
Volume 46, 2 March 2012
ISSN 1367-9120
ELSEVIER
Journal of
Asian Earth Sciences
(Formerly the Journal of Southeast Asian Earth Sciences)
Editor-in-Chief
BOR-MING JAHN

Journal of Asian Earth Sciences 46 (2012) 195–208

Contents lists available at SciVerse ScienceDirect

Journal of Asian Earth Sciences

journal homepage: www.elsevier.com/locate/jseaes

Late Quaternary geology of the Tunka rift basin (Lake Baikal region), Russia

Alexander A. Shchetnikov [a], Dustin White [b,c,*], Ivan A. Filinov [a], Nat Rutter [d]

[a] *Institute of the Earth Crust, Siberian Branch of the Russian Academy of Sciences, Irkutsk 664033, Russia*
[b] *Archaeology, University of Southampton, Avenue Campus, Southampton SO17 1BF, United Kingdom*
[c] *Institute of Archaeology, University of Oxford, Oxford OX1 2PG, United Kingdom*
[d] *Department of Earth and Atmospheric Sciences, University of Alberta, Edmonton, AB, Canada T6G 2E3*

ARTICLE INFO

Article history:
Received 18 January 2011
Received in revised form 13 December 2011
Accepted 17 December 2011
Available online 3 January 2012

Keywords:
Quaternary
Stratigraphy
Tunka rift
Lake Baikal
Siberia

ABSTRACT

The objective of this research is to obtain a better understanding of the evolution of the Tunka rift basin, part of the Baikal rift zone, and how it relates to the overall geologic history of the region, particular for the Quaternary period. The tectonically active Baikal rift zone began forming over 50 million years ago and continues today. In the Tunka basin, during the Oligocene and Middle Pliocene, relatively weak tectonic disturbances took place and thick accumulations of organic-rich sands, silts, and clays were deposited in lacustrine–marshy subtropical environments. Tectonism increased between the Miocene and Pliocene and thick units of coarse alluvium and floodplain sediments were deposited. During the Late Pliocene–Quaternary, tectonism formed basins that are now filled with a variety of coarse clastic materials. Early and Middle Pleistocene sediments are poorly exposed, covered by widespread Late Pleistocene deposits. Three Late Pleistocene sedimentary facies dominate: boulder–pebble gravels (proluvial, glacial fluvial, and alluvial sediments), alluvial sand, and loess-like sediments with associated slope deposits altered by post-depositional wind erosion. The relationship between these complexes, including radiocarbon and other chronological data and fauna and flora remains, indicates that they began forming c. 70000 yr ago. Paleosols, glacial deposits and cryogenic material indicate that at times the climate was cool or cold. During the early Late Pleistocene renewed tectonism took place causing increased deposition of coarse sediments. The middle Late Pleistocene deposits consist mostly of sandy, floodplain alluvium. By the end of the Late Pleistocene–Holocene, alluviation was reduced and replaced by a high degree of erosion and aeolian deposition.

1. Introduction

The Tunka rift basin is located in the southwest part of the Baikal rift zone in continental east Asia (Fig. 1). The basin exhibits graben features that are characteristic of the neotectonic and geomorphological development of the Baikal rift system as a whole. The geological formations can be regarded as examples that not only define a morphologic and tectonic type (Florensov, 1960; Sherman et al., 1973; Shchetnikov and Ufimtsev, 2004), but also include rare features (e.g. abrupt inclination of the rift bottom, c. 1000 m elevation over 200 km extent of the rift valley) not found in other intercontinental rift systems (Ufimtsev and Shchetnikov, 2002).

The objective of this study is to understand better the evolution of the Tunka rift by separating geologic events temporally and spatially that will aid in reconstructing the Quaternary history of the Tunka basin, and to the Lake Baikal region in general. Over the last few decades, sediment exposures in the basins of the Tunka rift valley have been studied and considered to be Pleistocene stratigraphic units, although no absolute dates were available. Recently, we have investigated key sections in the area and obtained new radiocarbon and thermoluminesence ages that allow us to establish the Late Quaternary chronology of the Tunka rift sedimentary fill. These data also enable reconstructions of depositional processes influenced by neotectonic activity. Below we give an overview of general geological research in the Tunka rift valley, provide details on geomorphology and Quaternary geology, present new data from Late Quaternary sections, and finally outline the general Cenozoic history of the Tunka rift basin.

2. Background to research

2.1. Previous work

In the 1950s, a number of large-scale drilling projects were initiated in the Tunka and surrounding basins of the Lake Baikal area (Fig. 2) and several stratotype sections of Cenozoic stratigraphic units were described (Logachev, 1958a,b; Florensov, 1960, 1969) (Fig. 3). Subsequent work on the Cenozoic geology of the Tunka basin was published by Mazilov et al. (1972, 1993), Adamenko et al. (1975, 1984), Popova et al. (1989), Kashik and Mazilov (1994), Ufimtsev et al. (2002, 2003) and Shchetnikov et al. (2009).

* Corresponding author.
E-mail address: dustin.white@arch.ox.ac.uk (D. White).

1367-9120/$ - see front matter © 2012 Elsevier Ltd. All rights reserved.
doi:10.1016/j.jseaes.2011.12.010

Other studies have focused more specifically on the conditions and characteristics of local sedimentation in the Tunka basin (Logachev, 1958b, 1974; Ravskii et al., 1964; Pavlov et al., 1976; Trofimov et al., 1995; Shchetnikov and Ufimtsev, 2004; Krivonogov, 1995; Hase et al., 2003; Ufimtsev et al., 2004b; Vogt and Vogt, 2007). Issues related to recent tectonics and geodynamics of the rift are examined in Sherman et al. (1973), Logachev and Mohr (1978), Logachev and Zorin (1992), Logachev (1984), Delvaux et al. (1997), Mats (1993), Parfeevets and San'kov (2006), Arzhannikova et al. (2005), and Shchetnikov and Ufimtsev (2004). As outlined in the work cited above, studies of the Tunka basin have traditionally focused on Paleogene–Neogene sediments, which make up the majority of the Cenozoic rift system. In the course of earlier drilling, Quaternary sediments up to 500 m thick (Sherman et al., 1973) were identified but received little study.

The development of the Baikal rift system, as well as the Tunka rift, is divided into two main taphrogenic stages (Logachev and Mohr, 1978; Logachev and Zorin, 1987; Kashik and Mazilov, 1994). Rift development was preceded by an extended Cretaceous–Eocene period of tectonic stability, where a peneplain developed after as much as 25 m of material was weathered and eroded (Mazilov et al., 1972, 1993; Pavlov et al., 1976). The first stage of development of the Tunka rift occurred during the Oligocene and Early Pliocene and was characterised by relatively weak tectonic activity, trough-like down warping of depressions, and the accumulation of thick carbon-rich fine-grained sediments (more than 1700 m) under humid and warm climatic conditions (Logachev, 1958a). The development of rich thermophilic broadleaf vegetation with an admixture of subtropical forms (e.g. magnolia, myrtle, and laurel) developed near the edges of the basins where with time coal developed. The central parts of the basins accumulated sands, silts, clays, and diatomaceous algae in lacustrine–marshy environments. During this period the Tunka rift had a connection to Lake Baikal, as evidenced by the presence of endemic Baikal fauna in Miocene sediments of the Tunka basins, such as fresh-water sponges from the Lubomirkiidae family (Martinson, 1948). By the end of the Miocene the connection to Lake Baikal ended (Shchetnikov and Ufimtsev, 2004). Lava flows were also common during this time with some layers up to 50–80 m thick (Logachev, 1993, 2003; Logachev and Mohr, 1978; Sherman et al., 1973) (Fig. 4).

