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The Influence of Geodynamic Factors on the Postsedimentary Lithogenesis of Jurassic Terrigenous Complexes of the Caucasus (Southern Dagestan)

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Abstract—The influence of geodynamic factors on the postsedimentary lithogenesis of terrigenous sequences during the replacement of the passive continental-margin regime to the active regime is considered with the Toarcian–Aalenian complexes of southern Dagestan as an example. It has been established that clayey rocks were lithified under conditions of subsidence lithogenesis, whereas the orogenic catagenesis of sandy rocks was related to stress initiated by the initial (amagmatic) subduction phase. The Kübler index has been used to reveal the postsedimentary zonality that is consistent with the intensity of folding and cleavage. The metagenesis zone fits the cleavage zone. The Rb–Sr and K–Ar datings of clayey rocks from the metagenesis zone yielded 180–190 Ma that can be interpreted as the timing of the maximal manifestation of postsedimentary lithogenesis.

INTRODUCTION

The aim of this communication is to decipher the spatiotemporal trend of postsedimentary lithogenesis under the influence of geodynamic factors with the Toarcian–Aalenian terrigenous sequences of southern Dagestan as an example. Simanovich and Yapaskurt (2002) have considered the general aspects of the issue of geodynamic types of the postsedimentary lithogenesis. They assume that the subsidence lithogenesis (diagenesis and catagenesis) versus metagenesis (anchimetamorphism) transition is related to changes in geodynamic regimes (folding, cleavage, and fluid-thermal environment). Metagenesis and metamorphism can overlap lithified rocks subjected to deep catagenesis or interrupt subsidence lithogenesis at the initial stage of catagenesis or even diagenesis. Hence, the influence of a plethora of factors on the postsedimentary lithogenesis is integrated by geodynamic regimes of specific sedimentary basins.

It is necessary to scrutinize the issue formulated above based on the study of reference models. Therefore, we chose the thick Jurassic terrigenous complex in the northeastern Caucasus as an example. Rocks of this complex accumulated in a passive continental margin. At the terminal Aalenian–initial Bajocian, these rocks underwent the early Alpine orogeny owing to changes in the geodynamic environment from the passive continental margin to the active one.

During the field works, we attempted to cover as much as possible all stratigraphic subdivisions of the Toarcian–Aalenian terrigenous complex and, first of all, collect specimens and samples from districts with different types of folding and cleavage. In total, we scrutinized 95 outcrops, where sandstone, siltstone, and mudstone specimens were picked up; 70 mudstone and shale samples were also taken for the extraction of clayey fractions (Fig. 1).

GEOLOGICAL SETTING AND LITHOLOGY OF TOARCIAN–AALENIAN COMPLEXES

The small-scale schematic geologic map of the study region (Fig. 1) presented in this work is based on materials of the geological survey carried out in 1985 by S.I. Syrovatskii and other geologists of the North Caucasian Industrial-Geological Association (scale 1: 50000). According to these geologists, the thickness of the Middle Jurassic sequence ranges from 2.8 to 5 km in different structural–facies zones of the region.

In general, the monotonous Toarcian and Aalenian complexes of the study region consist of rhythmic alternation of different proportions of sandstones, siltstone, and mudstone interlayers in different sequences of the Toarcian–Aalenian succession. According to (Gavrilov, 2002), the Lower–Middle Jurassic complex of the northeastern Caucasus formed under the influence of a large river system that carried out huge amounts of terrigenous material to the sedimentation basin. The accumulation of deltaic sediments was compensated by intense subsidence of the basin floor and eustatic fluc-

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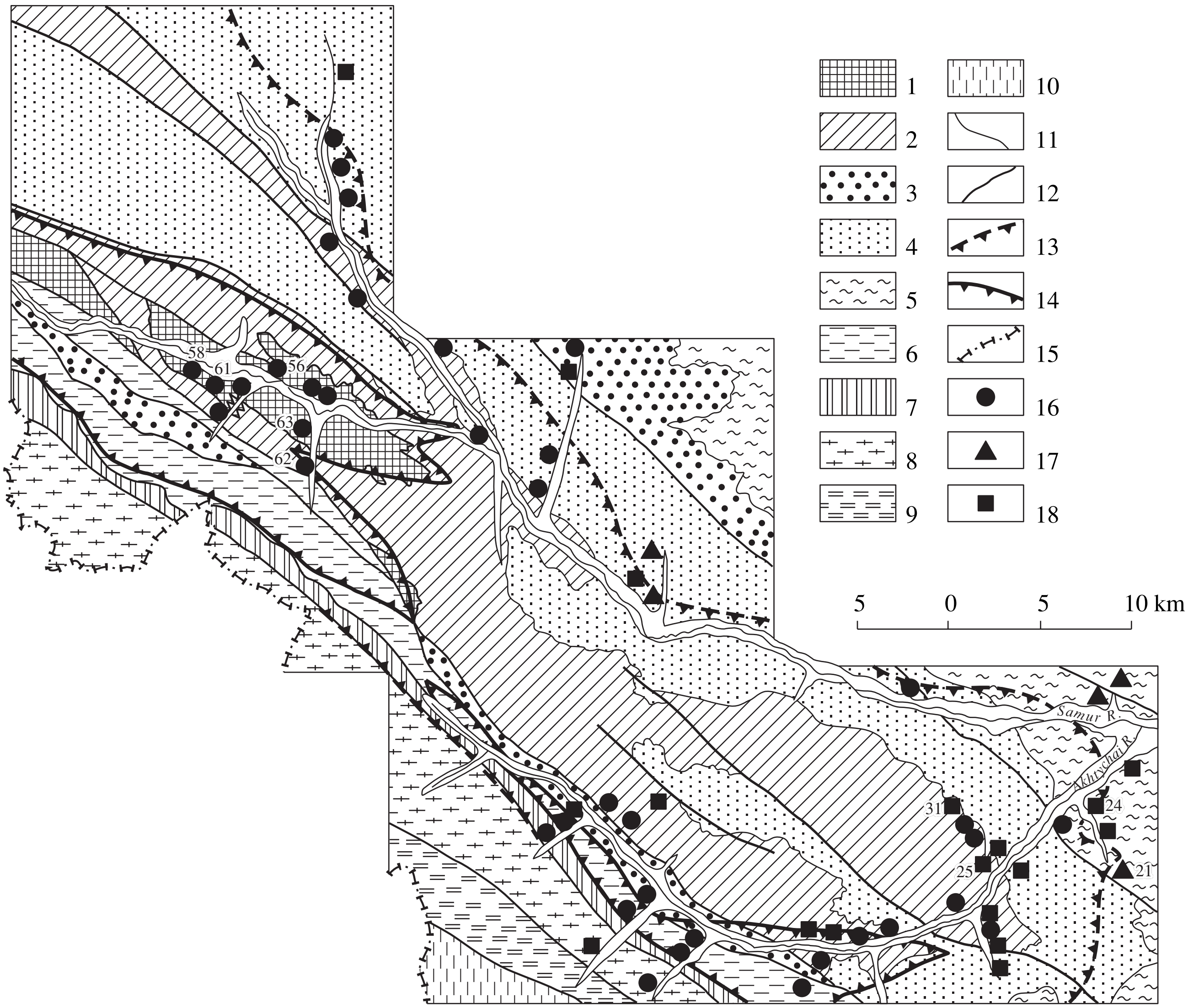

