Stress state and deformation of the Earth’s crust
in the Altai–Sayan mountain region

Yu.L. Rebetsky \(^{a, *}\), O.A. Kuchai \(^{b}\), A.V. Marinin \(^{a}\)

\(^{a}\) Schmidt Institute of Physics of the Earth, Russian Academy of Sciences, ul. B. Gruzinskaya 10, Moscow, 123995, Russia

\(^{b}\) A.A. Trofimuk Institute of Petroleum Geology and Geophysics, Siberian Branch of the Russian Academy of Sciences, pr. Akademika Koptyuga 3, Novosibirsk, 630090, Russia

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Abstract

We present the results of tectonophysical reconstruction of natural stresses of the Earth’s crust in the Altai–Sayan mountain region using cataclastic analysis of fault slips and seismic data on the focal mechanisms of earthquakes. This method allows one to obtain the parameters of the total stress tensor by invoking additional data: generalized experimental data on the brittle fracture of rocks, seismic data on the released stress of strong earthquakes, and data on the topography and density of rocks. Results of the tectonophysical reconstruction of stresses showed significant inhomogeneity of the stress state, which is manifested not only in the variation of the strike and dip of the principal axes of the stress state, which is manifested not only in the variation of the strike and dip of the principal axes of the stress tensor, determining changes in the geodynamic regime of the Earth’s crust, but also in the close location of the regions of high and low isotropic tectonic pressure in relation to the lithostatic pressure. The variance of the ratio of tectonic pressure to lithostatic pressure is in the range of 0.59–1.31, with an average value for the region close to unity. This paper discusses internal or external mechanisms capable of generating the stress field obtained by the tectonophysical reconstruction.

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Introduction

Tectonic overview. The fold-mountain Altai–Sayan region is part of the Ural–Mongolian (Central Asian) fold belt stretching from the Urals to the Pacific Ocean, and is situated between the fold-mountain systems of Central Asia (Tien Shan and Pamir) and near-platform Baikal rift systems. The present Altai–Sayan mountain region began to form in place of denudation plains and low mountains in the Cenozoic (more probably, since the Oligocene time of 33–25 Ma) (Milanovskii, 1996). Late Cenozoic activation manifested itself in the accumulation of Neogene–Quaternary (the age of the sediments is 25 Ma to present time) continental glacial, lacustrine, and alluvial sediments of the Zaisan, Chuya, Uvs Nuur, Junggar, and other basins (Delvaux et al., 1995; Milanovskii, 1996). Data of fission-track analysis also indicate the chronology of mountain building from the south to the north of the Indo-Australian plate: the Himalayas—55 Ma and the northern Tien Shan and Junggaria—12 Ma (Bullen et al., 2001; Jolivet et al., 1999, 2001; Dobretsov et al., 1995; Wang et al., 2004). For the mountain structures of the Altai (including the Gobi and Mongolian Altai) and Sayan, the beginning of the recent uplift is estimated at 5–3 Ma (Buslov et al., 2008; De Grave et al., 2002, 2008; Jolivet et al., 2007; Vassallo et al., 2007).

For the area shown in Fig. 1, the maximum height difference in topography with 2-min averaging is about 3500 m, and with averaging of 30 min, the average height of the terrain is 1500 m with deviations of ±500 m in zones of uplift and depression, respectively. With this averaging of topography for the area covering the Mongolian, Rudnyi, and Gornyi Altai, the Eastern and Western Sayans, and the Northern Mongolia block (see Fig. 1), basins with elevations less than 1500 m occupy areas of about 580,000 km\(^2\), mountain uplifts with elevations of more than 1500 m occupy an area of about 480,000 km\(^2\). Thus, here the regions of crustal uplifts and depressions are close in area. The crust is thickened—45–65 km with roots in the region of mountain uplifts and antroots in the crust of large depression basins.

Data on active faults (Arzhannikov, 2003; Arzhannikov and Arzhannikov, 2006; Imaev et al., 2006; Mirosh-
nichenko et al., 2003; Trifonov et al., 2002) show a complex system of reverse faults, thrusts (with a dip angle of the fault plane less than 30º), and strike-slips. Large strike-slip faults are also observed on the northern and southern margins of the Altai and Sayans: the Kuznetsk Altai fault and the Biysk and Katun faults.

Recent movements. The slip rate along most of the major faults of the Mongolian Altai and Sayans is estimated from GPS geodesy data (San’kov et al., 2007) to be 1 to 4 mm/yr. The horizontal displacement vector field indicates oblique convergence at a rate of about 6 mm/yr in Mongolian Altai blocks identified from fault systems. San’kov et al. (2003) and Timofeev et al. (2005) note a trend of increasing rate of horizontal movements to the southwest of the Siberian platform. Within the Altai–Sayan mountain region, significant deformations in the north–south direction are observed in the vicinity of the Uvs Nuur lake (10–8–10–9 yr–1). To the south and north of this region, the deformation sharply reduces. From calculations of the horizontal strain components based on GPS data (San’kov et al., 2006) using the triangulation method with a characteristic size of 300–400 km of the base triangles with the vertices at the GPS observation sites, the contraction axes are predominantly oriented northeast. GPS data on recent vertical displacements show rapid uplift rates for the Eastern Sayans and Katangai.

Seismic regime. Since 1963, all earthquakes with $M > 3.4$ in the Altai–Sayan region are recorded by a network of regional seismic stations. The location accuracy of the epicenters fits into 5–15 km. The focal depth in the Altai–Sayan region cannot be determined reliably because of the large distance between the seismic stations. The epicentral field of the earthquakes in the period of instrumental recording (from 1963 to 2003) indicates the overall stability of the observed pattern of distributions of seismic events (Blagovidova et al., 1986; Zhalkovskii et al., 1978). Zhalkovskii et al. (1995) have shown that the epicenters of a considerable part of the earthquakes with $M > 3.4$ are arranged in chains along ridges. According to Lukina (1996), the strongest earthquakes of the Altai and Western Sayans are confined to the nodes of intersection of young active north–south trending faults with faults of other trends or they occurred in zones of current faults (Kuznetsova et al., 1999).