A second taphrogenic stage occurred during the Late Pliocene–Quaternary and is characterised by an increased rate of tectonic movement, resulting in variations of relief in the Tunka region. In general, the basins acquired their present morphologies and began filling with relatively coarser material than earlier episodes, with deposits exceeding 1000 m in thickness. During the Middle Pleistocene, the area of sediment accumulation decreased but was later restored to its previous boundaries in the middle Late Pleistocene. This tectonic stage occurred under colder and drier climatic conditions (Kuimova and Sherstyankin, 2003) and coincided with the formation of mountain-taiga and trough-steppe landscapes. The sediments are comprised largely of alluvial, proluvial, volcanogenic–detrital, glaciofluvial, and glacial deposits and to a lesser extent, lacustrine–marshy units. In places, locally exposed Quaternary deposits contain fossil remains, discussed in more detail below.

2.2. Geomorphology

The Tunka rift basin and its bordering mountains, the Tunka Range to the north, the Khamar Daban Range to the south, and the Munku Sardyk Range to the west, form a geographical region known as the "Tunka Cis-Baikal", which also includes several smaller basins (from east to west: Bistraya, Tory, Tunka, Turan, Khoito Gol, and Mondin; Fig. 2A). From the north, the rift is bordered by a steep (up to 30–40°) fault scarp of the Tunka horst coinciding with the southern slope of the range (Fig. 2B). This fault is still active capable

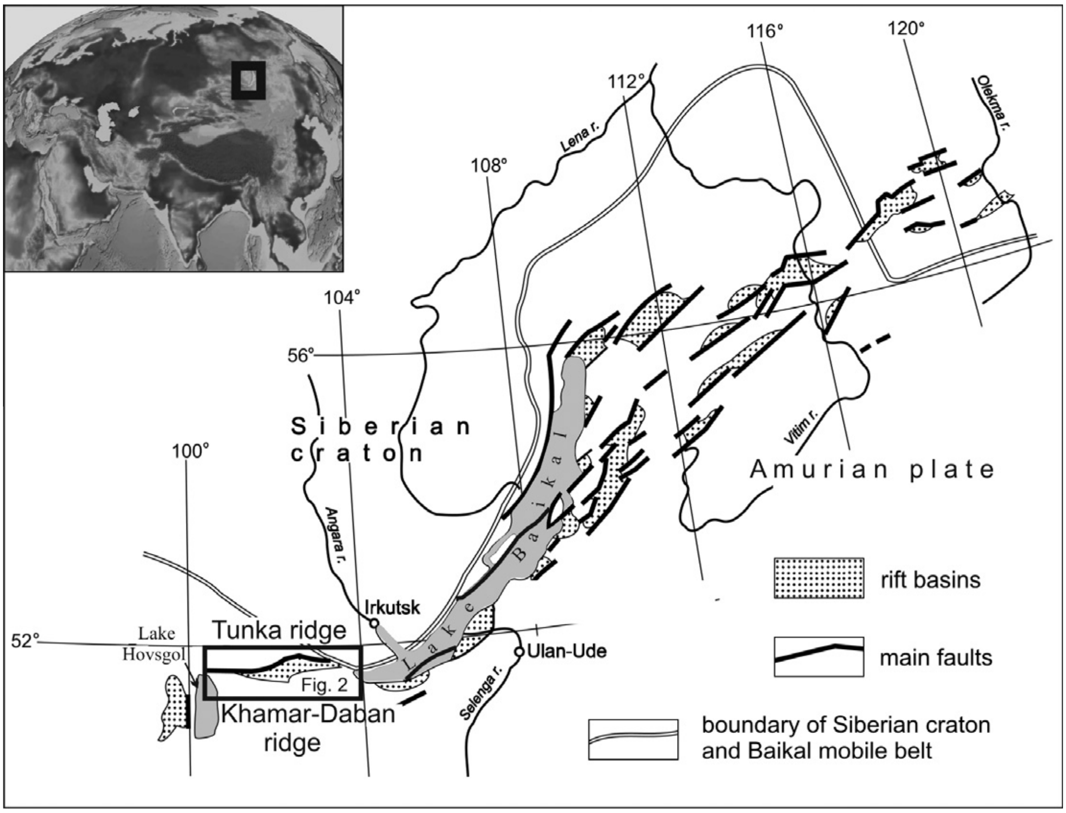

Fig. 1. Location map of the Tunka rift basin.

