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**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**

Anastasia Arzhannikova, J.F. Ritz, Christophe Larroque, Pierre Antoine, Sergey Arzhannikov, Aleksei Chebotarev, Jean-François Stéphan, Marc M. Massault, Jean-Luc Michelot

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1 **Cryoturbation versus tectonic deformation along the southern edge of the Tunka Basin**  
2 **(Baikal Rift System), Siberia: New insights from an integrated morphotectonic and**  
3 **stratigraphic study**  
4

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24

25

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

41

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

44

## 45 **1. Introduction**

46 The Tunka Basin is located in the southwestern part of the Baikal Rift System. As many of  
47 the Baikal Rift basins, it shows a morphological asymmetry characterized by a sharp and  
48 steep border on the northern side, and a smooth morphology on the southern one (Logatchev,  
49 1974; 2003). To the North, the south-dipping Tunka normal fault, located at the foothills of  
50 the Tunka mountain range (Fig. 1), corresponds to the main structure that controlled the

51 development of the Tunka Basin (Sherman et al., 1973). Along this fault, Quaternary  
52 landforms and sediments are affected by paleoearthquake surface ruptures attesting of its  
53 present activity (e.g. Khromovskikh et al., 1975; McCalpin and Khromovskikh, 1995;  
54 Chipizubov et al., 2003; Smekalin, 2008; Smekalin et al., 2013; Ritz et al., 2018;  
55 Arzhannikova et al., 2018).

56       Regarding the southern edge of the Tunka Basin, the question of a tectonic control with  
57 a south Tunka fault located at the foothills of the Khamar-Daban mountain range (Fig.1),  
58 presently active, remains a major question notably in terms of seismic hazard. Several studies  
59 propose the occurrence of one or several active faults (e.g. Sherman et al., 1973; Lukina,  
60 1989; Lunina and Gladkov, 2004), potentially the source of large historical earthquakes  
61 (Lukina, 1989, Radziminovich and Shchetnikov, 2013). However, there are no clear  
62 morphotectonic and/or paleoseismological observations attesting to those structures.

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

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

76

77 **2. Tectonic setting**

78 The Tunka Basin is a broad E-W trending emerged basin (200 km long, 10-30 km wide),  
79 connecting the Baikal Basin to the NE in Russia, and the Hövsgöl Basin to the SW in  
80 Mongolia. Along with the other basins of the Baikal Rift System, the Tunka Basin defines the  
81 northwestern border between the Amurian Plate and the Siberian Craton (Petit and Fournier,  
82 2005; Petit and Déverchère, 2006). The Tunka Basin is bounded by the Tunka and Khamar-  
83 Daban mountain ranges to the north and the south, respectively. These ranges differ in their  
84 geomorphological characteristics. The Tunka Range shows alpine-type reliefs with narrow  
85 ridges and peaks incised by glacial cirques and valleys. The highest summits reach more than  
86 3000 m asl, and are located less than 10 km from the southern foothills. The reliefs in the  
87 Khamar-Daban Range are smooth and present summit plateaus which an altitude of 500 to  
88 700 m lower than the Tunka summits. The northern slope of the Khamar-Daban Range dips  
89 gently to the Tunka Basin, with highest reliefs 20-25 km southwards from the foothills.

90 The Tunka Basin shows a complex inner structure with smaller depressions (from west  
91 to east: Mondy, Khoytogol, Turan, Tunka *sensus stricto* and Tory) separated by interbasin  
92 highs (Nilovsky and Elovsky spurs) (Fig. 1b). These depressions differ in size and depth of  
93 basement, with the greatest depths situated northward against the Tunka Fault (Sherman et al.,  
94 1973). The Tunka depression, located in the centre of the Tunka Basin, is the largest in size  
95 and sediments thickness. The mean topographic surface of the Tunka Basin dips towards the  
96 east from 1350 m asl to 650 m asl, and then shows a 700 m difference of height over the 170  
97 km length of the basin (Fig. 1b).