Fig. 1. Schematic geologic structure and postsedimentary transformation zones (supplemented with materials of S.I. Syrovatskii *et al.*). (1, 2) Lower Jurassic: (1) lower–upper Toarcian, (2) upper Toarcian; (3–8) Lower and Middle Jurassic: (3) Toarcian–Aalenian, Burshi Formation (in north) and Bezhta Formation (in south), (4) lower Aalenian, (5) lower–upper Aalenian, (6) upper Aalenian, (7) Aalenian–Bajocian, (8) Bajocian–Bathonian; (9) Upper Jurassic; (10) Upper Jurassic–Lower Cretaceous; (11) geological boundaries; (12) faults; (13) boundaries of orogenic catagenesis zones I and II (K.I. = 0.60); (14) boundaries of orogenic catagenesis zone I and metagenesis (K.I. = 0.42); (15) state boundary; (16–18) sampling points and minerals of the hydromica–sericite series: (16) Δd = 0–0.05, (17) Δd = 0.06–0.15, (18) Δd > 0.15.

tuations of the sea level, resulting in the regression or seaward advancement of the delta. This was responsible for the prominent cyclic structure of the Toarcian–Aalenian sequence. Gavrilov (2002) identified cycles of three orders. Cycle of the first order (hereafter, cycle I) is characterized by a thickness of a few hundreds of meters and caused by periodic subsidences of the basin floor and its compensation infilling. Cycle of the second order (cycle II) is related to the seaward advancement of sandy fans of the delta. Cycles I and II have a regressive structure. Cycle of the third order (cycle III) is marked by the centimeter-scale alternation of silty and clayey strata, presumably, caused by the seasonal (flood-related) compositional variations of the suspended material.

It is necessary to emphasize that the Toarcian–Aalenian sequences of the northeastern Caucasus accumulated in the passive continental-margin environment (Panov, 2001; Lomize and Panov, 2002).

Folding in the study region shows a distinct zonal pattern. On the whole, the intensity of folding and faulting increases from the northeast to southwest (Sholpo, 1964; Sholpo *et al.*, 1993).

In the northern areas of the study region, within the left bank of the Samur River and the Samur Range, one can outline a box fold zone where sectors with low-angle and horizontal bedding are separated by bands with high-angle bedding (fold limbs). The fold amplitude ranges from 500 to 600 m and their width reaches a few kilometers.

A considerable part of the study region belongs to the fold–block zone of the Lateral Range located at the watershed of the Samur and Akhtychai rivers and upper course of the Samur River. This zone incorporates symmetric and asymmetric, rounded and angular folds complicated by steep low-amplitude faults and numerous upthrust-type faults.

Figure 1 shows that the Toarcian–Aalenian sequences of the Bezhta fold zone are exposed in narrow tectonic blocks sandwiched between the Lateral Range fold–block zone and the fold zone of the Main Caucasian Ridge (hereafter, Main fold zone). These blocks include asymmetric, sometimes overturned, steep and dome-shaped folds often complicated by minor folds.

The southwestern area of the region incorporates the Main Ridge fold zone characterized by intense and intricate folding (asymmetric, ridge-shaped, overturned, occasionally isoclinal folds) complicated by numerous upthrusts and overthrusts.

Generally, the Toarcian–Aalenian terrigenous sequences of southern Dagestan are not cleaved even in canoe fold sectors. The intense cleavage (slate) zone, which is traced in the upper course of the Samur River (Toarcian sequence), grades with a steplike (*en echelon*) arrangement into upper Aalenian sequences in the upper and middle courses of the Akhtychai River. The metagenesis zone shown in Fig. 1 spatially fits the cleavage zone. Cleavage is not been manifested in the intensely deformed Bajocian sequences. This deformation is developed in the Bathonian sequences near the axial zone of the Main Caucasian Ridge.

PETROGRAPHIC FEATURES OF THE POSTSEDIMENTARY LITHOGENESIS

Sandstones. Irrespective of the age and deformation degree, sandstones have a homogeneous composition in the entire study region. They are observed as compact, mostly massive and less often fine-layered, brownish gray hard rocks that can be divided into the very fine-, fine, and medium-grained varieties. The sorting of clastic material varies from the common good grade to the less often medium grade. Clastic grains are poorly rounded. The clayey admixture is absent or insignificant. The sandstones have a rather homogeneous clastic component composition, % quartz 45–55, plagioclase 8–15 (K-feldspar grains are rare), rock fragments 40–50, and clastic micas 1–3. Accessory minerals are represented by the universal zircon and tourmaline and the subordinate sphene and apatite. Rock fragments are represented by lithoclasts (cherts, siliceous–micaceous rocks, metasedimentary fine-grained rocks, and very rare microlitic effusives).

Thus, according to Shutov's classification (1967), the studied sandy rocks can be referred to as lithoclastic quartz graywackes.

The sandy rocks are most informative for solving the issue of the relation of lithification with various geological processes. In undeformed sedimentary basins (subsidence lithogenesis and platformal zones), sandy rocks are lithified due to the dissolution of clastic grains under the lithostatic pressure with the formation of conform, incorporation, and microstylolitic contacts between grains. If the lithostatic and fluid pressure are approximately equal ($P_s = P_f$), the dissolved material (primarily, silica) is redeposited within the bed or adjacent layers as regeneration rims around the clastic quartz grains. The classic description of this process is given in (Kopeliovich, 1965).

Dissolution textures produced by intergranular pressure are also widespread in the studied Jurassic sandstones of southern Dagestan. One can see conform, incorporation, and less common microstylolitic contacts between grains (Fig. 2a). However, the dissolved substance was not redeposited inside the bed. This is evident from the absence of authigenic quartz (regeneration rims around the clastic grains) and other tectosilicates. The cross-cutting veinlets in sandstones consist of quartz and chlorite that are commonly concentrated in selvages (Fig. 2b). The veinlets characterized by vague boundaries represent solution filtration zones formed in weakened zones of insufficiently lithified rocks. One can also observe typical but rare late veinlets with distinct selvages related to the crushing of lithified rocks, but they always crosscut the filtration zone. These facts indicate that the pressure-induced dissolution and the subsequent removal of silica and other components beyond the beds took place in fluid ($P_s > P_f$) conditions (Marakushev, 1988) that differed from those in subsidence lithogenesis and are typical of the dynamic orogenic environment. Features indicating the dynamic environment during lithogenesis are observed even in weakly folded sectors of sandstones. Such zones contain plagioclase grains with traces of brittle deformation and quartz grains with signs of plastic deformation.

