Active faults of the northern Tien Shan: tectonophysical zoning of seismic risk

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Abstract

This study continues the work by Mikhail Gzovsky on geological (tectonophysical) criteria for seismic risk. It is suggested to perform seismic-risk zoning according to parameters of normal and shear stresses on fault planes converted from results of tectonophysical stress reconstructions. The approach requires the knowledge of both dip and strike of the respective fault segments. Slip geometry is estimated from stress tensor, assuming that it is directed along shear stress. The suggested approach is applied to faults in the northern Tien Shan, and the current stress parameters are reconstructed using source mechanisms of catalogued earthquakes recorded by the KNET seismological network of the RAS Science Station in Bishkek. Stress modeling is performed by the method of cataclastic analysis providing constraints on stress ellipsoids, as well as on relations between the spherical and deviatoric components of the stress tensor. Plotted on the Mohr diagram, the fault stress points allow estimating whether the respective fault segments are close to the critical state (brittle failure). The suggested seismic-risk zoning of faults in the northern Tien Shan reveals up to 25 km long hazardous fault segments.

Keywords: current stress; seismic risk; tectonophysical zoning; active fault; hazardous fault; Coulomb stress

Introduction

Detecting faults prone to generation of earthquakes is the key point in seismic risk assessment. Discrimination of fault segments in terms of potential hazardous slip or creep is important for safe and environment-friendly operation of transportation and pipeline systems. Currently seismotectonics and GPS measurements, as well as methods based on earthquake records, are most commonly used for these purposes. Hazardous faults are identified from signatures of present and past activity on the surface (rock falls, landslides, etc.) or in trenched faults. Seismotectonic analysis allows qualitative assessment of hazard in different fault segments while intensity of paleoearthquakes from trenching data provides quantitative support.

Land-base and remote GPS measurements likewise can reveal activity of faults but mostly within short observations periods restricted to ongoing faulting. Instrumental records of earthquakes highlight brittle failure processes deep in the crust, but seismic analysis applies to historic events which often have magnitudes much smaller than the maximum possible values. The latter are found from recurrence plots updated using information on historic and prehistoric events.

Tectonophysical zoning of active faults is an alternative approach, which was suggested by Gzovsky (1975) fifty or sixty years ago but has been almost forgotten. Gzovsky used the term geological criteria of seismicity referring to signatures of seismic hazard showing the most probable highest magnitude of shocks and their expected recurrence. In the present-day terminology, they are rather tectonophysical criteria, as we call them in the consideration below.

Tectonic and seismic activity occurs in areas of rugged terrain and structurally differentiated crust, with high-gradient slip rates, and also in fields of young volcanism (Gzovsky, 1975). The geological (tectonophysical) criteria of seismicity make basis for contouring areas of highest shear stress and high-gradient vertical motions within different periods: millions of years for neotectonic methods, thousands and tens of thousand years for geomorphology, and decades for GPS measurements. The results were used to predict stress and strain trends and to assess the energy of pending events and duration of critical stress periods (Gzovsky, 1975).
By tectonic stress Gzovsky meant primarily the stress measured directly in rocks with implications for principal stress directions (the method of conjugate par faults), while estimation of total crustal stress was considered as a problem resolvable in the future.

We develop the approach by Gzovsky proceeding from today’s advance in rock mechanic studies of critical preseismic stress.

Gzovsky’s approaches to seismic zoning

In his synthesis of seismic risk data available at that time, Gzovsky (1975) noted that geological (tectonophysical) criteria of seismicity sometimes contradicted real observations. Specifically, they would indicate high activity in the Alps where actually the seismicity was low, but low activity in the Hungarian depression which was shocked by several earthquakes of shaking intensity 8. To explain the paradox, Gzovsky hypothesized deceleration of mountain growth in the Quaternary with decreasing deviatoric stress and proposed to study ongoing crustal movements using GPS.

Some other paradoxical examples are high seismicity in the Garm valley in spite of tectonic and geomorphologic signatures of stability (Gzovsky, 1975) or low seismicity in the northwestern Greater Caucasus (zones of probable shaking intensity 6), though some parts of the shore between Sukhumi and Novorossiysk rather should belong to the intensity 8 zone according to geomorphology and tectonics. Gzovsky noted that the seismic process changed in time and crust movements had slower or faster rates, which tectonic and geomorphological data, averaged over long periods, failed to resolve.

Gambrustsev (1955) also mentioned that tectonic movements varied with time and earthquakes had greater intensity during reactivation of seismic sutures in the crust which would appear to be stable from parameters averaged over hundreds or thousands of years.

The earthquake magnitude was hypothesized (Gambrustsev and Belousov, 1960) to depend on the size of the high-stress zone and the magnitude of largest events to influence the frequency of earthquakes, as the duration of the pre-faulting maximum shear stress was related with the stress level. Therefore, the size of the zone of high potential energy of elastic strain was related to the length of seismicogenic faults. Gzovsky (1975) noted in this respect that triaxial compression impeded earthquakes while faults could be healed.