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

101 Quaternary coarse-grained sediments (Sherman et al., 1973; Mazilov et al., 1993). This  
102 difference in sedimentation has been interpreted as markers of a slow then fast stages in the  
103 rifting process (e.g. Logatchev and Zorin, 1987; Mats et al., 2001; Krivonogov and Safonova,  
104 2017). The late Pleistocene–Holocene formations are mainly composed of fluvio-glacial and  
105 alluvial deposits (boulder-pebbles, gravels, sands) and aeolian sediments (silts, loess) (Ravsky  
106 et al., 1964; Shchetnikov et al., 2012). Most of the coarse-grained material is accumulated  
107 within the Tunka Range piedmont, while the fine-grained sediments are drifted to the inner  
108 parts of the depressions. The lower areas are covered by numerous lakes and swamps due to a  
109 shallow permafrost that can be up to hundreds of meters thick (Logatchev, 1974). The Tunka  
110 Basin and the Tunka Range still show features associated with the Late Pleistocene glaciation  
111 such as glacial valleys, cirques, moraines and kame terraces. During the last glaciation, a 300  
112 m thick glacier occupied the westernmost part of the Tunka Basin (i.e. Mondy Valley)  
113 (Olyunin, 1965). In situ-produced  $^{10}\text{Be}$  dating of Mondy terminal-moraine yielded an average  
114 exposure age of 14 ka (Arzhannikov et al., 2015). During the last glaciation, an intense  
115 aeolian activity occurred and was associated with the deposition of loess, sandy loess and the  
116 formation of sand dunes and ridge-and-runnel topographic features (Ravsky et al., 1964).  
117 According to Shchetnikov et al. (2012), the maximum period of aeolian sedimentation in the  
118 Tunka rift basin occurred at the end of the Late Pleistocene and during the Early Holocene.

119 Cryogenic deformations such as ice wedges and other cryoturbation features were  
120 observed in the cross sections within the Upper Holocene alluvial sediments (Alexeev et al.,  
121 2014). Solifluction phenomena due to thawing permafrost are commonly observed in the  
122 superficial deposits (Ravsky et al., 1964).

123 Nowadays, the Tunka Basin corresponds to an emerged rift structure. It is located  
124 within a transition zone between the present-day compressional deformation of North  
125 Mongolia and the extensional one of the central Baikal Rift as attested by the focal

126 mechanisms of earthquakes (Melnikova and Radziminovich, 1998; Radziminovich et al.,  
127 2016). Morphotectonic and paleoseismic analyses show that the kinematics of faults  
128 controlling the Tunka Basin changed through time. From the Miocene to the Quaternary a  
129 transtensional regime dominated, with normal and left-lateral strike-slip faulting along the  
130 major fault zones, allowing the opening of basins and thick accumulation of loose sediments  
131 (San'kov et al., 1997). Then the strain regime was replaced by a transpressional regime in the  
132 Late Quaternary combining compressional and left-lateral strike-slip fault deformations  
133 (Larroque et al., 2001; Arzhannikova et al., 2005; Arjannikova et al., 2004; Arzhannikova,  
134 Melnikova, Radziminovich, 2007; Jolivet et al., 2013; Ritz et al., 2018). This recent  
135 transpressional strain regime would be the result of the northwards propagation of the  
136 collisional deformation associated with the convergence between India and Asia (Ritz et al.,  
137 2000). Evidence of recent inversion was observed along the Mondy Fault and its western  
138 extension northward of the Hövsgöl Lake (Larroque et al., 2001; Arzhannikova et al., 2003;  
139 Arjannikova et al., 2004) and along the East Tunka Fault and East Sayan Fault (Chipizubov et  
140 al., 2003; Smekalin, 2008; Shchetnikov, 2016; Ritz et al., 2018; Arzhannikov et al., 2018).

141         The question of an active fault located at the southern edge of the Tunka Basin  
142 remains undetermined. From geomorphological and structural considerations Sherman (1973)  
143 has proposed a fault bounding the Turan and Tunka depressions to the south at the foothills of  
144 Khamar-Daban Range (Fig. 1b). Lukina (1989) and Lunina and Gladkov (2004) extended this  
145 fault until the Baikal Lake. However, no evidence of deformations in the Upper Pleistocene-  
146 Holocene deposits has been described along this hypothetical structure. At the western end of  
147 the basin, evidences of tectonic scarps associated to the Mondy Fault are observed at the  
148 southern foothills of the Nilovsky Spur until its eastern limit (Geological map-East Sayan  
149 series, 1961; Arjannikova et al., 2004) (Fig. 1b). Moreover, two historical earthquakes were  
150 felt within this area in 1814 (Shimki Village, I=IX (MSK)) and in 1829 (Turan and Shimki