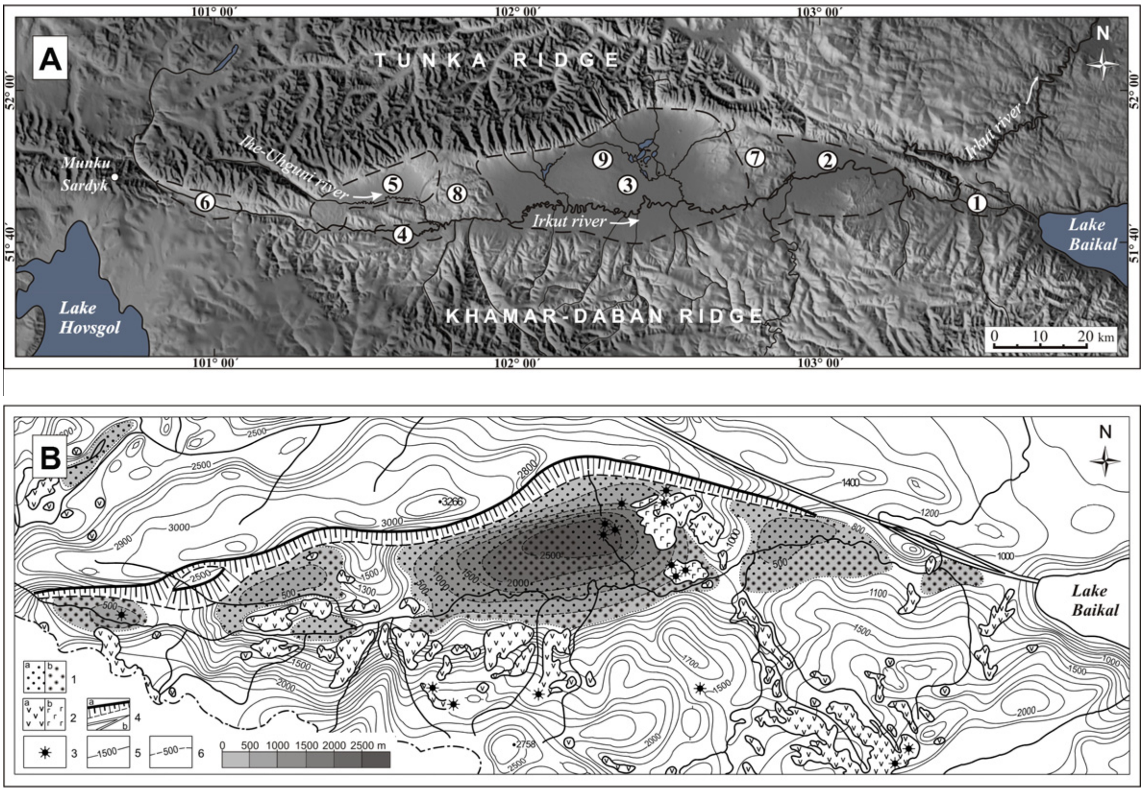

Fig. 2. The Tunka rift basin and bordering regions. A – Digital elevation model of the rift (based on SRTM digital elevation data). Numbers indicate basins (1 – Bystraya; 2 – Tory; 3 – Tunka; 4 – Turan; 5 – Khoito Gol; 6 – Mondin), inter-basin spurs (7 – Elovsky; 8 – Nilovsky), and anticlinal massif Badar (9). B – Tectonic topography with sediment thicknesses (m) (Logachev, 1974). 1 – Rift basins, including (a) the areas of current sedimentation and (b) areas affected by tectonism; 2 – basalt flows (a – Quaternary; b – Neogene); 3 – volcanoes; 4 – main faults (a – scarp of the main Tunka fault; b – Main Sayan fault); 5 – elevation contour line in m; 6 – isopachs of Cenozoic deposits.

of generating earthquakes up to magnitude 8 (Richter scale; Chipizubov et al., 2003). During the Holocene, this fault generated at least four earthquakes exceeding magnitude 7 (Chipizubov et al., 2003). Extensional deformation has formed linear systems of 'scarp-micrograbens' (McCalpin and Khromovskikh, 1995) which dissect the terraces of the river valleys. These systems are found throughout the entire foothills area of the Tunka Range. With an absolute height reaching 3284 m a.s.l., the Tunka Range also exhibits evidence of Late Pleistocene valley glaciation. Nearly all of the valleys oriented toward the rift depressions had glaciers that advanced beyond the foothills. The neighbouring Munku-Sardyk Range reveals the same features. Its highest peak, presently covered with an ice cap, reaches 3491 m a.s.l.

From the south, the Tunka rift is bounded by the Khamar Daban Range which rises to a height of 2500–2994 m a.s.l. and extends over 250 km from Lake Baikal to Lake Hovsgol (Fig. 1). Along the axis of the ridge there are shallow trough valleys and corries. The topography of the bottom of the Tunka basin is represented by an alteration of relatively lower inter-mountain valleys and low-mountain spurs with exposures of the crystalline basement. These inter-basin spurs are categorised as elements of the internal structure of the rift system (Florensov, 1960; Sherman et al., 1973; Logachev, 1974) and form topographic "connectors" between the chain of the Tunka and Khamar Daban ranges, creating intermediary hypsometric levels 150–600 m high between the basin bottoms and the ridges that border them. All of the spurs are cut by narrow antecedent valleys of the Irkut River (Shchetnikov et al., 1997), the main drainage axis of the Tunka rift.

The basins of the Tunka rift also display local tectonic inversions (Shchetnikov, 2008; Ufimtsev et al., 2009). These inversions directly impact the development of the rift structure and result in deformation of the basin bottoms. These areas, which include the dome-shaped anticlinal massif Badar located in the centre of the Tunka basin and its peripheral depression along the foothills of the Khamar Daban, are subject to erosion as are the Malaia Bistraya and Mondin basins, located at opposite ends of the rift, which are entirely inversed and dissected by erosion up to 150 m deep (Shchetnikov and Ufimtsev, 2004).

The basins in the Tunka rift vary in shape, although most are elongated. One important aspect of the basins is reflected in the differences of their hypsometric levels. Their altitude gradually increases westward, from 600 to 670 m a.s.l. in the Bistraya basin (454 m a.s.l.) to more than 1300 m a.s.l. in the Mondin basin. The bottom of the Bistraya depression is higher than the surface water of the neighbouring Baikal basin by approximately 200 m on average. This difference in base elevation occurs within 20 km, while the maximum elevation difference in relation to Lake Baikal reaches 900 m.

Late Pleistocene volcanic cinder cones also occur in the Tunka rift basin, rising 80–120 m above the valley floor. Those closer to the centre of the basin are almost entirely buried under approximately 70 000 yr of sediment accumulation with only their surfaces exposed (Ufimtsev et al., 1999). The intensity of tectonic subsidence in the centre of the basin is evidenced by the discovery of artifacts, believed to be Neolithic in age, nearly 12 m under the floodplain sediment (Lvov, 1924). Drilling data indicate autochthonous interlayers of entirely intact green hypnoeutrophic peat approximately 150–180 m below the surface. Within the inter-basin and inter-rift spurs, the terrace structure consists of up to nine rock cut terraces, with steps exceeding 100 m (Ufimtsev et al., 2004a).

The Tunka rift basins are occupied largely by valleys with relatively low sediment accumulation rates. The first type of valley is represented by alluvial formations of the Irkut River and its