Suture dissolution surfaces, which are widespread near foliation zones and within the adjacent sandy beds (Fig. 2c), can be interpreted by two mechanisms: the further development of dissolution textures in the course of rock lithification or the initial cleavage dissolution (Galkin, 1992). In the previous work (Simanovich, 1978), we have indicated the homologous nature of such processes and shown that intergranular differential sliding textures are often observed in sandstones within the foliation zone (Fig. 2d) (Simanovich, 1978). Recrystallization–granulation blastesis of quartz was observed in only one sandstone specimen (Fig. 2e) (Simanovich, 1978).

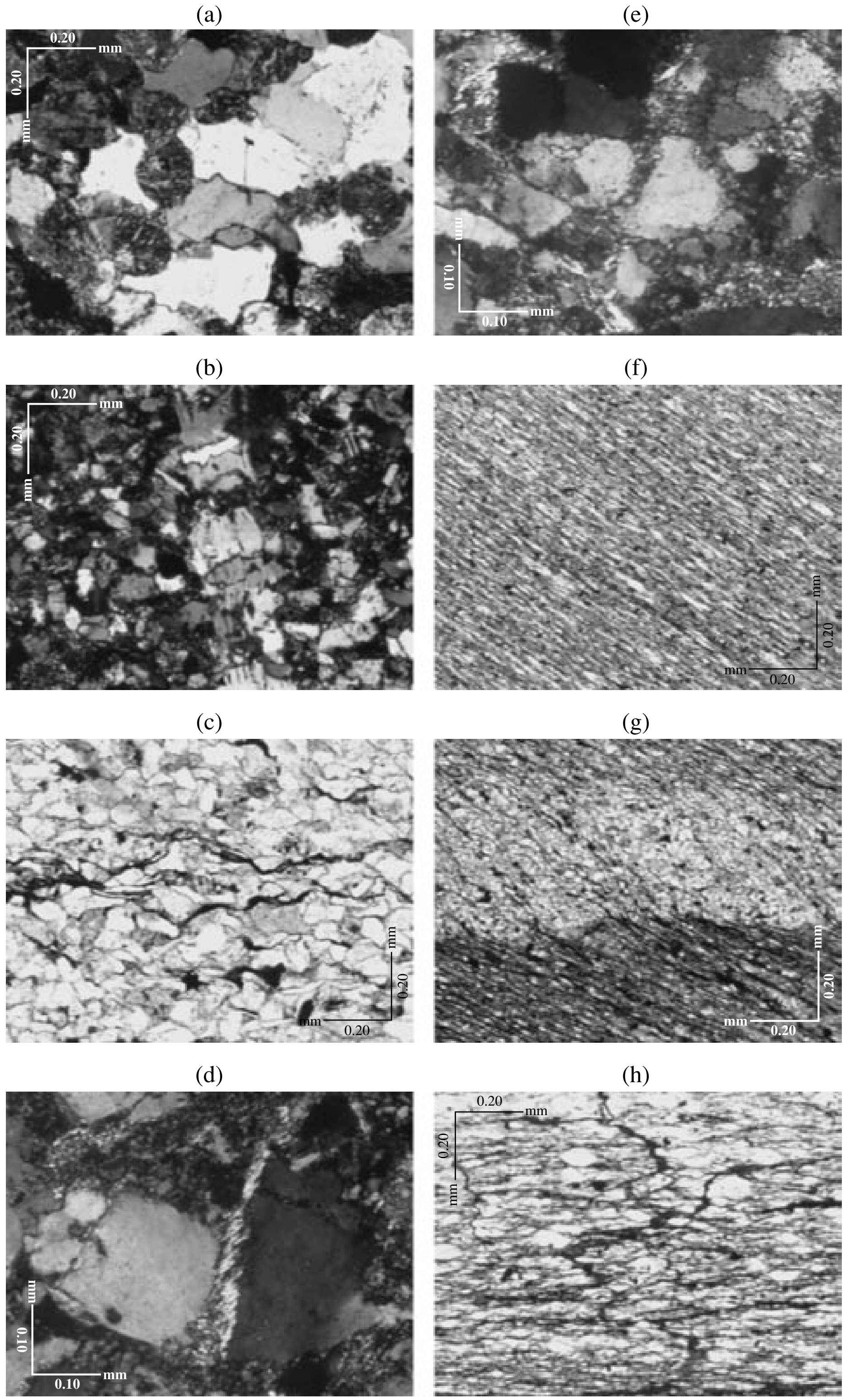

Fig. 2. Photomicrographs of thin sections. (a) Conformation texture, crossed nicols; (b) filtration zone, crossed nicols; (c) dissolution suture surfaces, parallel nicols; (d) differential sliding textures, crossed nicols; (e) recrystallization–granulation blastesis at quartz grain contacts, crossed nicols; (f) fine cleavage in slates, parallel nicols; (g) cleavage–bedding relationship, parallel nicols; (h) Z-shaped deformation of bitumoid (?) veinlet in cleaved rock, parallel nicols.

Thus, we have all grounds to assume that the Jurassic sandy beds were still unlithified by the time of the initiation of folding in the study region. They were lithified under orogenic compressive conditions over a sufficiently long time interval. Hence, the preorogenic subsidence catagenesis of sandy rocks, reported, for example, from rocks of the Verkhoyansk terrigenous complex (Simanovich and Yapaskurt, 2002) was not manifested in the Jurassic rocks of southern Dagestan. In general, transformations of the Jurassic sandy sediments and rocks under conditions of compression and high dynamic activity can be qualified as *orogenic catagenesis* (Luk'yanova, 1995; Simanovich and Yapaskurt, 2002).

Clayey rocks are observed as dark gray, occasionally almost black varieties composed of hydromicas and chlorite with a minor admixture of both dispersed and coalified organic material (often elongate fragments). Beyond the foliation zone, the clayey rocks have massive texture in homogeneous interlayers. If mudstones alternate with silt-rich strata, the rocks have fine-bedded and lenticular-bedded textures. In general, pure clayey interlayers virtually always contain 5–20% of the fine silt admixture.

In terms of the degree and character of lithification, as well as secondary structures, the clayey rocks appreciably differ from the sandy rocks that often remain friable even in deep zones under conditions of subsidence lithogenesis. The clays begin to consolidate immediately after diagenesis marked by the successive squeezing out of free and capillary waters. Porosity of the clayey rocks decreases to mere 5% at a depth of 4000 m (Lomtadze, 1955; Logvinenko, 1968). Therefore, one can suppose that the oldest (early Toarcian) mudstones underwent the maximal consolidation (lithification) during the subsidence of clayey rocks at the predeformation stage of the Jurassic sedimentary basin and the mudstones were significantly lithified by the time of initial folding.