151 villages I=VII-VIII, Mondy Village I=IX) (Kondorskaya and Shebalin, 1977) (Fig. 1a). After  
152 Lukina (1989) and Radziminovich and Shchetnikov (2013), these historical events were  
153 produced by rupture of the western part of the South Tunka Fault or by the Mondy Fault.  
154 Currently, the instrumental seismicity is much active in this western part than in the eastern  
155 part of the Tunka Basin (e.g. Solonenko et al., 1997; Radziminovich et al., 2013) (Fig. 1a).  
156 Among these instrumental earthquakes two events of magnitudes M4.5 and M5 occurred near  
157 Kyren Village in 1958 (Golenetsky, 1998).

158

### 159 **3. Methods**

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

174

### 175 **4. Geomorphology, Quaternary sediments and structures**



176 **4.1 Kyren site**

177 **4.1.1 Geomorphology of the Kyren area**

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

191 However, on an aerial picture, we have been puzzled by an E-W linear feature  
192 affecting the Holocene deposits forming the plain at the east of the right bank of the Algak  
193 River, to the south of Kyren Village (Fig. 3a, see Fig. 2a for location). In a satellite image,  
194 this feature is much less obvious, and seems to underline the northern edge of a ridge-and-  
195 runnel topography area (Fig. 3b). The satellite image also allowed pointing out some smaller  
196 W-E aligned “dome structures” (Figure 3b). To decipher whether these structures are tectonic  
197 features and could correspond to the hypothetical south Tunka fault or not, we carried out field  
198 works including trenching throughout the W-E aligned break in the slope profile and two  
199 “dome structures”. These investigations were completed by the study of the sedimentary

200 section located on the right bank of the Algak River (Figure 3b), where fluvial erosion has  
201 produced a 6 m high sub-vertical cliff in a meander concavity.

202

#### 203 **4.1.2- Stratigraphy of Quaternary deposits: the Algak River bank section**

204 From the base to the top, the Algak River bank section exposes fluvial, then fluvio-aeolian  
205 and aeolian deposits affected by a thick deformation horizon in their upper 1.5 m (Fig. 4).

206 From the base to the top, the Algak River profile can be subdivided into three sub-sequences  
207 (SS1 to SS3) overlying the uppermost part of a peat layer occurring at -6.1 m at the level of  
208 the present day River (Fig. 4):

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

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

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

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

245 According to the study of Algak River bank section it is clear that no other periglacial horizon  
246 with large involutions is present below 1.5 m from the surface. This observation is in good  
247 accordance with another study made along the Irkut River banks where a thick layer of 1.5 to  
248 2 m of highly deformed humic silts and sands associated with ice wedge pseudomorphs has  
249 been evidenced along several kilometres (Alexeev et al., 2014). In this study the cryoturbated

250 organic rich layers located between - 40 and -100 cm from the surface have been dated by  
251 radiocarbon analysis from 2 600 and 3 600 BP respectively (Alexeev et al., 2014). They  
252 correspond to a period of climatic instability starting with a very strong cooling event around  
253 4 500 BP followed by the alternation of several phases of climatic cooling and warming  
254 during the Subboreal (Groisman et al., 2013). According to the close correspondence between  
255 the thickness and the structure of the deformed layers in both the Upper part of the Irkut and  
256 Algak rivers banks sections it is proposed that the deformed layers of the Algak River profile  
257 in the Kyren area belong to the same period (Subboreal).

258

### 259 **4.1.3 - The ridge-and-runnel features**

260 **4.1.3.1 Morphology**In the flat area located to the East of the Algak River, and presently  
261 covered by steppe vegetation, the analysis of both Russian aerial photographs and satellite  
262 images allows evidencing numerous W-E elongated ridges (100 to 1000 m long) a few meters  
263 in high (Fig. 3b). They only occur in this area located to the East of the Algak River and not  
264 to the West in the alluvial plain between the Kyren and the Algak rivers. Features of the same  
265 type have also been observed further North, near Kyren Village (Fig. 3b), and are largely  
266 described in the Tunka Basin (Ravsky et al., 1964; Shchetnikov et al., 2012). They have been  
267 interpreted as resulting from aeolian deflation and transport by strong WE winds within the  
268 Tunka corridor (Shchetnikov et al., 2012).