He interpreted nucleation of large earthquakes in large faults as prolonged formation of small faults showing up as small earthquakes; in their turn, large earthquakes were understood as rapid motions along large faults as a result of pulse-like growth of originally isolated small faults (Gzovsky, 1975). The potential energy of elastic strain in the zone of high maximum shear stress drops dramatically as it becomes crosscut by a growing fault. Gzovsky (1975) considered a seismic source as an area where the potential elastic strain changed as a result of fault reactivation rather than as a tectonic suture (Dobrovolsky, 1991), an idea which agreed with the views by Benioff (1951), Bullen (1955) and Savarensky (1954). Gzovsky also mentioned the importance of faulting mechanisms in their relation to crustal stress.

The stress responsible for the ongoing seismic process can differ from that in which the respective faults were originally forming. Therefore, it is crucial to compare stress in active faults with the highest level and with stress estimates based on the amount of slip accumulated for hundreds of thousand or millions of years in the past.

Natural stress and strain: state of the art

Decades ago, when Gzovsky (1975) formulated the objectives of tectonophysics in seismic risk assessment, data of natural stress were collected by measurements of strain in crust and directly in mines during hydraulic fracture; by geomorphological and GPS measurements; and by methods of field tectonophysics and seismology. Geomorphological and GPS estimates gave rates of terrain changes and required the knowledge of relations between stress and gradients of motion. The measurements of joints in tectonophysics and fault plane solutions in seismology could show only principal directions but not the level of stress.

The situation has changed over the recent decades, both in methods and in amount of collected data. Remote sensing has allowed great progress in GPS measurements, with horizontal crust movements measured ten times more precisely than the vertical ones. Recent estimates of horizontal movements in Central Asian intracontinental orogens (Kuzikov and Mukhamediev, 2010; Sankov et al., 2011; Timofeev et al., 2009, 2013; Zubovich et al., 2007) were interpreted mostly in terms of regional evolution models, but they should be converted into gradients along some directions or rates of longitudinal strain. Strain can be also estimated by triangulation methods, which are quite well developed (Sankov et al., 2011; Tychkov et al., 2008). In continental active tectonic areas, lateral strain has annual rates of \( n \times 10^{-8} - n \times 10^{-9} \text{ yr}^{-1} \); higher rates are obtained at shorter base lines and smaller scales of averaging (Karmaleeva, 2012; Kuzmin, 2004): three or four orders of magnitude higher at tens to hundreds of kilometer lines, while strain is localized in vicinities of large faults.

Stress patterns have been largely documented for the upper 3 km of crust (Brady and Bzown, 2004; Brown and Hoek, 1978; Herget, 1973; Potvin et al., 2007; Zubkov et al., 2010). Much evidence was obtained previously on horizontal compression shortening exceeding the standard lateral pressure (Dimnik, 1926) and vertical (overburden) pressure. Currently it is known that (i) vertical stress varies geographically but approaches the overburden pressure; (ii) most of measurements show horizontal compression exceeding the vertical stress in areas of long-lasting uplift; (iii) most of measurements show stress about the standard of Dimnik (1926) in areas of long-lasting subsidence and nearly vertical maximum compression; (iv) horizontal compression axis is especially variable near the surface (from 0.3 to many times as high as the overburden pressure) but the variations decrease with depth.
(0.8 to 1.1 times the overburden pressure at a depth about 3 km); (v) principal stress directions are strongly variable at some localities, and the principal direction of compression is hard to estimate at large horizontal stress.

The methods for studying natural stress in mines have not changed much while tectonophysical studies have advanced considerably since the 1960s when conjugate shears was the only method (Gzovsky, 1975). The new method of cataclastic analysis (MCA) of discontinue displacements (Rebetsky, 2001) allows estimating stress tensor components, strain increment, and mechanic properties (including strength and yield) on the basis of plasticity theory. The MCA procedure consists of four stages. It begins with independent parallel calculations of principal stress directions and the Lodé-Nadaï coefficients for stress tensors and seismotectonic strain increment (Rebetsky, 2001) proceeding from basic energy postulates of mechanics. The second stage aims at estimating the relative effective isotropic pressure and the magnitude of highest shear stress that remained unknown at the first stage (Rebetsky, 2003). Modeling stems from experimental results of Byerlee (1968) showing that partly healed old joints of different directions can reactivate as shear stress on their surfaces exceeds the surface friction. At the third MCA stage, calculations apply to effective cohesion averaged on the scale corresponding to that of the reconstructed stress (Rebetsky, 2009a,c), using stress release in the sources of largest local earthquakes as additional data. Finally, tectonic and pore fluid pressure values are estimated which are related with effective pressure known from the previous stage, assuming that vertical stress approaches the overburden pressure (Sibson, 1974); an updated version of the latter hypothesis is based on the vertical momentum conservation in the thick-plate approximation (Rebetsky, 2007a).

Thus, the MCA approach can provide constraints on the total stress tensor and fluid pressure, and thus give a new perspective of detecting active faults and grouping them in terms of seismic risk.

