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Seismic Activity of the Khambinskii Fault (Southwestern Transbaikalia)

O.P. Smekalin [a], A.V. Chipizubov [a], N.A. Radziminovich [a], V.S. Imaev [a,b]

[a] *Institute of the Earth's Crust, Siberian Branch of the Russian Academy of Sciences, Lermontova 128, Irkutsk, 664033, Russia*

[b] *Diamond and Precious Metal Geology Institute, Siberian Branch of the Russian Academy of Sciences, ul. Lenina 39, Yakutsk, 677027, Russia*

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Abstract—The paper presents results of a seismogeological study based on analysis of seismic data and historical facts about the seismic activity of the Khambinskii fault zone. According to the data obtained, a genetic type of dislocations on conjugate faults (Gusinoe Ozero and Orongoi paleoseismogenic structures) is related to reverse faults with a strike-slip component. Geophysical studies of the Gusinoe Ozero structure have determined the dip of the fault plane toward the mountain framing of the depression and its outcrop at the bottom of the seismic scarp. The significant seismic potential of the Khambinskii fault is responsible for the maximum intensity of shocks in the nearby cities and settlements of southeastern Transbaikalia. The seismic fault activity has been confirmed by the historical earthquakes of 1856 and 1885, the $M = 5$ earthquake that occurred on 2 October 1980, and at least two prehistoric earthquakes. The latest of the latter occurred no earlier than ~4 ka and had $M = 7.0$–7.3, while the earliest was even more intense and took place in the first half of the Holocene, no later than ~6 ka.

Keywords: Khambinskii active fault, Gusinoe Ozero paleoseismogenic structure, seismic potential, shallow-depth exploration, trenching, Transbaikalia

INTRODUCTION

It was the discovery of unconsolidated proluvial deposits that initiated detailed investigations in the Khambinskii fault zone. This paper presents the study results for two sites. The first site is situated opposite the southern extremity of Lake Gusinoe (Gusinoe Ozero). Dislocations of this site are designated as the Gusinoe Ozero paleoseismogenic structure (PSS). The second site is located along the northeastern flank of the Khambinskii fault branching into separate branches like a horsetail structure. One of the branches is represented by a seismic scarp to be described in this study for the first time. The remoteness of the sites from each other and opposite directions of horizontal slips are indicative of two different paleoseismogenic structures. The structure located along the northwestern flank of the Khambinskii fault is conventionally called as the Orongoi PSS.

High seismic potential of the Khambinskii fault (magnitude (M) of not less than 7) was determined after the finding of the Gusinoe Ozero PSS in 1965. The findings of the first investigations are presented in a number of publications (Solonenko, 1968; Khromovskikh and Lastochkin, 1970; Lastochkin, 1982). They present the morphokinetic parameters of the structure obtained using the tools and investigation techniques of that time (Solonenko, 1977). Next decades were marked by new publications devoted to the geological structure, tectonics and seismicity of the Gusinoe Ozero depression and its mountain framing, where the seismic activity of faults (controlling the edges of the depressions) was detected (Golenetskii et al., 1982; Bulnaev, 2006; Lunina and Gladkov, 2009; Lunina et al., 2010; and others).

The purpose of this paper is to present the results of the recent Khambinskii fault studies, including the data on the newly discovered Orongoi PSS in the northeastern part of the fault. The investigations were based on the latest achievements in the field of paleoseismology (McCalpin, 2009). The analysis of dislocation morphology, trenching, near-surface geophysical methods, historical data on seismicity, and remote sensing data allowed determining the time interval of the Gusinoe Ozero dislocation formation, change our understanding on the kinematics of its slips, and redefine the lenght of the Khambinskii fault fragment activated in the late Cenozoic.

Substantiation of the Khambinskii fault seismic potential and its associated zone of probable earthquake sources (PES) must be of great practical importance. The Khambinskii fault, among other faults on the eastern shore of Baikal, appears to be a source of maximum possible earthquakes in the Ulan-Ude territory. Being the most densely-populated part of the republic, southern Buryatia finds itself in the risk zone with its developed infrastructure, industrial and energy production sectors.

✉ Corresponding author.
E-mail adress: smekalin@crust.irk.ru (O.P. Smekalin)

GEOLOGICAL AND TECTONIC CHARACTERISTICS OF THE KHAMBINSKII FAULT

The Khambinskii fault formed in the Dzhida–Vitim suture zone in the Mesozoic at the beginning of the early-destructive stage, which was marked by large-scale generation of depression structures which are still topographically distinguished in the form of intermountain basins. In present-time settings, the Dzhida–Vitim zone is considered to be a southeastern border for the Selenga complex of metamorphic cores formed in the Mesozoic (Sklyarov et al., 1994, 1997; Mazukabzov et al., 2011 and others). In the Cenozoic, the Baikal rift and shearing zone was formed at the Mesozoic regional structures as a result of the remote impact of the Indo-Asian collision. The faults of the Dzhida–Vitim suture reactivated at the edge part of this zone (De Grave et al., 2007; Buslov, 2012).

According to the newest structural geometry, this suture zone separates the Baikal and Hentiyn-Daur megadomes (Fig. 1). These are regional positive structures developing on bedrocks passing through their own orogenic (granite-arch) stages of different duration and time of transition to the destructive phase of orogenesis (Komarov, 1996). The Baikal dome was subjected to the greatest degree of destruction, where the rift process developed most intensively along the crest line.

In the Cenozoic, inversion of vertical displacement was the most typical for the Transbaikalia depressions in the Hentiyn-Daur megadome framing. It led to a reduction of the sedimentation area, but the Jurassic and Cretaceous deposits were involved in the uplift with the formation of hilly and low-altitude relief in the place of former sedimentary basins from the time of the Mesozoic activation (Shatalov, 1977).

However, at the western side of the Gusinoe Ozero depression, the outcrop of the Khambinskii fault practically coincides with the occurrence borderline of the Mesozoic–Cenozoic deposits (with minor exception). Thus, it indirectly confirms activity of the fault for the whole period of its existence but with different intensity.

In topographical representation, the Khambinskii fault is delineated with a scarp extending up to 200 km with some discontinuities. It is trending from the SW to the NE along fore fronts of southern, southeastern and eastern offsets of the Khamar-Daban Range, from the Gegetui depression to the Orongoi depressions (Fig. 2). The fault is not evenly active along the whole strike. According to (Levi et al., 1996), the Cenozoic differentiated displacement took place across

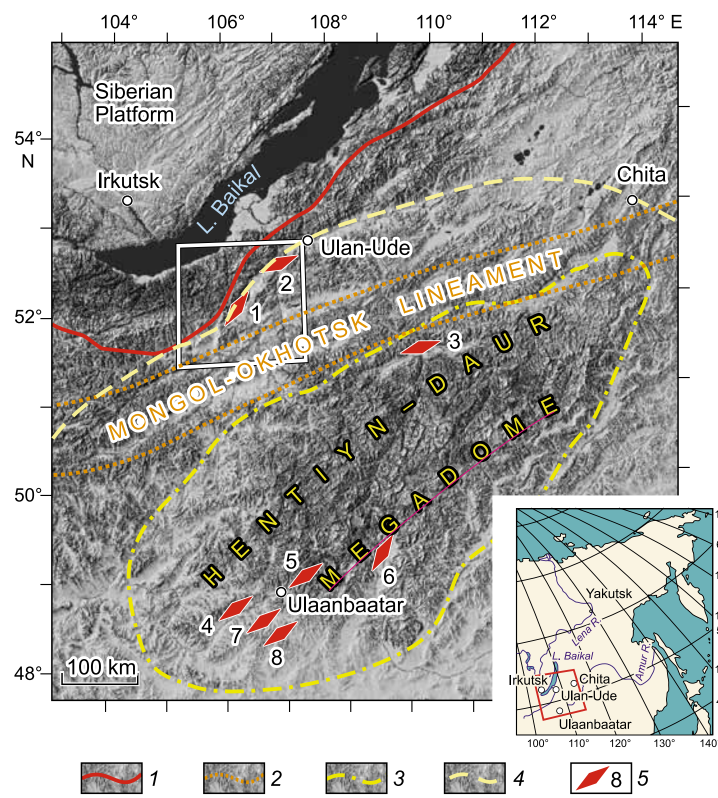

Fig. 1. The tectonic structures of southwestern Transbaikalia. *1*, Dzhida–Vitim fault according to (Khrenov, 1988); *2*, lines delineating the zone of active faults of the Mongol-Okhotsk lineament according to (Nikolaev, 1986); *3*, *4*,outer core outline (*3*) and the bottom of (*4*) the Hentiyn-Daur megadome, according to (Shatalov, 1977); *5*, paleoseismogenic structures within the limits of the Hentiyn-Daur megadome: Gusinoe Ozero (1), Orongoi (2), Chikoi (3), Khustai (4), Gunzhin (5), Kerulen (6), Sharkhai (7), Avdar (8).The image is obtained with the use of the 3" SRTM DEM digital relief model. The area shown in Fig. 2 is delineated with the white rectangle.

