sequences from the northern and southern areas of the considered profile. For instance, deposits of the Aalenian Koirakh and Fortuakh formations (block I) lack any signs of stress influence, whereas rocks of synchronous Gudoshauri Formation, on the contrary, show intense cleavage, although there are no grounds to suggest that these sections subsided to different depths. As seen in Fig. 2, and above-mentioned characteristics, these rocks substantially differ in mineral composition. In the considered case, one can suggest, with sufficient confidence, that stress deformations represent a principal factor responsible for mineral and structural differences between these sequences.

The share of influence assessment of geostatic and stress loads in secondary rock transformation of the most ancient rocks (Kistinka and Tsiklauri formations) in the Lower–Middle Jurassic sequence presents a real challenge. The important role of stress in alterations of the Kistinka Formation rocks is evident from the well-developed cleavage, as well as from the presence of authigenic mica oriented along cleavage planes. The Na-mica—paragonite assemblage appears at different levels of the Kistinka Formation rocks, but is most abundant near the contact with granite-gneiss of the Dar'yal massif. This is likely related to the compression strain. According to Afanas'ev (1958), the Paleozoic Dar'yal massif was primarily composed of biotite–quartz–oligoclase plagiogranites and later converted into gneiss. Dynamic metamorphism apparently caused destruction of Na-rich minerals (oligoclase and others) and released Na that migrated with fluids into the granitoid-overlying Kistinka sequence. Simultaneously, the Kistinka Formation experienced recrystallization of its sedimentary material, which resulted in the incorporation of Na into newly formed micas in the form of autonomous phase (paragonite or mixed-layer paragonite–muscovite mineral).

It is also clear that basal layers of the Liassic sequence subsided, by the beginning of the Bajocian, to a significant depth and rocks suffered geostatic load, which represents one of the main catagenetic factors. Thus, both stress-metamorphism factors and factors of cata- and metagenesis jointly contributed to rock transformations.

At the same time, there is some indirect evidence that cleavage development commenced rather early, at the stage of the Liassic sequence accumulation. It was noted above that cleavage has a pre-folding origin. Folding processes are assumed to occur as early as in the late Liassic time (Koronovskii *et al.*, 1997). It is interesting also that dikes, which intrude upon deposits of the Tsiklauri Formation and some of which served as feeding channels during submarine lava effusion, generally form a system of parallel bodies oriented along the general strike of the Greater Caucasus structure. It seems that dikes intruded at the extension stage along a system of fissures, whose orientation was earlier determined by compression forces perpendicular to the strike of the Greater Caucasus. As was noted by Azhgirei (1968), post-Aalenian (i.e., Bajocian) dikes of the Sadon area, located west of the Terek River, cross the bedding and lack signs of dynamic metamorphism, whereas middle Liassic dikes, synchronous with the eruptive centers, experienced very strong dynamic metamorphism.

Thus, it can be concluded that the Jurassic terrigenous sequence was intermittently subjected to dynamic loads as early as the initial stage of its formation, i.e., this process was synsedimentary and, consequently, much contributed to the alteration of sediments deposited by that time. Repeated stress events in the Late Jurassic and post-Jurassic time determined the formation of cleavage in younger rocks and affected earlier cleavage as well.

We suppose that in the Liassic–Aalenian trough of the Greater Caucasus basin these intermittent stress events (alternating with periods of extension) played a more important role in transformations of Lower–Middle Jurassic deposits than did the geostatic factor.

The temperature factor is another important role presenting a problem that arises when estimating secondary rock alterations. Some aspects of its influence on postsedimentation transformations of the Lower–Middle Jurassic sequences were considered earlier (Gavrilov and Tsipurskii, 1987, 1988; Gavrilov *et al.*, 1992).

An assumption on the multiple stress events agrees well with the complicated folded structure and intense cleavage development in oldest Jurassic rocks (Fig. 2). The pulsational patterns of deformation development, secondary structures, and chemical–mineralogical alterations are consistent with the model of intermittent compression–extension strains in the ancient sedimentary basins proposed by (Cloetingh, 1986; and others). Due to compression strains, the strengthening and acceleration of chemical–mineralogical transformations in the Liassic–Aalenian sequence were probably responsible for the substantial reworking of sediments accumulated by the time of structural reorganization in the Greater Caucasus at the beginning of the Liassic time. As a result, the sedimentary material, which commenced transportation into the Bajocian–Bathonian basins from the Caucasian axial rise (within the Central Caucasian segment), was characterized by properties peculiar of substantially altered rocks, i.e., a prevalence of the mica polytype $2M_1$ (Fig. 2).

Stress events determined the general trend of chemical–mineralogical rock alterations, which is particularly reflected in transition from a relatively multicomponent clay mineral association in sediments to a mainly two-component sericite–chlorite association, i.e., in unification of the mineral rock composition. It should be emphasized that strains in different blocks could noticeably vary, resulting in complicated patterns of the stress-facies distribution along the profile.

CONCLUSION

The formation of chemical–mineralogical and textural–structural characteristics of the Lower–Middle Jurassic terrigenous sequence was a multievent process. The mineral composition of clayey rocks formed at the stages of sedimentogenesis and diagenesis, as well as at the stage of later intense transformations. An analysis of clayey rocks allows the initial lithology of Jurassic sequences to be reconstructed in high confidence: sediments of the Sinemurian–early Pliensbachian basin (Kistinka Formation) were characterized by the development of the mica–hydromica–kaolinite association, whereas the late Pliensbachian–Aalenian sediments were marked by the hydromica ($1M > 2M_1$)–mixed-layer illite/smectite minerals–kaolinite–chlorite association; in the southern part of the basin, the latter was probably supplemented by biotite. Rock transformations at the stage of subsidence and stress-metamorphism resulted in the formation of a mainly sericite ($2M_1$)–chlorite association in clayey rocks. The Kistinka Formation rocks contain paragonite, in addition to sericite of the muscovite type.

Regular and consistent changes in mineralogical characteristics (mineral associations, mica and chlorite polytypes, crystallinity degree of clay minerals, and others), and intensity of cleavage development, suggest that stress was the leading role in the formation of chemical–mineralogical and textural–structural properties of rocks in the area of distribution of clayey shales with cleavage. Substantial alterations of the Liassic–Aalenian rocks probably occurred as early as at the deposition stage. Therefore, when pre-Bajocian tectonic reorganization transformed the trough into rises, the latter supplied Bajocian–Bathonian basins with the terrigenous material characteristic of rocks that already experienced intense secondary alterations. Later, repeated stress-related metamorphic events resulted in a higher intensity of cleavage development, and in the formation of superimposed structures in the Lower–Middle Jurassic sequences, as well as in the formation of cleavage in younger deposits.

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