In the northern and northeastern areas (left bank of the Samur River, the Samur Range, and its foothills), the younger (Aalenian) clayey rocks only swell under the influence of the atmospheric precipitation; i.e., they were not completely transformed into mudstones, resulting in the appearance of the so-called "black water" in the left tributaries of the Samur River. The clayey rocks are observed as nonswelling varieties (mudstones) in the field of older (early and late Toarcian) sediments that evidently subsided to deep zones prior to the sedimentary basin deformation. The Toarcian mudstones are characterized by the columnar jointing oriented parallel to fold axes. This peculiar cleavage is likely to be developed during the crushing and sliding of the brittle mudstones between the lithified sandstones layers during the folding. However, it should be noted that this cleavage is never observed in thin sections.

In the foliation zone (Fig. 1), the clayey rocks containing some fine silt admixture are intensely cleaved and the distance between the cleavage surfaces is 0.0*n* mm (Fig. 2f). Macroscopically, they are observed as dark gray or almost black lamellar rocks (slates) that are easily split along the cleavage. Primarily, the clay members were characterized by the heterogeneous grain size composition and the presence of silt lenses and interlayers. Therefore, the foliation (cleavage)/primary bedding angle (usually, 5°–30°) is easily recognized in thin sections (Fig. 2h), suggesting the synorogenic or postorogenic origin of cleavage. The sandy interlayers, which occur in the cleavage zones, are also slightly inclined relative to the cleavage surface, but they are not virtually cleaved. As mentioned above, conform textures in the sandy interlayers are supplemented with the suture deformation surfaces, differential sliding surfaces, and occasional recrystallization–granulation blastesis of quartz.

Thus, the cleavage was superimposed on the already deformed, nearly upright standing terrigenous rock beds. The intense cleavage developed under the impact of lateral compression was restricted to the clayey rocks sandwiched between the more competent (lithified) sandy rock beds. The degree of contraction perpendicular to the cleavage surface can be estimated based on Fig. 2h showing the Z-shaped deformation of a veinlet (probably, bitumoid) that crosscut the rock before the beginning of compression.

CLAY MINERALS AND POSTSEDIMENTARY ZONALITY

We studied 70 samples of <1-μm fraction of mudstones and clayey siltstones from different sectors of the study area. The investigation by the routine procedure on a DRON-2 diffractometer showed that the clay minerals are represented by a mixture of dioctahedral micaceous minerals and trioctahedral chlorite. Figure 3 shows that they contain reflections with *d* equal to approximately 10, 5, and 3.33 Å (micas) and 14, 7, 4.75, and 3.52 Å (chlorites). In order to determine the degree of mixed-layer composition of the micaceous minerals, we applied the following method recommended by Omel'yanenko *et al.* (1982). One determines the first basal reflection of micas recorded in natural and glycerin-saturated samples. Then, the perfection degree of micaceous minerals is determined using the formula $\Delta d = d_{nat} - d_{sat}$. According to Omel'yanenko *et al.*, micaceous minerals with Δd = 0.15 Å should be referred to as sericite. Nearly all micaceous minerals in the studied samples fit this Δd value; i.e., they correspond to sericite. Figure 1 shows localities containing the natural sericite (Δd = 0.0–0.05 Å), sericites with a minor content of expanding layers (Δd = 0.06–0.15 Å), and micaceous minerals with a higher content of the smectite component (Δd > 0.15 Å).

The Kübler index (K.I.) is another important parameter that makes it possible to determine the transformation degree of clayey rocks in the study area. This parameter previously known as crystallinity index

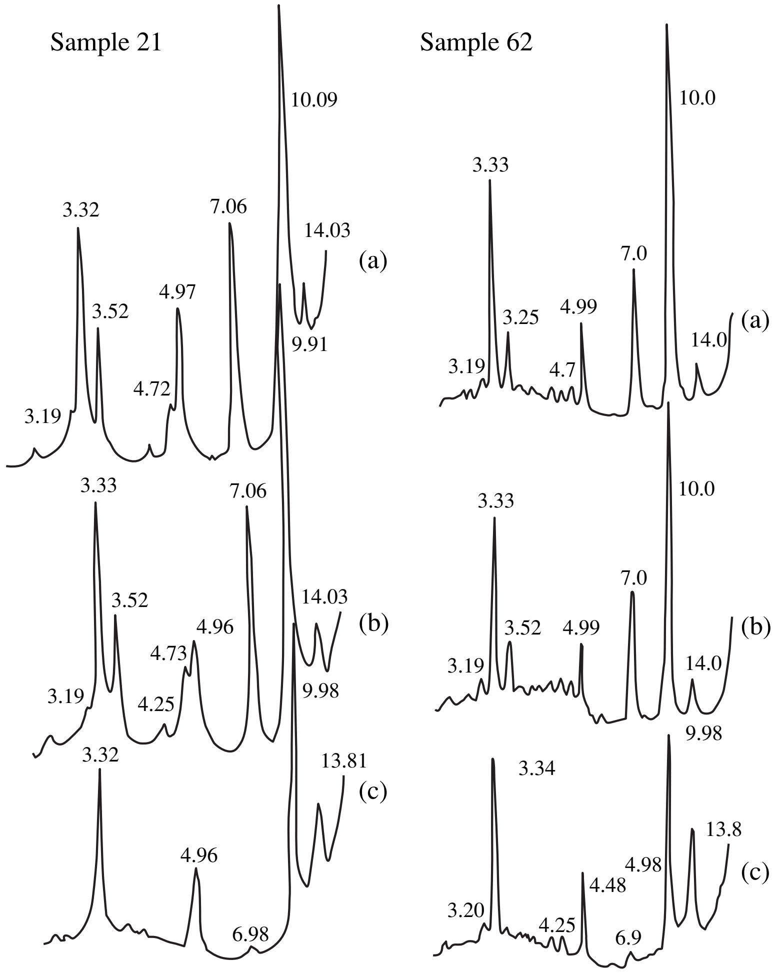

Fig. 3. Typical diffractograms of clayey fractions (<1 μm) from different zones postsedimentary transformation: (sample 21) orogenic catagenesis I (K.I. = 0.82, Δd = 0.18); (sample 62) metagenesis (K.I. = 0.32, Δd = 0). (a) Natural samples; (b) glycerin-saturated; (c) calcined at 550°C.