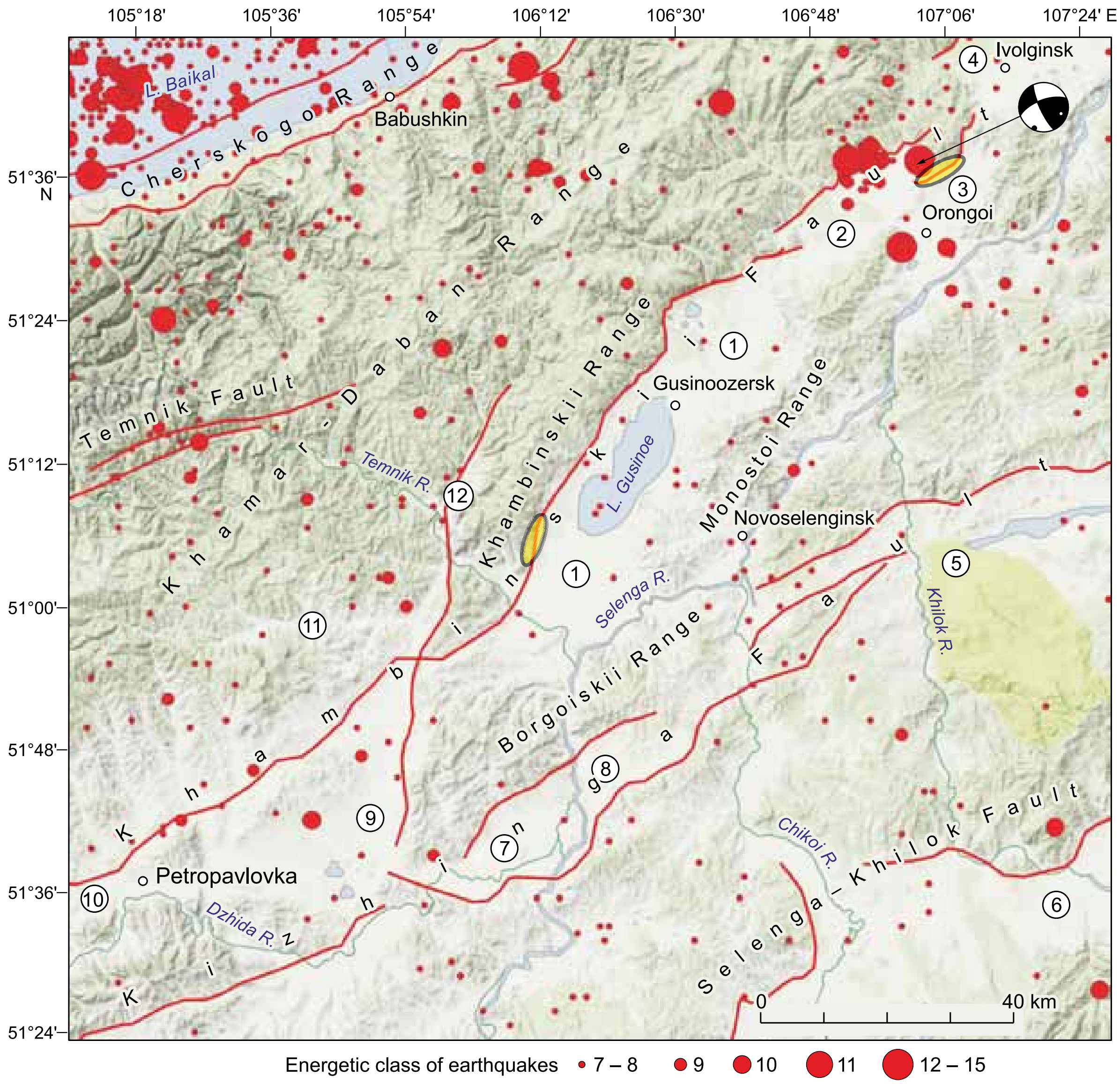

Fig. 2. Seismicity and active faults of southwestern Transbaikalia. Epicenters of the earthquakes with corresponding energy classes for the period from 1935 to 2015 are indicated according to the catalogue of Baikal Branch of the Federal Research Center of Unitary Geophysical Service, RAS. The focal mechanism of the 1980 Orongoi earthquake with $M = 5$ (in the lower hemisphere) is shown as in (Golenetsky et al., 1982). The faults are indicated in accordance with the results of field works and the authors' interpretation of remote techniques materials, geological and topographic maps. The circled figures correspond to the following depressions: 1, Gusinoe Ozero, 2, Upper Orongoi, 3, Lower Orongoi, 4, Ivolga, 5, Tugnui, 6, Chikoi–Khilok, 7, Ust'-Dzida, 8, Boldog, 9, Borgoi, 10, Gegetui; 11, Iroi, 12, Udunga. The yellow ellipses indicate the sites of detailed seismological investigations. The landscape (Google) image is used in Fig. 2.

the fault in the segment of 60 km. Structurally, this interval covers the junction zone from the Khambinskii Range to the Gusinoe Ozero depression and is represented in the form of a stepwise scarp with the summarized vertical displacement amplitude reaching up to 300 m and gradients of vertical movement up to $0.7–1.0 \cdot 10^{-8}$ a yr^{-1}. According to S.V. Lastochkin, the amplitude was accumulated during the neotectonic Pliocene–Quaternary stage, from which about 50 m are referred to the upper Pleistocene–Holocene. The uplifted side of the fault (the eastern slope of the Khambinskii Range) has over a dozen of falls and landslides; therefore, it could be considered as one of the indicators of higher seismic activity of the fault (Lastochkin, 1982). The Gusinoe Ozero depression is located on the downside of the fault and filled with middle–upper Mesozoic deposits and Neocene conglomerates.