Recent studies carried out on the updated network of seismic stations of Altai have shown that the epicenters of seismic events are mostly concentrated on the borders of mountain ranges and basins (Emanov et al., 2005a,b; Gol’din et al., 2005), although some earthquakes, sometimes strong enough, occur in the crust of intermontane valleys and basins (Zaisan and Ureg Nuur earthquakes). Most of the earthquakes (see Fig. 1) are confined to the crust of uplifts and ridges. Intraplate fold-mountain orogens are characterized by the existence of large crustal regions in the form of “seismically silent”
depressions. Thus, for the Great Lakes, Uvs Nuur, and Tuva basins, the number of earthquakes is a few units in different catalogs, with a few hundred earthquakes in the crust of adjacent ridges and uplifts. From the standpoint of the brittle fracture mechanism, this suggests the greater strength of the rocks of crustal depressions or an increased ratio of the effective confining pressure to differential stresses.

**Basis of tectonophysical analysis of stresses**

**Initial seismic data.** The state of stress of the Earth’s crust of the Altai and Sayan orogens was studied using seismic data on the focal mechanisms of earthquakes (Gol’din and Kuchai, 2007; Kuznetsova et al., 1999; Zhalkovskii et al., 1995). Stress reconstruction was performed on the basis of seismic data on earthquake focal mechanisms obtained from records of the local network of seismic stations. The catalog of focal mechanisms with $M_b > 3$ included 308 events during the period from 1963 to 2003 (see Fig. 2). The depth of earthquake foci is set equal to 15 km since there are no reliable determinations of the source depth. Experimental data for determining the focal mechanisms used in the present study were earthquake records from seismic stations of the Altai–Sayan Seismological Expedition, Expedition at the Institute of the Earth Crust of the SB RAS, Kazakh Seismological Expedition, Kazakh National Nuclear Center, and Mongolia, and data on the sign of compressional displacements published in the Seismological Bulletins.

Preliminary analysis of this catalog showed that the distribution density of the earthquake epicenters and their magnitude range allowed stress reconstruction with a typical lateral averaging size of 50–70 km and to a depth of 30–40 km (the crust as a whole). In the calculation of the stress parameters, the initial seismic data were processed at the nodes of a $25^\circ \times 0.25^\circ$ grid located at a depth of 15 km. All procedures of generating homogeneous samples of earthquake focal mechanisms were completed for 640 grid nodes with not less than six earthquakes in each of such samples. The grid nodes corresponded to 640 quasi-homogeneous domains, which were assigned the obtained parameters of the natural stress tensor (Rebetsky et al., 2012).

We note that Zhalkovskii et al. (1995), Kuznetsova et al. (1996, 1999), Trifonov et al. (2002), and Gol’din and Kuchai (2007) analyzed seismotectonic deformations of the study region based on earthquake focal mechanisms using the method of Riznichenko (1985). However, the strain averaging scale in their calculations was much larger, which was responsible for a loss of a number of important features characteristic of the crust of intraplate orogens.

**Tectonophysical method of reconstruction of natural stresses.** The state of stress of the crust of the Altai and Sayans was studied using the method of cataclastic analysis (MCA) of fault slips (Rebetsky, 1999, 2003, 2005, 2007a, 2009a,b). In the MCA, as in a number of well-known methods (Angelier, 1984; Carey-Gailhardis and Mercier, 1987; Gephart and Forsyth, 1984; Gushchenko, 1975, 1979, 1996; Nikitin and Yunga, 1977), the basis for the calculation of stresses are seismic data on the focal mechanisms of earthquakes. In contrast to the above-mentioned methods, the MCA algorithm includes procedures for calculating not only the parameters of the stress ellipsoid and increments of seismotectonic strains, but also the relative values of the spherical and deviatoric components of the stress tensor. These capabilities of the MCA are due to the fact that the computation algorithm employs experimental observations of rock fracture (Coulomb criterion of brittle fracture) (Brace, 1972; Byerlee, 1978; Mogi, 1964; Stavrogin and Protosenya, 1992) extended to real fractured rocks in the form of a fracture zone in Mohr’s diagrams (Rebetsky, 2003, 2005, 2007a). In addition, the MCA uses data on the stress released in the foci of strong earthquakes, topography, and large intracrustal inhomogeneities. Based on these additional data, the effective brittle strength of mountain ranges, the pore fluid pressure average for the crust, and stress magnitudes are evaluated.
An important feature of the method is the identification of quasi-homogeneously deformed domains within which a homogeneous sample of earthquake focal mechanisms is generated. The MCA algorithm incorporates procedures of identifying such earthquakes based on the energy principles of the theory of plasticity (dissipation of elastic strain energy at faults). In the study region, all procedures of generating homogeneous samples of earthquake focal mechanisms were completed for 640 quasihomogeneous domains, with not less than six earthquakes in such samples.

Another important feature of the MCA is a simultaneous calculation of parameters of both the stress tensor and the seismotectonic strain increment tensor. In this method, as in the methods of Angelier (1984) and Gushchenko (1979), localization is performed on unit hemispheres of the areas of the possible emergence of the principal stress axes \( \sigma_1 \) and \( \sigma_3 \) (summation of the compression and expansion quadrants of earthquake focal mechanisms), which corresponds to the orientation of these axes for which earthquakes in a homogeneous sample would only lead to a decrease in the internal energy stored in elastic deformations (Bott, 1959). The choice of a unique solution from the entire set of possible positions of the principal stress axes is solved in the MCA by finding the maximum energy of dissipation of elastic strains on the seismotectonic strain-increment tensor calculated for the earthquakes from the homogeneous sample.

Note that the estimation of crustal stresses from fault slip data is the inverse problem of tectonophysics (Gzovskii, 1975), which explains the use of the term the reconstruction of natural stress. Solution of this inverse problem of tectonophysics based on data on the focal mechanisms of earthquakes during the last decades is determined by the result of reconstruction of the present stress field corresponding to the stress tensor parameters acting at the present time.