(C.I.) was widely used for the compilation of the map of postsedimentary zonality (Yang and Hesse, 1991; Hesse and Dalton, 1991; and Warr *et al.*, 1991). Recently, researchers have revealed that this index depends not only on the crystallinity degree of micaceous minerals, but also on several other parameters, such as the average number of layers in coherent scattering zones (i.e., the natural crystallinity of micas), variations in the distribution of the number of layers in the coherent scattering zones, and the average percentage of smectite interlayers in micaceous minerals (Jaboyedoff *et al.*, 2001). Therefore, it has been recommended to discard the term “crystallinity index” and return to the term “Kübler index” (*Report…*, 2002). In practice, the Kübler index turned out to be an extremely useful parameter that allows one to outline postsedimentary alteration zones in both area and cross section.

The Kübler index is determined by measuring the width of 10 Å peak of the micaceous mineral in natural samples at half-hight of the peak (in °Δ2θ). Since the micaceous mineral is represented by sericite with a low content of expanding interlayers in nearly all of the studied samples, the Kübler index in our case is significantly governed by the coherent scattering zones, i.e., the crystallinity degree of micas.

Figure 1 shows the K.I. values and isolines based on these values. The gradation used in this map is based on (Hesse and Dalton, 1991). The map reflects the following zonality: (1) orogenic catagenesis I (K.I. ≥ 0.60); (2) orogenic catagenesis II (0.60 > K.I.> 0.42); and (3) metagenesis (anchimetamorphism (K.I. ≤ 0.42).

It is evident from the map that the postsedimentary zonality matches the dislocation degree of layers and the rock alteration degree intensifies from the north to south. It is worth noting that the cleavage zone fits the zone delineated by isoline of K.I. = 0.42, i.e, the metagenesis (anchimetamorphism) zone. This zone encompasses the Toarcian rock domain (in west) and grades into the upper Aalenian sequences in *en echelon* style (in southeast). The K.I. value increases to 0.6–0.8 further to the south in the cleavage-free Bajocian rocks.

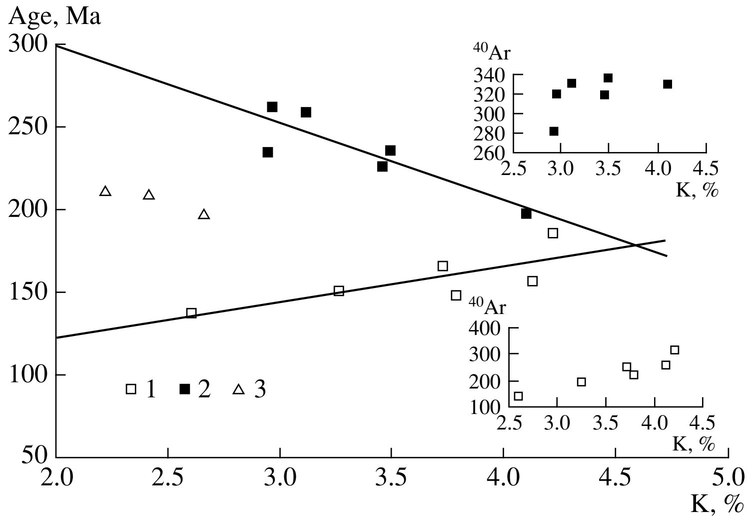

Fig. 4. The calculated K–Ar age vs. K content relationship. (1) Whole-rock samples and fine sample fractions from the metagenesis zone; (2) fine sample fractions from the catagenesis zone; (3) whole-rock samples from catagenesis zone. Insets show dependence of the radioactive Ar content (rel. units) from the K content.

REFLECTION OF STAGES OF POSTSEDIMENTARY LITHOGENESIS IN Rb–Sr AND K–Ar SYSTEMS OF CLAYEY ROCKS

From the point of view of the isotopic geochronology, clayey and other siliciclastic rocks represent primarily inequilibrium systems. Therefore, the isotopic dating of stratigraphic subdivisions based on these rocks for the determination of sedimentation timing is an almost unresolvable issue. Nevertheless, the existing methods of the perception and interpretation of the real results of isotopic datings occasionally allow one to assume the real fixation of a certain age. This is possible if the studied rocks underwent postsedimentary transformations with the respective rearrangement of isotopic systems. If the intensity of such transformations was sufficient and the transformation was short-lived, the transformation age can be fixed in isotopic systems. It is, however, necessary to emphasize that prediction of the possibility of obtaining such dates before the beginning of investigation is impossible. The result, positive or negative, is determined only after the investigation. Interpretation of the data obtained as the reflection of a certain age should be based on the measurement data. The relation of age value with specific geological event is a separate task, the resolution of which needs the collective efforts of experts in different fields. Naturally, the success of investigation largely depends on the choice of the object.

The application of isotopic methods for the study of the timing and factors responsible for the transformation of Jurassic rocks in the Caucasus appears to be promising (Bujakaite *et al.*, 2003). Several circumstances were taken into consideration during the choice of the studied sample collection. Two series of Toarcian–Aalenian rock samples subjected to different degrees of secondary alterations comprised the base of our investigations. One series characterized the metagenesis zone based on the low I.K. value (≤ 0.42) of micaceous minerals and the presence of cleavage in rocks. Another series included the less altered clayey rocks of orogenic catagenesis zone in terrigenous complexes.

The dimension of clayey rocks was an additional circumstance that was taken into consideration. The fine-grained clay minerals are most readily subjected to postsedimentary transformations. Alterations of the coarser fractions more strongly depend on the intensity of transforming factors. Therefore, isotopic measurements were carried out using the fine fraction (<1 μm) and whole-rock samples with the share of fine fraction not exceeding 10%.

Locations of the sampling for isotopic investigation and results of the measurements are shown in Fig. 1 and the table, respectively.

K–Ar measurements. The K–Ar system of clayey rocks is commonly very sensitive to postsedimentary transformations, first of all, owing to the difference in geochemical properties of the parental potassium isotope and its decay product (Ar). The K content in samples versus the calculated K–Ar age plot (Fig. 4) vividly shows the principle distinction of the K–Ar system behavior in rocks altered to different degrees.