RESEARCH METHODS

Since 1975, the time of early Khambinskii fault investigations, the methodological procedures and tools platforms have been substantially expanded as well as the theoretical

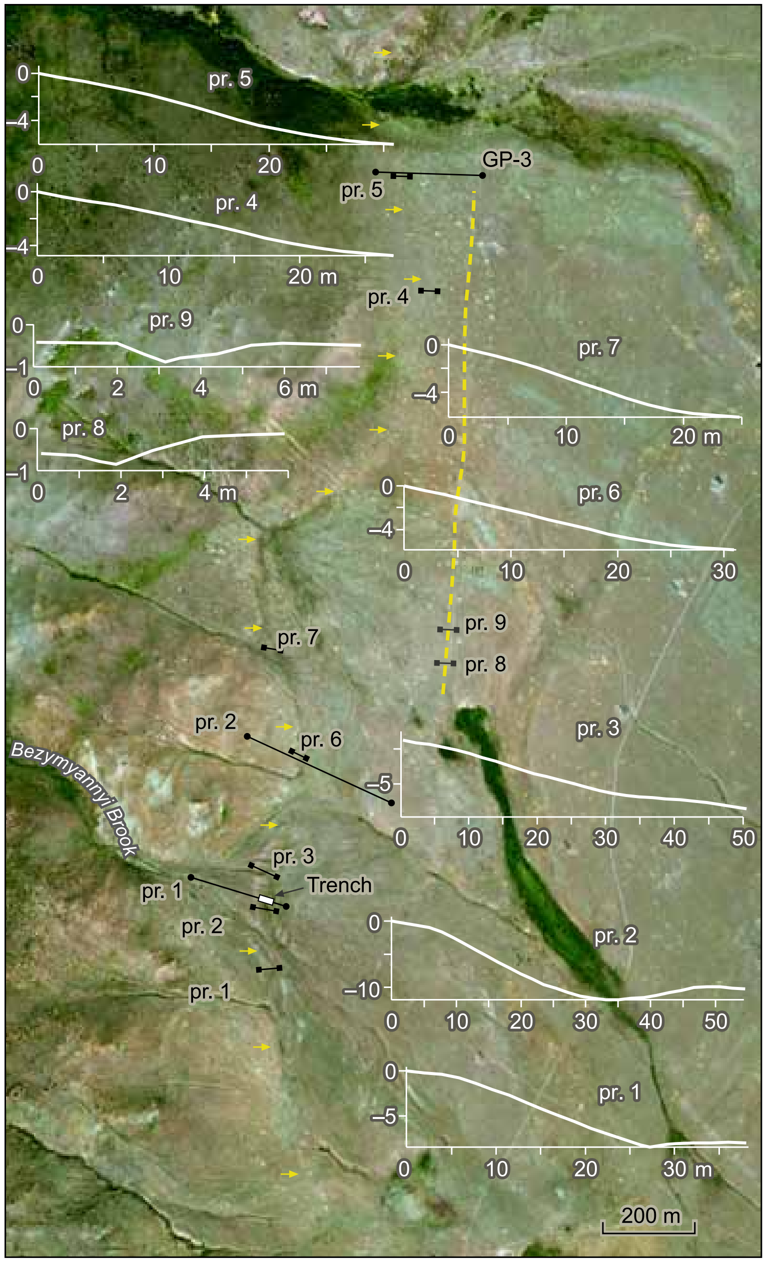

Fig. 3. Gusinoe Ozero PSS: geophysical mapping (lower ellipse in Fig. 2). The yellow arrows represent the main line of dislocations; the dashed line indicates the predicted tension fracture. The white contours correspond to topographic profiles; their site location is labeled with the short black segments. The ends of the long black segments correspond to the first and the last GP stations. The white rectangular marks the trench. The relief image is retrieved from: /landsatlook.usgs.gov/viewer.

bases that have been supplemented with modern achievements in the field of paleoseismology. Trenching and near-surface geophysical techniques have found a wide application. The tools of morphometry and collection of remotely sensed data have been upgraded. The techniques of determining the absolute age of dislocations and predicting paleoearthquakes have been developed (McCalpin, 2009).

The S-Digit Minidigital inclinometer (Geo Fennel) and the Leica DISTO D510 laser rangefinder were used to determine the displacement amplitude and a relative dislocation age over the scarp heights. Profile orientations (including geophysical ones) were carried out with the Garmin eTrex 30 GPS-receiver. Linear dimensions of scarps, gills and slides were determined using a digital elevation model (3" SRTM DEM). The age was measured by radiocarbon dating of the specimens sampled from the buried humus soil exposed in the ditch at the bottom of the seismic scarp.

To determine the dip and strike of the deep fault planes, geophysical investigations were conducted on three profiles. The following works were done on each profile: CMRW (complex method of refracted waves) seismic survey (with use of the Seismolog-24 seismic acquisition system), geoelectric survey (with use of the Skala-64 electroprospecting station) with modified axial dipole sounding (ADS) and ground penetrating radar techniques (OKO-II fitted with a Triton antenna assembly).

GUSINOE OZERO PSS

Morphological features of Gusinoe Ozero PSS. For several decades, morphometry has been the only technique applied to studying the quantitative parameters of seismogenic deformations in the Baikal region. Therefore, the first studies of the Gusinoe Ozero structure substantiated its seismic activity only with detailed data on dislocation morphology and characteristics of secondary earthquake (EQ) indicators, such as land falls, landslides etc. (Solonenko, 1968; Lastochkin, 1982). The structure was described as a large fault cutting through the unconsolidated proluvial cones of the Bezymyannayi Brook valley and southwards, as a small erosive trench cut with the total length of 600–700 m and the vertical displacement amplitude of 4 m (Fig. 3). On the surface, the dislocation is an asymmetric trench of 10 m in width; its uphill side of no more than 5–6 m in height (25°–30° steepness), and the downhill side up to 2 m in height (10°–15° steepness). Hanging mudflow channels are distinguished on the uplifted fault side, and they cut through the uphill trench side up to a depth of 1.0–1.5 m. Only one mudflow channel makes a cut on the trench side almost up to the bottom depth. The dislocations with the displacement amplitudes up to 1 m are observed at a distance of 2.5 km.

Sinistral displacements with 23–24 m amplitude are defined by talweg S-bendings and the left sides of erosive cuts along the fault line (scarp) in the debris cone of the Bezymyannyi Brook valley and northwards from it (Fig. 4). The axes of the debris cone apex on the opposite fault sides are displaced at a distance up to 50 m.

A tension fracture up to 1 km long, is traced eastwards at a distance of 300 m from the main fault and parallel to it. It varies in width from 0.5 to 2 m and in depth from 0.2 to 0.6 m with 2 × 3 m edgewise bulges at a depth of 0.2–0.4 m (Lastochkin, 1982). Hence, the upper crests of the opposite trench slopes lie either at the same level, or the downhill side of the fault is lofted (Fig. 3, profiles 8, 9). The present dislocation layout might have formed since the late Pleistocene (not earlier than 100–120 ka).

Trenching of the Gusinoe Ozero PSS. In 2000, A.V. Chipizubov and A.P. Serebrennikov dug a trench over the scarp to the proluvial cone surface of the Bezymyannyi Brook valley. The place for trenching had been chosen for soil sampling to determine the absolute age of the observed dislocations. The ditch exposed the lowest part of the scarp and a drainless trench at its bottom, where humus argillaceous-sandy clay deposits might have accumulated. For technical reasons, the scarp was not exposed along its entire length—from the lower crest to the upper one. Consequently, the main plane of the fault displacement lied beyond the borders of the exposed structure section, whose outcrop was expected in the middle of the scarp height. Only one of the feathering faults was exposed at the eastern edge of the ditch section. It is likely to be the lower boundary of the thrust fault comprised of schistose mass (F layer) and the bottom land-waste layer with clear strata interfaces parallel to the thrust fault plane.

The main part of the exposed section (Fig. 5) is dominated by two crushed stony-block layers (B and D), separated with a land-waste-schistose layer (C), and overlaying recent soil (A). The strata boundaries follow the surface pro-

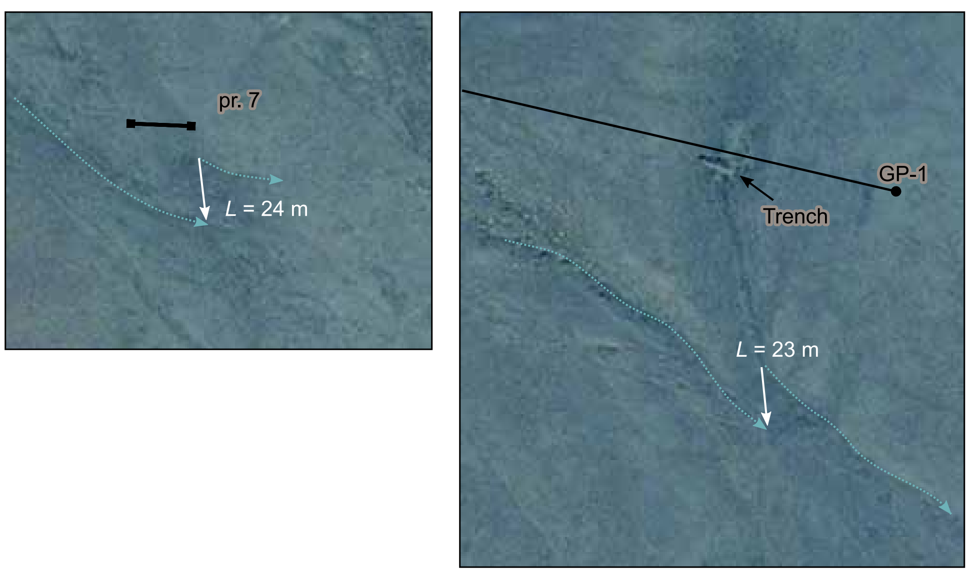

Fig. 4. Magnified fragments (see Fig. 3) illustrating fault deformations of the Gusinoe Ozero PSS. The fragments are bound to the map via profiles and trench positions.