Parameters of tectonic stress ellipsoid

**Orientation of the principal stress axes.** The reconstruction of the ellipsoid stress parameters and seismotectonic strain increments in the first stage of the MCA (Rebetsky, 1999, 2007a) makes it possible to construct maps of orientations of the principal stress axes and to zone the crust according to the type of stress tensor (ellipsoid) and the geodynamic regime determined by the type of stress state. Figure 3a shows projections of the dipping axes of the principal stresses of the algebraically minimum stress \( \sigma_3 \) on a horizontal plane. Note that in most of the study region, these axes have a fairly flat south-dipping orientation (see the circle diagram). The average dip angle of these axes is 26°. The orientation of the other principal axis \( \sigma_1 \) has a roughly east–west strike with a broader change in the dip angle (see Fig. 3b). Along with this statistically representative distribution of the strike and dip angles of the principal stress axes, there are many local areas where the orientation of the axes undergoes significant deviations.

In the eastern part of the Gorny Altai, the subhorizontal position of the axis of maximum compression \( \sigma_3 \) is confined mainly to mountain areas and uplifts (Katun and Dzhebash anticlinoraria), and its subvertical position to basins and depressions (Tuva and Uvs Nuur basins). In the Western Sayans, the orientation of the axes of this principal stress is extremely unstable in both strike and dip. We suggest that this nature of the stress state is associated with the dominance of intermontane valleys and basins over mountain uplifts in this region of the crust. In the crust of weakly seismic intermontane depressions, earthquakes frequently have fault-type mechanisms, which, combined with the reverse-fault and strike-slip fault mechanisms in the crust of mountain ridges, leads to instability of stress reconstruction.

In the central zone of the Eastern Sayans, the maximum compression axes have a stable subhorizontal position and change strike from north-northwest in the north to northeast in the south. Further south, in the East-Tuva mountain uplift and the Tuva–Northern Mongolia block, they dip more steeply to the northeast. These two regions are separated by a narrow zone stretching along the Sangilen plateau with a roughly east-west orientation of the axis of maximum compression. In mountain areas of the Mongolian Altai, the maximum compression axes again take a subhorizontal position with a north–south strike. In the very east of the study region, these axes again have a mosaic orientation, which changes from roughly east–west direction to roughly north–south direction.

From the circle diagrams presented in Fig. 3, it follows that the subhorizontal (less than 30°) dip of the maximum compression axes is observed for 60% of the domains, and 12% of the domains has a steep (more than 60°) dip of these axes. Additional analysis showed that in 75% of the domains of stress reconstruction in the areas of uplifts, the orientation of the maximum compression axis is subhorizontal (dip less than 45°). In 75% of the domains of stress reconstruction in areas of large intermontane depressions or foredeeps, the orientation of the maximum compression axis is subvertical (dip of more than 45°).

**Type of geodynamic regime and the type of stress tensor.** Based on the orientation of the principal stress axes considered above, it is possible to perform geodynamic zoning of the study region. Figure 4 shows a diagram of division of the crust into six geodynamic types (types of stress state) obtained from an analysis of the position of the vector toward the zenith relative to the principal stress axes (Rebetsky, 2007a).

Figure 5 shows the field of the parameter which determines the type of geodynamic regime, which for the study region provides the full range of possible states from horizontal compression (the \( \sigma_1 \) axis is subvertical, see Fig. 3b) and shear (the \( \sigma_2 \) axis is subvertical) to extension (the \( \sigma_3 \) axis is subvertical, see Fig. 3a), with horizontal shear for the intermediate type. The Gorny Altai is dominated by domains of horizontal shear and its combination with extension. Local areas of horizontal extension are observed for the crust of the Chuya and Kurai intermontane depressions and near the western and the eastern ends of the South Chuya fault. The crust of the Western Sayans, dominated by basins and large intermontane depressions (Tuanka and Uvs Nuur basins, and...
Achit Nuur and Khemchik–Tuva depressions), is mainly subjected to horizontal extension and its combination with shear. To the south toward the Mongolian Altai and to the east toward the Eastern Sayans, the type of geodynamic regime is replaced by horizontal compression and its combination with shear. A vast region of horizontal extension is observed for the Tuva–Northern Mongolia block. The Zaisan basin is in a horizontal shear setting.

As seen from the diagram in Fig. 5 (the number of domains with different types of geodynamic regime), the state of horizontal shear is the most representative (about 30% of the domains), and the states of horizontal compression and extension are typical of a similar number of domains—about 20 and 25%.

After the first stage, the Lode–Nadai coefficient is calculated, which determines the type of stress ellipsoid and, hence, the type of stress tensor. In 70% of the Earth’s crust domains, its values are close to pure shear \( |\mu_{\sigma}| \leq 0.2 \). Among the exceptions with the largest area are the Tuva and Uvs Nuur basins, where there are several crustal regions with the stress tensor near uniaxial compression.

Stress-strain parameters in a coordinate system attached to the axis toward the zenith. The thickness of the seismogenic layer—the Earth’s crust within which the stress parameters are determined—is much smaller than the lateral dimensions of the study region. Neglecting the surface and bottom topography of the crust, we can speak of stress distribution in a plate with the characteristic lateral dimensions related to thickness as 30–50 to 1 (a thin plate). In this case, it is convenient to represent the stress tensor parameters averaged for the plate crust by the normal stresses of maximum and minimum compression acting in the lateral

Fig. 3. Projections of the principal stress axes: \( a \), algebraically minimum stress (maximum compression) \( \sigma_3 \); \( b \), algebraically maximum stress (deviatoric tension) \( \sigma_1 \). The axes are oriented in the dip direction: the circle indicates the origin of the axis at a dip angle of more than 15º, and the other end indicates the dip direction. Shorter axes correspond to larger angles of dip. If a circle is in the middle of the axis, this means that the dip angle is less than 15º. In the lower right corner of the figure shows circle diagrams defining the representativeness of individual ranges of values, i.e., the number of events in sectors of 20º and 10º, respectively, for the strike azimuth and dip angles of the principal stress axes. The asterisk marks the location of the 27.12.2011 and 26.02.2012 Tuva earthquakes.
direction and by the shear stresses applied to the plate bottom. We note that such shear stresses should be referred to as subduction stresses reflecting the interaction between the crust and the mantle.