### Active faults in the northern Tien Shan

MCA data on natural stress in the northern Tien Shan (Fig. 1) obtained by a joint team of the United Institute of the Physics of the Earth, laboratory of tectonophysics (Moscow) and the Science Station (Bishkek) will be used further for tectonophysical zoning of active faults. We found and digitized 102 active faults (Fig. 2) with constant dips, each with multiple straight segments, marked by the coordinates of starting and final points. The length of these segments (0.01–0.02 deg) are mainly a few kilometers or rarely 10–15 km. The mapped faults show strike variations, on both short (a few kilometers) and long (tens of km) base lines, according to varying strikes of basins and ranges and other terrain elements. The accuracy of strike estimates can be improved by averaging over 10 km.

Note that digitized azimuths of fault segments are necessary but insufficient for tectonophysical analysis in which the inferred 3D stresses are projected the onto a surface (fault plane): both strike and dip of faults have to be known, especially, in the case of dipping fault planes. In this respect, constant dips were assigned to fault segments of different lengths on the basis of additional geomorphological analysis of geological maps and field measurements (Fig. 3).

The mapping faced several problems. One is the lack of fault maps of a due scale to be used for reference, while the available composite geological and tectonic maps (1:500,000 and smaller) are obsolete and often disagree with more detailed maps. Thus, special care was taken to draw correctly the continuous lines of main faults. More difficulty came from discontinuous exposure of faults impeding field measurements. Poorly exposed and closely spaced parallel faults could be...
mistaken for a single structure or, vice versa, one long fault could be misinterpreted as multiple faults.

Another problem was that the study has focused on active faults, which are much fewer than inactive ones in the existing maps; furthermore, different maps assume different durations of activity required to consider a fault as active.

Some 1:50,000 geological survey maps and most of smaller-scale maps lack indications of dip directions and average angles, but show only directions and geometry of slip (thrust, normal, reverse or strike slip) or more often nothing but fault lines.

One more thing, not critical but annoying, is the lack of a regional catalog of fault names which would be helpful to pick data from different texts and figures.

Note that the maps of Figs. 2 and 3 were compiled with reference to other maps and text publications. They are, namely, more than sixty sheets of geological and tectonic maps of Kyrgyzstan and southern Kazakhstan of different scales; the map of Late Pleistocene–Holocene active faults of the Pamir–Tien Shan region (Makarov, 2005); the tectonic map of Kazakhstan and surroundings (Bespakov, 1975) and several other 1:500,000 maps (Chedia, 1988; Igemberdiev, 1982; Tursungaziev and Petrov, 2008; Zhukov, 1988); eight sheets of the 1:200,000 geological map of the USSR, northern Tien Shan series, compiled by Grishchenko and Turbina in 1964; Otkhotnikov and Novikov in 1962; Chabdarov et al. in 1967; Burtman et al. in 1960; Pomazkov and Burov in 1958; Zakharov and Zakharova in 1958; Pomazkov in 1962; and Lasovsky and Mozolev in 1961, all joined in the collection of Bespalov et al. (1960–1973); some later versions of unpublished 1:200,000 maps K-43-IX and K-43-XV compiled by Zhukov (1975); a later version of the 1:200,000 geological map published as an open file (Mikolaichuk and Apayarov, 2004).

Mountains flanking the easternmost Chu basin were mapped using the 1:100,000 unpublished structural map by Efremov et al. compiled in 1989, along with analysis of 1:50,000 unpublished geological and hydrological maps (few pieces from the archives of the Geological Survey of Kyrgyzstan) by Zakharov et al.; Levchenko et al.; Bondar et al.; Semenov et al.; Galanin et al.; Khristov et al.; Rubtsov et al.; Morozov et al.; Bondar; and Chernyavskaya et al. compiled in the 1960s through 2001.

Faults were considered active if they crossed Cenozoic rocks (or structures). In the case of controversy in fault locations among different authors, more trust was given to later references, larger-scale maps, and personal observations.

The fault planes with ambiguously estimated dips were characterized by average values in the case of minor variance; if the difference exceeded 30°, the fault was divided into segments, and the missing field-measured dips were calculated approximately according to horizontal topography elements crosscut by the fault. In a few cases, unknown steeply dipping sides of faults were inferred from geology and tectonics and estimated by analogy with the surrounding areas. We tried to reduce subjectivity-related biases, which are inevitable in these conditions.

The fault dips were corrected for possible increase with depth and the results were used together with data on stress in the upper 10 km of seismogenic crust. The accuracy of dip
estimates, though not very high (10°–20°), was sufficient for estimating reliably the stress parameters. The effect of dip accuracy was assessed by additional stress calculations in which dips were allowed to be 10° higher or lower than in Fig. 3 (see below).

The map of Fig. 3 provides a better idea of faulting in the northern Tien Shan than the map of A. Mikolaičuk published by Kalmetieva et al. (2009) and than that of Delvaux et al. (2001). Most of N–W faults dip south- or northward (Fig. 3) and some N–S faults dip to the east or west at 20° to 80° (relative to the horizon), most often at 50°–70°.

Most of faults in Fig. 3 dip beneath ranges (the Suusamyr basin, southern border of the Chu basin, etc.), except for those in the southern slope of the Kendyktas Range and the northern slope of the Kastek Range. In the central Kyrgyz Range and near the eastern end of the Chu basin, there are parallel faults dipping to the north and to the south, i.e., faults of different slip geometries may be spatially close, which is important for estimating principal stress directions.

