Cryoturbation versus tectonic deformation along the southern edge of the Tunka Basin (Baikal Rift System), Siberia: New insights from an integrated morphotectonic and stratigraphic study

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AbstractThe Tunka Basin is a broad, emerging basin situated between the Baikal Lake to the east and the the Hövsgöl Lake to the west. The basin is bounded to the north and to the south by the Tunka and the Khamar-Daban mountain ranges, respectively. The Tunka normal fault, located at the southern foothills of the Tunka mountain range, is the main structure that controlled the development of the Tunka Basin during the Neogene. Paleoearthquake-surface ruptures attest of its present activity; and show that its western and eastern terminations are undergoing a tectonic inversion characterized by left-lateral-reverse deformations. The southern edge of the Tunka Basin is classically interpreted as being tectonically controlled. In this paper, we present the results of a geomorphological and stratigraphic analysis within its southwestern and southeastern parts suggesting that there is no active fault affecting the foothills of the Khamar-Daban mountain range. The different features observed in the Quaternary deposits are interpreted to be the result of periglacial processes induced by alternating episodes of permafrost aggradation and degradation during the Holocene. Our study concludes that the Khamar-Daban Range and the Tunka Basin are uplifting together, and that the Tunka and Mondy faults are the two main triggers of regional earthquakes.

Keywords: Baikal Rift System, Tunnka Basin, Quaternary deposits, morphological and stratigraphic analyses, periglacial processes, cryoturbation

1. Introduction

The Tunka Basin is located in the southwestern part of the Baikal Rift System. As many of the Baikal Rift basins, it shows a morphological asymmetry characterized by a sharp and steep border on the northern side, and a smooth morphology on the southern one (Logatchev, 1974; 2003). To the North, the south-dipping Tunka normal fault, located at the foothills of the Tunka mountain range (Fig. 1), corresponds to the main structure that controlled the development of the Tunka Basin (Sherman et al., 1973). Along this fault, Quaternary landforms and sediments are affected by paleoearthquake surface ruptures attesting of its present activity (e.g. Khromovskikh et al., 1975; McCalpin and Khromovskikh, 1995; Chipizubov et al., 2003; Smekalin, 2008; Smekalin et al., 2013; Ritz et al., 2018; Arzhannikova et al., 2018).

Regarding the southern edge of the Tunka Basin, the question of a tectonic control with a south Tunka fault located at the foothills of the Khamar-Daban mountain range (Fig.1), presently active, remains a major question notably in terms of seismic hazard. Several studies propose the occurrence of one or several active faults (e.g. Sherman et al., 1973; Lukina, 1989; Lunina and Gladkov, 2004), potentially the source of large historical earthquakes (Lukina, 1989, Radziminovich and Shchetnikov, 2013). However, there are no clear morphotectonic and/or paleoseismological observations attesting to those structures.

During the last glacial period, the late Quaternary sediments of the Tunka Basin are affected by intense and widespread cryoturbation processes due to the periglacial environment close to large mountain ice caps (Alexeev et al., 2014). These processes have induced surficial deformation that has been interpreted as seismites (Glagkov and Lunina, 2010). In such environments, the question of distinguishing deformations caused by tectonic processes from those induced by periglacial processes is a major concern (e.g. Audemard and de Santis, 1991; Mc Calpin, 1996; Van Vliet-Lanoe et al., 2004; Obermeier et al., 2005; Baize et al., 2007).

In this paper, we present the results of a detailed morphotectonic and stratigraphic analysis along the southern edge of the Tunka Basin. Our observations allow: 1) describing several features of particular relevance for discussing their origin (i.e. seismogenic vs cryogenic origin); and 2) discussing whether or not there is an active fault bounding the Tunka Basin to the south. Our results have implications about the recent geodynamic evolution of the Tunka Rift, and the regional seismic hazard.

2. Tectonic setting

The Tunka Basin is a broad E-W trending emerged basin (200 km long, 10-30 km wide), connecting the Baikal Basin to the NE in Russia, and the Hövsgöl Basin to the SW in Mongolia. Along with the other basins of the Baikal Rift System, the Tunka Basin defines the northwestern border between the Amurian Plate and the Siberian Craton (Petit and Fournier, 2005; Petit and Déverchère, 2006). The Tunka Basin is bounded by the Tunka and Khamar-Daban mountain ranges to the north and the south, respectively. These ranges differ in their geomorphological characteristics. The Tunka Range shows alpine-type reliefs with narrow ridges and peaks incised by glacial cirques and valleys. The highest summits reach more than 3000 m asl, and are located less than 10 km from the southern foothills. The reliefs in the Khamar-Daban Range are smooth and present summit plateaus which an altitude of 500 to 700 m lower than the Tunka summits. The northern slope of the Khamar-Daban Range dips gently to the Tunka Basin, with highest reliefs 20-25 km southwards from the foothills.

The Tunka Basin shows a complex inner structure with smaller depressions (from west to east: Mondy, Khoytogol, Turan, Tunka *sensus stricto* and Tory) separated by interbasin highs (Nilovsky and Elovsky spurs) (Fig. 1b). These depressions differ in size and depth of basement, with the greatest depths situated northward against the Tunka Fault (Sherman et al., 1973). The Tunka depression, located in the centre of the Tunka Basin, is the largest in size and sediments thickness. The mean topographic surface of the Tunka Basin dips towards the east from 1350 m asl to 650 m asl, and then shows a 700 m difference of height over the 170 km length of the basin (Fig. 1b).