Figure 6a shows the trajectories of the maximum compression axes acting in the lateral direction, which have a more ordered orientation than the projections of the dip axes of the principal stresses (Fig. 3). This figure highlights the north–south and northwest strike of the maximum lateral compression axes in the crust of the Zaisan basin, in the Gorny and Mongolian Altai, in the Kurtushiba anticlinorium, and in the central zone of the Eastern Sayans. This orientation of the compression axes corresponds to the strike of the fold and anticline hinges within the Katun ridge and in the southern part of the Kurtushiba anticlinorium. Some correspondence exists in the region with the northeast trending axis of maximum lateral compression at the junction of the Central Tuva basin with the central zone of the Eastern Sayans. In addition, Fig. 6a shows the deviatoric component of the maximum lateral compression normalized by the maximum shear stress. Note similarity of the data in Fig. 6a and Fig. 5.

In the areas of the geodynamic regime of horizontal compression determined this study in accordance with the diagram in Fig. 4, the largest compressive stresses act in the horizontal direction, and for the horizontal extension regimes, the deviatoric stresses in the lateral direction are tensile. Attention is also drawn to the Zaisan and Junggar basins, which is subjected to high lateral compressive stresses along with horizontal shear.

The orientation of the underthrust shear stresses indicates that the maximum impact of the underlying mantle on the crust is observed in the Gorny Altai, in the Kurtushiba anticlinorium, in the central part of the Mongolian Altai, and within the Eastern Tuva highland and the Sangilen ridge (Fig. 6b). Here the shear stresses on horizontal planes are more than half the maximum shear stress. In these areas, the subduction shear stresses are mainly oriented from east to west and from northeast to southwest. Increased values of these stresses are more frequently found in the crust of uplifts (Gorny and Mongolian Altai, Tuva–Northern Mongolia blocks), mountain ranges, and reduced values occur in basins (Zaisan, Tuva).

![Fig. 4. Types of geodynamic regimes](image)

Fig. 5. Type of geodynamic regime according to the diagram in Fig. 4. The diagram shows the number of domains for which stress data with different types of geodynamic regimes were available. The asterisk marks the position of the 27.12.2011 and 26.02.2012 Tuva earthquakes.
From the diagram of the distribution of the number of domains with different values of the azimuth, which determines the direction of subduction shear stresses \( \tau_z \), it is evident that the predominant direction of these shear stresses is the range of azimuths of 180° to 360° with the maximum near 270°–290°. This orientation determines the predominant west-east direction of sliding of the crust relative to the mantle.

**Total stress tensor**

**Reduced stress.** In accordance with the MCA algorithm, in the second stage of reconstruction, the spherical and deviatoric components of the stress tensor are evaluated up to normalization (reduced stress) by an unknown value of the internal cohesion \( \tau_f \), which is the averaged strength parameter of rock masses (Rebetsky, 2005). The scale of strength averaging corresponds to the scale of stress averaging, which, for this region, is equivalent to a linear scale of 50–70 km. Because of the scaling factor, the averaged strength of rock masses may differ significantly from the values obtained for rock samples (20–100 MPa (Byerlee, 1978)). Therefore, within the MCA, an algorithm for estimating \( \tau_f \) was developed which is implemented in the third stage of reconstruction. Selection of the fault plane of each earthquake from this sample was based on the criterion proposed in (Rebetsky, 2003, 2005, 2007a, 2009a,b), according to which the true fault plane of the two nodal planes is the one that provides a larger value of Coulomb stress.

The results of the second stage of reconstruction showed that the reduced effective confining pressure (the difference between the rock pressure and the fluid pressure in the fracture-pore space \( p^* = p - p_I \) for \( p = -\left(\sigma_1 + \sigma_2 + \sigma_3\right)/3 \) is...
nonuniformly distributed in the study region (Fig. 7). Extensive areas of increased effective pressure extend along the Kobdo fault within the Gornyi Altai and the Alash fault in the Eastern Sayans. In the Eastern Sayans, they are associated with the Belin-Busingol fault system trending roughly north-south and the east-west Obuchev fault. In the Mongolian Altai, there is a change from increased effective pressure to low pressure from north to south.

It should be noted that increased values of $p^*/\tau_f$ are observed for the crust regions subjected to both horizontal extension (areas along the Kobdo fault for the Gornyi and Mongolian Altai) and horizontal compression (the central part of the Eastern Sayans). Moreover, in a number of areas subjected to a more stable (in the area of occurrence) horizontal compression regime with a northwest trending $\sigma_3$ axis (the Zaisan basin, the southern part of the Mongolian Altai, and the northern margin of the Eastern Sayans), decreased effective pressure is observed.

In general, the reduced maximum shear stress is distributed similarly to the effective pressure. In areas where increased values of $p^*/\tau_f$ (Fig. 7) are observed, there are also increased values of $\tau/\tau_f$. In areas of reduced values of $p^*/\tau_f$, the values of $\tau/\tau_f$ are also reduced. This distribution is related to the assumption adopted in the MCA that the stress state in highly seismic areas is nearly limiting, which is expressed as tangency of the large Mohr circle to the outer envelope of (brittle fracture point) in the Mohr diagram. Since the limiting state for rocks is determined by the Coulomb–Mohr relation between the spherical stress tensor and the stress deviator (Rebetsky, 2003) (normal and shear stresses in the brittle fracture plane), a consequence of this is the relationship between the effective pressure and the maximum shear stress (Fig. 8). The ratio $\tau/p^*$ for the cloud of points in Fig. 8 for large values of $p^*/\tau_f$ is close to 0.5, which corresponds to the angle of internal friction with the coefficient $k_f = 0.5$ adopted in the calculation of the second stage of stress reconstruction (Rebetsky, 2007a).

It should be noted that the strongest earthquakes ($M_b \geq 5$) of the study region occur in areas of lower pressure and are not observed in areas of increased effective pressure (see Fig. 8). This is consistent with the idea of the greater efficiency of brittle fracture in areas of smaller confining compression (Rebetsky, 2007a,b,c; Rice, 1982).