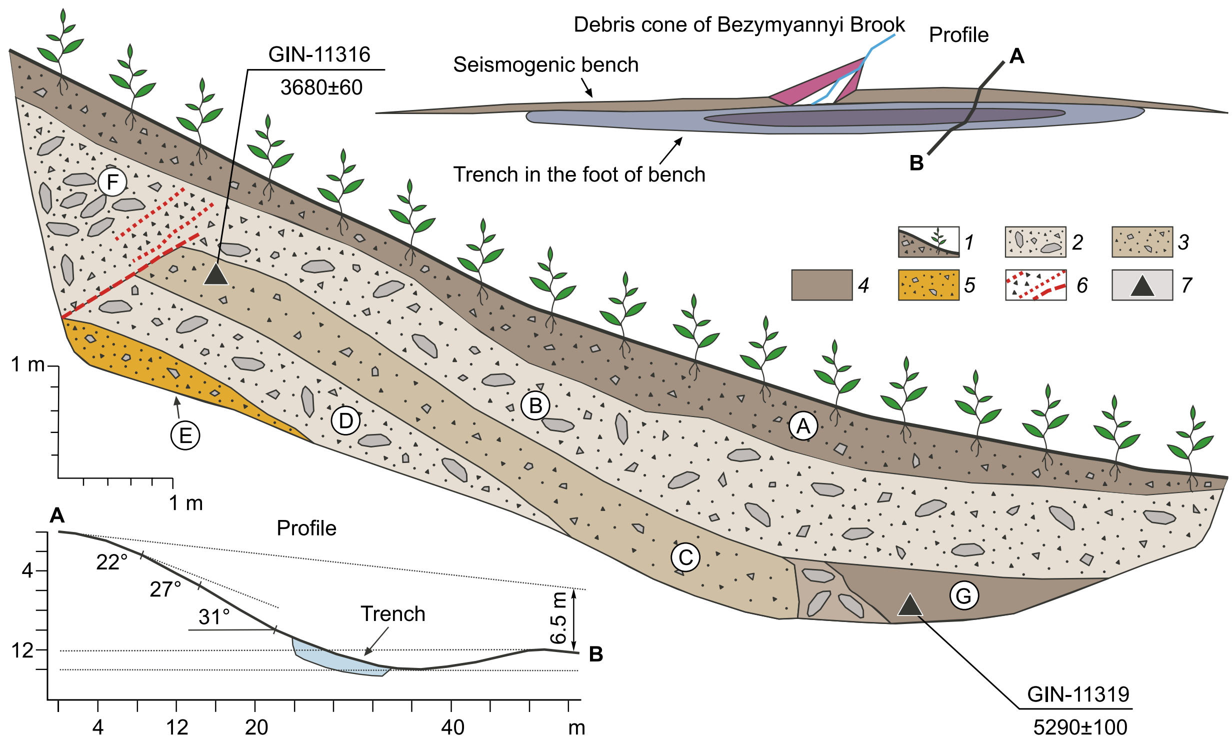

Fig. 5. Cross-section of the northern wall of the trench. The top view: Trench surface plan and its topographic profile relative to the Bezymyannyi Brook erosive cut. The bottom view demonstrates a profile that goes over the seismogenic scarp, and the values of the main morphometric parameters: *1*, ground line and present-time topsoil; *2*, schistose-land-waste-sandy mixture; *3*, land-waste-sandy mixture with inclusions of riprap stones; *4,* humus sabulous-clayey-argillaceous sand deposits; *5*, proluvium: schistose deposits, land-waste, reddish-colored sand; *6*,seismogenic fault zone with the land-waste layer and the contact line across one plane of displacement; *7*, sampling places for radiocarbon dating with age and specimen number indications. The circled letters correspond to the layers indices, used in the article.

file that gives indirect evidence of the deposits accumulation in the already existed trench. Alternation of strata is owing to the periodical change of accumulation modes and sediments distribution. The crushed stony-block layers correspond to the time of a larger material displacement as a result of slope processes activation that had been probably caused by the scarp growth together with the formation of blank surface—the source of colluvium. Accumulation of these layers might have taken place during seismic displacements and for some time after.

Layer C, separating seismogenic deposits, is not so thick when compared to the others. There is no block material in its composition, and it is much darker than underlying and overlapping layers, probably, because of high humus content in the filler. At the bottom of the scarp, the landwaste-schistose layer is in contact with sandy-argillaceous layer G. Layer G is deposited at the bottom of the drainless trench and comprised of a thin material washed out from layers C and D. The well-sorted material of layer G is indicative of its long formation period between seismic events. According to radiocarbon dating, the time interval between seismic events is equal to several thousand years.

The bottom site of the stratigraphic sequence is comprised of reddish land-waste-schistose proluvium (layer E), which differs from the rest part of the formation not only by its color but also by its density and hardness degrees. This layer might be referred to the upper Pleistocene. Before the formation of visible deformations in the cross-section, the stratum superface had been the surface of the alluvial cone.

Based on the proposed deformation analysis, we came to the conclusion that the seismogenic scarp formed as a result of at least two seismic events. A less strong event happened after the burial of the humus layer (C) with the radiocarbon dating of the sample being 3680 ± 60 years (the calibrated age of GIN-11316 determined using the Calib 7.1 software (Stuiver et al., 2017) lies in the interval of 2207–1897 cal BC). The other, stronger event occurred before soil formation in the closed basin of the near-fault trench (layer G, radiocarbon age 5290 ± 100 years, calibrated age 4344–3943 cal BC). In relation to our time, the first earthquake happened not earlier than 4224 years ago, and the second one took place not later than 5960 years ago (the dates were determined by adding 2017 years to the values of the calibrated age). The vertical displacement amplitude of the last paleoearthquake does not exceed 1.0–1.5 m (vertical distance between the subfaces of layers D and B) at the total displacement amplitude of 6.5 m across the relief (see the cross-section in Fig. 5). The displacement amplitude of the previous paleoearthquakes (one or several) equals 4.5 m. The flat upper part of the scarp (22°–27°) supposedly corresponds to them, but the steepest middle part of the scarp (31°) conforms to the last seismic event. The magnitude of the last paleoearthquake could have reached up to 7.3, the number has been derived from the curves of empiric dependence of magnitude on the component of vertical displacement (Strom, 1993; Strom and Nikonov, 1997; Chipizubov, 1998). Being determined by two events, the displacement recurrence interval is less than 2000 years.

Seismological interpretation of geophysical data. To determine the attitude parameters of the deep fault planes, three geophysical profiles (GP) were exposed, their position is shown in Fig. 3. The most informative technique applied on each profile proved to be the electrical-resistivity method with modified axial dipole sounding (ADS). The results obtained for the three profiles are presented below. The cross-sections obtained by CMRW and ground penetrating radar techniques are presented only for GP-1.

