

Postsedimentary Lithogenesis of Terrigenous Complexes in Fold Systems: Rock Fabric and Cleavage

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Abstract—The simplest intergranular cleavage and foliation of clayey rocks are typical of postsedimentary lithogenesis of folded terrigenous complexes at passive continental margins if the rocks do not contain metamorphic minerals. Cleavage zones coincide with metagenesis (anchimetamorphism) zones. Cleavage is often superimposed on the rock fabric formed in the process of subsidence lithogenesis (e.g., in the Verkhoyansk terrigenous complex). In other cases, cleavage in terrigenous rocks is regarded as the final stage of the development of their microfabric formation related to folding (e.g., Jurassic sequence of southern Dagestan).

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INTRODUCTION

The postsedimentary lithogenesis is a succession of processes that start with the first symptoms of diagenesis and terminate with the appearance of metamorphic minerals and the establishment of mineral equilibrium at elevated temperatures and pressures.

According to the concept developed in the second half of the 20th century (Kossovskaya and Shutov, 1955, 1971; Logvinenko, 1968), the grade of sedimentary rock transformation gradually increases from diagenesis to metamorphism: diagenesis → initial catagenesis → deep catagenesis → metagenesis → metamorphism. Under the influence of ideas of Pustovalov (1956), Kossovskaya and Shutov (1976) concluded that the stage of regional epigenesis (catagenesis) prepares nonequilibrium systems of sedimentary rocks for the gradual transition to the initial metamorphism. Logvinenko (1968) and many other researchers held this viewpoint. Based on the study of the fold system of the western Verkhoyansk region, Kossovskaya and Shutov (1976) postulated a gradual elimination of specific petrographic features of the Verkhoyansk terrigenous complex in the course of its subsidence to deep zones, successive transition to biotite and staurolite grade of metamorphism.

Dunoyer de Segonzac (1970) and Weaver (1984) also deemed that the grade of sedimentary rock transformation is directly related to the depth of their subsidence. This is consistent with the concept of subsidence metamorphism developed by Winkler (1967), Vernon (1976), and other petrologists.

On the basis of original data and analysis of literature, we have recently developed a principally different concept (Simanovich and Yapaskurt, 2002; Simanovich et al., 2004): the character of postsedimentary lithogen-

esis of terrigenous complexes is mainly governed by the type of parental sedimentary basins (SB) and diverse manifestations of their subsequent geodynamic evolution (Table 1). The postsedimentary transformations of undeformed SB (subsidence lithogenesis) and deformed SB (orogenic catagenesis and metagenesis) differ not only in the presence or absence of strains, but also in variations of the fluid regime. As was stated by Marakushev (1988), the lithostatic and fluid pressures in undeformed SB are approximately equal. This situation hampers metamorphic reactions of dehydration and decarbonation. The development of tectonic dislocations creates a specific (metamorphogenic) regime of sedimentary rock transformation, because the lithostatic pressure becomes much greater than the fluid pressure in this case.

ROCK FABRIC FORMATION AND LITHIFICATION IN THE PROCESS OF SUBSIDENCE LITHOGENESIS

Since this paper is aimed at elucidation of the genesis of fabric and microfabric of terrigenous rocks in fold systems, it is necessary to unravel the processes that predated the folding and deformation of SB, because they were mainly responsible for the subsequent transformations under the influence of stress.

The rate of lithification of terrigenous sediments is largely controlled by their grain size. Kopeliovich (1965) suggested that the clayey fluid-plastic sediments rather rapidly grade into the semisolid state at a depth of a few meters. He proposed to designate this transition as a boundary between diagenesis and epigenesis (catagenesis). The onset of lithification of the overlying clayey sediments was proposed as this boundary for loose sandy sediments. As overlying sed-