**Determination of stress magnitudes.** In the third stage of the MCA, the internal cohesion averaged for the corresponding scale level is evaluated to transform from reduced stresses to their absolute values. This procedure can be performed if there are data on the dynamic parameters (released stress) of a strong earthquake source in the study region. In the Altai–Sayan region, such an earthquake was the 2003 Chuya earthquake (CE) ($M_w = 7.3$). For this earthquake, the Harvard University Web site gives the seismic moment and seismic wave energy $M_0 = 10^{20}$ Nm and $E_s = 4.8 \times 10^{15}$ Nm, respectively, which allows the released stress to be evaluated from the formulas of (Kostrov, 1975). According to our calculations, the released stress $\Delta \tau_n$ for the CE was about 4 MPa (in the calculations, the shear modulus of the crust was taken as $\mu = 3 \times 10^4$ MPa). Note that these values are consistent with the results of the analysis of the released stress performed in (Timofeev et al., 2005).

On the other hand, from the results of the first two stages of reconstruction using the MCA for the CE, it is possible to calculate the released stress up to the unknown normalizing coefficient ($1/\tau_f$), and thus, data on the released stress $\Delta \tau_n$ of the CE make it possible to evaluate $\tau_f$. To estimate the relative quantity $|\Delta \tau_n/\tau_f|$, we used data on the stress parameters of five quasi-homogeneous domains located near the CE source (see Table 1).
To calculate the average value of \( \frac{\Delta \tau_n}{\tau_f} \), we used domain Nos. 2, 3, and 4 from Table 1 since their total area corresponded to the size of the CE source. For these domains, the average normalized value of the released shear stress is 0.68. The above data on the released stress \( \Delta \tau_n \approx 4 \) MPa make it possible to determine the effective (average for an averaging scale of 50–70 km) internal cohesion, which for the investigated region of the crust was about 6 MPa \( \left( \frac{\Delta \tau_n}{\tau_f} = 0.667 \right) \).

Note the following important fact: The released stress obtained in our calculations is lower than the internal cohesion. This is not a contradiction because the presence of surface anisotropy leads to the fact that the direction of displacement along the fault often does not coincide with the shear stress direction on the fault before its activation (the coincidence of these directions corresponds to the well-known postulate introduced in (Bott, 1959; Wallace, 1951)). A consequence of this discrepancy should be a decrease in the stress release.

Conversion of the stress in the crust of the study region from relative to absolute values gives a range of 7 to 120 MPa for the maximum shear stress \( \tau \) and a range of 4.5 to 240 MPa for the effective pressure \( p^* \). This stress level is also 3–4 times higher than the corresponding parameters obtained for seismic areas of the Earth’s crust near the boundaries of oceanic lithospheric plates (Rebetsky, 2009a,b,c).

Figure 9 shows histograms of the distribution of the number of domains with different effective pressures and maximum shear stresses. It is seen that the range of the effective isotropic pressure of 20–60 MPa and the range of the maximum shear stress of 18–30 MPa are the most representative.

Tectonic and fluid pressure. As shown above, after the third stage of the MCA, it is possible to estimate the deviatoric stress (the maximum shear stress) and the effective normal stress (effective confining pressure). However, it would be extremely important to separately estimate the confining pressure in the solid rock skeleton and the fracture pore fluid pressure. This requires an additional equation for each of the

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<th>Latitude</th>
<th>( \sigma_1 )</th>
<th>Azimuth</th>
<th>Dip</th>
<th>( \sigma_2 )</th>
<th>Azimuth</th>
<th>Dip</th>
<th>( \sigma_3 )</th>
<th>Azimuth</th>
<th>Dip</th>
<th>( \mu \sigma )</th>
<th>( p^*/\gamma_f )</th>
<th>( \tau/\gamma_f )</th>
<th>( \Delta \tau_n/\gamma_f )</th>
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<td>0</td>
<td>270</td>
<td>90</td>
<td>-0.04</td>
<td>9.19</td>
<td>4.97</td>
<td>-0.56</td>
<td>0.21</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Note. \( p^*/\gamma_f \) is the relative effective pressure; \( \tau/\gamma_f \) is the relative maximum shear stress; \( \Delta \tau_n/\gamma_f \), \( \Delta \tau_s/\gamma_f \) are the relative internal cohesions.
domains with the data from the previous three stages of reconstruction. To this end, in the fourth stage of reconstruction of the MCA, use is made of the well-known proposition (Sibson, 1974; and others) that the vertical stress $\sigma_z$ is close to the weight of the rock column (in the calculations, the lithostatic pressure of one kilometer of rocks was assumed to be 27 MPa), which in (Rebetsky, 2009b,c), based on the theory of thick plates and shallow shells, is extended to account for the fact that almost always there is a deviation of one of the principal stress axes from a strictly vertical orientation.

Figure 10a shows the ratio of the fluid pressure to the lithostatic rock pressure calculated at the grid nodes at a depth of 15 km with the surface topography taken into account. Because of the averaging scale, the fluid pressure is the average for the crust as a whole. The ratio $p_{fl}/p_{lt}$ varies from 0.17 to 1 with a mean of 0.86 ($p_{lt}$ is the lithostatic pressure). The largest areas of its maximum values (0.95–1.0) are found in the crust of the Mongolian Altai, and regions of smaller area are also present in the crust of the Western and Eastern Sayans and in small-area localities of Gorny Altai. In these regions, the relative effective confining pressure $p^*/\tau_f$ takes intermediate and minimum values (see Fig. 7).

Using data on the fluid pressure and the effective internal cohesion, we can transform from the relative effective pressure $p^*/\tau_f$ (Fig. 7) to the absolute tectonic pressure $p$. It is most convenient to analyze the areal distribution of this parameter taking into account the lithostatic pressure. Figure 10b shows the distribution of the ratio of the tectonic pressure to the lithostatic pressure $p/p_{lt}$ taking into account the surface topography at a depth of 15 km. It is essential to note that the average tectonic pressure for the study region close to the lithostatic pressure ($p/p_{lt} = 0.995$), and the entire range of the ratio $p/p_{lt}$ is 0.59–1.31. The maximum values of $p/p_{lt}$ are found in the crust of the Eastern Sayans, and the minimum in the crust of the Western Sayans. On average, in the crust of the Eastern Sayans, the confining pressure is relatively lower than that in the Western Sayans and the Mongolian Altai. Comparison of the ratio $p/p_{lt}$ for the Earth’s crust of basins and intermontane valleys with mountain ranges and uplifts shows that there is no unique relationship of the high and low values of $p/p_{lt}$ with relief.