The Tunka Basin is filled with Oligocene to Quaternary deposits up to 3 km thick. The drilling data show that the lower part corresponds to Oligocene–early Pliocene fine-grained volcano-sedimentary rocks, while the upper part corresponds to late Pliocene–

Quaternary coarse-grained sediments (Sherman et al., 1973; Mazilov et al., 1993). This difference in sedimentation has been interpreted as markers of a slow then fast stages in the rifting process (e.g. Logatchev and Zorin, 1987; Mats et al., 2001; Krivonogov and Safonova, 2017). The late Pleistocene–Holocene formations are mainly composed of fluvio-glacial and alluvial deposits (boulder-pebbles, gravels, sands) and aeolian sediments (silts, loess) (Ravsky et al., 1964; Shchetnikov et al., 2012). Most of the coarse-grained material is accumulated within the Tunka Range piedmont, while the fine-grained sediments are drifted to the inner parts of the depressions. The lower areas are covered by numerous lakes and swamps due to a shallow permafrost that can be up to hundreds of meters thick (Logatchev, 1974). The Tunka Basin and the Tunka Range still show features associated with the Late Pleistocene glaciation such as glacial valleys, cirques, moraines and kame terraces. During the last glaciation, a 300 m thick glacier occupied the westernmost part of the Tunka Basin (i.e. Mondy Valley) (Olyunin, 1965). In situ-produced ^{10}Be dating of Mondy terminal-moraine yielded an average exposure age of 14 ka (Arzhannikov et al., 2015). During the last glaciation, an intense aeolian activity occurred and was associated with the deposition of loess, sandy loess and the formation of sand dunes and ridge-and-runnel topographic features (Ravsky et al., 1964). According to Shchetnikov et al. (2012), the maximum period of aeolian sedimentation in the Tunka rift basin occurred at the end of the Late Pleistocene and during the Early Holocene.

Cryogenic deformations such as ice wedges and other cryoturbation features were observed in the cross sections within the Upper Holocene alluvial sediments (Alexeev et al., 2014). Solifluction phenomena due to thawing permafrost are commonly observed in the superficial deposits (Ravsky et al., 1964).

Nowadays, the Tunka Basin corresponds to an emerged rift structure. It is located within a transition zone between the present-day compressional deformation of North Mongolia and the extensional one of the central Baikal Rift as attested by the focal mechanisms of earthquakes (Melnikova and Radziminovich, 1998; Radziminovich et al., 2016). Morphotectonic and paleoseismic analyses show that the kinematics of faults controlling the Tunka Basin changed through time. From the Miocene to the Quaternary a transtensional regime dominated, with normal and left-lateral strike-slip faulting along the major fault zones, allowing the opening of basins and thick accumulation of loose sediments (San'kov et al., 1997). Then the strain regime was replaced by a transpressional regime in the Late Quaternary combining compressional and left-lateral strike-slip fault deformations (Larroque et al., 2001; Arzhannikova et al., 2005; Arjannikova et al., 2004; Arzhannikova, Melnikova, Radziminovich, 2007; Jolivet et al., 2013; Ritz et al., 2018). This recent transpressional strain regime would be the result of the northwards propagation of the collisional deformation associated with the convergence between India and Asia (Ritz et al., 2000). Evidence of recent inversion was observed along the Mondy Fault and its western extension northward of the Hövsgöl Lake (Larroque et al., 2001; Arzhannikova et al., 2003; Arjannikova et al., 2004) and along the East Tunka Fault and East Sayan Fault (Chipizubov et al., 2003; Smekalin, 2008; Shchetnikov, 2016; Ritz et al., 2018; Arzhannikov et al., 2018).

The question of an active fault located at the southern edge of the Tunka Basin remains undetermined. From geomorphological and structural considerations Sherman (1973) has proposed a fault bounding the Turan and Tunka depressions to the south at the foothills of Khamar-Daban Range (Fig. 1b). Lukina (1989) and Lunina and Gladkov (2004) extended this fault until the Baikal Lake. However, no evidence of deformations in the Upper Pleistocene-Holocene deposits has been described along this hypothetic structure. At the western end of the basin, evidences of tectonic scarps associated to the Mondy Fault are observed at the southern foothills of the Nilovsky Spur until its eastern limit (Geological map-East Sayan series, 1961; Arjannikova et al., 2004) (Fig. 1b). Moreover, two historical earthquakes were felt within this area in 1814 (Shimki Village, I=IX (MSK)) and in 1829 (Turan and Shimki villages I=VII-VIII, Mondy Village I=IX) (Kondorskaya and Shebalin, 1977) (Fig. 1a). After Lukina (1989) and Radziminovich and Shchetnikov (2013), these historical events were produced by rupture of the western part of the South Tunka Fault or by the Mondy Fault. Currently, the instrumental seismicity is much active in this western part than in the eastern part of the Tunka Basin (e.g. Solonenko et al., 1997; Radziminovich et al., 2013) (Fig. 1a). Among these instrumental earthquakes two events of magnitudes M4.5 and M5 occurred near Kyren Village in 1958 (Golenetsky, 1998).

3. Methods

In our study, we used morphotectonic and stratigraphic analyses. Morphotectonic analyses were based at first on a remote sensing study of Google Earth satellite images, aerial photographs and TanDEM-X digital elevation models with resolution of 12 m/pixel, of the foothills of the Khamar-Daban Range to determine whether or not there is an active fault along the southern margin of the Tunka Basin. These remote sensing analyses were then completed by field investigations within locations where the occurrence of morphotectonic markers were in question. Those included the detailed analysis of the topography and the morphology of the markers, as well as the analysis of the stratigraphy of deposits and their potential deformations through trenches studies. Age constraints of stratigraphic units were provided by the radiocarbon dating of samples containing organic matter. The radiocarbon samples were analyzed at the French National Platform LMC14 of the National Service of INSU “ARTEMIS”. Dendrochronologically calibrated calendar age ranges were calculated using the program Calib 7.1 Radiocarbon Calibration (Stuiver et al., 2019) with 2 standard deviations uncertainty. The age ranges are rounded off to the nearest decade.