Table 1. Geodynamic types of postsedimentary lithogenesis

SB types	Stages of postsedimentary lithogenesis		Source
	undeformed SB	deformed SB	
Ancient and young platforms	Diagenesis ↓ Catagenesis		Kopeliovich, 1965 <i>Osadochnye basseiny ...</i> , 2004 Simanovich, 1978, 2000 Simanovich and Yapaskurt, 2002
Rifts and aulacogens	Diagenesis ↓ Catagenesis ↓ Subsidence (?) metamorphism	Metagenesis (anchimetamorphism)	Logvinenko, 1968 Simanovich, 2000 Reverdatto et al., 1992, 1995
Passive continental margins	Diagenesis ↓ Catagenesis	Orogenic catagenesis ↓ Metagenesis (anchimetamorphism) ↓ Metamorphism	Luk'yanova, 1995 Simanovich and Yapaskurt, 2002 Simanovich et al., 2004 Kossovskaya and Shutov, 1955, 1976 Simanovich and Yapaskurt, 2002 Simanovich et al., 2004 Yapaskurt, 1999 Korikovskii, 1995
Active continental margins	Diagenesis ↓ Catagenesis (zeolite facies of metamorphism)	Accretionary diagenesis ↓ Glaucophane-type metamorphism	Chamov and Kurnosov, 2001 Coombs, 1960

Table 2. Porosity of rocks and their behavior in water, modified after (Logvinenko, 1968)

Rock	Porosity, %	Behavior in water	Thickness of overlying sediments
Clay from alluvium of the Volga River	50	Readily soaks in water	A few meters
Carboniferous clay, Moscow region	37	Soaks in water	100–200 m
Middle Carboniferous mudstone from the Donets Basin	10–12	Hardly soaks in water	>4 km
Middle Carboniferous mudstone from the Donets Basin	1–3	Does not soak in water	>7 km

iments are deposited, porosity of clayey rocks markedly decreases and the rate of their lithification increases (Table 2).

Thus, subsidence lithogenesis beginning from the depth of a few meters down to 7–10 km results in the progressive and directed compaction of clayey rocks and their transformation into unsoaking mudstones.

If sandy rocks are not cemented with carbonate, clayey, gypsum, and other materials at the early stage of transformation, they behave in a quite different manner under these conditions. Kopeliovich (1965) pointed out that loose sand beds occur in Tertiary, Mesozoic, and, occasionally, even Paleozoic sections of the platform cover. Loose sands were also reported from boreholes

that penetrated the Riphean sequences (V.V. Kostyleva, private communication).

However, subsidence of sands to a relatively shallow depth (a few kilometers) leads more often to their transformation into the hard sandstone as a result of the development of pressure solution (hereafter, dissolution) structures and the consequent regeneration of clastic quartz grains (Heald, 1955, 1956; Lerbekmo and Platt, 1962; Kopeliovich, 1965; Simanovich, 1966, 1978; Logvinenko, 1968; Yapaskurt, 1999). Kopeliovich described three (conformal, microstylolitic, and incorporation) types of dissolution structure related to lithostatic pressure, which gets many times intensified at the contact between clastic grains (Fig. 1a). Thereby, a considerable amount of silica and other components is transferred to solution. Since the volume of clastic grains is proportional to the cube of their diameter, one can approximately estimate the volume of dissolved material (Kopeliovich, 1965). In particular, the decrease of particle diameter by 5, 10, and 20% corresponds to solution of 14.2, 27.1, and 48.8%, respectively, of their initial volume.

Under approximately equal lithostatic and fluid pressures, a concentration gradient of SiO_2 and other components appears between the strained grain solution domains and the least pressure zones, i.e., in pores, where the substance is precipitated. In quartzose, oligomictic, and arkosic sandy rocks, this substance is most often represented by quartz that regenerates the clastic grains. In polymictic sandy rocks (graywackes), which commonly contain up to 50% of clastic quartz, regeneration of clastic grains is developed to a much lesser extent (Yapaskurt, 1992). As a result, the sandy rocks are lithified at the stage of deep catagenesis.