The upper line in the plot approximates points corresponding to fine (<1 μm) sample fractions in the catagenesis zone. One can see that the calculated age inversely correlates with the K content in samples. This relationship can easily be explained by the gradual saturation of clay minerals with K, which is typical for the first alteration stage of clayey sediments. Thus, the age partially inherited from rocks of the provenance decreases with increasing maturity of the clayey mate-

Rb–Sr and K–Ar characteristics of Jurassic clayey rocks, Dagestan

Sample no.	Kübler index	ppm		$^{87}Rb/^{86}Sr$, at.	$^{87}Sr/^{86}Sr$, at.	K, %	^{40}Ar, mm^3/g	T (K–Ar), Ma
		Rb	Sr					
Metagenesis zone, whole-rock samples (Figs. 4, 5; upper line)								
56		132.9	157.7	2.441	0.71603	2.60	0.0142	137
58		180.7	77.5	6.762	0.72751	3.78	0.0224	148
61		146.5	82.06	5.173	0.72374	2.93		
62		146.1	81.42	5.200	0.72344	3.20		
63		134.4	122.4	3.180	0.71738	2.81		
Metagenesis zone, fraction <1 μm (Fig. 4, lower line; Fig. 5, lower line)								
56	0.29	212.6	189.1	3.254	0.71754	3.26	0.0197	151
58	0.30	263.3	70.46	10.84	0.73675	4.21	0.0317	186
61	0.30	234.2	77.40	8.775	0.73096	3.72	0.0249	166
62	0.30	248.3	54.77	13.16	0.74263			
63	0.27	234.2	115.8	5.860	0.72330	4.12	0.0260	157
Catagenesis zone, whole-rock samples (Fig. 5, upper line)								
21		95.30	143.0	1.930	0.71447	2.21	0.0190	211
25		134.0	124.8	3.111	0.71803	2.65	0.0212	197
Catagenesis zone, fraction <1 μm (Fig. 5, upper line)								
21	0.82	200.8	178.9	3.250	0.71721	3.10	0.0331	259
25	0.44	220.7	150.8	4.240	0.72073	2.93	0.0282	235
Samples with only K–Ar datings, whole-rock samples								
24	0.70					2.40	0.0204	209
Samples with only K–Ar datings, fraction <1 μm								
24	0.70					2.95	0.0320	262
27	0.51					3.48	0.0336	236
28	0.42					4.09	0.0329	198
31	0.52					3.45	0.0319	226

Notes: The K content was determined in the Chemical Analytical Laboratory of the Geological Institute, Moscow (I.V. Kislova, analyst) with an accuracy of <1%; the Ar content, by the isotopic dilution method with an error of ±2–2.5%. The age was calculated based on the following constants: $^{40}K/K = 1.167 \times 10^{-4}$ mol/mol, $\lambda_e = 0.581 \times 10^{-10}$ yr^{-1}, $\lambda_\beta = 4.962 \times 10^{-10}$ yr^{-1} (Dalrymple, 1979). The determination error was ±1% for $^{87}Rb/^{86}Sr$ and ±0.0002 for $^{87}Sr/^{86}Sr$.

rial. At the same time, the table and upper inset in Fig. 4 show that the radiogenic Ar content remains virtually constant in all samples. One can hardly suppose that all radiogenic Ar was retained in the clayey fraction since the accumulation of sediments. The accumulation of K in the clayey fraction is related to its structural rearrangement often accompanied by the partial loss of Ar. The loss is possibly compensated in the samples by Ar accumulated after the termination of catagenesis. Therefore, the equality of contents should be considered an incidental phenomenon. Samples with the maximal K content are closest to the terminal stage of catagenesis. In our case, Sample 28 with the K content of 4% and the calculated age of approximately 200 Ma fits the condition formulated above. The catagenetic processes probably terminated slightly later than this dating.

The lower line in Fig. 4 approximates data points of significantly less altered rock samples from the metagenesis zone. They show a positive correlation of the K content with the calculated K–Ar age and, hence, the dependence of the K–Ar system on the metamorphic event that was responsible for the loss of radiogenic Ar in the samples. This is seen well in the lower inset of Fig. 4. The loss inversely correlated with the perfection degree of mineral structure. In general, the K-rich clay minerals have a more perfect structure. Therefore, the age calculated for Sample 58 (186 Ma) with the highest K content should best match the timing of the initial metagenesis and the termination of transformations in the catagenesis zone.

We believe that the intersection of two approximating lines should correspond to the termination of catagenetic transformations in the northern zone and the

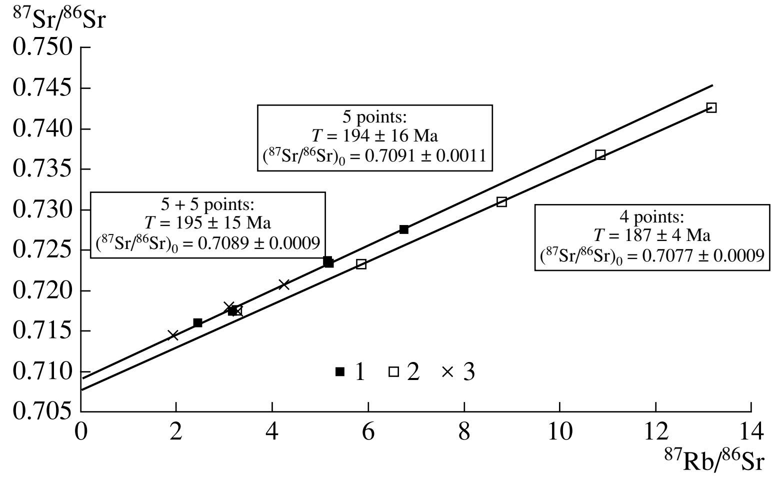

Fig. 5. Rb–Sr isochron plots based on whole-rock samples and fine rock fractions from metagenesis zone. Boxes show results of the calculation of direct relations based on different numbers of points. (1) Whole-rock samples from metagenesis zone; (2) fine fractions from metagenesis zone; (3) whole-rock samples and fine fractions from catagenesis zone.

initiation of metagenetic alterations in the southern zone. This episode corresponds to 180 Ma with a significant and indeterminate error.

Figure 4 also shows the positions of three data points of whole-rock samples from the catagenesis zone. The younger age of whole-rock samples, relative to the calculated age of their fine fractions, has no trivial explanations. At the same time, it is hardly expedient to propose complex models for the specific transformation of rock lithology based on a low-representative material.

Rb–Sr measurements. All results of the Rb–Sr measurements are given in the table and shown in the isochron plot of Fig. 5. Data points corresponding to five samples of the fine fraction of rocks in the metagenesis zone define a slope corresponding to 178 ± 9 Ma and the initial Sr isotope ratio $(^{87}Sr/^{86}Sr)_0$ = 0.7090 ± 0.001 (MSWD = 2.9).One of the five points with the lowest Rb/Sr ratio (Sample 56) is located rather far away from the straight line. The remaining four points almost perfectly lie on the line corresponding to T = 187 ± 4 Ma and $(^{87}Sr/^{86}Sr)_0$ = 0.7077 ± 0.0005 (MSWD = 0.2; Fig. 5, lower line). In our opinion, these data are more significant than the data calculated from five samples. The elimination of data point of Sample 56 from the calculations is justified by two circumstances. First, Sample 56 is marked by the lowest Rb/Sr ratio due to the very low Sr content (table) and, hence, a specific mineral composition. Second, this point lies well on the line of five data points of whole-rock samples from the metagenesis zone (Fig. 5, upper line). The calculation based on five points yields an age of 194 ± 16 Ma and $(^{87}Sr/^{86}Sr)_0$ = 0.7091 ± 0.0011 (MSWD = 3). Thus, both lines have virtually similar slopes. Therefore, if we take into consideration the scatter of data, the discrepancy of age values calculated from the whole-rock samples and fine fractions are statistically insignificant. The higher $(^{87}Sr/^{86}Sr)_0$ value in the whole-rock samples probably reflect a certain inheritance of the radiogenic Sr from the primary rocks.