4. Geomorphology, Quaternary sediments and structures

4.1 Kyren site

4.1.1 Geomorphology of the Kyren area

The area studied near Kyren is located within the central part of the Tunka Basin at the foothills of the Khamar-Daban Range (Fig. 2). Contrary to what is observed along the Tunka Fault, bounding the basin to the north, this southern boundary corresponds to a smooth slope comprised between 800 and 900 m in altitude. Topographic profiles performed across this morphological feature do not show evidence of scarp, but instead a smooth regular convex profile (Fig. 2d, profile 1). Assuming that a possible fault was running along this inflexion line, we carefully analysed the morphology of the three largest alluvial fans at the outlets of the Haragun, Haribiati and Algak-Kyren rivers, searching for a recent fault scarp. Detailed field observation of the outlet areas did not reveal any abnormal topographic variation, as shown by profiles 2 and 3 (Fig. 2d). Moreover, we did not observe uplifted or even clearly stepped fluvial terraces within the drainage network inside the Khamar-Daban Range. On the contrary, the longitudinal incision profiles of the various rivers are continuous throughout the different sedimentary depressions in the Tunka Basin.

However, on an aerial picture, we have been puzzled by an E-W linear feature affecting the Holocene deposits forming the plain at the east of the right bank of the Algak River, to the south of Kyren Village (Fig. 3a, see Fig. 2a for location). In a satellite image, this feature is much less obvious, and seems to underline the northern edge of a ridge-and-runnel topography area (Fig. 3b). The satellite image also allowed pointing out some smaller W-E aligned “dome structures” (Figure 3b). To decipher whether these structures are tectonic features and could correspond to the hypothetic south Tunka fault or not, we carried out field works including trenching throughout the W-E aligned break in the slope profile and two “dome structures”. These investigations where completed by the study of the sedimentary section located on the right bank of the Algak River (Figure 3b), where fluvial erosion has produced a 6 m high sub-vertical cliff in a meander concavity.

4.1.2- Stratigraphy of Quaternary deposits: the Algak River bank section

From the base to the top, the Algak River bank section exposes fluvial, then fluvio-aeolian and aeolian deposits affected by a thick deformation horizon in their upper 1.5 m (Fig. 4). From the base to the top, the Algak River profile can be subdivided into three sub-sequences (SS1 to SS3) overlying the uppermost part of a peat layer occurring at -6.1 m at the level of the present day River (Fig. 4):

1) SS1 sub-sequence (-6.10 to - 4.8 m) is first made by finely laminated greyish sandy to silty deposits corresponding to fluvial overbank deposits including thin organic debris layers. In the upper two third the sedimentation becomes more and more organic, two peat layers occur, and numerous well-preserved wood pieces, trunk remains, and even some *in situ* tree stumps are observed. Radiocarbon ages obtained from two wood samples show that this subsequence starts at the end of the Late glacial and that the tree stump horizon, associated with the uppermost peat layer, dates from the Early Holocene (Preboreal). Some involutions within unit 15 can be related to syngenetic freeze-thaw processes (deep seasonal frost) but no permafrost related features have been observed in the sediments deposited during this period. Without any palynological analysis it is difficult to describe more precisely the landscape but the large size of the wood remains and tree stumps recovered in this unit (> 20 cm in diameter) seems to indicate a forested taiga environment, at least developed in the alluvial plain, as described by Groisman et al. (2013) reporting the occurrence of “spruce and larch forests along the rivers”.

2) SS2 sub-sequence (-4.8 to -1.5 m) is made by three metres of laminated sandy silts including a thin (10 cm thick) dark grey organic layer (marshy soil, unit 9) and some thin

syngenetic frost cracks at the base in unit 10 (depth: 0.2-0.3 m). Organic debris is still present but definitely less abundant compared to SS1. This part of the sequence results from a period of reactivation of fluvial activity in a more open (and colder?) environment in which aeolian processes are more and more intense to the top. By reference to the synthesis of Groisman et al. (2013) on the Holocene climate and environments in Siberia, deposits corresponding to the Boreal and Atlantic periods, and corresponding to the most forested period of the Holocene (until about 6 000 BP), should be represented here by organic rich deposits including a lot of wood and tree remains. According to the facies of SS2, the Boreal and Atlantic periods are clearly lacking in the Algak River profile likely owing to the discontinuous character of fluvial sedimentation.

3) SS3 sub-sequence (- 1.5 to 0) is made by a succession of highly deformed beds of sandy silts and organic sandy silts layers (unit 5), then by homogeneous sandy deposits covered by a thin humic top soil horizon. This part of the sequence clearly shows the succession of at least two different events of aeolian sandy silt deposition alternating with episodes of intense deformation (load casting and diapiric features). According to the local climatic context (sporadic permafrost), these deformations are interpreted as cryoturbation features developed in the active layer of a permafrost by periglacial loading (Harris et al., 2000) or differential frost heave (Van Vliet Lanoë, 2004). Other outcrops observed along the banks of the Algak River show the large extension of this thick and complex deformation horizon in the Kyren area.