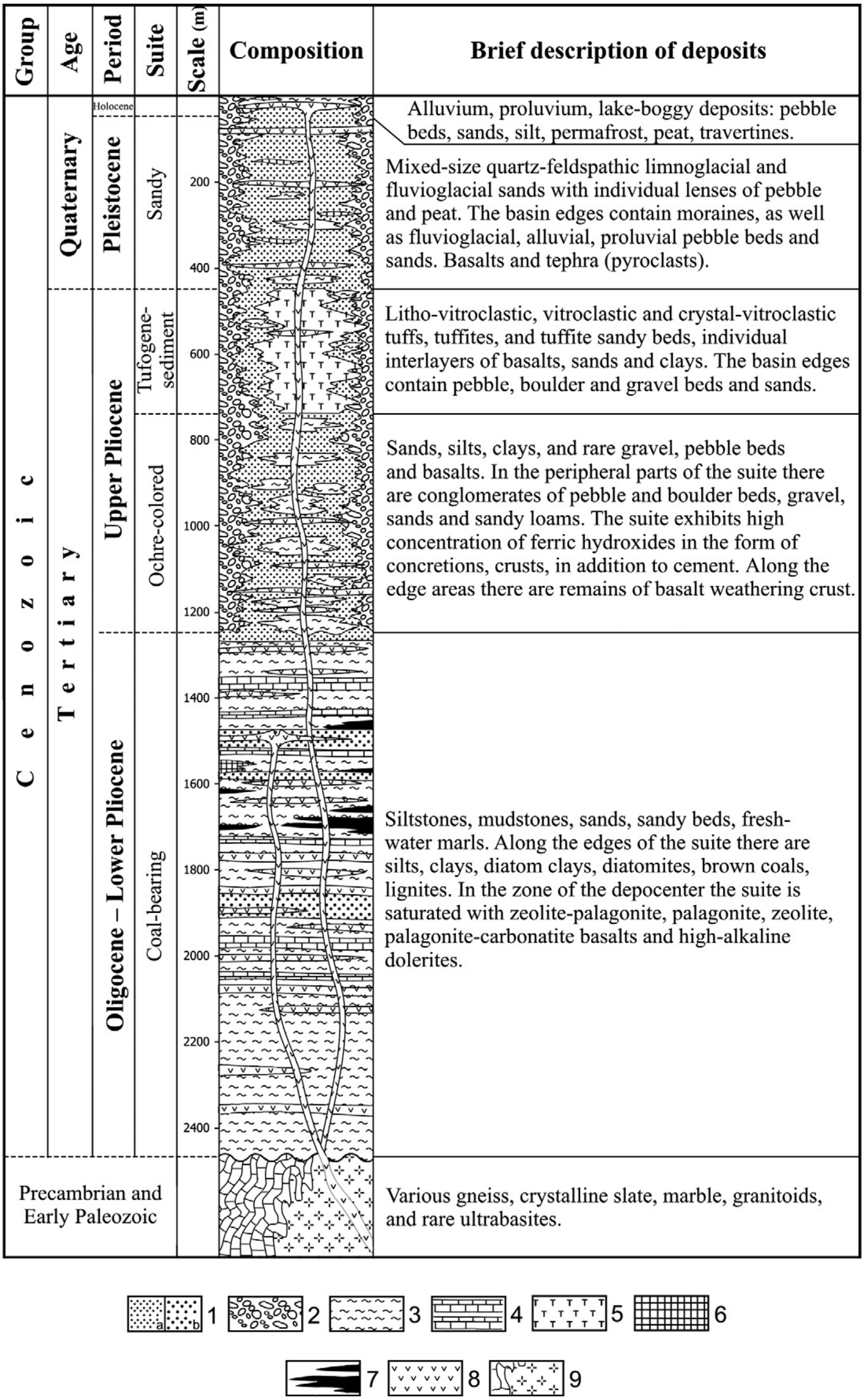

Fig. 3. Stratigraphic column of the Tunka rift basin (after Logachev, 1974 and Mazilov et al., 1993): 1 – sands and gravels (a), sandstones (b); 2 – boulders–pebble beds and conglomerates; 3 – siltstones and mudstones; 4 – fresh-water marls; 5 – tuffs, tuffaceous sandstones; 6 – diatomaceous shales and diatomites; 7 – brown coals and lignites; 8 – basalts and tephra; 9 – crystalline bedrocks.

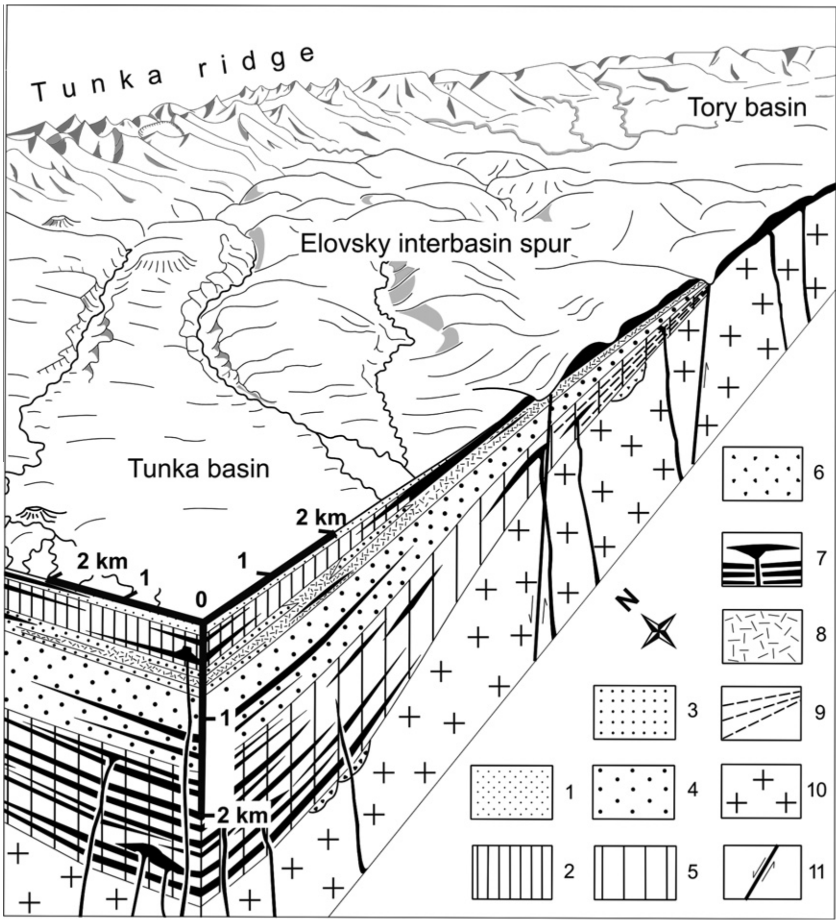

Fig. 4. Structural diagram of the central part of the Tunka rift basin (modified from Logachev, 1974). 1 – Late Pleistocene and Holocene deposits; 2 – Early and Middle Pleistocene sandy facies; 3 – Late Pliocene–Early Pleistocene Akhalik Formation; 4 – Pliocene Anosov Formation; 5 – Oligocene–Miocene Tankhoi (coal-bearing) Formation; 6 – Cretaceous weathering crust; 7 – basalts; 8 – pyroclasts; 9 – coal layers; 10 – crystalline basement; 11 – faults.

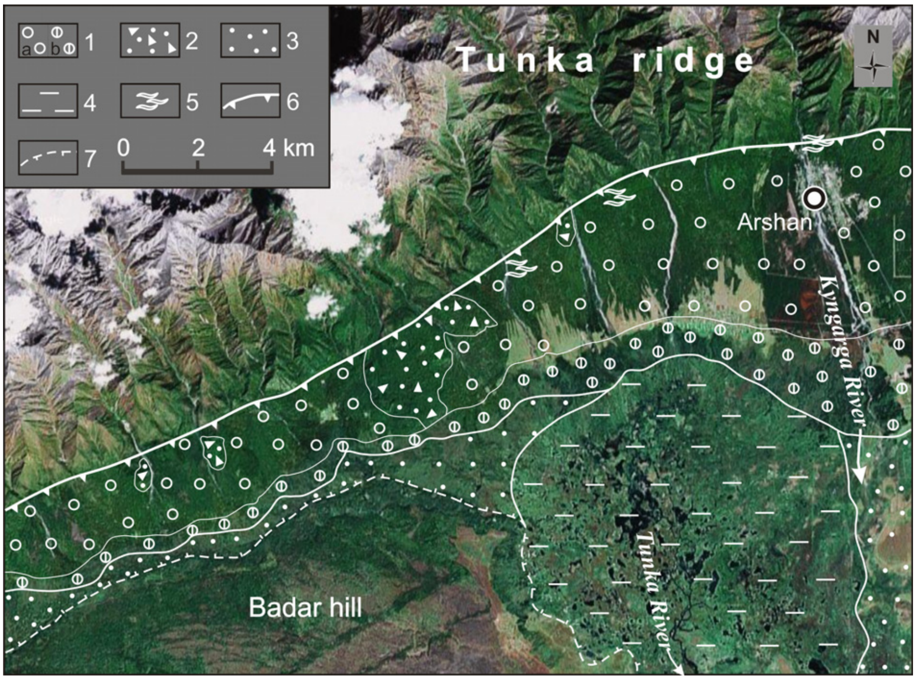

Fig. 5. Satellite image (Landsat 7, 2000) of the northern edge of the Tunka rift basin and surroundings. 1 – Belt of integrated alluvial fans (a – Holocene; b – Late Pleistocene); 2 – eroded fragments of moraine deposits (Late Pleistocene); 3 – alluvial floodplain (Holocene); 4 – lake-bog plain in area of intensive contemporary subsidence (Holocene); 5 – area of travertine accumulation (Holocene); 6 – Arshan palaeo-seismic dislocation represented by "scarp-micrograben" system (Late Pleistocene–Holocene); 7 – anticlinal massif composed of sands (Late Pleistocene).