In this study we determine fault dips using field geological data and topographic maps, from changes in curvature of fault projections on the surface. Dip angles are taken from geological reports and field data.

Crustal stress in the northern Tien Shan and theoretical background for tectonic zoning

Seismological data. The stress pattern of the northern Tien Shan was reconstructed using the catalog of earthquake mechanisms recorded by the local KNET network run by the RAN Science Station (Bishkek), with reference to our previous results (Rebetsky and Sycheva, 2008; Rebetsky et al., 2012). The catalog includes more than one thousand $M = 1.16$ to 5.40 events (Sycheva et al., 2003, 2005; Sycheva and Kuzikov, 2012) that occurred from 1994 through 2012 (Fig. 4). For stress modeling we used the mechanisms of $M = 1.5–3.5$ earthquakes, which corresponded to the regional scale of stress averaging on 0.05° × 0.05° grids. The grid spacing was about 5–6 km in longitude and latitude, according to the 10–15 km linear averaging scale inferred from the density and magnitudes of the catalogued events. The modeling was applied to overlapping 10-km thick horizontal layers with their middle at the depths 5, 10, 15, and 20 km, but the consideration below is confined to the upper 10 km for which stress parameters were obtained in 286 domains.

Principal stress directions (maximum deviatoric extension and maximum compression, Fig. 5a, b) were estimated at the first stage of MCA stress reconstruction (Rebetsky, 1997, 2001, 2007a) as three Euler (triple orthogonal vectors) angles or as dip directions and angles, along with the Lodé-Nadaï coefficient $\mu_\sigma$, to an accuracy of 5°–10° and 0.1–0.2, respectively, in stable areas but at a worse accuracy in areas of variable earthquake mechanisms.

The patterns obtained at stage I of MCA modeling (Fig. 5) were discussed earlier (Rebetsky and Sycheva, 2008; Rebetsky et al., 2012). Note that stage I yields four out of six stress tensor components, which represent deviatoric stress accurate to the values normalized to maximum shear stress. However, even these results provide additional information on faults if their dip angles are known, at least approximately (besides the strike). With three Euler angles and the dip and strike of faults, one can estimate the directions of shear stress applied to the fault plane. Only the magnitude of this stress depends on the maximum shear stress (which remains unknown after the first stage of modeling) while its directions presumably coincide with the relative displacement of the fault walls (Bott, 1959; Wallace, 1951). The latter hypothesis is used below to infer the slip geometry of faults.

Fig. 3. Map of faults in northern Tien Shan (compiled by S.I. Kuzikov, see text). Dip angles, in degrees (published for the first time), and directions are shown on hanging walls. Histogram in top left corner shows dip distribution of fault segments.
Of course, the estimation is possible only for the faults that run across or near the crust domains characterized by stress data. Depth-dependent variations of fault coordinates were taken into account by extending the faults to a depth of 5 km along the known dip and using the next grid node (no farther than 15 km or three times the characteristic grid spacing).

**Reduced stress.** The deviatoric stress components normalized by the unknown cohesion strength of rocks $\tau_f$ can be estimated from earthquake mechanisms and are called reduced stress (Rebetsky, 2005, 2007b). Figure 6 shows the lateral distribution of reduced effective confining pressure ($p^* = p - p_f$ is the pressure with regard to the fluid pressure $p_f$, according to (Terzaghi, 1943)); such patterns are also available for reduced values of maximum shear stress and principal stresses.

The cohesion $\tau_f$ of rocks averaged over a few tens of kilometers is constant within a zone of regional geodynamic setting. Thus comparison becomes possible for reduced stresses obtained at different grid nodes, as well as on planes of different fault segments, and can discriminate zones of high and low stress (Fig. 6).

The state of faults depends on stress in their vicinities, and this dependence shows up in fault morphology. The shear stress $\tau_n$ that controls convergence of fault walls is higher when the fault plane is closer to the plane of maximum shear stress. On the other hand, this shear stress should overcome friction that depends on effective (Terzaghi, 1943) stress $\sigma_n^*$ normal to the fault, i.e., reactivation of faults depends on the relation between normal and shear stresses on the fault plane.

**Fault strength.** The Coulomb stress ($\tau_C = \tau_n + k_f \sigma_n^*$ at $\sigma_n^* < 0$ and $k_f = 0.6$) includes shear and effective normal stresses, and its critical value reflects the rock strength (yield). The proximity of stress to the critical value can be judged from the Mohr diagram (Fig. 7). The large and small Mohr circles refer to the effective principal stress components $\sigma_1^*$, $\sigma_2^*$, $\sigma_3^*$ while points between them (light gray shade) represent effective normal and shear stresses at randomly oriented planes in the crust. The zone between the lines of brittle strength (yield) and minimum friction (dark gray shade) corresponds to the state of faults prone to activity (slip) as the Coulomb stress $\tau_C$ at the given fault segment reaches the surface cohesion $\tau_f^*$ ($0 \leq \tau_f^* \leq \tau_f$).