According to the study of Algak River bank section it is clear that no other periglacial horizon with large involutions is present below 1.5 m from the surface. This observation is in good accordance with another study made along the Irkut River banks where a thick layer of 1.5 to 2 m of highly deformed humic silts and sands associated with ice wedge pseudomorphs has been evidenced along several kilometres (Alexeev et al., 2014). In this study the cryoturbated organic rich layers located between - 40 and -100 cm from the surface have been dated by radiocarbon analysis from 2 600 and 3 600 BP respectively (Alexeev et al., 2014). They correspond to a period of climatic instability starting with a very strong cooling event around 4 500 BP followed by the alternation of several phases of climatic cooling and warming during the Subboreal (Groisman et al., 2013). According to the close correspondence between the thickness and the structure of the deformed layers in both the Upper part of the Irkut and Algak rivers banks sections it is proposed that the deformed layers of the Algak River profile in the Kyren area belong to the same period (Subboreal).

4.1.3 - The ridge-and-runnel features

4.1.3.1 Morphology In the flat area located to the East of the Algak River, and presently covered by steppe vegetation, the analysis of both Russian aerial photographs and satellite images allows evidencing numerous W-E elongated ridges (100 to 1000 m long) a few meters in high (Fig. 3b). They only occur in this area located to the East of the Algak River and not to the West in the alluvial plain between the Kyren and the Algak rivers. Features of the same type have also been observed further North, near Kyren Village (Fig. 3b), and are largely described in the Tunka Basin (Ravsky et al., 1964; Shchetnikov et al., 2012). They have been interpreted as resulting from aeolian deflation and transport by strong WE winds within the Tunka corridor (Shchetnikov et al., 2012).

This type of morphology is well known in the loess area of Central Europe where they are locally named “Greda” (Léger, 1990). They appear as for example along the right bank of the Danube River where they can reach more than 1 km long and 50 m high (Jipa, 2014). More recently the study of large Greda structures located in western Europe on the edge of the Odenwald plateau, connected to the right bank of the Rhine graben close to Heidelberg (Germany), evidenced the “onion peel” structure of the stratigraphic units making these elongated ridges (Antoine et al., 2001). They have been built by the rapid accumulation (14 m between about 30 and 17 ka (Moine et al., 2017)) of sands and silts deflated then transported as low level suspension from the braided alluvial plain of the Rhine during storms from N-NW. These features, characterized by rapid accumulation of sandy loess showing laminated facies made by the alternation of sandy and silty layers up to 1 cm thick are typical of aeolian deposition in flat surfaces located immediately downwind of important sand and silt sources as braided periglacial rivers (Antoine et al., 2001). In the case of the Kyren area, the source of the silty material was the braided plain of the Irkut River located upwind of the ridge area and where active deflation processes occurred during the end of the Last glacial and the Younger Dryas, according to ^{14}C ages from trench 1 (see below) and in good agreement with the data published by Shchetnikov et al., (2012).

However, some of these small elongated domes (with length comprised between 10 and 25 m) showing steep slopes, likely do not only result of pure aeolian processes. In order to understand the origin of these features we opened three trenches in the plain located at the east of the Algak River. Figure 5 shows the sites where they were opened: the first site (Fig. 5a) corresponds to an E-W smooth topographic scarp across which we opened trench 1 (Fig. 6). The second site (Figs. 5b and 5c) corresponds to E-W “oval dome structures”. There, we opened two trenches (Figs. 7 and 8, respectively).

4.1.3.2 - Stratigraphy and deformations

Within trench 1, we observed stratified (laminated?) silty-loess units and thin organic-rich paleosols affected by ductile deformations (Figs. 6a, b). In the southern part of the trench, the units are upturned along a 30-40 cm thick steeply dipping deformation zone delimited by highly deformed organic-rich horizons (dark brown silts). Within the upper part of the trench, just below the modern soil, the humic horizon bounding this deformation zone to the south, defines a ~50 cm wedge (Fig. 6c). In the northern part of the trench, stratigraphic units are strongly folded and overturned towards the north. In some places, small brittle structures are associated with folding. A thin organic palaeosol horizon, interstratified within the deformed units, yielded an age between 11 201 and 12 078 calBP (Table 1), corresponding to the extreme end of the Younger Dryas or more likely to the very early Holocene, a period that is definitely more favourable to soil development (Groisman et al., 2013). We interpret the structure observed in the southern part of trench 1 as a deformed ice wedge pseudomorph characterized notably by the two highly crumpled humic horizons, which delimit it.

Figure 7 shows the features observed within trench 2 excavated across one of the small “dome structures” (see Fig. 3 for location). In this section, silty loess and silty-clayey units describe a knee fold with a steep flank towards the north. In front of the fold, stratigraphic unit are overturned as well as a humic palaeosol horizon that is sheared along a ~30° south-dipping ductile plane. Radiocarbon dating of a charcoal collected inside the sheared palaeosol yielded an age between 11 204 and 11 716 calBP (Table 1) indicating that the formation of this humic horizon must be allocated to the Early Holocene (Preboreal) as in trench 1 (Fig. 7b). At the back of the fold, where stratigraphic units are tilted southwards, we observe brittle structures (mainly small steep reverse faults) associated with folds. Another charcoal, collected from the beige silty unit just below the present soil horizon (30 cm depth), yielded a very young radiocarbon age of about 500 years calBP (Fig. 7a, Table 1). This age is likely not connected with the age of the sediment (translocation of a modern charcoal?).