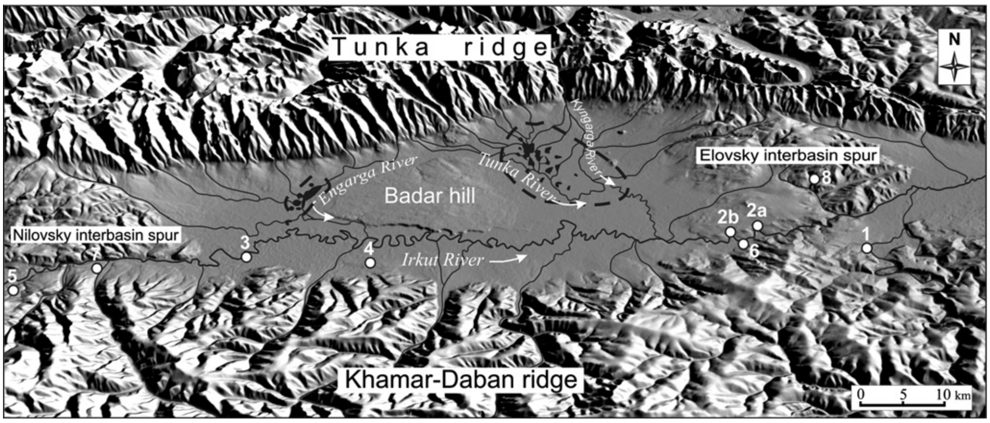

Fig. 6. The central part of the Tunka rift basin (based on SRTM digital elevation data). Dashed line indicates areas of recent subsidence. Numbers identify locations of stratigraphic sections: 1 – Slavin Yar; 2a – Belyi Yar I; 2b – Belyi Yar II; 3 – Shimki; 4 – Kyren; 5 – Turan; 6 – Zaktui; 7 – B. Zangistan; 8 – Elovka.

tributaries, the major stream system of the Tunka rift basin. The valley width is up to 5–7 km, with a commonly marshy floodplain. The plain exhibits two terraces located within or near the inter-valley spurs. The terraces are 8–12 m and 14–18 m high, with deformed surfaces. The higher terrace formed during the Late Pleistocene (70000–15000 yr BP) and the lower terrace at the end of the Late Pleistocene–Early Holocene (Shchetnikov and Ufimtsev, 2004; Ufimtsev et al., 2003). In the central parts of the major basins, where the intensity of recent tectonic subsidence increases, the height of the terraces located above the floodplain decreases, gradually pinching out. The second type of valley is represented by tilted piedmonts, which includes a series of merged detrital fans formed by various streams and ranging between 3 and 5 km in width. Along the foothills of the Tunka Range the structure of tilted piedmonts also includes eroded terminal glacial moraines.

3. Results

The discussion that follows is divided into two parts. The first part deals with the general surficial Quaternary geology of the Tunka rift valley. The second part is a description of key cross-sections studied in the field, including the reconstruction and chronology of Quaternary events.

3.1. Quaternary deposits of the Tunka rift basin

Quaternary deposits of the Tunka rift form three sedimentary facies, including boulder–pebble gravels, sand, and loess. The boulder–pebble gravels are represented at the marginal parts of larger basins (i.e., at the base of mountains) and also occupy the bottom of relatively small depressions or basins (Bistraya and Mondin). In the central parts of the larger basins (for example the Tunka and Tory) sands replace the boulder–pebble gravels. In the contact zone of sand and boulder–pebble gravels these sediments are commonly interbedded. Overlying the boulder–pebble gravels and sand is loess that covers wide areas of the Tunka rift valley.

3.1.1. Boulder–pebble gravels

Boulder–pebble gravels are represented by coalescing alluvial fans that are found, for example, at the base of the fault scarp of the Tunka Range (Fig. 5). This is typical proluvium (grey-coloured, poorly sorted, with varying degrees of roundness), with a gradual facies change toward the centre of the basin, from block-boulders to pebble-debris with light grey, fine-laminated sand. The formation of alluvial fans, up to 500 m of sediment thickness, is divided into two stages: Late Pleistocene and Holocene (Shchetnikov, 1999). During the Late Pleistocene, large alluvial fans formed with a significant portion of glacially-derived material, whereas during the Holocene the amount of debris accumulation decreased substantially with smaller cones forming on the surface of larger ones. Alluvial fans in the mountain valleys also include glacially-derived sediment dating to 110000–55000 yr BP and 22000–10000 yr BP (Shchetnikov and Ufimtsev, 2004), which are represented by washed out fragments of terminal moraines. At the mouth of the Ihe-Uhguni River where it enters the Khoito Gol basin (Fig. 2), there is an undisturbed terminal moraine that dammed a system of glacial lakes. In general, however, the occurrence of glacial deposits in the Tunka rift basin is limited.

In the western Tunka rift, boulder–pebble deposits are found in basins formed by various tectonic events. On elevated levels, boulder–pebble gravels with thin, up to 1 m lenses of sand, form a series of river terraces. These terraces occur at a height of up to 150 m on the slope of the antecedent valley of the Irkut River, which cuts through the bench of crystalline rocks between the Mondin and Khoito Gol basins. Based on the available thermoluminescence dates, the age of the upper terrace is approximately 70000 yr BP, whereas the lower ones are between 120000 and 100000 yr BP (Ufimtsev et al., 2003). Incised into the bedrock, under the alluvial deposits, is a series of rock-cut terraces with boulder–pebble gravels, suggesting that sediment deposition was characterised by gradual infilling of an erosive cut. Boulder–pebble gravels are found in almost all of the inter-basin spurs of the Tunka rift. In the southern part of the Tunka rift basin, south of the Irkut River, the largely alluvial sediments contain lenses of boulder–pebble gravel associated with alluvial fans of the river valleys which flow from the Khamar Daban Range.

3.1.2. Sands

The second type of Quaternary sediment in the Tunka rift basin is represented by sand deposits which are dominated by floodplain alluvium. Overall, the sandy deposits display minimal variation in terms of grain size and mineral composition, and often include lenses of fine pebble gravel and occasionally with buried soils that have been dated to the last interstadial period (55000–22000 yr BP). The sand units are also characterised by a homogenous light yellow-grey colour and polymictic composition. These layers are commonly thin, up to a few centimeters thick, and flat lying, with some current cross-bedding usually found near the basin boundaries. Some sands contain cold water diatoms including *Melosira islandica*, *Melosira scabrosa*, *Cocconeis placentula* and *Didymosphenia geminate* (Florensov, 1964), as well as shells of terrestrial molluscs,

Table 1
Radiocarbon and thermoluminescence dates reported from study sites in the Tunka rift basin.