The Coulomb stresses of faults can be positive or negative. The former refer to stress points within the field of brittle failure in the Mohr diagram, between the lines of yield and minimum friction (Fig. 7). The closer the point to the yield line (C), the greater the hazard (white points) and the higher the shear stress released in slip. The points near the line of minimum dry friction are within the field of the least probable activity (open circles). Negative values show that the point lies within the elastic field, i.e., to the right of the minimum friction line (triangles).

The fault segments were divided into three groups according to reduced Coulomb stresses $\tau_C/\tau_f$ ($\tau_C = \tau_n + k_f \sigma_n^*$): (i) shear stress $\tau_n$ at the fault plane is below the minimum friction ($\tau_C \leq 0$); (ii) shear stress is above the minimum friction but the Coulomb stress is not very high ($0 < \tau_C/\tau_f \leq 0.8$); (iii) the Coulomb stress is high $0.8 < \tau_C/\tau_f \leq 1$. In the consideration below, the fault segments of the three groups are called, respectively, inactive, active, and hazardous.

**Division of active faults in the northern Tien Shan**

**Slip geometry of active faults** can be inferred from relative magnitudes of shear stresses along the dip and strike of the footwall fault plane (Fig. 8). Namely, stress projection on the fault dip exceeding that on the strike corresponds to the normal
Fig. 5. Stress pattern of northern Tien Shan (upper crust, 0 to 10 km) obtained by MCA, stage I. Horizontal projections of minimum (a) and maximum (b) principal stresses and Lodé-Nadai coefficients (c). Insets in top left corners show rose diagrams of principal stresses (a, b) and Lodé-Nadai coefficients (c). See text for explanation.
or reverse slip geometry, with right- or left-lateral strike-slip components; thus, the concept of reverse slip includes low-angle thrusts and steeply dipping faults. The setting with greater shear stress along the strike than along the dip corresponds to right- or left-lateral strike slip, with or without reverse or normal slip components. The stress-based division applies only to the fault segments for which stress data are available from areas within 15 km around; otherwise, faults are drawn as thin lines (Figs. 9–11).

The relative magnitudes of shear stresses along the dip and strike of fault segments used for their classification are measured by the $\phi$ angle: $60^\circ$–$120^\circ$ and $-60^\circ$–$-120^\circ$ for reverse and normal slip, respectively; $120^\circ$–$150^\circ$ and $-120^\circ$–$-150^\circ$ for reverse and normal slip, respectively, with a right-lateral component; $30^\circ$–$60^\circ$ and $-30^\circ$ to $-60^\circ$ for reverse and normal slip, respectively, with a left-lateral component; and the ranges $-30^\circ$ to $0^\circ$ and $0$ to $30^\circ$ as well as $150^\circ$ to $180^\circ$ and $-150^\circ$ to $-180^\circ$ for left- and right-lateral strike slip, respectively (Fig. 8). Thus the fault segments form four groups: (i) mostly reverse slip with a minor right- or left-lateral strike-slip component (Fig. 9a); (ii) mostly normal slip with a minor right- or left-lateral strike-slip component (Fig. 9b); (iii) mostly right-lateral strike slip with a minor reverse or normal component (Fig. 9c); (iv) mostly left-lateral strike slip with a minor reverse or dip-slip component (Fig. 9d). Reverse slip is obviously present in zones subject to shortening (Fig. 5a, b), while dip-slip and strike slip geometries are typical of extension and shear settings, respectively.

Faults over most of the study area, especially in its central and northeastern parts, show reverse slip with a minor right- or left-lateral component (Fig. 9a) and strike mainly in the ENE–WSW directions. The assumption of $10^\circ$ steeper dips in reverse faults makes them generally less representative. The situation improves if the faults are reidentified as those including segments with a right- or left-lateral strike slip prevalent over reverse slip, with $10^\circ$ shallower dips, respectively.
The faults with prevalent normal slip are much fewer; they are located mainly in the east and west of the study area (Fig. 9b) and vary in strike from ENE–WSW to ESE–WNW. All faults in the central part of the northern Tien Shan have reverse slip geometry. Segments with reverse slip are much longer than those of dip-slip. Unlike the case of reverse-slip faults, normal faults become more representative when assumed to have 10° steeper dips but the assumption of lower dip angles does not make them less representative.

Right-lateral strike-slip faults with minor reverse and dip-slip components are commensurate in length with prevalently reverse-slip faults (Fig. 9a). They strike mainly in the WNW–ESE direction and are rather evenly distributed over the study area (Fig. 9c), except for zones along 74°05′ and 75°10′ E. The reverse slip component is present in a greater part of right-lateral strike-slip segments and in almost all left-lateral segments. The latter are located mainly in the east and in the west of the area (Fig. 9d), are most often shorter than 15 km, and strike in the NE–SW or ENE–WSW directions.

Faults with prevalent strike slip become slightly more representative at the assumption of 10° steeper dips. The reidentified fault segments are short and connect those of prevalent reverse slip. Note that our division may slightly differ in depth, as the stress patterns were analyzed to depths within 10 km. Thus, the identified slip geometry may slightly differ from that observed on the surface, though the difference cannot be as dramatic as leading to a wrong choice between dip-slip and reverse slip or between right and left strike slip.