Despite the prolonged study of dissolution structures and the conjugated regeneration of clastic grains, many aspects of this problem remain unclear. For example, it is evident that the abundance of micaceous detritus in sandy rocks obviously intensifies the solution of clastic quartz and other grains, probably, owing to the development of alkaline medium in the process of catagenetic breakdown of biotite. A complex, festoon-shaped corrosion of clastic grains with their two-fold or even stronger contraction may often be seen at contacts with decomposed mica grains. Such relationships are often called gravitational corrosion structures.

Interbeds with abundant micaceous detritus, which are distinguished by intense development of dissolution structures, may alternate with thin interbeds consisting of well-sorted and washed clastic quartz grains. These interbeds are devoid of dissolution structures and cemented by authigenic quartz formed as a result of the removal of silica from over- and underlying interbeds, where solution of quartz is provided by catalytic effect of the decomposed micaceous detritus (Simanovich, 1966; Simanovich and Kostyleva, 1994).

It is noteworthy that the dissolution structure is formed not gradually with subsidence but virtually

instantaneously (in geologic sense) at a certain stage of catagenesis. Otherwise, it would be difficult to explain the retention of high alkalinity and potassium gradient over a long time (Simanovich and Kostyleva, 1994; Yapaskurt, 1999). It remains unclear what a triggering mechanism leads to the spontaneous pressure solution in the course of subsidence metamorphism and why the ancient sand buried at a considerable depth remains friable.

Long microstylolitic sutures, which crosscut the older structures, are occasionally observed in conformal-regeneration sandstones (Simanovich, 1966). In some sense, they are similar to cleavage zones, because they contain a significant amount of insoluble residue (organic matter, clays, and ore minerals).

Thus, structurization and lithification of clayey and sandy rocks during subsidence lithogenesis differ both in kinetic of the process and final results that govern the subsequent behavior of rocks during dislocations in the SB.

CLEAVAGE IN UNMETAMORPHOSED TERRIGENOUS ROCKS

Many definitions and classifications of cleavage are known. In classic monographs (Belousov, 1986; Azhgirei, 1966), cleavage is classified on the basis of genetic attributes and attributed to folding and faulting. Kirillova (1962) suggested that adsorption of water during the lateral compression of clay minerals (smectites) leads to their swelling and splitting into thin sheets. Goncharov (1988) also believed that the cleavage formation is related to advection in the process of folding. He defined flow cleavage (axial-plane folding), fracture cleavage, and shear cleavage, and so on. Based on morphological features, fan-shaped, reverse fan-shaped, plicated, and other types of cleavage were recognized (Gusev, 1979).

In later publications, the preference was given to morphological attributes of cleavage as not only microstructures but also structures (hereafter, fabric) of deformed rocks. Galkin (1988) distinguishes the following types of cleavage in unmetamorphosed and slightly metamorphosed rocks:

(1) **Foliation** is a rock fabric manifested as parallel arrangement of flat or flattened grains (mica, chlorite, talc, and others) in rocks. Foliation (or schistosity) is developed in metamorphic rocks (phyllites and crystalline schists). According to Talitskii (1992), fine cleavage in mudstones with flattened clay minerals may also be referred to as foliation.

(2) **Intergranular cleavage.** Aggregates of clay minerals, micas, ore matter, and fine-dispersed organic matter make up threadlike near-parallel zones between clastic grains. These zones may be sporadic or very frequent. They occur along boundaries of almost each clastic grain and often merge with one another to split the rock into small lenses. Single clastic grains or their