Site	Stratigraphic context	Depth (m)	Uncalibrated ^{14}C yr BP	Method	Material	Identification	Lab ref. no.
Anchuk	Alluvial sand	2.3	20 390 ± 120[h]	^{14}C AMS[i]	Bone	*Mammuthus primigenius* Blum.	OxA-24817
B. Zangistan	Buried soil	2.7	32 570 ± 340[h]	^{14}C AMS[i]	Bone	*Spiroceros kiakhtensis*	OxA-19193
B. Zangistan	Buried soil	2.7	34 300 ± 550[h]	^{14}C AMS[i]	Bone	*Spiroceros kiakhtensis*	OxX-22519
B. Zangistan	Buried soil	2.7	34 800 ± 600[h]	^{14}C AMS[i]	Bone	*Spiroceros kiakhtensis*	OxA-22518
Belyi Yar I	Buried peat	22.5	44 200 ± 4500[h]	^{14}C	Wood charcoal		IGAN-3370
Belyi Yar II	Buried peat	13.4	26 250 ± 300[b]	^{14}C	Organic remains		SOAN-577
Belyi Yar II	Alluvial sand	15.0	35 440 ± 1860[c]	^{14}C	Organic remains		SOAN-3144
Belyi Yar II	Alluvial sand	15.8	40 860 ± 480[d]	^{14}C	Organic remains		SOAN-141
Elovka	Loess-shaped sandy loam	1.2	18 350 ± 75[h]	^{14}C AMS[i]	Bone	*Leo spelaea* Goldfuss	OxA-20672
Elovka	Loess-shaped sandy loam	4.5	22 000 ± 8300[g]	TL	Quartz sand		BurGIN 191
Kyren	Alluvial sand	11.0	31 500 ± 2300[g]	TL	Quartz sand		BurGIN 104
Shimki I	Alluvial sand	7.5	9260 ± 80[h]	^{14}C AMS	Wood fragment		TO-10549
Shimki II	Buried peat	6.1	11 180 ± 70[f]	^{14}C	Organic remains		GIN-8091
Slavin Yar (Zun-Murin)	Buried soil	8.0	37 790 ± 310[a]	^{14}C AMS	Wood charcoal		TO-13278
Slavin Yar (Zun-Murin)	Buried soil	11.0	45 810 ± 4070[a]	^{14}C	Wood charcoal		IGAN-3133
Tibelti	Buried soil	2.3	9280 ± 40[e]	^{14}C	Organic remains		SOAN-1596
Tibelti	Alluvial sand	13.0	31 860 ± 370[e]	^{14}C	Organic remains		SOAN-1583
Tibelti	Alluvial sand	14.8	40 060 ± 820[e]	^{14}C	Organic remains		SOAN-1592
Turan	Alluvial sand	6.9	58 000 ± 10 000[g]	TL	Quartz sand		BurGIN 202
Turan	Alluvial sand	25.0	76 000 ± 9000[g]	TL	Quartz sand		BurGIN 201
Zaktui	Loess-shaped loam	2.3	33 090 ± 250[h]	^{14}C AMS[i]	Bone	*Mammuthus primigenius* Blum.	OxA-21014
Zaktui	Loess-shaped loam	2.3	33 190 ± 240[h]	^{14}C AMS[i]	Bone	*Mammuthus primigenius* Blum.	OxA-21015
Zaktui	Loess-shaped loam	2.3	35 560 ± 300[h]	^{14}C AMS[i]	Bone	*Crocuta spelaea*	OxA-19719
Zaktui	Loess-shaped loam	2.3	36 000 ± 800[h]	^{14}C AMS[i]	Tusk	*Mammuthus primigenius* Blum.	OxA-88[j]

[a] Shchetnikov et al. (2009).
[b] Adamenko et al. (1975).
[c] Kul'chitsky et al. (1994).
[d] Firsov et al. (1985).
[e] Logachev (1981).
[f] Trofimov et al. (1995).
[g] Ufimtsev et al. (2003).
[h] New dates reported in this study.
[i] Indicates ultra-filtration samples.
[j] Indicates sample with low collagen yield (5.71 mg collagen from 330 mg original weight).

mostly *Vallonia* ex gr. *pulchella*, *Gastrocopta theeli*, and *Vertigo* cf. *modesta* (Ufimtsev et al., 2002; Shibanova, 1996). The subaqueous malacofauna is represented only by individual samples of *Anisus acronicus* (Ufimtsev et al., 2002). In addition, the sands contain fossil remains of Late Pleistocene mammals adapted to cold conditions, including *Mammuthus primigenius* Blumenbach, *Coelodonta antiquitatis* Blumenbach, *Cervus* sp., *Capreolus* sp. and *Bison priscus* (Shchetnikov and Ufimtsev, 2004).

The western part of the Tunka rift basin is dominated by Holocene floodplain sandy loam and loam, with alluvial sand commonly found along the southern edge of the rift valley. The loamy deposits often contain numerous buried soil horizons between 7 and 15 cm thick. Overall, recent floodplain sediment is up to 7–9 m thick and overlies Late Pleistocene sands and pebble-beds estimated to be approximately 50 000 yr BP (Ufimtsev et al., 2003). Sediment deposition occurred during periods of short-term and intensive subsidence in lake-like widening of the river channels. A similar situation is currently observed in the northern part of the Tunka basin, where on the north side of the Irkut River in the valleys of the Engarga and Tunka Rivers, areas of contemporary subsidence are covered with numerous small lakes (Figs. 5 and 6).

Whereas Late Pleistocene floodplain and lake deposits developed predominantly in widening of river channels, there were no large lakes in the Tunka valley by the end of the Late Pleistocene and during the Holocene. The basins are typically filled with channel and floodplain alluvium, mostly composed of sand. One of the factors that prevents, and prevented in the past, the formation of large lakes in the Tunka basin is the significant inclination of its surface – 900 m over the course of 200 km. This is further compounded by the elevated eastern edge of the rift valley above Lake Baikal.

3.1.3. Loess

The third dominant sediment type is blanket loess composed predominantly of sandy loam and loam with variable loess-like characteristics. Blanket loess, commonly with gruss and gravel, forms beneath talus cones such as those at the base of the northern slope of the Khamar Daban Range. Loess is also found in the smaller river valleys that drain low mountain inter-basin spurs. Along the bottom of the rift basins these sediments are present in the form of relatively thin (a few meters) surface layers covering lower terraces above the floodplain. The deposits are normally deformed by cryoturbated involutions. The formation of these cryoturbation features probably occurred during late glacial time and the beginning of the Holocene (Shchetnikov and Ufimtsev, 2004).