The line of GP-1 goes near the trench and crosses the deformed proluvial cone of the Bezymyannyi Brook valley which is the most distinguishable site of the Gusinoe Ozero dislocation. The proluvial cone and thin sediment plume uprising along the valley are composed of boulder-block mass. Being dry, these deposits have higher resistance in comparison to the water-saturated fractured weathered rocks of the valley rock bed. The ADS profile shows a steady increase in thickness of high-resistivity deposits from the ridge side down the slope and sudden growth of thickness under the scarp bottom (Fig. 6, lower profile). The relatively low side of the fault is composed of loose high-resistivity formations along the whole depth of the section.

A narrow low-resistivity zone cuts through the layer of high-resistivity deposits and reaches the surface at the bottom of the seismogenic scarp. The occurrence of this zone is related to the main plane along which the fault sides are displaced. Relatively low apparent resistivity is detected in this zone because of its humidity caused by surface waters accumulated in the closed trench at the bottom of the scarp. The retaining of moisture results in filling the space between blocks with a finely-dispersed substance carried with surface fluids.

The analysis of the apparent resistivity section provides evidence of two outcropping main ruptures. The first one is steeply inclined in direction of the ridge and exposed at the bottom of the seismogenic scarp. The plane zone of the second rupture is less contrasting by its electrical characteristics in comparison with enclosing sediments. It is gently dipping as the first rupture, and they are likely to converge at some depth. The systems of small linear trenches lying parallel to the main scarp could be related to the fault outcrop. The above described fracture found at the northeastern fault flank or a "tensile fracture" is an obvious example. Its detailed parameters are given in the paper (Lastochkin, 1982).

The refracting boundary is well-determined by the velocity profile, with its attitude depth tending to increase gradually in the direction to the depression from 14 m at the left fault edge up to 22 m at the right edge. This refracting boundary is more contrasting (with maximum gradients of velocity variations) at the uplifted side of the fault and it becomes discontinuous in the vicinity of the scarp. This refracting horizon characterizes the contact between unconsolidated sediments and the rock bed.

The cross-section constructed from the results of the radar survey draws the attention to the great number of the subvertical and diagonal transverse contacts mostly inclined in the direction to the depression. They may correspond to the limits of singular cones of fragmentary material in loose

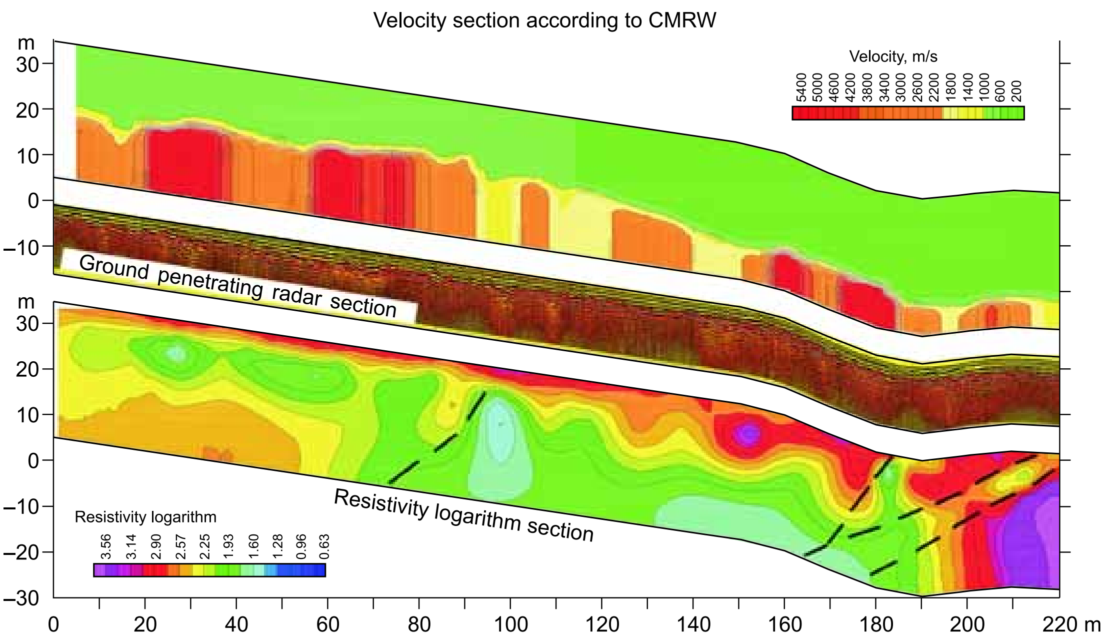

Fig. 6. Interpretive cross-sections obtained by the complex of geophysical methods for PR-1 profile.

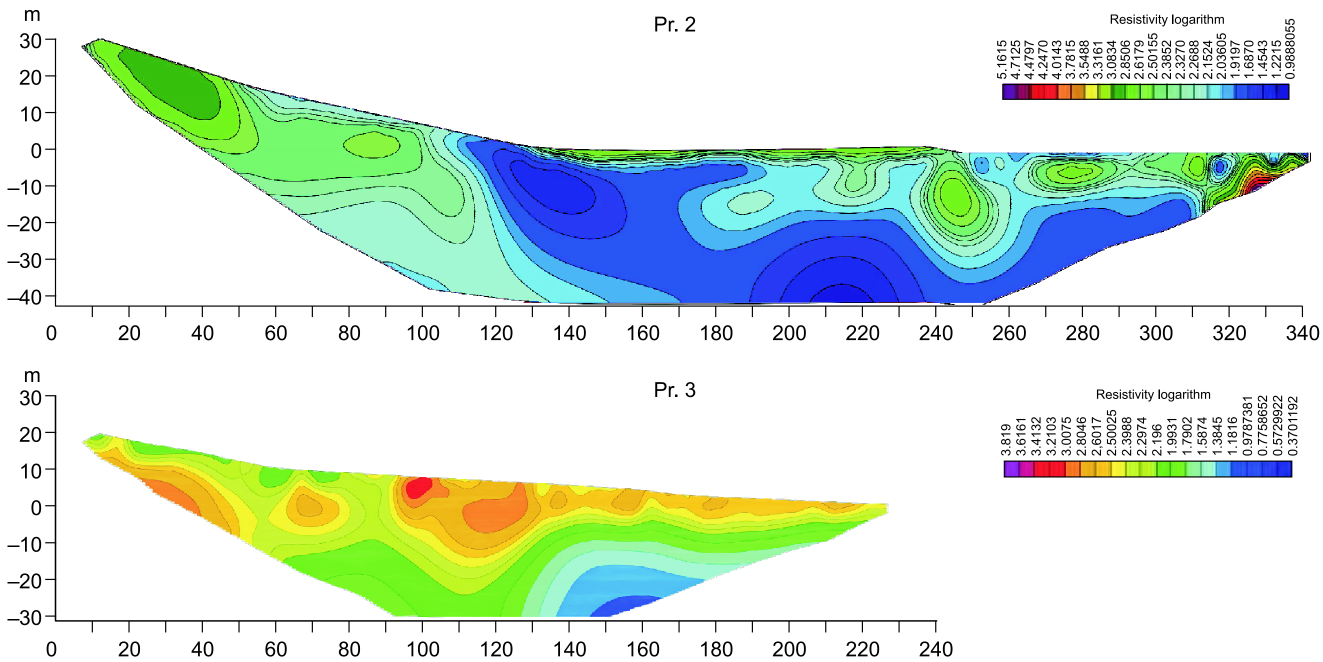

Fig. 7. Interpretive resistivity cross-sections obtained by the modified axial dipole sounding (ADS) method.

proluvial deposits without any direct relation to the deep structure.

A study comprising three methods has been performed on geophysical profiles (GP) 2 and 3. Figure 7 represents only the geoelectrical sections as the most informative and providing basic material for seismological interpretation. Unlike profile 1 obtained across the surface of the deformed deposits cone, GP-2 and GP-3 cover the suture zones of the bedrock slope and piedmont alluvial plain. The left parts of the sections correspond to the uplifted sides of the faults, but the left ones to the downsides. We notice a regularity of the resistivity distribution on both profiles. The resistivity increases with the depth on both profiles for the uplifted fault side. The downside of the fault possesses the opposite characteristics: its apparent resistivity tends to fall with increase in depth. The phenomenon is explained by the proximate position to the watertight stratum (bedrock) on the uplifted fault side. On the contrary, it is related to the dryness of near-surface unconsolidated sedimentary deposits on the low fault side.