Figure 11 shows the relationship between the maximum shear stress $\tau$ and the relative value of $p/p_{lt}$. Note also that the point cloud is generally symmetric about the vertical line corresponding to the confining lithostatic pressure, i.e., the areas of confining compression and tension are similar in the number of domains and the maximum shear stress. From the data obtained, as the fluid pressure increases, the maximum shear stress decreases by a factor of 5–6 but the amplitude of the variation of the relative tectonic pressure remains unchanged (deviations from lithostatic values). Thus, we do not see a preferential distribution of overlithostatic compression in the crust of the study region and an increase in the deviatoric stress in the crust regions where the confining pressure is very high. This is not evidence in favor of a unified external loading mechanism, but indicates that the mechanism responsible for the formation of the current stress field has the same energetic effect on the areas of confining compression and tension.

Tuva earthquakes of 2011 and 2012

When this article was submitted for publication, two strong earthquakes occurred near the town of Kyzyl: 27.12.2011 with $M_b = 6.6$ and 26.02.2012 with $M_b = 6.7$ (depths of 12–15 km). The epicenters of these Tuva earthquakes fell into the region (see Fig. 2) for which the parameters of the natural stress tensor were obtained from seismic data on the focal mechanisms of earthquakes for the period of 1963–2003. According to our data (Fig. 3), here the orientation of the axis of maximum compression and the axis of maximum deviatoric
tension is north-northeast (azimuth 53°–58°, dip 6°–6.18°) and
south-southeast (azimuth 138°–135°, dip 26°–27°), respec-
tively, which is in good agreement with the earthquake focal
mechanism available at the site.

As follows from Fig. 7, both of these earthquakes occurred
in the crustal region where the reduced effective pressure
\( \left( \frac{p^*}{\tau_f} \right) \) varies in the range of 10–15, which corresponds to
absolute values of the effective pressure of 60–90 MPa. Thus,
as already noted (see Fig. 6 and Fig. 7), these two earthquakes,
as well as the 2003 Chuya earthquake, occurred where the
confining pressure in the Earth’s crust is relatively low (for
the crust Altai and Sayans, the maximum effective pressure
is close to 200–240 MPa).

The strong earthquakes occurring in the Altai and Sayans
in the past 10 years suggest that hazardous regions of the crust
are those with an effective pressure of 50–100 MPa (for a
reduced pressure of 10–15 in Fig. 7) extended along existing
active faults. The length of such regions determines the
expected earthquake magnitude (Rebetsky, 2007b,c).

Stress formation mechanism for the crust of the Altai
and Sayan

We will briefly repeat the most important results of the
stress reconstruction: In the crust of mountain ranges and
uplifts, the greatest compression in a large number of cases is
subhorizontal; in the crust of basins, depressions, and large
intermontane valleys, the maximum compression is more often
subvertical; the subduction shear stresses on horizontal planes
(with the normal to the center of the Earth) are oriented from
east to west and are found mainly in the crust of mountain

Fig. 10. Distributions of relative values of fluid pressure \( \frac{p_{fl}}{p_{lt}} \) (a) and tectonic pressure \( \frac{p}{p_{lt}} \) (b) at a depth of 15 km in the Earth’s crust. The lithostatic pressure is calculated with the topography taken into account. The diagram in the top inset shows the number of domains with the corresponding ratios of \( \frac{p_{fl}}{p_{lt}} \) (a) and \( \frac{p}{p_{lt}} \) (b). Value \( \frac{p}{p_{lt}} \) shown in standard deviation from mean 0.995.
ranges and uplifts; the confining tectonic pressure changes at a depth of 15 km from 0.6 to 1.3 of the lithostatic pressure with a mean close to the lithostatic pressure.

The above results of reconstruction of natural stresses require special analysis to identify the mechanisms responsible for their formation. Currently, the most discussed mechanism of generation of differential stress in the orogens of Central Asia should include: Collision of the Indian plate, producing long-range stresses of roughly north–south compression (Molnar and Topponier, 1975); planetary rotation, forming north-south oriented distributed forces and leading to the shift of the continental plates to the equator (Stovas, 1975); rise of the abnormally heated mantle from the depth to the base of the continental lithosphere (Artyushkov, 1993); volumetrically distributed intralithospheric forces associated with the gravitational spreading of the continents (topographic forces) (Bird, 1998; Richardson et al., 1975) and the impact caused by the sliding of the lithosphere off oceanic ridges and movement of its cold and denser part into trenches (where the mantle is passive). This latter mechanism is often combined with the concept of an active mantle, whose flow, due to thermo-gravitational convection, interacts with the continental lithosphere (the movement of the Indian plate is a consequence of this flow) via shear stresses.

In this section we will try to understand how the results of reconstruction of natural stresses provide an understanding of the mechanism of their generation. We first note that all the above loading mechanism produce additional stresses in the crust relative to the initial gravitational stress state caused by the action of the weight of rocks. The loading mechanism related to the pressure of the Indian plate from south to north is close in final result to the mechanism of impact caused by the movement of the East Siberian plate from north to south under the action of planetary rotation. The difference in the formation of additional stresses of horizontal compression is the distributed nature of the active forces for the second loading mechanism and the relative displacements in a coordinate system attached to the center of mass of the Earth.

According to the model of Artyushkov (1993), the epiplateform orogeny of the Altai and Sayan is due to the severe decompression of the heavy basic rocks preserved in the lithosphere after the separation of heavy eclogite from the crust and its sinking into the mantle (the first stage post-folding uplift). This decompression should give rise to a buoyancy

![Fig. 11. Ratio of the maximum shear stress \( \tau \) (MPa) to the relative confining pressure \( p/p_L \) for domains with different fluid pressures.](image-url)
force in the subcrustal lithosphere that produces an uplift dome. The consequence of this loading will be the appearance of tensile horizontal stresses in the crust, which are superimposed on the initial gravitational state present here.

In models of the influence of topographic forces (Artushkov, 1973; Frank, 1972), orogens are treated as passive structures which are assigned the role of relaxation of the gravitational potential difference caused by the presence of relief (not only of the surface but also the base of the lithosphere) and internal heterogeneities. Under this concept, only processes at the boundaries of the oceanic plates (spreading and subduction zones) and mantle flow in collision zones can produce active loading of the continental lithosphere (Bird, 1998; Richardson et al., 1976; Sandiford et al., 1995). Continental crust areas far enough away from these boundaries are subjected to stresses associated with the gravitational state.