Due to its geographic position, the Tunka rift basin constitutes a sub-latitude wind corridor. The valley is dominated by eastern winds but western winds are the most powerful. Sands therefore move from west to east, the direction opposite of the average local wind. The major period of aeolian sand deposition was between 29 000 and 8000 yr ago (Shchetnikov and Ufimtsev, 2004). During this time, dunes formed on the Badar alluvial plain, on the western slopes of the Elovsky and Nilovsky ridges, and on the northern side of the Khamar Daban Range. At present, aeolian sand deposition occurs at the edge of pine groves, and on the steppe lands along valley bottoms and near the western slope of the Nilovsky ridge in the Khoito Gol basin, partly caused by recent anthropogenic impact on the natural landscape.

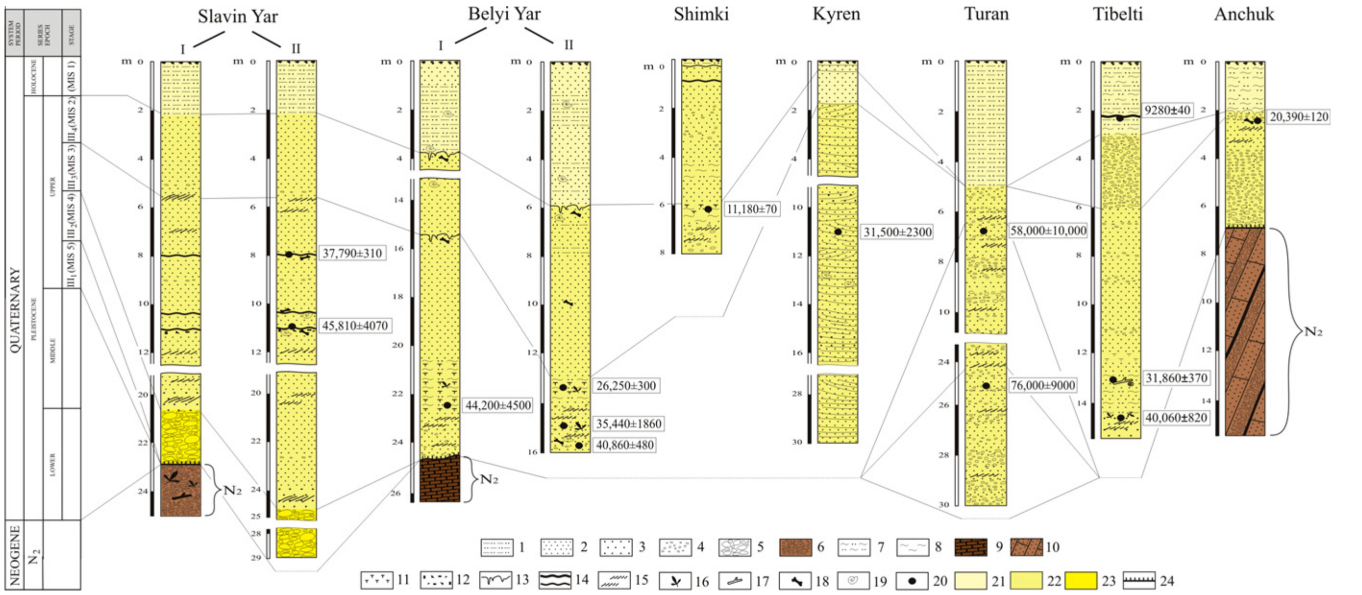

Fig. 7. Stratigraphic columns of Quaternary sediments of the Tunka rift basin. 1 – Sandy loam; 2 – medium and fine-grained sands; 3 – coarse sand; 4 – gravel with fine pebbles; 5 – boulder–pebble beds; 6 – boulder–pebble conglomerates; 7 – loams; 8 – clays; 9 – tuffaceous sandstone; 10 – interlayering of sandstone and conglomerates; 11 – peat; 12 – concentration of wood coal; 13 – cryoturbated horizons; 14 – soil horizons; 15 – cross-bedding; 16 – detrital organic remains; 17 – large wood fragments; 18 – mammal bones; 19 – concentrations of mollusc shells; 20 – location of dating samples; 21–23 – lithological facies: loess (21), sand (22) and boulder–pebble gravels (23); 24 – non-depositional unconformity.

Within the loess complex, special mention should be given to the homogeneous grey loam and sandy loam deposits with relatively thin and barely visible wave-like layering, which morphologically reflects wind ripples which are also currently being formed in the Tunka Cis-Baikal region. These deposits are located at the base of the inclined piedmont plains bordering the slope of the Khamar Daban Range. The loams are of pale grey colour with homogeneous structure and columnar jointing. The vertical walls are stable and by all characteristics resemble the Chinese loess of Shansi and Ordos (Derbyshire, 1983).

Blanket loess deposits in the Tunka rift basin also preserve remains of Late Pleistocene mammals, predominantly from the "mammoth-steppe" fauna (*M. primigenius*, *C. antiquitatis*, *B. priscus*, *Cervus elaphus*, *Capreolus pygargus*, *Equus hemionus*, *Alces* sp., etc.). As part of a recent collaborative dating program with colleagues from the Natural History Museum and Durham University in the United Kingdom, we obtained using ultra-filtration methods nine new radiocarbon AMS dating results from faunal bone samples from the Tunka rift basins (Table 1). An adult *M. primigenius* vertebra found at a depth of 2.3 m in stratified alluvial sediments at Anchuk returned an age of 20390 ± 120 ^{14}C yr BP (OxA-24817), and *M. primigenius* remains found at a depth of 2.3 m in the loess section Zaktui yielded ages of 33 090 ± 250 ^{14}C yr BP (OxA-21014) and 33 190 ± 240 ^{14}C yr BP (OxA-21015). In addition, remains of *Crocuta spelaea*, an occurrence rare in the Pribaikalye and Trans-Baikal areas, have also been recovered at Zaktui and dated to 35560 ± 300 ^{14}C yr BP (OxA-19719). This is the first reported age of this species in eastern Siberia. Also found in association with these fauna at Zaktui was a juvenile mammoth tusk, which yielded an age estimate of 36000 ± 800 ^{14}C yr BP (OxA-88**), generally consistent with the adult *Mammuthus* and *Crocuta* specimens dated from this bone-bearing horizon. In addition, *Spiroceros kiakhtensis* remains found by A. Fedorenko in 1980 within a buried soil in the loess-loam section at B. Zangistan were recently dated for the first time in Russia by AMS. The ages of 32570 ± 340 ^{14}C yr BP (OxA-19193), 34800 ± 600 ^{14}C yr BP (OxA-22518), and 34300 ± 550 ^{14}C yr BP (OxA-22519) are in good agreement with dating results from other Late Pleistocene loess sections in the Tunka basins. Finally, remains of *Leo spelaea*, another particularly rare species for the Late Pleistocene Baikal region, were found at a depth of 1.2 m in the loess sediments at Elovka and dated at 18350 ± 75 ^{14}C yr BP (OxA-20672), whereas the age at the base of this section is estimated to be c. 22 000 yr BP (Ufimtsev et al., 2003).

3.2. Key Late Quaternary stratigraphic sections in the Tunka rift basins

The basins of the Tunka rift contain only a few natural exposures of Quaternary sediments of pre-Holocene age. These include Slavin Yar, Belyi Yar, Tibelti, Anchuk, Shimki, Kyren, and Turan stratigraphic sections (Figs. 6 and 7), described here in detail. These profiles allow investigation of the terrace complexes of the Tunka valley which are commonly disturbed by tectonic activity and erosion. While new chronological information is now available for each of these sections (Table 1), the resolution of these temporal data still pose limitations in accurate cross-correlation of sedimentary units. Nevertheless, these results provide important new evidence for the Late Quaternary evolution of the Tunka rift basin.