On GP-3, the rolling gradient transition region between the values of 1.99 and 2.39 stretches along the whole section being gently inclined in the direction to the ridge. Such clearly defined linear boundaries are not present on GP-2. The zones with maximum gradients either lie close to the surface or delineate local regions with different orientations.

The geophysical cross-sections point at distinct differences in the side structures of the Khambinskii fault. The boundary line between the fault sides is not always clearly delineated because the pattern of geophysical parameters depends not only on the rock composition, but on their porosity, permeability and saturation degrees. Thrust faulting kinematics of the Gusinoe Ozero PSS can be confirmed with the use of GP-1 and GP-3. It is difficult to take a straightforward decision about the direction of the fault plane dip.

ORONGOI PSS

The Orongoi PSS comprises the seismogenic fault recently found and investigated on the border between the Lower Orongoi depression and the Slyudinskaya interdepression junction. The fault is extending northeast, but closer to the southeast extremity becomes sublatitudinal and disappears under the sediments of the Upper Orongoi depression (Fig. 2), which is why this fault can be considered as one of the Khambinskii fault branches. The seismic dislocation is represented with a scarp at the bottom of the mountain slope; a rupture and a vertical/horizontal displacement of the smoothed debris cone surfaces in deep erosion trench cuts. The length of the surface dislocation is at least 5 km. In a plane view, the fault line is not straight and designated with the sequence of arches or directed lines turned with a convex side towards the depression to be the evidence of a thrust with the dip of the fracture plane eastwards from the depression. The ends of each arch are aligned with the adjacent erosion cuts. One of these arches is displaced towards the depression against the intermediate axial line of the structure. At this site the uplifted fault side is complicated by a secondary fracture with the vertical displacement amplitude of 0.4–0.5 m (see profiles 10 and 11, Fig. 8). This secondary fracture is parallel to the main fault scarp and may be a back arc of a wedge-shape block pressed out under conditions of lateral compression. The mechanism of the

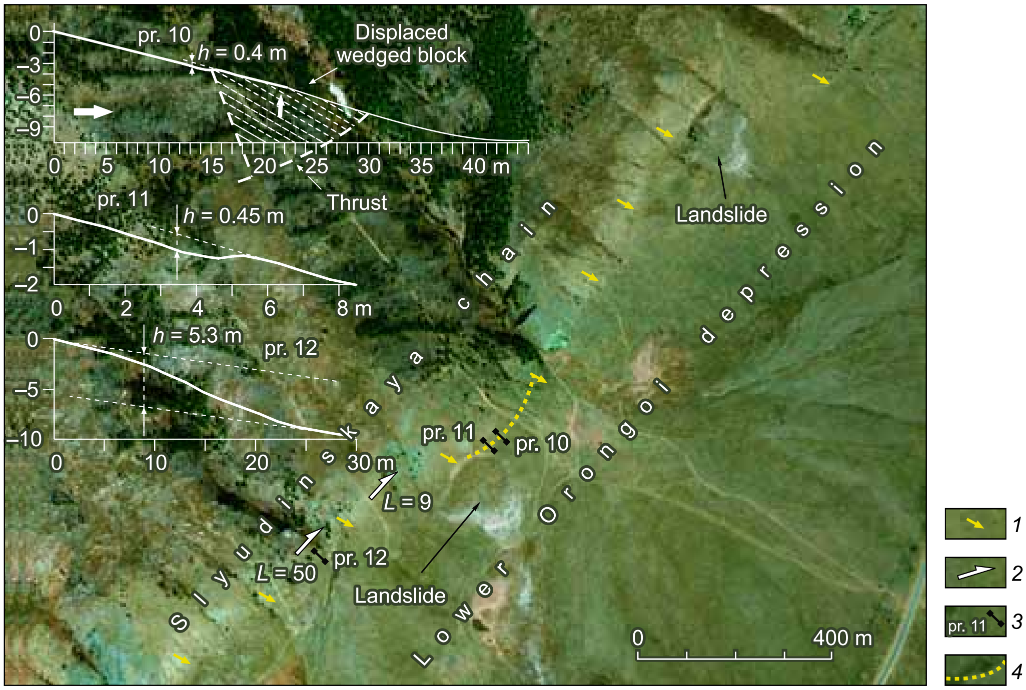

Fig. 8. Geophysical mapping of the northeastern flank of the Khambinskii fault (upper ellipse in Fig. 2). *1,* main dislocation line; *2*, near the mouth fault deformations with direction and displacement amplitude indications, m; *3*, topographic profiles position; *4*, fault on the uplifted side. The relief image is retrieved from: /landsatlook.usgs.gov/viewer.

wedges displacement at the bottom of the slope is attributed to the wide spread of fractures with the southwestern dip in the bedrocks.

The indicators of the sinistral displacement are defined in the mouth reach of two erosion trough cuts. One of the cuts (labeled with a black arrow with index L = 50 in Fig. 8) has a knee of the right (southern) side (in accord with the mechanism of dextral slip), which results in asymmetric expansion of the valley with preserved relics on the postmature surface of the debris cone. In conditions of the dextral slip on the fault, the scarp of the left side became a barrier for substance ablation along the valley, then eroded and smoothed. At present, the left side of the valley is almost linear at the place where the fault crosses it. The amplitude of the dextral displacement is 50 m at the vertical amplitude of 5.3 m (Fig. 8, profile 12). The latter value is a bit lower than that of the vertical displacement of Bezymyannyi Brook cone surface.

The analogous deformation of the relict surface of the debris cone is observed in the mouth reach of the second trough cut at a distance of 200 m to the northwest. However, the amplitude of the vertical dextral displacement is much lower and equals to 9 m. The difference in amplitudes of horizontal displacement is primarily due to the age of the erosive valleys. The first valley is older than the second one, and it is expressed in their length. The first one is 1400 m long, but the second valley reaches only 320 m in length.

The Orongoi structure is represented by rupturing linear deformations. Besides, it is comprised of small rockslides in proluvial-deluvial deposits of the piedmont plain. Lateral dimensions of the landslides reach up to 130–150 m. The break-away walls are on the same line with the seismogenic scarp at the bottom of the slope and distinguished from it by a steep dip angle, while their height does not exceed several meters.

Activity of other segments of the northern termination of the Khambinskii fault has been demonstrated by the linear contacts of the Neocene deposits and Proterozoic granitoids from the Khamar-Daban Range, found on the northern side of the Upper Orongoi depression, where the fault is well defined in the relief and changes its direction in the strike from latitudinal to east–northeastern. At present, Neocene deposits are involved in the uplift of the Khamar-Daban piedmont in the left bank area of the Orongoi River. The Neocene is characteristic of the southern side of the Upper Orongoi depression. Within the inner limits of the depression, small-area hills and pyramidal inselbergs are observed lying on the Jurassic–Cretaceous bottom of the depression.

The occurrence of formerly thick Neocene deposits in the Upper Orongoi and Gusinoe Ozero depressions and their absence in the Lower Orongoi and Ivolga depressions indicate the existence of a tectonic barrier in the pre-Quaternary age. The barrier was *in situ* the current junction between the Orongoi depressions, which takes the form of the Slyudinskaya chain being the northeast extension of the Monostoi

Range. At present the chain is cut through with the narrow valley of the Orongoi River. The bottom of the northeastern slope of this chain extends along the contact of the Mesozoic–Cenozoic deposits of the Upper Orongoi depression and Proterozoic granitoids from the Monostoi Range.