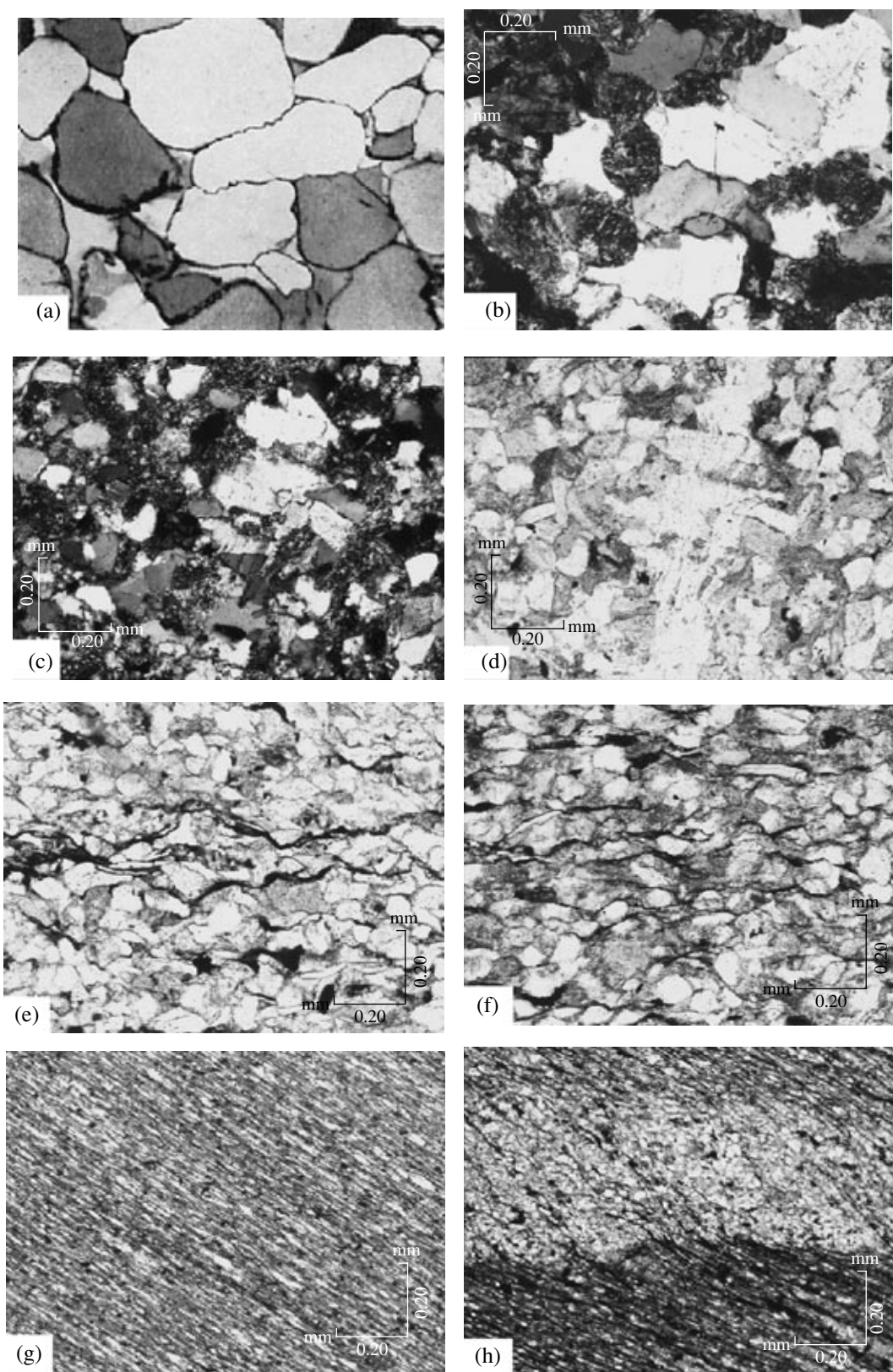


Fig. 1. Photomicrographs of thin sections. (a) Conformal-regeneration fabric of the Shoksha quartzitic sandstone, Karelia, crossed polars, magn. 60; (b–h) southern Dagestan: (b) conformal fabric, crossed polars; (c) filtration zone, crossed polars; (d) the same, plane light; (e) sutures of solution surface, plane light; (f) intergranular cleavage, plane light; (g) foliation (planar cleavage), plane light; (h) relationships between cleavage and bedding, plane light.

groups separated by cleavage zones are called *microlithons* (Galkin, 1988).

(3) **Plicative cleavage** developed in rocks with foliation or intergranular cleavage is a superimposed type. It may be seen under a microscope that planes of the older microfabric are deformed into microfolds or microflexures.