To the above loading mechanisms in the crust of the orogens, we should add the mechanism of formation of residual stresses in uplift areas due to denudation of the surface (Makaveev, 1982; Neizvestnov, 2000; Ponomarev, 1990, 2007). This mechanism is associated with the relaxation of elastic strain due to changes in the volume of rocks at great depths. A detailed algorithm for estimating these residual stresses was developed in (Rebetsky, 2008a,b,c), and this type of stress is not manifested in any of the above mechanisms.

**Initial gravitational stress state.** In the first stage of our analysis, we estimate the average (for the region) values of the confining tectonic pressure and the maximum shear stress at a depth of 15 +1.5 km (the stress calculation depth with the average relief height taken into account) for the case where the crust of the study region is affected only by body forces in the absence of external influence (initial gravitational stress state, IGSS). Note that for IGSS, the axis of maximum compression \( \sigma_3 \) has a subvertical orientation, its values are close to the weight of the overlying rocks, and the ratio of tectonic pressure to lithostatic pressure varies from 0.56 (purely elastic state) to 1 (complete relaxation of deviatoric stresses) (Rebetsky, 2008a,b,c).

Assuming that the absolute average values of \( \sigma_3 \) (the compressive stresses are negative) to be equal to the average lithostatic pressure \( \bar{p}_{lt} \) at a depth of 15 km, we estimate the other two principal stresses of equal magnitude \( \sigma_1 = \sigma_2 \). To determine \( \sigma_1 \), we will use the Coulomb–Mohr failure criterion written for the invariants of the stress tensor:

\[
\sigma - k_j (p - \bar{p}_{lt}) \leq \tau_j \quad \text{for} \quad \tau = (\sigma_1 - \sigma_3)/2
\]

and

\[
\bar{p} = -(\sigma_1 + \sigma_2 + \sigma_3)/3. \tag{1}
\]

To calculate \( \sigma_1 \) from (1), we use the above data on the fluid pressure (Fig. 11a) assuming that the average \( \bar{p}_{lt} = 0.9 \bar{p}_{lt} \). The average lithostatic pressure of a rock column 16.5 km thick is determined as \( \bar{p}_{lt} = 446 \) MPa (\( p = 2.7 \) g/cm³), and in accordance with the results of the calculation, the strength parameters should be taken as \( \tau_j = 6 \) MPa and \( k_j = 0.5 \). Then, using the equality in (1), we obtain \( \sigma_1 = -412 \) MPa, from which the average confining tectonic pressure \( \bar{p} = 423 \) MPa and the maximum shear stress \( \tau = 17 \) MPa.

The above average tectonic pressure yields \( \bar{p}/\bar{p}_{lt} = 0.948 \), which is less than the average value for the region \( p/\bar{p}_{lt} = 0.995 \) obtained by stress reconstruction using the MCA (see Fig. 10b). The value of \( \tau \) is also significantly less than the average maximum shear stress from our calculation (Fig. 9b), which is about 34 MPa. Note that from the results for the reconstruction of natural stress, \( \tau \) was about 36 MPa for uplift areas, and in the areas of subsidence, it was about 31 MPa.

Suppose that the approximation of the confining tectonic pressure to the lithostatic value is achieved by relaxation of deviatoric stresses, accompanied by an increase in the horizontal compression (Rebetsky, 2008a,b), without the possibility of reindexing of the principal axes. This relaxation of deviatoric stresses will result in an even smaller value of \( \tau \), which will be absolutely inconsistent with the results of the calculations (Fig. 11). Thus, the relaxation of deviatoric stresses of the IGSS corresponding to the ideas of the gravitational spread of mountain uplifts under the action of topographic forces does not yield results similar to the natural data.

Increase in the confining tectonic pressure with a simultaneous increase in the deviatoric stress is possible if we assume that there are some internal or external factors that determine the increase in the lateral compression. Such external factors may be, first of all, the lateral pressure from the East Siberian platform and Hindustan (Molnar and Topponer, 1975) and the residual stresses of the IGSS in uplift areas after partial denudation of the surface—an internal factor (Rebetsky, 2008a,b). In both of the above cases, the result is the generation of additional stresses of horizontal compression, which are added to the stresses of the IGSS. Note also that in this case, the mechanism of formation of mountain uplifts after Artushkov is excluded from consideration because it produced horizontal extension additional to the IGSS, which is responsible for a deviation of the model stress state from that obtained in the tectonophysical reconstruction of stresses.

**Additional horizontal compression due to external lateral loading (collision of the Indian plate).** In order that this additional horizontal compression increase the ratio \( \bar{p}/\bar{p}_{lt} \) to a value of 0.995 under the factor of external influence, it is necessary to apply a roughly north–south lateral pressure of 50 MPa (the additional load is calculate for the purely elastic response of the geoenvironment under complete roughly east–west constraint and a Poisson’s ratio of 0.25). In this case, the total (due to the IGSS and the additional compression) stress acting in a roughly north–south direction will be \(-462\) MPa, and this will be the maximum compression \( (\sigma_3) \). The stress \( \sigma_1 = -425 \) MPa will act in the east–west direction, and the intermediate principal stresses \( \sigma_2 = -446 \) MPa in the vertical direction. Thus, the average geodynamic regime of the orogen will be horizontal shear.

However, the obtained average additional lateral compression is not sufficient for the averaged stress state to become critical. Under the additional north–south compression, the average maximum shear stress slightly increased \( \tau = 18.5 \) MPa, but the confining pressure increased even more significantly \( \bar{p} = 444 \) MPa, which does not provide the critical
value—the equality in (1). The establishment of the critical state is needed to explain the high seismic activity in the region.