The fault is well defined in the geological substrate through the wide spread of tectonites. Consequently, the Upper Orongoi depression appears to be an active structure with the potential of realizing tectonic slips on the faults in several directions, where the northeastern direction appears to be the main one. However, the sinistral fault displacement has been determined for this plane, which contradicts our conclusions on the kinematic type of the Orongoi dislocation. The reverse direction of the horizontal slip in the focal mechanism is likely to point at partial return of the geological environment to the initial state in the time of weak and relatively moderate earthquakes (with $M \leq 5$), as contrasted with the strongest ones.

HISTORICAL AND INSTRUMENTAL SEISMICITY

Historical data given in annals and catalogues do not allow locating earthquake epicenters in sparsely populated areas with acceptable accuracy, and the earthquakes are usually assigned to the nearest seismogenic structure. From all the historical earthquakes recorded in the territory of Selenga Dauria (Bazarov, 1968) only two (of 1856 and of 1885) could be referred to the Gusinoe Ozero segment of the Khambinskii fault in the Dzhida–Vitim suture zone. The most complete data on the macroseismic effects of these earthquakes are given in the catalogue compiled by I.V. Mushketov and A.P. Orlov (1893).

Analysis of the macroseismic effects of the earthquake that occurred November 8 (20) 1885, was done by different authors and predictably resulted in different estimates of the epicenter location and earthquake energy (Fig. 9). According to A.V. Chipizubov, the earthquake source must have been referred to one of the faults of the Dzhida–Vitim suture zone being the most active structure in the region. In compliance with the parameters obtained from macroseismic data (Chipizubov, 2016), the magnitude of the earthquake lies in the range of $M = 6.4 \pm 0.3$, at which intensity of 7–8 are possible at a distance up to 40 km.

The earthquake that occurred on May 11 (22), 1856 was maximally felt in Kyakhta and Selenginsk. The epicenter correlation with one these settlements presupposes magnitude overestimation due to the increase of the shocks radius to measure intensity of 5–6. If the epicenter is placed in the middle of the segment connecting these two settlements, the radius of strong shocks will decrease up to 40 km, and the value of the magnitude will be $M = 5$ (Chipizubov, 2009). Taking into account the uncertainty of the epicenter position

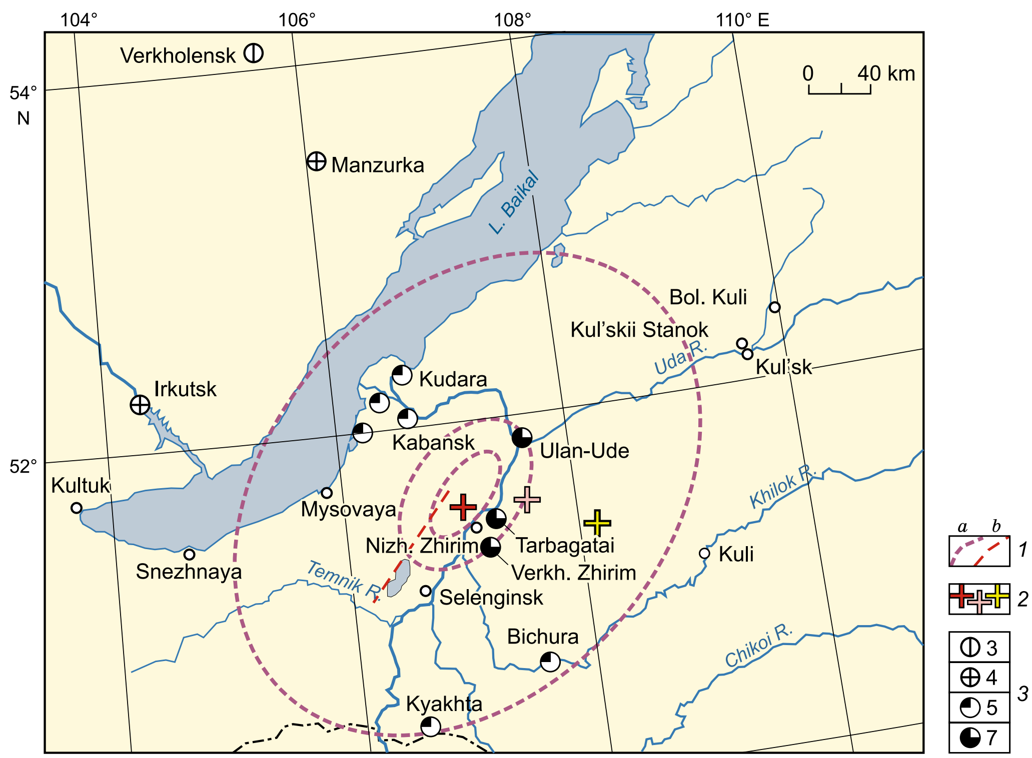

Fig. 9. Isoseist diagram of the November 8 (20), 1885 earthquake according to (Chipizubov, 2016). *1*, theoretical isoseists of 8, 7, and 5-intensity effects for the earthquake with $M = 6.5$ (*a*) and the northeastern boundary of the Gusinoe Ozero depressions system (*b*); *2*, earthquake epicenters according to different informational sources: yellow, (The New Catalogue …, 1977), (electronic catalogues: Catalogue of Earthquakes of Northern Eurasia (CENE) and Special Catalogue of Earthquakes of Northern Eurasia), pink, Earthquake Catalogue of the Baikal Zone, red, (Chipizubov, 2016); *3*, sites, figures, points.

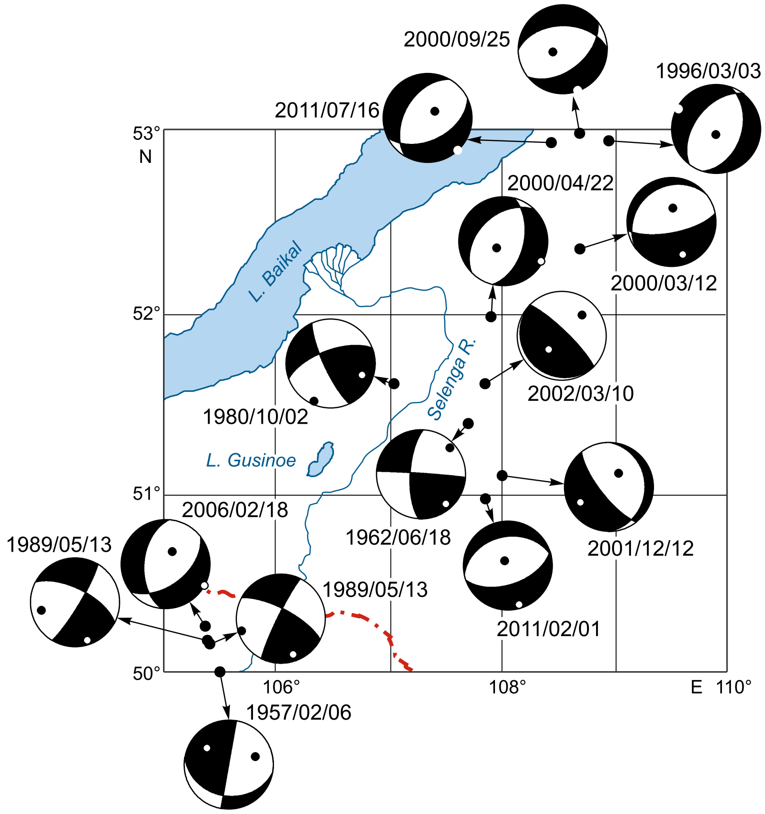

Fig. 10. Focal mechanisms (lower hemisphere) of the earthquakes of western Transbaikalia. The white dots mark tension axis projections on the sphere; the black dots—compression axis projections on the sphere (according to (Solonenko et al., 1993; Seredkina, Mel'nikova, 2013; Radziminovich et al., 2016)), Earthquakes of Northern Eurasia (Annual Materials of FRC UGS RAS) and Earthquakes of Russia, Global Centroid Moment Tensor Program (http://www.globalcmt.org).

and relatively moderate magnitude, this earthquake could be genetically assigned to one of three active fault zones: Khambinskii, Kizhinga, Selenga–Khilok (Fig. 2). According to (The new catalogue…, 1977) and the electronic Catalogue of earthquakes of Northern Eurasia (CENE), the epicenters of other two historical earthquakes (May 21, 1865 and April 11, 1856) with magnitudes $M = 5.5$ are located northwest from the Gusinoe Ozero depression closer the axial line of the Khamar-Daban Range.