Galkin (1988) regards the formation of intergranular cleavage as a deformation realized under conditions of lateral compression (stress) by means of selective dissolution, diffusion removal of certain components from rock, and residual concentration of other (insoluble) components within cleavage zones. The distance between cleavage zones decreases with the increase of the clay material content, and the cleavage zones become flatter.

The development of cleavage is not virtually accompanied by change of metamorphic grade; i.e., this is an isofacies process (Luk'yanov and Luk'yanova, 1987; Luk'yanova and Luk'yanov, 1987). However, cleavage zones often include metamorphic minerals, e.g., phengite and other micas of polytype 2M. Galkin (1988) has shown that compositions of chlorite in microlithons and cleavage zones are appreciably different.

Structures of rear zones of clastic grains in cleaved rocks are of great interest. Such terms as *pressure shadow*, *stress shadow*, *extension cavity*, and *overgrowth apron* are often used in the literature (Kossovskaya and Shutov, 1955; Simanovich, 1978; Galkin, 1988; Yapaskurt, 1999). The pressure shadows mark the least strained and undeformed zones. The inclusion/cement boundary is often detached, and the newly formed cavity is filled with a material transported by solution (overgrowth apron composed of quartz, calcite, chlorite, and other minerals).

Autonomous development of cleavage and folding (Galkin, 1988) is a very important conclusion. If the stability of strain field orientation is retained over a certain time span of the horizontal contraction of beds before, during, or after the folding, conditions favorable for the formation of cleavage are created in rocks. Hence, folding is not responsible for the cleavage formation. Moreover, even the tight (e.g., keel-shaped) folding is not accompanied by cleavage in many cases.

According to Talitskii (1992), however, the popular notion of cleavage formation after folding contradicts the factual data. Folds without cleavage are common in foldbelts. Conversely, the development of cleavage without the coeval folding is also known. Talitskii believes that cleavage is formed prior to bending at the stage of longitudinal contraction of beds. Thus, the cleavage generates the essential contrasting strains in beds and creates conditions favorable for their bending. In the course of bending, strain passes from the level of grains to the level of beds, and cleavage is rearranged (the planes are turned into direct and reverse fans). Cleavage superimposed on folds can be formed at the

stage of general flattening. Such cleavage is commonly developed parallel to axial planes of folds.

According to (Galkin, 1988), convincing evidence for the formation of cleavage in unconsolidated sediments is lacking, and the cleavage fabric probably appears only in consolidated lithified rocks.

CONFORMAL STRUCTURES OF OROGENIC CATAGENESIS AND CLEAVAGE IN FOLDED TERRIGENOUS SEQUENCES

In order to correctly interpret the formation sequence of microstructures and microtextures (hereafter, microfabric) in terrigenous rock complexes of fold systems, we should elucidate the processes and factors of their transformation before the deformation stage, and above all, to determine the rate and timing of lithification during subsidence lithogenesis. As was shown above, the rate and degree of lithification of clayey and sandy rocks at the stage of deep catagenesis may be markedly different. The character of lithification of sandy interbeds is probably controlled by their facies type.

According to Gavrilov (2005), the Lower–Middle Jurassic terrigenous complex in the northeastern Caucasus was formed by a large drainage system that transported huge masses of terrigenous material to the SB. Sedimentary sections in delta show a distinct cyclic character with alternation of sandstone, siltstone, and mudstone in various proportions. The sandy rocks (lithic quartz graywackes) are well sorted and washed (they are almost completely devoid of clayey cement). This circumstance was probably the main cause of long-term retention of loose sands in the course of subsidence lithogenesis. Lithification of sandy rocks (orogenic catagenesis) occurred later at the deformational stage of SB evolution.

The sequence of postsedimentary transformation of microfabric of terrigenous rocks is described in the recent paper devoted to postsedimentary lithogenesis of Jurassic terrigenous complexes in southern Dagestan (Simanovich et al., 2004). Lomize and Panov (2002) established the sequence of transformation of the Caucasus passive continental margin into the active margin. The regional hiatus in the late Aalenian–early Bajocian was a response to the general uplift of the continental margin, closure of rifts, and development of transverse folding and faulting. According to these authors, the stratigraphic break, which defines the interval between initial and final phases of compression, is estimated at 178–175 Ma ago. Processes of compression were related to the onset of subduction (more precisely, its amagmatic phase) that lasted for ~3 Ma.