Our estimates show that the limiting relation in condition (1) is achieved if the average additional north–south compression is about 108 MPa. In this case, the stress $\sigma_3$ acting in a roughly north–south direction will be $-520$ MPa, and the smaller compressive stress acting in the east–west direction reach $-441$ MPa, which is close to the value of the intermediate principal stress $\sigma_2$. In this case, $\tau = 40$ MPa, i.e., slightly higher than the calculated average values (see Fig. 10b) (34 MPa), and $\tilde{p}/\tilde{p}_H = 1.05$, which is significantly higher than the values obtained in our calculations (Fig. 11).

The estimates of the average stress values show some problems in explaining the natural stress state by the mechanism of additional horizontal compression. However, these problems increase if we take into account local inhomogeneities of the stress state that due to the presence of relief. In the crust of ridges, relief heights over 1500 m will lead to an increase in the tectonic pressure (the geodynamic regime for the average stress state is horizontal shear), which should lead to increased compression at cracks and fractures and, hence, reduced seismic activity even under the assumption of constancy of maximum shear stress. On the other hand, the tectonic pressure in the crust of valleys should conversely reduce, which should lead to an increase in the seismic activity in the crust of valleys compared with the crust of ridges, which is not confirmed. To fit the model stress state due to external lateral loading to the natural state, it is necessary to introduce local factors that reduce the confining local compression in the crust of mountain ridges and increase this compression in the crust of basins and intermontane valleys. The analysis made here taking into account the local features of the stress state associated with the relief should be extended to the mechanism of tectonic stress formation due to the influence of topographic forces.

**Additional horizontal compression due to the residual stresses of the IGSS.** The residual strains of the gravitational stress state can be calculated by setting the amplitudes of erosional transfer from uplift areas (ridges) to subsidence areas (depressions). Formulas for such estimates are given in (Rebetsky, 2008a,b,c). Here we perform inverse estimation. Let us determine the amplitudes of erosion that can explain the average crustal stresses the closest to the reconstructed values. This analysis should take into account that uplift areas (ridges) are in close proximity to the subsidence areas of basins and large intermontane valleys. Due to this proximity, the condition of lateral constraint is satisfied only in one direction—along the strike of ridges. We assume that the compressive stress acting in the other lateral direction (in the direction of valleys and basins) corresponds to the pure IGSS, $-412$ MPa. From this assumption and condition (1), we find that the maximum compression stresses oriented subhorizontally along the strike of the ridges (the Mongolian Altai and Eastern Sayans, see Fig. 3) reach a value $\sigma_3 = -464$ MPa at residual stresses of $-52$ MPa. In this case, $\tau = 26$ MPa (slightly below the average values (Fig. 9b, 34 MPa) and $\tilde{p}/\tilde{p}_H = 0.989$, which is very close to the values obtained in our calculations (Fig. 11).

Assuming a higher compressive stress level both along the laterally and in the direction of valleys, for example, $-420$ MPa (average of the vertical and horizontal stresses for the purely gravitational stress state), we have $\sigma_3 = -480$ MPa at a residual stress of $-68$ MPa. In this case, $\tau = 30$ MPa, and $\tilde{p}/\tilde{p}_H = 1.005$, which is quite close to the values obtained in our calculations (Figs. 9 and 11). Residual stresses of $52$–$68$ MPa required for the establishment of this level of tectonic stress can arise at $3.4$–$4.2$ km erosion of the surface (Rebetsky, 2008a,b,c). For a period of $5$–$3$ Ma, corresponding to the time of the last orogeny of the Altai and Sayan, these amounts of erosion corresponds to its speed of $0.8$–$1$ mm/yr, which is consistent with available estimates of the denudation of orogens (Ollier, 1984).

Note that the occurrence of additional compressive stresses in the crust of the ridges that act in the direction of valleys should lead to a decrease in the differential stresses and an increase in the effective confining pressure in the crust of the valleys. The result of this should be a departure of the critical stress from the critical state defined by expression (1), and a decrease in the probability of brittle fracture. As noted in the introduction, statistical data show that the seismic activity of valleys is significantly lower than the seismicity of the crust of uplifts and ridges; i.e., the predicted stress state is consistent with the observed seismicity distribution patterns.

**Conclusions**

The tectonophysical calculations of the present state of stress of the crust of the Altai–Sayan region show that the interpretation of the present state of orogens must take into account the residual stresses of horizontal compression occurring in rocks due to their elastic unloading in areas of rapid erosion and denudation (uplift areas) (Rebetsky et al., 2008). The role of this stress-generating mechanism is particularly important for intraplate orogens, which are characterized by high differentiation of vertical movements (uplift in ridges and subsidence in intermontane depressions), pronounced block divisibility (Bogachkin, 1981), and significant surface erosion. In blocks in which additional stresses of lateral compression were accumulated at relatively great depths only under gravitational stress (Rebetsky, 2008a,b,c) and which are moved closer to the surface, erosional unloading can be followed by a change in the principal stress index. This gives rise to the geodynamic regime of horizontal compression. In this case, intermontane valleys and large basins, which accumulate a considerable part of eroded material, undergo continuous subsidence of the paleosurface and are therefore subjected to horizontal extension (Ponomarev, 1990, 2007). These phenomena determine the coexistence of the horizontal compression regime in the crust of uplift areas and the horizontal extension regime in the crust of large intermontane valleys, basins, and depressions. This combination of geodynamic regimes is observed not only for the Altai–Sayan.
orogen, but also for the North Tien Shan (Rebetsky et al., 2010).

In our calculations, the alternation of areas with different geodynamic regimes (Fig. 5) and the block nature of the distribution of the orientation of subduction shear stresses (Fig. 6b) correspond to a linear scale of 500–700 km and 50–100 km respectively, which reflects the dependence of the present state of stress of the Earth’s crust in the study region on the deep crustal and mantle heterogeneities. We suggest that these heterogeneities cause upward and downward movements in the crust and mantle, providing the basis for the formation of residual stresses in the crust. The stress field formed by this mechanism exceeds in magnitude the background stresses determined by the external conditions of lateral loading from adjacent areas (pressure of the Indian plate). Our estimates show that for the Altai–Sayan orogen, this effect of this factor is not more than 5–10% of the stresses caused by other factors (residual stresses of the IGGSS). However, due to the long period (tens of millions of years) of the interaction of the Eurasian and Indian plates, it forms the structure of the orogen, determining the strike directions of ridges, trough axes, fold hinges, and major faults.

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