269 This type of morphology is well known in the loess area of Central Europe where they are  
270 locally named “Greda” (Léger, 1990). They appear as for example along the right bank of the  
271 Danube River where they can reach more than 1 km long and 50 m high (Jipa, 2014). More  
272 recently the study of large Greda structures located in western Europe on the edge of the  
273 Odenwald plateau, connected to the right bank of the Rhine graben close to Heidelberg  
274 (Germany), evidenced the “onion peel” structure of the stratigraphic units making these

275 elongated ridges (Antoine et al., 2001). They have been built by the rapid accumulation (14 m  
276 between about 30 and 17 ka (Moine et al., 2017)) of sands and silts deflated then transported  
277 as low level suspension from the braided alluvial plain of the Rhine during storms from N-  
278 NW. These features, characterized by rapid accumulation of sandy loess showing laminated  
279 facies made by the alternation of sandy and silty layers up to 1 cm thick are typical of aeolian  
280 deposition in flat surfaces located immediately downwind of important sand and silt sources  
281 as braided periglacial rivers (Antoine et al., 2001). In the case of the Kyren area, the source of  
282 the silty material was the braided plain of the Irkut River located upwind of the ridge area and  
283 where active deflation processes occurred during the end of the Last glacial and the Younger  
284 Dryas, according to  $^{14}\text{C}$  ages from trench 1 (see below) and in good agreement with the data  
285 published by Shchetnikov et al., (2012).

286 However, some of these small elongated domes (with length comprised between 10 and 25 m)  
287 showing steep slopes, likely do not only result of pure aeolian processes. In order to  
288 understand the origin of these features we opened three trenches in the plain located at the  
289 east of the Algak River. Figure 5 shows the sites where they were opened: the first site (Fig.  
290 5a) corresponds to an E-W smooth topographic scarp across which we opened trench 1 (Fig.  
291 6). The second site (Figs. 5b and 5c) corresponds to E-W “oval dome structures”. There, we  
292 opened two trenches (Figs. 7 and 8, respectively).

293

#### 294 **4.1.3.2 - Stratigraphy and deformations**

295 Within trench 1, we observed stratified (laminated?) silty-loess units and thin organic-rich  
296 paleosols affected by ductile deformations (Figs. 6a, b). In the southern part of the trench, the  
297 units are upturned along a 30-40 cm thick steeply dipping deformation zone delimited by  
298 highly deformed organic-rich horizons (dark brown silts). Within the upper part of the trench,  
299 just below the modern soil, the humic horizon bounding this deformation zone to the south,

300 defines a ~50 cm wedge (Fig. 6c). In the northern part of the trench, stratigraphic units are  
301 strongly folded and overturned towards the north. In some places, small brittle structures are  
302 associated with folding. A thin organic palaeosol horizon, interstratified within the deformed  
303 units, yielded an age between 11 201 and 12 078 calBP (Table 1), corresponding to the  
304 extreme end of the Younger Dryas or more likely to the very early Holocene, a period that is  
305 definitely more favourable to soil development (Groisman et al., 2013). We interpret the  
306 structure observed in the southern part of trench 1 as a deformed ice wedge pseudomorph  
307 characterized notably by the two highly crumpled humic horizons, which delimit it.

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

320 A third trench was dug across another small dome structure (Fig. 8), in which silty to sandy  
321 loess units describe a symmetric ~ 4 m wide isopaque anticline in the central part of the  
322 trench. On the southern flank of the fold, several steep north-dipping reverse faults affect the  
323 units (Fig. 8a). To the north and south of the anticline, units are overturned and sheared  
324 northwards and southwards downslope, respectively. Within the upper part of the trench, the

325 above-described structures are eroded and then overlaid by a yellow silty horizon then by the  
326 organic horizon of the topsoil. A small trench (trench 3b, Fig. 9) dug perpendicularly to trench  
327 3 (see location on the Fig. 3b) allowed analysing the features in 3D. The characteristics of the  
328 deformations exposed in trench 2, and the three-dimensional observations made at the  
329 intersection of the two profiles in trenches 3 and 3b (Figs 8 & 9), have shown that the  
330 deformation of the various sandy loess layers resulted from a slow and localised uplift  
331 process. Some features are also produced by a slow collapse of the units (normal faults).