The postsedimentary history of Toarcian–Aalenian terrigenous sequences of the southern Dagestan is divided into three periods: (1) subsidence lithogenesis in the course of progressive rifting and deepening of the



Fig. 2. Cleavage (the upper Aalenian, southern Dagestan).



Fig. 3. Sandstone bed within the cleaved sequence of slates, southern Dagestan.

basin, (2) orogenic (synfolding) lithogenesis (compression), and (3) period after the early Alpine orogeny.

In the course of subsidence to a depth of 5–7 km, the clayey rocks (including the clayey siltstones) were compacted, lithified, and transformed into unsoaking mudstones. Indeed, the lower and upper Toarcian clayey rocks occur as very compact mudstones with pencil jointing oriented along fold axes. This type of jointing can be produced as a result of synfolding brittle deformations of the compact mudstones that are sandwiched between the sandy rock beds subjected to bending. The upper Aalenian clayey rocks, which apparently subsided to a much shallower depth, retained their ability to partial soaking.

Dissolution (conformal and less frequent incorporation and microstylolitic) structures related to the intergranular pressure are abundant in the sandstones (Fig. 1b). These fabrics are indistinguishable in morphology from similar fabrics in the sandstones transformed at the stage of deep catagenesis. No indications of the directed gravitational corrosion under lithostatic load are noted. The linear dimension of clastic grains diminishes by 15–20%, indicating that 25–40 vol % of the clastic material passed into the dissolved state (Kopeliovich, 1995). However, the dissolved substance (silica, above all) was not redeposited within the bed. This is evident from the complete absence of authigenic quartz as regeneration rims on clastic quartz grains that occupy 40–60 vol %.

The sandstones are crosscut by quartz veinlets with an admixture of chlorite. Contours of these veinlets are diffuse (Figs. 1c, 1d). The veinlets likely mark the zones of filtration through weakened sectors of incompletely lithified rocks. Distinct thin quartz veinlets of the second generation crosscut the diffuse zones of filtration. These observations suggest that, relative to con-

ditions of subsidence lithogenesis, solution structures produced by lateral stress, which fosters the removal of silica and other components beyond the beds, appeared at different fluid pressures ($P_s > P_f$) (Marakushev, 1988), which are typical of dynamic processes accompanied by folding.

Suture surfaces of solution (Figs. 1e, 1f)—probably, initial manifestations of intergranular cleavage—are developed near and within cleavage and schistosity zones.

Thus, we have all grounds to suggest that Jurassic sandy rock beds were not lithified at the beginning of folding. They were lithified as a result of folding and compression over a rather long period, probably, in the course of discrete events.

In the cleavage zone (Fig. 1), clayey rocks with an insignificant admixture of fine silt are intensely foliated, and distance between cleavage surfaces is 0.0n mm (Figs. 1g, 2). These rocks are observed as dark gray or almost black slates easily splitting along cleavage planes. In thin sections, one can easily define relationships between foliation (cleavage) and primary bedding (Fig. 1h). As a rule, they make up a small angle (5–30°), indicating the syn- or postfolding origin of the cleavage. Sandy interbeds in the cleaved zones are also slightly inclined to the cleavage surface (Fig. 3). These sandstones are hardly affected by cleavage. As was mentioned above, conformal structures are supplemented with solution sutures in these rocks.

Thus, already deformed, nearly overturned beds of terrigenous rocks were affected by cleavage. Intense cleavage was formed under conditions of lateral compression almost exclusively in clayey interbeds that are sandwiched between the beds of more competent lithi-

fied (in the process of orogenic catagenesis) sandy rocks.