Comparison of inferred slip geometry with available geological data. It is important to compare the inferred slip geometries (Fig. 9) for faults in the upper 10 km of crust with published surface geological data. Our results agree with the geometry of slip according to the mechanism of the $M = 8.3$ Kemin earthquake of 1911 which occurred in the Chon-Kemin zone (northern Issyk Kul region), in the northeastern part of the study area. The earthquake mechanism was interpreted previously as reverse slip with a minor left-lateral strike-slip component (Bogdanovich et al., 1914; Kuchai, 1969). The Chon-Kemin zone distinguished by Shults (1948) and Makarov (1977) consists of ENE faults that dip southward (Bachmanov et al., 2008) and have a reverse-slip geometry with a left-lateral strike-slip component. In the west, the zone joins WNW–ESE faults with their geometry interpreted as right-lateral strike slip (Abdarakhmatov et al., 2001a), which is consistent with our predictions of right-lateral strike slip with or without a reverse component.

Our interpretation of the South Kochkor fault geometry (Fig. 9a) is in line with signatures of reverse and right-lateral slip in prehistoric earthquakes as ~25 km long ruptures in bedrock found in the southern Kochkor basin along the northern slope of the Terskei Range (Korjenkov, 1999).

The results for the Shamsi-Tyunduk fault dipping to the south agree with previous geological estimates: a reverse slip with a right-lateral strike-slip component (Makarov, 2005).

The main reverse fault in the Uzunbulak–Oikain zone (Figs. 3 and 9), which begins at the southwestern termination of the Kochkor basin, dips to the north according to Delvaux et al. (2001) and Kalmetieva et al. (2009). Our modeling also supports the SE dip inferred by Bachmanov et al. (2008) for its largest pinnate fault, though we suggest a higher angle: 60° (Fig. 3) instead of 30°.

There are, however, some cases of disagreement, either in the inferred slip geometry or in dip angles. For example, WSW faults in the Kastek Range have a left-lateral strike slip component according to Delvaux et al. (2001) but reverse,
Fig. 9. Slip geometry of faults according to current stress in upper 10 km of crust (Fig. 5). 

- **a**: reverse slip,
- **b**: normal slip,
- **c**: right-lateral strike slip,
- **d**: left-lateral strike slip.

Mixed vertical and horizontal components of motion are shown as well: reverse and dip slip with strike-slip components. Thin lines are faults for which no stress data is available.
normal, and right-lateral strike slip components in our results; the change from reverse to normal slip may be due to quite a high dip angle of 60° and nonverticality of the principal stress.

Another example concerns faults of the Karakol zone, which extend from the northwestern end of the Kochkor basin along the Karamoinok Range in the north and change their strike from NW to W–E near the Suusamyr basin. Our interpretation of right-lateral strike slip with reverse and normal components (Fig. 6) disagrees both with left-lateral strike slip suggested by Bachmanov et al. (2008) and Kalmetieva et al. (2009) and north-dipping reverse slip according to Delvaux et al. (2001). A large fault running in the north of the zone across the southern Kyrgyz Range appears as a W–E left-lateral strike slip in (Kalmetieva et al. (2009). We suggest strike slip with a dip-slip component for a small segment of an oblique fault in the Karakol zone (Fig. 9) dipping at a high angle of 70°. This is a case of a vertical shear setting (Yunga, 1990), when the maximum and minimum principal stresses are at ~45° to the vertical and minor variations of this angle can lead to either reverse or dip slip.

Like the Karakul fault, the faults in the northern border of the Suusamyr basin are interpreted as having a right-lateral strike slip component by V. Trifonov (Makarov, 2005) but as left-lateral strike slip faults by Kalmetieva et al. (2009). Our results agree rather with the former interpretation.

In a few cases, our estimates of fault dips differ from geological data. This is, specifically, a higher dip angle of 50° (Fig. 3), corresponding to deeper crust, we infer for the Issyk-Ata fault which was reported (Korzhenkov et al., 2012) to thrust northward at 190° and to dip at 19° near the surface.

For the South Kochkor fault, we suggest a dip angle smaller than that derived from seismotectonic and seismic data (Omuraliev et al., 2009): 40° (Fig. 3) against 60° and 45° above and below the 17.5 km depth, respectively.

Thus, the results of our modeling for the current deformation in the northern Tien Shan are generally consistent with published data of neotectonic motions observed on the surface.

**Stress at fault segments.** Reduced shear stress |\(\tau_n|/\tau_f| and effective normal stress |\(\sigma_n^*|/\tau_f| values (Fig. 10a, b) were calculated at stage II of the MCA procedure with regard to pore fluid pressure. Note that extension is considered positive in MCA, as in the classical mechanics. Normal and shear stresses often vary along faults according to their strike and dip variations; stresses may also change from one grid node to another.

High shear stress (Fig. 10a), along with normal compression (Fig. 10b), appears over the longest segments of W–E faults at the junction of the Kyrgyz and Karamoinok Ranges. An NW fault in the central Kyrgyz Range experiences low compression normal to the fault plane, which is expected to produce local rupture and springs on the surface. Locally, extension is inferred as well for this segment, but it should be taken with caution, as the stress is estimated at depths to 10 km, which is the accuracy limit of the method.