A third trench was dug across another small dome structure (Fig. 8), in which silty to sandy loess units describe a symmetric ~ 4 m wide isopaque anticline in the central part of the trench. On the southern flank of the fold, several steep north-dipping reverse faults affect the units (Fig. 8a). To the north and south of the anticline, units are overturned and sheared northwards and southwards downslope, respectively. Within the upper part of the trench, the above-described structures are eroded and then overlaid by a yellow silty horizon then by the organic horizon of the topsoil. A small trench (trench 3b, Fig. 9) dug perpendicularly to trench 3 (see location on the Fig. 3b) allowed analysing the features in 3D. The characteristics of the deformations exposed in trench 2, and the three-dimensional observations made at the intersection of the two profiles in trenches 3 and 3b (Figs 8 & 9), have shown that the deformation of the various sandy loess layers resulted from a slow and localised uplift process. Some features are also produced by a slow collapse of the units (normal faults).

According to the characteristics of the deformation and to the climatic context of the area during the Holocene (Groisman et al., 2013) we interpret the various "dome structures" and associated soft deformations as resulting from the growing of thick ice lenses a few meters below the present day surface (not reached in the trenches). This type of features is extremely common in areas located at the boundary between continuous and discontinuous permafrost as today in Canada (Northwest Territories) and where mineral permafrost mounds or cryogenic mounds (lithalsas, according to Harris, 1993) have developed during Late Holocene in alluvial fine-grained sediments (Wolfe et al., 2014). In these environments, the development of ice lenses within the soil (segregation ice lenses) is mainly linked to a flux of unfrozen groundwater pumped by cryo-suction processes (Allard, et al., 1996; Pissart, 2002).

In the Kyren plain the formation of these permafrost mounds has led to the progressive deformation and folding of the surficial periglacial sandy loess deposits. Brittle structures observed on the flanks of the anticline are linked to episodes of growth when the entire sedimentary cover was frozen, while the intensely sheared and overturned units observed on both flanks of the fold correspond to solifluction features that formed during successive thawing episodes. In trench 3, a thermokarst erosion feature showing sharp and irregular erosion boundaries and a very heterogeneous infilling (including angular blocks of loess transported as frozen blocks) results from the thawing of a former ice wedge (Fig. 9b, unit 4).

This feature is a marker of a rapid event of permafrost degradation that could also be at the origin of the periglacial load casting processes observed in the uppermost part of the sequences exposed along the Algak River banks.

4.2 Tuyana site

The Tuyana site is situated at the junction zone between the Khamar-Daban Range and the Elovsky Spur at the eastern termination of the Tunka depression (Fig. 10 and Fig. 1 for location). There, the Irkut River incises the southern part of the Elovsky Spur over a length of ~7 km before reaching the western termination of the Tory depression. As observed within the Kyren site, the southeastern edge of the Tunka depression with the Khamar-Daban Range defines a smooth and sinuous break of slope comprised between 750 m and 800 m in altitude. However, the 140 m deep incision of the Irkut River through the Elovsky Spur suggests that uplift is occurring in the area.

Figure 11 shows the TanDEM-X digital elevation model of the area where the Irkut River starts incising the Elovsky high. In this area, the DEM seems to define a ~50 m width, 12 m deep, E-W linear gutter, affecting the foothills of a NS interfluve of Khamar-Daban Range northern flank (Figs. 11a, b).

Located in the western part of this area, the archaeological excavation of the Tuyana Upper Palaeolithic site (Kozyrev et al., 2014; Shchetnikov et al., 2019) (Fig. 11c) represents an important observation window. At that place the thickest sedimentary sequences (up to 4 m) are located in depressions, whereas minimum thickness (≤ 1 m) is typical for the most elevated parts of the topography. All the stratigraphic profiles exposed by the archaeological excavation of the Tuyana site (10 000 m^2) were investigated to analyse the structures affecting the sediments.

The profiles show a 3.5 m thick sequence composed by pale-yellow aeolian silts (loess); laminated silty colluvial deposits and at least one thick dark brown humic palaeosol horizon. The whole is overlying the weathered granitic bedrock. The sedimentation is characterized by subaerial processes, under conditions of alternating humid and cryoarid climates (Shchetnikov et al., 2019). Significant climate changes are recorded within the sequence by the presence of buried soils and the widespread occurrence of cryogenic deformations. The oldest deposits, including a buried palaeosol dated at about 36 ka calBP by radiocarbon, belong to the Karga complex (Middle Pleniglacial / MIS 3) (Shchetnikov et al., 2019).

During field investigations cryogenic and gravitational deformations represented by cryoturbations, flow (gelifluxion) structures and associated fractures where identified. From the top of the sequence, the first cryogenic deformations are especially well exposed in the topsoil (Holocene) by load casting and diapiric features forming a continuous and widespread cryoturbation horizon (Shchetnikov et al., 2019) (Fig. 12 a,b). This cryoturbated horizon shows the same characteristics and has likely the same age than the one highlighted in the upper 1m of the Algak River bank section in the Kyren area (see: 4.1.2).

Another kind of cryogenic deformation is represented by ice wedge pseudomorphs forming wedge-shaped structures filled with silty sediments. According to the radiocarbon ages published by Shchetnikov et al. (2019), a succession of several former ice wedge networks is recorded in the Tuyana sequence between the Karga period and the Holocene (MIS 3 to 1) (Fig. 12b, c). At the very base of the sequence, the surface of the weathered bedrock is locally penetrated by deformed ice wedge pseudomorphs and eroded by a system of small channel-water streams (Fig. 12). We interpret these channels as the result of thermokarst erosion following a rapid degradation of a former ice wedge network preceding the deposition of the silty sediments.