Data points of the fine fractions and two whole-rock samples (table, 21 and 25) taken in the catagenesis zone also lie on this line. The fine fractions of these samples have basically different values of the Kübler index (0.44 and 0.82). Nevertheless, their inclusion in the isochron relationship calculation based on five whole-rock samples of the metagenesis zone does not virtually change the result, indicating the relative significance of this parameter for the estimation of the Rb–Sr system stability.

Since slopes of both lines are similar, we also calculated the general linear correlation for all 14 data points and obtained the minimal age of 176 ± 10 Ma and $(^{87}Sr/^{86}Sr)_0$ = 0.7098.

Thus, both Rb–Sr and K–Ar results make it possible to estimate the timing of the main stage of metagenetic and catagenetic transformation of clayey rocks in the study region. One can conditionally accept that the maximal transformation took place in the interval of 180–190 Ma. The more detailed subdivision of transformation stages outlined by results of the isotopic dating needs the inclusion of additional analytical data.

Our estimates of the timing of rock transformations are slightly higher than the timing of subduction in the study region corresponding to the late Aalenian–early Bajocian (Lomize and Panov, 2002), i.e., approximately 175 Ma (Harland *et al.*, 1990). We do not have sufficient grounds to explain the discrepancy noted above.

GEODYNAMICS AND POSTSEDIMENTARY LITHOGENESIS

Researchers have proposed different concepts on the tectonic evolution and origin of folding in the Jurassic complexes of the Greater Caucasus (Sholpo *et al.*, 1993 and others). We suppose that the isotopic results mentioned above are consistent with the geodynamic model proposed by Lomize and Panov (2002) who examined the succession of geological events during the transformation of the Caucasian passive continental margin into the active phase. They revealed a universal hiatus at the late Aalenian–early Bajocian related to the general uplift of the continental margin, disappearance of riftogenic structures, and compression-induced fold-and-fault deformations. These deformations indicate that the continental margin uplift at the late Aalenian–early Bajocian was caused by the compression oriented parallel to its strike. The stratigraphic interval encompassing the *Graphoceras concavum* and *Sonninia sowerby* faunal zones defines the timing of stress appearance and disappearance corresponding to 178 and 175 Ma of the time scale proposed in (Gradstein *et al.*, 1994).

These events were related to the onset of subduction or, more precisely, its amagmatic phase. According to (Lomize and Panov, 2002), the subducting oceanic lithosphere, which leaned against the edge of the opposite plate, could be pushed down and subducted only under the impact of a strong horizontal compression that accompanied the subduction during the entire amagmatic stage (approximately 3 Ma).

Evidently, postsedimentary lithogenetic processes in the Toarcian–Aalenian rock complexes of southern Dagestan were developed in the course of the dramatic tectonic events described above and significantly promoted by these events.

The postsedimentary history of the Toarcian–Aalenian rocks in southern Dagestan can be divided into three periods: (1) subsidence lithogenesis in the course of the progressive rifting and basin subsidence; (2) orogenic (synorogenic) lithogenesis (compression); and (3) period after the termination of the early Alpine orogeny.

Processes of diagenesis during the accumulation and subsidence of terrigenous sediments are scrutinized in (Gavrilov, 1982, 2002; and others). These transformations are expressed in the intense nodule formation (siderite), the redistribution of elements between layers with different grain size compositions, and the migration of bicarbonates of Fe, Mn, and Ca from the clayey sediments to the sandy ones.

Evidently, clayey siltstones and other clayey rocks appreciably condensed and lithified in the course of burial to a depth of 5–7 km. The clays are transformed into the dense nonswelling mudstones because of the loss of water and decrease of porosity, which equals to only 1–3% at a depth of 7–10 km (Logvinenko, 1968). Field observations support these assumptions. The upper Aalenian clayey rocks in the northern part of the study region underwent the minimal subsidence and are capable to swell; i.e., they were not completely transformed into mudstones. In contrast, the lower and upper Toarcian clayey rocks are observed as very compact mudstones with the typical columnar jointing that is developed during the brittle (synorogenic) deformations of rocks.

The sandstones and coarse siltstones represented by the relatively well-sorted and rewashed varieties generally lack the clayey and other cements. Therefore, they remained in the unlithified (friable) state during the lithogenesis. Otherwise, the intense development of synorogenic dissolution structures mainly under the impact of stress (lateral pressure) would have been impossible. Processes of orogenic (more precisely, synorogenic) lithogenesis promoted the lithification of sandy rocks owing to the intense development of conform textures (Fig. 2a). In this process, silica and other dissolved components were completely removed from the transforming sequences; i.e., they were not redeposited in pore spaces. As was emphasized above, this circumstance is related to the variation in the fluid regime of the postsedimentary lithogenesis.

The contribution of orogenic catagenesis to the transformation of clayey rocks is not sufficiently clear. As mentioned above, micaceous minerals in the clayey rocks are represented by sericite with a relatively low Kübler index. Micaceous minerals with the minimal perfection are found in the upper Aalenian swelling clayey rocks of the northern area. One thing is certain: the transformation degree of the clayey rocks within the Toarcian–Aalenian domain increases from the northeast to southwest. The metagenesis zone with K.I. < 0.42 fits the cleavage zone. This phenomenon is evidently related to stress.

Metagenetic transformations of clay minerals are superimposed on the older catagenetic alterations of the clayey material. This statement is supported by opposite trends of the K–Ar datings for rocks of the catagenesis and metagenesis zones (Fig. 4). This figure shows that metagenesis of the clayey rocks and the syngenetic cleavage developed after the termination of their catagenesis approximately 180 Ma ago.

Processes of the orogenic catagenesis and metagenesis of the Toarcian–Aalenian sandy and clayey rocks probably evolved in the discrete-continuous manner under conditions of strong compression. This process was accompanied by the transverse contraction of the basin, intensification and complication of the fold-and-fault dislocations, and intensification of the lithification of rocks (primarily, sandstones). The lithification of the Toarcian–Aalenian sediments gradually produced a rigid block, the frontal part of which was cleaved as result of collision with the Transcaucasian Massif. According to our estimates, this event took place 180–190 Ma ago, probably, at the end of the amagmatic phase of subduction. We assume that the age of 180 Ma is the most precise indicator of the termination of events related to the first (amagmatic) phase of subduction.