Inasmuch as normal compression results from friction that impedes shearing, the shear-to-normal stress ratio can be interpreted in terms of seismic risk. This applies to fault segments of high shear stress at the junction of the Kyrgyz and Karamoinok Ranges and the central Kyrgyz Range (Fig. 10a), as well as to some faults where both shear and normal compression stresses are low. Additional calculations with the assumption of 10° shallower dips show that the faults become more representative, with higher |\(\sigma_n^*|/\tau_f| ratios, almost fully at the account of faults north of the Djumgal Range and the Sandyk Mountains.

**Pore fluid pressure,** which contributes largely to the critical stress, can be estimated by the MCA algorithm using regional average cohesion |\(\tau_f|, which is either calculated from released stress (Rebetsky, 2009a,b,c) or assumed a priori, for
Fig. 10. Zoning of faults according to stress. a, Reduced shear stress $|\tau_n|/\tau_f$; b, reduced normal stress $|\sigma_n^*|/\tau_f$ (compression is negative); c, shear-to-normal stress ratio $\tau_n/|\sigma_n^*|$. Thin lines are faults for which no stress data is available.
instance, as 6 MPa obtained for the crust of the Altai–Sayan area (Rebetsky et al., 2013). With this assumption and a rock density of $\rho = 2.7 \text{ g/cm}^3$, the pore fluid pressure for the faults of the area is from 0.37 to 0.99 of the overburden pressure (Fig. 11).

The pore fluid pressure reaches its limit ($0.95 < \frac{p_{fl}}{p_{lt}} < 1$) in quite many faults but is rather low and close to the hydrostatic pressure ($\frac{p_{fl}}{p_{lt}} \approx 0.4$) in some faults. Increase in pore pressure can trigger slip even in faults (active or inactive) that are not hazardous at present. They are, namely, a large group of fault segments within an N–NW zone along 74°20′ E, where the pore pressure is 0.4–0.7 of overburden pressure (Fig. 11). The zone is located at the western termination of a large cluster of stress values in the central Northern Tien Shan; stress estimates west of this zone are much fewer, and the stress pattern is generally less stable. There are two more local zones of 0.6–0.8 pore/overburden pressure ratios, slightly higher than in the zone along 74°20′ E, where the faults can become hazardous upon a 10% pore pressure increase: one northwest of Lake Issyk Kul and the other in the southern slope of the Kastek Range.

According to our experience (Rebetsky, 2007a, 2009b; Rebetsky and Marinin, 2006a,b; Rebetsky and Tatevosian, 2013), shear stress is medium or even low in areas where large earthquakes are being nucleated. The reason is that effective normal stress in these segments is low and the released energy is spent more on rupture growth than on resisting friction (Rebetsky, 2007b; Rice, 1980). In this respect, the segments under low normal compression (Fig. 10b) that fall within the zone of brittle failure (high Coulomb stress) are hazardous as well.

Identification of hazardous faults from Coulomb stress.

The patterns of normal and shear stress, along with pore fluid pressure data, have implications for division of faults according to seismic risk. In order to check the accuracy of the division based on Coulomb stress, we performed calculations assuming different fault dips: those as in Fig. 3 and 10° steeper or shallower (Fig. 12), and obtained rather long (15–25 km) hazardous segments in all cases.

The longest 25 km fault segment, with stress about the critical value ($0.8 < \frac{\tau_{C}}{\tau_{f}} < 1$), along the southern slope of the Dzugal Range and Sandyk Mountains (Fig. 12b; dip angles are same as in Fig. 3) can generate $M = 6.5–7.0$ earthquakes. Shorter hazardous fault segments (10–15 km) at the junction of the Kyrgyz and Karamoinok Ranges (northward fault dips), in the central Kyrgyz Range, and in the southern slope of the Kastek Range, can generate $M = 5.5–6.5$ events, about the largest possible magnitude for the area.

There are also several shorter segments of critical state near the Bishkek RAS Science Station, the longest one (12 km) at the junction of the Chu basin and the northern slope of the Kyrgyz Range (Fig. 12b), within the GPS network of the Science Station. Still shorter (within 5 km) hazardous segments fall within the GPS network along N–S profiles east of the Science Station. Monitoring of crust deformation along these profiles would mitigate the risks. Nearly critical stress has been inferred also for two 15 km long fault segments in the eastern end of the Chu basin along the southern slope of the Kastek Range.