The geomorphological approach of the area shows typical features of gravitational displacements such as landslides and mudflows associated with permafrost degradation events on slopes (Fig. 13). In the Tuyana sections, folded and lenticular structures as well as doubled or laterally beveled layers are widespread. Figure 13 (b,c) shows the deformations associated with the movement of large soil masses down slope inducing the formation of various folded structures and small cracks whose vertical extension is strictly limited to deformation horizons.

Eventually, no tectonic faults crossing the deposits were observed in the various stratigraphic sections of the Tuyana Upper Pleistocene site where the whole deformation structures results from alternating phases of permafrost development and decay.

Figure 14 shows a 60-m long profile, subdivided in 6 sections, across the above-mentioned E-W linear gutter observed in the DEM (see Fig. 11b for location). The sediments correspond to alternated carbonated loam, fine sandy loam and buried soils. The thickness of sediments is up to 1.2 m above the weathered granitic bedrock. From a nearby section, within the western part of the archaeological site, Shchetnikov et al. (2019) dated 2 buried soils, which are 0.4 and 0.7 m deep, at 6310-6210 and 7720-7580 calBP, respectively. Thus, we considered that the section shown in Figure 12 is Late Pleistocene-Holocene in age. Liquefaction features are observed within the silty and organic-rich horizons. Small flame structures and folds can also be observed within the upper horizons, while larger flame structures with sometimes injection of weathered bedrock are observed in lower horizons.

As in Kyren, the features described above correspond to cryogenic structures and not as tectonic ones. We therefore consider that the E-W linear feature observed on the DEM (Fig. 11a, b) corresponds to an artefact due to the brutal change within the tree coverage. After Google earth satellite imagery, this brutal change seems to be linked to an ancient tree cutting band.

5. Discussion

Our geomorphological and stratigraphic analysis of the Holocene soft-sedimentary deposits and associated deformations along the southern edge of the Tunka Basin shows localised features (ice wedge pseudomorphs) and large-scale horizons of periglacial deformations (cryoturbations) resulting from alternating phases of permafrost aggradation and decrease during the Holocene. Within both studied sites (Kyren and Tuyana), we did not observe any tectonic features as for instance a fault scarp that would run in front of the Khamar-Daban Range. On the contrary, the foothills of this range correspond to a smooth and sinuous break of slope.

In Kyren site, morphotectonic and stratigraphic investigations show that the large scale E-W trending ridge-and-runnel topography observed on the right bank of the Algak River at the foothills of the Khamar-Daban Range corresponds to elongated aeolian ridges. These features, known as “Greda” in the Central Europe loess zone are typical of aeolian sandy silt deposition in periglacial environment close to braided river sources and have been described in many other places elsewhere in the Tunka Basin (e.g. Ravsky et al, 1964). Their W-E alignment is a marker of dominant palaeo-wind direction at the origin of deflation and transport of silts and sands during cold periods. We conclude that these morphologic features do not have a tectonic origin.

Besides, some remarkable elongated dome structures (between 10 and 25 m long) locally occur in this Greda area. Parfeevets and San’kov (2006) interpreted these dome structures as tectonic features, controlled by a NNE-SSW compression. They considered that the loess units define an anticline structure, which was affected by brittle revere faults (Parfeevets and San’kov, 2006). However, in the various trenches and cross sections performed during our field works all the features observed can be related to the growing of segregation ice lenses producing upward deformation in the Late-glacial sandy silts of the area. The observed brittle deformation is simply due to the growing of the structure which generates internal contraction strength inside itself.

Finally, the 6 m high section described from the right bank of the Algak River shows that deformations and involutions are restricted to the upper 1.5-1.8 m of the sandy silts. The same type of structures resulting from the alternation of cold and warm events during the Subboreal have been described along several kilometres of the Irkut river banks by Alexeev et al. (2014). Although not directly dated, the ice wedge pseudomorph we observed in trench 1 strongly suggests the importance of cryogenic processes that appear widespread between the Kyren and Tuyana sites (Alexeev et al., 2014). In Tuyana site, similarly, we did not observe any indices suggesting the occurrence of active tectonic structures within the Upper Pleistocene-Holocene deposits. As within the Kyren site, the observed structures correspond to cryogenic features as the flame structures and folds that are associated with cryoturbation processes affecting the deposits during the Holocene, or the ice wedge pseudomorph affecting the Upper-Pleistocene deposits.

Therefore, our study strongly suggests that there is no evidence for active fault (i.e. South Tunka Fault) along the southern part of the Tunka Basin. This statement is attested by the fact that we observed neither horizontal displacement of channels nor uplifted fluvial terraces along the drainage network of the Khamar-Daban Range and its foothill. Indeed, the morphology and organisation of fluvial terraces systems represent a valuable marker reflecting recent vertical and horizontal tectonic motion. For example, slow uplift is necessary to build stepped terraces staircases whereas stability or relative subsidence produces thick stacked terraces (Antoine et al., 2000; Veldkamp et al., 2000; Bridgland et al., 2007).