Probably, collisions at the contact of the Transcaucasian Massif with the passive continental margin produced the orogenic zone corresponding in space to the Lateral Range (within the territory of southern Dagestan). Terrigenous rocks within this zone were completely lithified. During the Bajocian, this orogen contacted with a southern marine basin that accumulated erosion products of the orogen.

CONCLUSIONS

Results of our investigations have revealed the mechanism of synorogenic lithification of sandy rocks (orogenic catagenesis). The clayey rocks underwent subsidence lithogenesis during the preorogenic period. Isotope data indicate that metagenetic transformations of the clayey rocks in cleavage zones are superimposed on the subsidence lithogenesis. Zones of postsedimentary lithogenesis have been mapped using the Kübler index. Isotope data indicate that the age of 180 Ma is the most precise estimate of the termination of folding and postsedimentary lithogenesis related to the first (amagmatic) subduction phase. We believe that collisions at the contact of the Transcaucasian Massif with the passive continental margin were responsible for the formation of the early Alpine orogen (Lateral Range).

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REFERENCES

Bujakaite, M.I., Gavrilov, Yu.O., Gertsev, D.O., Golovin, D.I., and Kushcheva, Yu.V., The K–Ar and Rb–Sr Isotopic Systems in Rocks from the Jurassic Terrigenous Complex of the Greater Caucasus, *Litol. Polezn. Iskop.*, vol. 38, no. 6, pp. 613–621 [Lithol. Miner. Resour., 2003, vol. 36, no. 6, pp. 522–529].

Dalrymple, G.B., Critical Tables for Conversion of K-Ar Ages from Old to New Constants, *Geology*, 1979, vol. 7, no. 11, pp. 558–560.

Galkin, V.A., Stress Analysis in the Petrostructural Study of Rocks, *Vestn. Mosk. Gos. Univ., Ser. 4: Geol.*, 1992, no. 4, pp. 35–40.

Gavrilov, Yu.O., Authigenic Jurassic Rock Formations in Western Dagestan, *Litol. Polezn. Iskop.*, 1982, vol. 17, no. 2, pp. 32–46.

Gavrilov, Yu.O., Dynamics of the Formation of the Jurassic Terrigenous Complex in the Greater Caucasus: Sedimentology, Geochemistry, and Postsedimentary Transformations, *DSc (Geol.–Miner.) Dissertation*, Moscow: Mos. Gos. Univ., 2002.

Gradstein, F.M., Agterberg, F.P., and Ogg, J.G., *et al.*, Mesozoic Time Scale, *J. Geophys. Res.*, 1994, vol. 99, no. B12, pp. 24051–24074.

Harland, W.B., Armstrong, R.L., Cox, A.V., *et al.*, *A Geologic Time Scale 1989*, Cambridge: Univ. Ontario Press, 1990.

Hesse, R. and Dalton, E., Diagenetic and Low-Grade Metamorphic Terranes of Gaspe Peninsula Related to the Geological Structure the Taconian and Acadian Orogenic Belts, Quebec Appalachians, *J. Metamorph. Geol.*, 1991, vol. 9, no. 6, pp. 775–790.

Jaboyedoff, M., Bussy, F., Kübler, B., and Thelin, Ph., Illite "Crystallinity" Revisited, *Clay Clay Miner.*, 2001, vol. 49, no. 2, pp. 156–167.

Kopeliovich, A.V., Epigenesis of Ancient Strata in the Southwestern Russian Platform, *Tr. Geol. Inst. Akad. Nauk SSSR*, Moscow: Nauka, 1965, issue 121.

Logvinenko, N.V., *Postdiageneticheskie izmeneniya osadochnykh porod* (Postdiagenetic Alterations of Sedimentary Rocks), Leningrad: Nauka, 1968.

Lomize, M.G. and Panov, D.I., Geodynamics of the Tethyan by the Beginning of Emplacement of the Bajocian Porphyritic Series (Amagmatic Subduction Phase), *Tr. Geol. Inst. Akad. Nauk Gruzii. Nov. ser.* 2002, issue 117, pp. 358–369.

Lomtadze, V.D., Formation Stages of Clay Rock Properties during Their Lithification, *Dokl. Akad. Nauk SSSR*, 1955, vol. 102, no. 4, pp. 819–822.

Luk'yanova, V.T., *Katagenez v orogennykh oblastyakh* (Katagenesis in Orogenic Areas), Moscow: KMK Ltd, 1995.

Marakushev, A.A., *Petrogenezis* (Petrogenesis), Moscow: Nedra, 1988.

Omel'yanenko, B.I., Volovikova, I.M., Drits, V.A., *et al.*, The Meaning of the Notion of Sericite, *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1982, no. 5, pp. 69–87.

Panov, D.I., The Northern Margin of the Tethys during the Jurassic and Cretaceous (Caucasus Segment), Abstracts of Papers, *7-ya mezhdunarodnaya konferentsiya po tektonike plit im. L.P. Zonenshaina* (7th Int. Conf. on Plate Tectonics in Memory of L.P. Zonenshain,), Moscow: Nauchnyi Mir, 2001, pp. 270–272.

Report of the Association International pour l'Etude des Argiles (AIPEA) Nomenclature Committee for 2001: Order, Disorder and Crystallinity in Phyllosilicates and the Use of the "Crystallinity Index", *Clay Clay Miner.*, 2002, vol. 50, no. 3, pp. 406–409.

Sholpo, V.N., *Tipy i usloviya formirovaniya skladchatosti Slantsevogo Dagestana* (Types and Formation Conditions of Folding in Slates of Dagestan), Moscow: Nauka, 1964.

Sholpo, V.N., Rogozhin, E.A., and Goncharov, M.A., *Skladchatost' Bol'shogo Kavkaza* (Folding in the Greater Caucasus), Moscow: Nauka, 1993.

Shutov, V.D., Classification of Sands, *Litol. Polezn. Iskop.*, 1967, vol. 2, no. 5, pp. 86–103.

Simanovich, I.M., Quartz in Sandy Rocks, *Tr. Geol. Inst. Akad. Nauk SSSR*, 1991, Moscow: Nauka, issue 314.

Simanovich, I.M. and Yapaskurt, O.V., Geodynamic Types of Postsedimentary Lithogenesis, *Vestn. Mosk. Univ., Ser.4: Geol.*, 2002, no. 6, pp. 20–31.

Warr, L.N., Primmer, T.J., and Robinson, D., Variscan Very Low-Grade Metamorphism in South-West England: A Diastermal and Thrust-Related Origin, *J. Metamorph. Geol.*, 1991, vol. 9, no. 6, pp. 751–764.

Yang, C. and Hesse, R., Clay Minerals as Indicators of Diagenetic and Anchimetamorphic Grade in Over-Thrust Belt External Domain of Southern Canadian Appalachians, *Clay Miner.*, 1991, vol. 26, no. 2, pp. 211–231.