In the northern Issyk Kul area, hazardous fault segments are no longer than 5 km. The currently active Toguz-Bulak fault (Fig. 12b) appears nonhazardous according to the Coulomb stress estimated for a dip of 70° (Fig. 3) but hazard is predicted for its western segment for a smaller dip of 60° (Fig. 12a). A similar situation occurs along the southern slope of the Kastek Range. Our modeling shows that the Toguz-Bulak fault is the most active in its western segment, where it changes its strike from W–E to NW, but it becomes progressively less active in the eastern direction. The hazardous segment of the fault (Fig. 12a) corresponds to its ~15 km long western flank. This result agrees with seismotectonic data (Korjenkov et al., 2011) from the Lake Issyk Kul area indicating that the fault has been inactive in the Quaternary, and amount of motion along it decreases eastward.
Currently there are no hazardous faults in the area of the Belovodsk events (Fig. 12): two large earthquakes of $M = 6.4$ and $M = 6.9$ in 1865 and 1885, respectively, and an $M = 6.9$ event of 1838 near the eastern end of the Chu basin, on the northern slope of the Kungei Range (Kalmetieva et al., 2009). However, the stress is critical in a 10 km long NE segment 10–20 km southwest of the zone (Fig. 12b). The assumption of $10^\circ$ higher or lower dips does not change the predicted...
number of hazardous faults in this zone. Note that there is controversy about the causative fault of the Belovodsk earthquakes, attributed previously to the Issyk-Ata fault (Chedia et al., 2000; Korjenkov and Nikonov, 2011) and then to the Chonkurchak fault (Korzhenkov et al., 2012), by the same authors. Therefore, relating historic and prehistoric earthquakes to specific faults is problematic.

We also predict high activity of the Issyk-Ata fault (Fig. 12a) with the assumption of a dip 10° lower than in Fig. 3. The fault generated the Balasagun earthquake of 1475, of shaking intensity 8–9 (Korzhenkov et al., 2012), with its epicenter in the Shamsi River catchment, in the eastern Chu valley, at the northern slope of the Kyrgyz Range. The Shamsi Tyundyuk border fault, inactive in the Holocene (Korzhenkov et al., 2012), was an unlikely candidate for being the cause of the earthquake. The activity of the Issyk-Ata and Ulunbulak faults confirmed by our modeling (Fig. 11a) was also inferred by Abdarakhmatov et al. (2001b).

Although no long hazardous fault segments have been revealed in the northwestern Issyk Kul area shocked by the $M = 6.9$ earthquake of 1838 (Kalmetieva et al., 2009), there is a tangle of faults comprising short (3–7 km) hazardous segments (Fig. 12). Minor changes in stress or pore fluid pressure, which is currently low (Fig. 11), may trigger motion on these segments with variable dips.

Our predictions of hazard from a complex source with several fault planes in the northwestern Issyk Kul area are in line with the seismotectonic evidence (Korzhenkov et al., 2011) of at least four $M = 7$ medieval earthquakes (in the 5th, latest 7th, 9th, and 14th centuries), with slip on two different planes in the event of the latest 7th century.

Our results indicate activity of the Kyzylloi fault, which generated the $M = 7.3$ Suusamyr event of 1992 near the western border of the study area; the activity is higher in the south than in the north.

Thus, only 20–30 % of crustal faults appear to be active in the present stress field and only about 20 % of their segments are hazardous and prone to large-scale brittle failure in earthquakes. The fault pattern forms over tens or hundreds of million years, and only a small portion of these faults accommodates dissipated elastic energy accumulated during mountain growth (Dobretsov et al., 2013).

Conclusions

In his studies of geological (tectonophysical) criteria of seismic hazard, Gzovsky proceeded from stress and strain of upper crust expressed in velocity gradients of tectonic motions, tectonic topography, and volcanism, which are commonly used in assessment of seismotectonic activity (Reisner et al., 1993; Rogozhin, 2000; Ulomov, 2008), as well as stress and strain trends.

We develop the ideas of Gzovsky (Rebetsky, 2007b,c, 2011) and suggest a pioneering approach for detection of earthquake nucleation zones proceeding from stresses estimated on the basis of earthquake mechanism data. It consists in relating fault planes with slip corresponding to the current stress in rocks and in using the Coulomb rather than shear stress for estimating seismic hazard.

The approach stems from the modern views of the relation between stress and brittle failure which differ in many aspects from those of the 1950–1960s. Now it has become clear that shear strength of rocks depends on maximum shear stress and cohesion, as well as on friction. That is a reason why hazardous seismic effects observed in nature may differ from the theoretical predictions by the method of Gzovsky.

Another possible reason for the discrepancy between the theory and the practice is that the seismic process is uneven in time. The tectonic and geomorphic features (vertical slip rates) used in estimates of seismic risk were based on gradients of crustal movements averaged over long times, which makes it impossible to judge whether the current activity corresponded to a period of slower or faster movements. Therefore, stress inferred from earthquake mechanisms represents the ongoing phase of the geological process. The problem of averaged slip rates is resolved in our tectonophysical approach to seismic risk division of faults.

Note in conclusion that principal stress directions are used for analysis of seismotectonic and geodynamic settings, as well as for explaining the mechanism of faulting (Rebetsky, 2007b,c; Rebetsky and Alekseev, 2014; Rebetsky and Polets, 2014; Rebetsky et al., 2012, 2013). According to the modern views of the earthquake physics, reduced stresses have regular patterns in nucleation zones of large earthquakes (Rebetsky, 2007, 2009b; Rebetsky and Marinin, 2006a,b; Rebetsky and Tatevosian, 2013). Thus, crustal stress has implications for three basic problems in geosciences: faulting mechanisms, earthquake source physics, and seismic hazard.

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