On the contrary, we observe continuity and no abrupt change in the bedrock longitudinal profiles of rivers throughout the different sedimentary depressions in the Tunka

Basin. Combined with the morphotectonic studies that demonstrated recent vertical component inversion (from normal to reverse) along the left-lateral strike-slip Mondy fault and along the north-Tunka fault (i.e; Larroque et al., 2001; Arjannikova et al. 2004; Ritz et al., 2018; Arzhannikov et al, 2018), this observation suggests that the Khamar-Daban Range and the Tunka Basin are uplifting together. Inside the Tunka Basin, the uplift is observed within the fluvial geomorphology, notably within the Bystraya river alluvial fan, in the eastern termination of the basin, which shows an incision of ~66 m, and within the Neogene sediments of the basin which show an incision of ~110 m (Mats et al., 2002; Arzhannikov et al, 2018). In the Tory depression, the Irkut River displays incised meanders with terrace heights up to 100 m above the present river bed (Shchetnikov, 2017). In the western part of the basin, the Irkut River terraces display a staircase morphology in the hanging wall of the Mondy fault (i.e. the southern compartment of the Mondy fault) both into the fluvio-glacial deposits and the bedrock, indicating the uplift of the Khamar-Daban block together with the Mondy depression. These observations also suggest that the main regional structures controlling this uplift are the north-Tunka and Mondy faults.

From our analysis, there was no tectonic activity along the southern border of the Tunka Basin during the Holocene, and therefore this topographic border does not represent a major source of seismic hazard. On the other hand, the Mondy, north-Tunka and Sayan faults correspond to the main active faults westwards the Baikal Lake, and have produced destructive earthquakes with magnitude close or larger to 7 during the Holocene (Treskov and Florensov, 1952; McCalpin and Khromovskikh, 1995; Delouis et al., 2002; Chipizubov et al., 2003; Ritz et al., 2018). Those faults are those representing a strong seismic hazard for industries and cities (mainly Irkutsk) developed in this former periglacial area.

6. Conclusion

Our study allows characterizing precisely several morphological structures extending 5 x 5 km and located in the westernmost Tunka Basin at the foothill of the Khamar-Daban Range. The structures are developed in a Holocene soft-sedimentary formation of roughly 6 m-thick and mainly composed by sands and silts of fluvio-aeolian origin. Several cryoturbation features and the known local climatic context attest to a periglacial environment during the sedimentation.

The main structures we observed correspond to E-W striking elongated ridges of several hundred meters long and few meters high. The trenches opened trough these ridges display folded, sometimes overturned, stratas, and several networks of reverse faults which preclude a pure aeolian origin related to transport by strong E-W wind within the Tunka corridor. Taking into account the 3D reconstitution of the folds and faults networks thanks to several perpendicular trenches we show that these structures are soft deformations resulting from the growing of thick ice lenses a few meters below the present-day surface. They do not have a tectonic origin. All these observations combined with the absence of recent morphotectonic indices along the Tunka-Kharmar-Daban boundary leads to the conclusion that there is no evidence of tectonic activity along the south-Tunka Basin. The seismic hazard of the western Baikal zone is therefore mainly linked to the activity of the 3 major faults: Sayan, north-Tunka and Mondy which are reactivated by the far-field India-Asia collision.

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Figures captions:

Figure 1. (a) Landsat image LT513502 of the Tunka basin with microseismicity (1960–2018, magnitude 2-6, catalogue of the regional seismological network, Baikal Division of Geophysic Survey of the Siberian Branch of the RAS, Irkutsk). Pink circles : historical earthquakes : 1742 (I = VII, estimated magnitude 7.5), 1814 (I = V, estimated magnitude 6.4), 1829 (estimated magnitude 7), 1950 (Mw 6.9), from Shebalin and Leydecker, 1997; Radziminovich and Shchetnikov, 2005; 2013; Chipzubov, 2017; Delouis et al., 2002. Focal mechanisms in the south Tunka basin are from Delouis et al. (2002). (b) Digital elevation model of the Tunka Basin with main active faults and their kinematics (1-normal, 2-reverse, 3-strike-slip): TF – Tunka fault, MSF – Main Sayan Fault, MF – Mondy Fault, "STF ?" – assumed South Tunka Fault.

Figure 2. (a) Google earth satellite image (Landsat) of the Kyren area within the southern central part of the Tunka Basin. (b) Morphological interpretation: 1 Holocene alluvial deposits, 2 Upper Pleistocene sand deposits, 3 Upper Pleistocene fluvio-glacial deposits, 4 Holocene alluvial fans, 5 Rivers, 6 Former meanders of the Irkut River, 7 Lakes, 8 Ridges, 9 River terraces, 10 Morphological scarps. (c) 3D view of the Haragun alluvial fan. (d) Topographic profiles (after TanDEM-X DEM, 12 m resolution); the dashed-line defines the ground surface, while the high frequency low amplitude signal observed above it corresponds to vegetation (trees between 5 and 15 m high).

Figure 3. Aerial picture (a) and Google Earth satellite image (Digital Globe) (b) of the site studied to the south of Kyren village. Red arrows show dominant wind directions. Location of the 3 trenches opened: trench 1 (51°39,803´N; 102°08,086´E), trench 2 (51°39,662´N; 102°08,579´E), trench 3 (51°39,653´N; 102°08,517´E).

Figure 4. Algak river bank section: (a) Field picture of the river bank section. (b) Stratigraphic description: 1) Dark grey-brown humic topsoil horizon. 2) Grey-brownish silty sands with numerous root tracks. 3) Little dark grey-brown sandy humic horizon; 4) Greyish homogeneous silty sands showing a strongly deformed lower boundary with unit 6 (downward protrusion features / load casting) and deformed "balls" from organic unit 6. 5) Strongly deformed sandy silt beds and dark grey-brown organic (peaty) horizon (main cryoturbation horizon). 6) Homogeneous light-grey sandy silts with deformed upper boundary (upward protrusion). 7) Light-grey sandy silts with horizontal wavy stratification (fluvio-aeolian). 8) Light-grey sandy fluvial silts with mainly planar stratification, locally cross-bedded (fluvial to fluvio-aeolian). 9) Silto-organic grey-brown horizon (alluvial meadow soil). 10) Yellowish-grey loamy-sands with some organic debris underlying the stratification and little frost-wedges. 11) Peaty horizon with large wood remains / tree stump in situ). 12) Laminated light brownish-grey silts with thin (millimetric) organic debris layers. 13) Peaty horizon with wood remains. 14) Laminated yellowish-grey silt with thin (millimetric) organic debris layers. 15) Yellowish-grey sandy silt with numerous organic debris and involutions (small scale periglacial deformations). 16) Grey-green laminated sandy loam with fine organic layers and some scattered wood remains. 17) Brownish peaty horizon (lower boundary not observed).