Metagenetic transformations of clay minerals in cleavage zone are superimposed on the older transformations. This is evident from the opposite chronological trends revealed by the K–Ar dating of rocks from the zones of orogenic catagenesis and metagenesis (Simanovich et al., 2004). These data testify that metagenesis of the clayey rocks genetically related to cleavage occurred after completion of the orogenic catagenesis approximately 180 Ma ago.

Orogenic catagenesis and metagenesis in both Toarcian–Aalenian sandy and clayey rocks apparently developed in a discrete (discontinuous-continuous) environment of strong compression that was accompanied by transverse contraction of basin, complication of fold-thrust dislocations, and the consequent intensification of lithification of rocks (mainly, sandstones). Lithification of the Toarcian–Aalenian sediments fostered the gradual formation of a block, and its frontal part was cleaved during collision with the Transcaucasus Massif lying above the subducted oceanic plate.

The cleavage zone occupies no more than 25% of total area (Simanovich et al., 2004). The boundary of this zone virtually coincides with the Kübler index isoline equal to 0.42. According to (Yang and Hesse, 1991; Hesse and Dalton, 1991), this value corresponds to the lower boundary of anchimetamorphism or metagenesis (Simanovich et al., 2004).

CONCLUSIONS

Thus, we considered an example of the formation and transformation of fabric and microfabric of terrigenous rocks before and during deformations of the SB. Probably, geodynamic settings in SBs of other types (Table 1) and their subsequent tectonic evolution generated diverse relationships between structures related to subsidence lithogenesis and deformations. In particular, our preliminary data indicate that deformations in accretionary wedges (SBs of active continental margins) are not accompanied by pre- or synfolding cleavage.

As has been established previously (Simanovich and Andriyanov, 1994; Simanovich and Yapaskurt, 2002), flyschoid sequences of the Verkhoyansk terrigenous complex in the southern Verkhoyansk region also underwent two stages of postsedimentary transformation. Subsidence lithogenesis of the undeformed SB lasted for about 100 Ma and led to the complete lithification of both clayey and sandy terrigenous rocks. The superimposed Late Mesozoic zonal metamorphism was caused by collision at the boundary between the Upper Yana–Kolyma Foldbelt and the Siberian Platform. These tectonic processes produced the extended thermal dome (Korikovskii, 1995), which is depicted at the present-day erosion level by biotite and staurolite isogrades (Simanovich and Andriyanov, 1994). Intensely

cleaved rocks of the metagenesis zone occur at flanks of the metamorphic aureole. Cleavage and related assemblage of deformational microfabric are superimposed on the older conformal-regeneration fabric formed in the course of subsidence lithogenesis.

In general, structurization of terrigenous rocks at the stage of postsedimentary lithogenesis, when the deformed sequences still lack metamorphic mineral assemblages, are characterized by the following common features:

(1) In terms of the degree and rate of lithification, the clayey rocks (including silty mudstones) are distinguished from the sandy rocks, which often remain friable before the deformational stage.

(2) As a rule, folding is not accompanied by cleavage. Cleaved zones are characterized by the simplest intergranular cleavage with complete retention of the primary bedding (S_0 and S_1). Schistosity cleavage (flattening) is typical of clayey rocks.

(3) The structural transformation of grained (sandy) rocks is caused by the development of dissolution structures both before the deformational stage (lithostatic pressure) and at the deformational stage (lateral compression, stress). At first glance, the conformal fabric developed in the process of folding does not differ from the solution fabric related to subsidence lithogenesis under the influence of lithostatic pressure.

(4) Mobilization and redeposition of dissolved substances in the course of subsidence lithogenesis, orogenic catagenesis, and metagenesis are drastically distinct because of the principally different fluid regimes. At the stage of deep catagenesis, the dissolved substance (mainly, silica) is redeposited in pores of rocks. Silica released during the synfolding (orogenic) catagenesis of sandy rocks is removed beyond the rock beds along filtration zones that are formed simultaneously with the lithification of rocks. The dissolved substance is mobilized during the formation of intergranular cleavage and accumulated as overgrowth aprons in pressure shadows or removed from the beds.

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