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

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

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

353

#### 354 **4.2 Tuyana site**

355 The Tuyana site is situated at the junction zone between the Khamar-Daban Range and the  
356 Elovsky Spur at the eastern termination of the Tunka depression (Fig. 10 and Fig. 1 for  
357 location). There, the Irkut River incises the southern part of the Elovsky Spur over a length of  
358 ~7 km before reaching the western termination of the Tory depression. As observed within the  
359 Kyren site, the southeastern edge of the Tunka depression with the Khamar-Daban Range  
360 defines a smooth and sinuous break of slope comprised between 750 m and 800 m in altitude.  
361 However, the 140 m deep incision of the Irkut River through the Elovsky Spur suggests that  
362 uplift is occurring in the area.

363 Figure 11 shows the TanDEM-X digital elevation model of the area where the Irkut  
364 River starts incising the Elovsky high. In this area, the DEM seems to define a ~50 m width,  
365 12 m deep, E-W linear gutter, affecting the foothills of a NS interfluvium of Khamar-Daban  
366 Range northern flank (Figs. 11a, b).

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



374 The profiles show a 3.5 m thick sequence composed by pale-yellow aeolian silts  
375 (loess); laminated silty colluvial deposits and at least one thick dark brown humic palaeosol  
376 horizon. The whole is overlying the weathered granitic bedrock. The sedimentation is  
377 characterized by subaerial processes, under conditions of alternating humid and cryoarid  
378 climates (Shchetnikov et al., 2019). Significant climate changes are recorded within the  
379 sequence by the presence of buried soils and the widespread occurrence of cryogenic  
380 deformations. The oldest deposits, including a buried palaeosol dated at about 36 ka calBP by  
381 radiocarbon, belong to the Karga complex (Middle Pleniglacial / MIS 3) (Shchetnikov et al.,  
382 2019).

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

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

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

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

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

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

424

425 **5. Discussion**

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

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

443 Besides, some remarkable elongated dome structures (between 10 and 25 m long)  
444 locally occur in this Greda area. Parfeevets and San’kov (2006) interpreted these dome  
445 structures as tectonic features, controlled by a NNE-SSW compression. They considered that  
446 the loess units define an anticline structure, which was affected by brittle reverse faults  
447 (Parfeevets and San’kov, 2006). However, in the various trenches and cross sections  
448 performed during our field works all the features observed can be related to the growing of

449 segregation ice lenses producing upward deformation in the Late-glacial sandy silts of the  
450 area. The observed brittle deformation is simply due to the growing of the structure which  
451 generates internal contraction strength inside itself.

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

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

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

474 Basin. Combined with the morphotectonic studies that demonstrated recent vertical  
475 component inversion (from normal to reverse) along the left-lateral strike-slip Mondy fault  
476 and along the north-Tunka fault (i.e; Larroque et al., 2001; Arjannikova et al. 2004; Ritz et al.,  
477 2018; Arzhannikov et al, 2018), this observation suggests that the Khamar-Daban Range and  
478 the Tunka Basin are uplifting together. Inside the Tunka Basin, the uplift is observed within  
479 the fluvial geomorphology, notably within the Bystraya river alluvial fan, in the eastern  
480 termination of the basin, which shows an incision of ~66 m, and within the Neogene  
481 sediments of the basin which show an incision of ~110 m (Mats et al., 2002; Arzhannikov et  
482 al, 2018). In the Tory depression, the Irkut River displays incised meanders with terrace  
483 heights up to 100 m above the present river bed (Shchetnikov, 2017). In the western part of  
484 the basin, the Irkut River terraces display a staircase morphology in the hanging wall of the  
485 Mondy fault (i.e. the southern compartment of the Mondy fault) both into the fluvio-glacial  
486 deposits and the bedrock, indicating the uplift of the Khamar-Daban block together with the  
487 Mondy depression. These observations also suggest that the main regional structures  
488 controlling this uplift are the north-Tunka and Mondy faults.