Figure 5. Field pictures of the morphological features analysed within the Kyren site where we opened the trenches: (a) Southwestern view of a break of slope (indicated by yellow triangles) at the western tip of the E-W linear feature observed in the aerial picture; (b) and (c) Northeastern and southwestern views of the aligned domes structures (indicated by red triangles), respectively.

Figure 6. Trench 1 (a) log of the western wall: 1 Dark brown modern soil, 2 Light grey stratified silty-loess units, 3 Beige finely stratified silty units with oxidation marks, 4 Palaeosol affected by solifluction, 5 Loamy organic-rich horizons, 6 Fractures, 7 Location and age of collected samples. (b) Northwestern view of the trench. (c) View of the steeply north dipping structure interpreted as an ice wedge pseudomorph.

Figure 7. Trench 2 (a) Log of the western wall (the northern part of the trench is the eastern wall that had been turned to appears as the continuity of log): 1 Dark brown modern humic soil, 2 Silty light brown reddish palaeosol, 3 Homogenous grey-bluish fine silty-loess unit, 4 Finely stratified silty-clayey unit, 5 Silty unit with fine clayey-silty interstratifications, 6 Clayey-silty triangular unit (ice wedge), 7 Palaeosol affected by solifluction, 8 Fractures, 9 Location and age of collected samples. (b) Northwestern view of the southern part of the trench, (b) Southwestern detailed view of a reverse fault, (c) Western view of the sampled palaeosol within the northwards overturned units in the northern part of the trench.

Figure 8. Trench 3 (a) Log of the eastern wall: 1 Dark brown upper horizon of modern soil, 2 Light brown silty lower horizon of modern soil, 3-6 - light to dark grey stratified silty - loessy units with oxidations marks and clay-rich lower parts, 7 Small humic soil horizon affected by solifluction, 8 Fractures. (b) Photo mosaic of the eastern wall.

Figure 9. (a) Stratigraphic profile of trench 3b: 1 - Dark brown-grey humic sandy silt (Ah upper horizon of the topsoil). 2- Dark brown sandy silt with granular structure and numerous root tracks (lower horizon of the topsoil). 3 - Grey-brown sandy loess with numerous bioturbations (root tracks). 4- Greyish sandy loess including numerous blocks of reworked organic silt (infilling of a little thaw channel). The very sharp, angular and strongly erosive boundary between 4 and 5 is typical for thermokarst erosion processes. 5 - Sequence of grey-brown sandy loess, locally finely laminated and including little frost-cracks (cryo-desiccation). The laminations and the frost-cracks are clearly underlined by thin sand beds (≈ 1 mm). 6 - Greyish, compact sandy loess, with numerous oxidised orange patches and bands iron oxide), and ferro-manganic nodules (0,5 to 1mm) (tundra gley horizon). 7 - Greyish, sandy loess, lightly coarser than 5, with some orange iron oxide patches (deep horizon of tundra gley 5). (b) 3D-view of the deformations observed at the junction between profiles of trenches 3 and 3b.

Figure 10. (a) Google Earth satellite image (Landsat) of the Tuyana studied area within the easternmost extremity of the Tunka depression. (b) Morphological interpretation: 1 Holocene alluvial deposits, 2 Upper Pleistocene sand deposits, 3 Upper Pleistocene fluvio-glacial deposits, 4 Rivers, 5 Ancient meanders of the Irkut River, 8 Ridges.

Figure 11. (a) Digital Elevation Model (TanDEM-X, contour lines every 5m) of the foothills of the Khamar-Daban Range, where the Irkut River incises the Elovsky Spur; (b) Enlargement of the studied area; the black dashed rectangle corresponds to the Tuyana archaeological site. (c) Southern view of the Tuyana archaeological site.

Figure 12. Examples of cryogenic deformations in the late Pleistocene-Holocene deposits of the archaeological site of Tuyana (see the position of sections within Figure 11b).

Figure 13. Examples of gravitational deformations in the Tuyana archaeological site area. (a) digital elevation model with landslide location. (b,c) cross sections of the Late Pleistocene deposits deformed by slope masse movements (see the position of sections within Figure 11b). The yellow dashed lines show the zone of unconformity formed as a result of erosion in the Younger Dryas (Shchetnikov et al., 2019). Radiocarbon ages according to (Shchetnikov et al., 2019). (d) photo of the surface of the landslide.

Figure 14. Eastern view of 6 consecutive N-S sections (a-f) of the Tuyana archaeological site (see the position of sections within Figure 11b).

Table 1. Radiocarbon (^{14}C BP) and calibrated ages (calBP) from Kyren profiles. Dendrochronologically calibrated calendar age ranges were calculated using the program Calib 7.1 Radiocarbon Calibration (Stuiver et al., 2019) with 2 standard deviations uncertainty. The age ranges are rounded off to the nearest decade.