The epicenters of instrumentally recorded earthquakes of western Transbaikalia are scattered around the area forming somewhere isometric clusters (Fig. 2, northwestern flank of the Upper Orongoi depression). On October 2, 1980, one of the strongest earthquakes in western Transbaikalia took place on the southern slope of the Slyudinskaya chain separating the Orongoi depressions from the Ivolga depression. Its magnitude was close to 5 ($K = 13$), and the intensity of shaking in the epicenter reached 7 on the MSK-64 scale. A weak foreshock ($K = 7$) preceded the earthquake. Few aftershocks were recorded after the main one.

Being instrumentally detected, the earthquake epicenter is maximally approximated to the Orongoi PSS. When being determined by macroseismic data, the epicenter is located southwards almost in the center of the segment connecting the instrumental epicenter and the Orongoi Village. In accordance with an isoseist diagram (Golenetskii et al., 1982), isoseismal areas of 5-6-intensity are represented with ellipsoids whose long axis is oriented in latitudinal direction, i.e., at significant angle with the Khambinskii fault strike. The mechanism of the earthquake source was determined by the standard method using polarity of *P*-wave arrivals. The data on polarity were processed when being collected from 39 stations of Baikal, Altai–Sayan and Mongolian networks and also from remote stations. Since the nodal planes have been well constrained, the determination of the mechanism is accepted as reliable. The earthquake was caused by strike-slip displacement under the subhorizontal compression and extension axes, which were oriented in SW–NE and SE–NW directions, respectively. Based on the epicenter location and seismic intensity distribution, the earthquake was determined to be caused by the slip on one of the faults of the interdepression junction.

DISCUSSION

There exist different views on the fault kinematics in the framing of the Gusinoe Ozero depression (Danilovich, 1949; Florensov, 1965; Lastochkin, 1982; Lunina and Gladkov, 2009; and others). One of the latest studies points at the dominating role of normal fault deformations in the depression's structure, and thrust deformations are interpreted as the effects of horizontal slips (Lunina, 2016). The earlier studies put forward the main argument in favor of normal fault kinematics, i.e., the occurrence of the trench at the bottom of the scarp and the proluvial cone of the Bezymyannyi Brook valley (Solonenko, 1968; Lastochkin, 1982). However, some factors are in favor of its uniqueness: negligible length of the trench and the absence of similar negative topographic forms at other sites of the Gusinoe Ozero PSS, including seismic dislocations in Transbaikalia and northern Mongolia.

Over the recent decade, investigators have obtained data on the thrust mechanism of seismogenic dislocations formation in the territory of Southwestern Transbaikalia and northern Mongolia within the limits of the Hentiyn-Daur megadome (Demberel et al., 2010; Imaev et al., 2012; Ferry et al., 2010; Dujardin et al., 2014; Smekalin et al., 2016; and others). Most seismogenic dislocations of thrust genesis are confined to the side bottoms of the Mesozoic–Cenozoic depressions; that appears to be indicative of their development in conditions of subhorizontal compression. These conditions are markedly different from those ones for embedding and intensive development of the depressions in the middle and late Mesozoic.

Geological data indicate that new structures formed mainly in conditions of horizontal compression. Firstly, the sedimentation area sharply reduced in the Cenozoic compared with the Mesozoic in Transbaikalia and the adjacent territory of northern and central Mongolia. In particular, it concerns the external framing of the Hentiyn-Daur megadome (for instance, ridges and highlands lying southwards and southeastwards next to Hentiyn, spurs of the Malkhan Range, etc.). The stripes of the Mesozoic deposits (no more than a dozen kilometers wide) are uplifted above current bottoms of the depressions. The difference in altitude for the Mesozoic deposits across the depressions sides and their bottoms reaches 100 m and more. In the Mesozoic–Cenozoic time, reduction of the depression areas was caused by the side uplift; in turn, it simultaneously led to decrease in the basement subsidence rate. The negligible thickness of Cenozoic deposits (no more than a few dozens of meters) validates this hypothesis compared with thickness of the Cretaceous and Jurassic deposits ranging from 1000 to 2000–2500 m.

Secondly, the Neocene sandstones, gravel-stones, conglomerates that partially cover the sides, and hills-inselbergs on the bottoms of the Gusinoe Ozero and Upper Orongoi depressions accumulated under conditions of intermountain closed basins. The cause of the basins formation was not the subsidence of the depressions bottoms, but rapid growth of interdepressions junctions, which led to the appearance of everlasting tectonic dams. One of such uplifts was the Slyudinskaya interdepression junction between the Upper Orongoi and Lower Orongoi depressions, which turned into an ablation obstacle for of the weak material from the Gusinoe Ozero and Upper Orongoi depressions. In the early Quaternary—the time when the Orongoi River cut through the junction—the Neocene deposits had been eroded before the Cretaceous and Jurassic rocks. This fact is indicative of the positive effect of vertical movements on the Cenozoic sedimentation process.

Thrust kinematics of the faults at the bottom of the depression sides and inducing conditions of subhorizontal compression are often not consistent with seismic data. There is a small number of focal solutions for Transbaikalia (Fig. 10). Their analysis reveals that they are different for the area to the southeast from the Baikal rift zone. Here, there are mechanisms both so-called Baikalian-type, which are normal faulting on planes oriented in SW–NE, and strike-slip and reverse faulting. The compression axis in the earthquake sources is mainly oriented in SW–NE direction, and correspondingly, the axis of tension has SE-NW orientation. Such an orientation results in sinistral slips on the planes with the latitudinal strike, and dextral diplacements on planes oriented in meridian direction. All three strongest earthquakes (Kyakhta (February 6, 1957, M = 6.4; May 13, 1989, M = 5.8), Orongoi) took place due to the action of compression and tension subhorizontal axes. Nearly all the studied dislocations within the borders of Hentiyn-Daur (Gunzhin, Khustai, Avdar, Sharkhai, Mogod, Kerulen, and others) appear to be either thrust faults or strike-slip faults or their combination (Demberel et al., 2010; Ferry et al., 2010; Imaev et al., 2012; Al-Ashkar et al., 2013; Dujardin et al., 2014; Smekalin et al., 2016; and others). At present the Gusinoe Ozero and Orongoi PSSs could be on the list of the structures whose thrust with strike-slip kinematic has been determined by the indicators set as a result of morphometric, trenching and geophysical investigations.

CONCLUSIONS

The latest seismological investigations in the Khambinskii fault zone have allowed obtaining new data on the parameters of the Gusinoe Ozero and Orongoi PSSs. Absolute values of interval ages for two Holocene paleoearthquakes have been determined. The upper limit for the time-interval of the first earthquake is 5960 years. The second temporally closest seismic event happened not earlier than 4224 years ago.

When determined for the first time, the kinematic parameters of the Orongoi PSS have exposed 5 km-segment of one of the northeastern flank branches of the Khambinskii fault. The amplitudes of right-lateral strike-slip of the heterochronous erosion valleys, which are crossed with the fault, are equal to 50 and 9 m. The maximum amplitude of the thrust (vertical) fault component is 5.3 m. The epicenter of the Orongoi earthquake (October 2, 1980, M = 5) is in close proximity to the exposed fault fragment. The presented paleoseismic, historical and seismological data indicate high seismic activity of the tectonic faults on the western side of the Gusinoe Ozero–Orongoi depression system. In consideration of the seismic potential for the Khambinskii PES zone, the obtained data highlight the necessity to estimate seismic hazards to Ulan-Ude and settlements situated within the limits of these depressions and their surroundings.

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