489 From our analysis, there was no tectonic activity along the southern border of the  
490 Tunka Basin during the Holocene, and therefore this topographic border does not represent a  
491 major source of seismic hazard. On the other hand, the Mondy, north-Tunka and Sayan faults  
492 correspond to the main active faults westwards the Baikal Lake, and have produced  
493 destructive earthquakes with magnitude close or larger to 7 during the Holocene (Treskov and  
494 Florensov, 1952; McCalpin and Khromovskikh, 1995; Delouis et al., 2002; Chipizubov et al.,  
495 2003; Ritz et al., 2018). Those faults are those representing a strong seismic hazard for  
496 industries and cities (mainly Irkutsk) developed in this former periglacial area.

497

## 498 **6. Conclusion**

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

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

517

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531

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725

726 **Figures captions:**

727 **Figure 1.** (a) Landsat image LT513502 of the Tunka basin with microseismicity (1960–2018,  
728 magnitude 2-6, catalogue of the regional seismological network, Baikal Division of  
729 Geophysic Survey of the Siberian Branch of the RAS, Irkutsk). Pink circles : historical  
730 earthquakes : 1742 (I = VII, estimated magnitude 7.5), 1814 (I = V, estimated magnitude 6.4),  
731 1829 (estimated magnitude 7), 1950 (Mw 6.9), from Shebalin and Leydecker, 1997;  
732 Radziminovich and Shchetnikov, 2005; 2013; Chipzubov, 2017; Delouis et al., 2002. Focal  
733 mechanisms in the south Tunka basin are from Delouis et al. (2002). (b) Digital elevation  
734 model of the Tunka Basin with main active faults and their kinematics (1-normal, 2-reverse,  
735 3-strike-slip): TF – Tunka fault, MSF – Main Sayan Fault, MF – Mondy Fault, “STF ?” –  
736 assumed South Tunka Fault.

737

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

746

747 **Figure 3.** Aerial picture (a) and Google Earth satellite image (Digital Globe) (b) of the site  
748 studied to the south of Kyren village. Red arrows show dominant wind directions. Location of  
749 the 3 trenches opened: trench 1 ( $51^{\circ}39,803'N$ ;  $102^{\circ}08,086'E$ ), trench 2 ( $51^{\circ}39,662'N$ ;  
750  $102^{\circ}08,579'E$ ), trench 3 ( $51^{\circ}39,653'N$ ;  $102^{\circ}08,517'E$ ).

751

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

770



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

776

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

782

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

791

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

796

797 **Figure 9.** (a) Stratigraphic profile of trench 3b: 1 - Dark brown-grey humic sandy silt (Ah  
798 upper horizon of the topsoil). 2- Dark brown sandy silt with granular structure and numerous  
799 root tracks (lower horizon of the topsoil). 3 - Grey-brown sandy loess with numerous  
800 bioturbations (root tracks). 4- Greyish sandy loess including numerous blocks of reworked  
801 organic silt (infilling of a little thaw channel). The very sharp, angular and strongly erosive  
802 boundary between 4 and 5 is typical for thermokarst erosion processes. 5 - Sequence of grey-  
803 brown sandy loess, locally finely laminated and including little frost-cracks (cryo-  
804 desiccation). The laminations and the frost-cracks are clearly underlined by thin sand beds ( $\approx$   
805 1 mm). 6 - Greyish, compact sandy loess, with numerous oxidised orange patches and bands  
806 iron oxide), and ferro-manganic nodules (0,5 to 1mm) (tundra gley horizon). 7 - Greyish,  
807 sandy loess, lightly coarser than 5, with some orange iron oxide patches (deep horizon of  
808 tundra gley 5). (b) 3D-view of the deformations observed at the junction between profiles of  
809 trenches 3 and 3b.

810

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

815

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

820

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

823

824 **Figure 13.** Examples of gravitational deformations in the Tuyana archaeological site area. (a)  
825 digital elevation model with landslide location. (b,c) cross sections of the Late Pleistocene  
826 deposits deformed by slope masse movements (see the position of sections within Figure  
827 11b). The yellow dashed lines show the zone of unconformity formed as a result of erosion in  
828 the Younger Dryas (Shchetnikov et al., 2019). Radiocarbon ages according to (Shchetnikov et  
829 al., 2019). (d) photo of the surface of the landslide.

830

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

833

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