3.2.1. Slavin Yar

The Slavin Yar section (Figs. 6–8) is the most extensive of the Quaternary sedimentary exposures in the Tunka rift basin (Shchetnikov et al., 2006, 2009). The section is located on the east side of the Zun-Murin River (51°42′15″N, 102°39′30″E), approximately 11 km from its confluence with the Irkut River (Fig. 9). The thickness of the section, comprised mostly of alluvial sediments, is approximately 30 m and extends for more than 1 km.

The section includes four major sediment units (bottom to top): (1) lithified brown-ochre boulder–pebble gravels with interlayers and lenses of sandstone; (2) light-grey pebble–boulder gravels; (3) light-brown sands and sandy loam; and (4) loess.

(1) These gravels overlie crystalline bedrock and are exposed at the southwest edge of the section about 2 m above the present average water level of the Zun-Murin River. The gravels can be traced for a distance of about 200 m, characterised by northeast dipping beds which pinch out upstream. A main feature of these deposits is their distinct brown-ochre colour caused by authigenic ferric hydroxide. The stony matrix is well rounded but poorly sorted, with clast sizes averaging

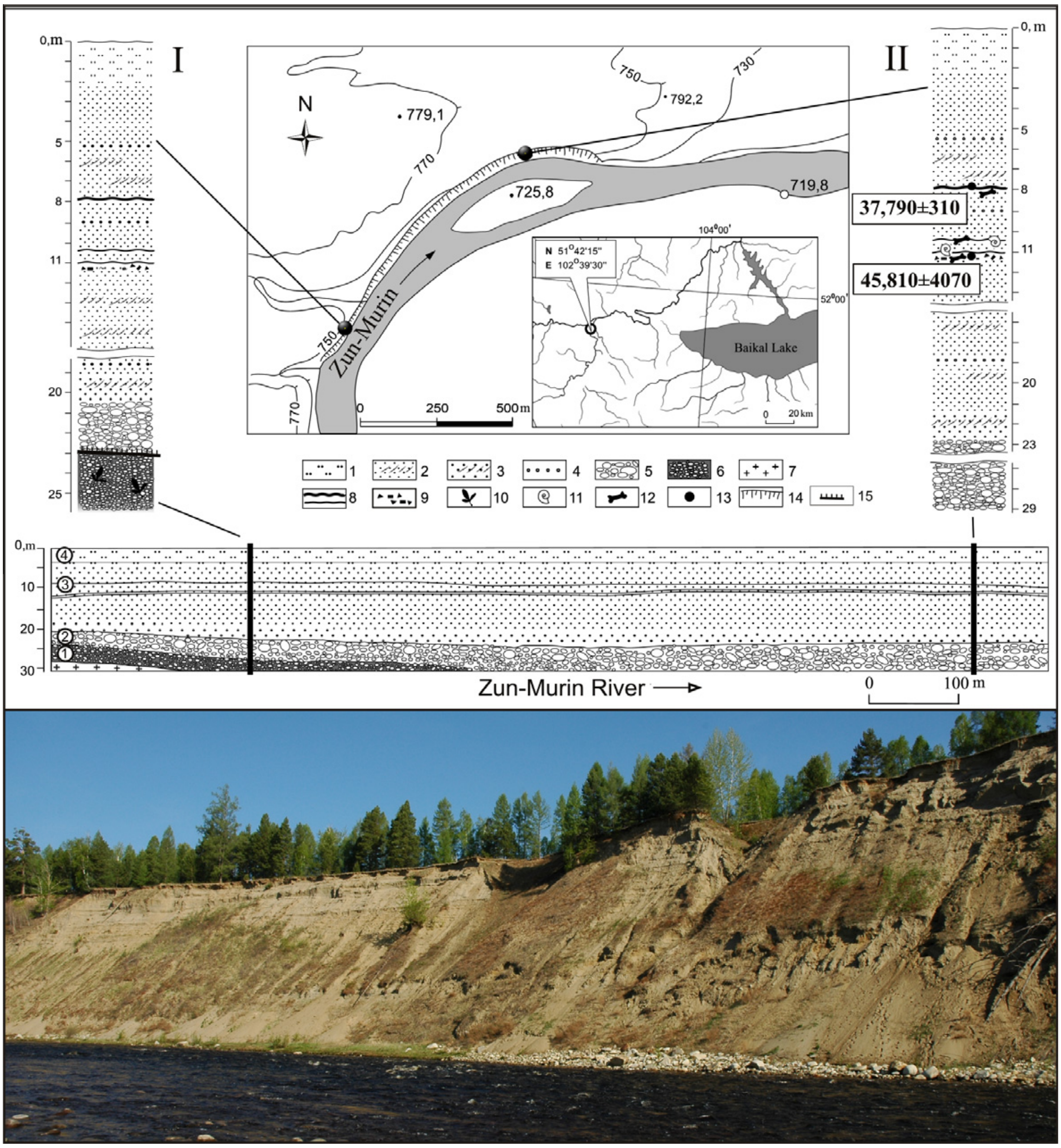

Fig. 8. Slavin Yar section. 1 – Sandy loam cover loess; 2 – mixed sand sizes; 3 – coarse sands with elements of channel cross-stratification; 4 – interlayers of gravel with fine pebbles; 5 – boulder–pebble beds; 6 – boulder–pebble conglomerates; 7 – crystalline bedrock; 8 – soil horizons; 9 – concentration of wood coal; 10 – large wood fragments; 11 – concentration of shells; 12 – mammal bones; 13 – radiocarbon samples; 14 – section exposure; 15 – non-depositional unconformity. Numbers in circles indicate: (1) brown-ochre boulder–pebble gravels; (2) light-grey pebble–boulder gravels; (3) sands; (4) loess.

10–15 cm in diameter, comprised of more than 80% basalt. The gravels also contain numerous inclusions of detrital vegetation, and lignitised wood (*Pinus* sp.) fragments including trunks up to 30 cm in diameter. The pollen spectrum is dominated by *Pinus sibirica, Pinus silvestris, Picea* sect *Eupicea*, and *Corylus* sp., along with the presence of *Tsuga* sp. Assuming rapid deposition and short distance transport, the composition of the flora indicates a climate at the time of sediment formation considerably milder than present conditions. The sediment characteristics of authigenic ferric hydroxides and brown-ochre colour, high degree of lithification, presence of *Tsuga* pollen, and coarse mechanical composition correlates well with the Late Pliocene Anosov (ochre) series of Cis-Baikal (Logachev, 1958a).

(2) The upper boundary of the brown-ochre boulder–pebble gravels forms a non-depositional unconformity with overlying, tightly consolidated light-grey pebble–boulder gravels with lenses of large-grained, obliquely laminated sands (material analogous to the contemporary deposits of the Zun-Murin River). The bed thickness of these deposits is 1–2 m in the southeast and increases towards the northeast to 6–7 m. In general, the gravels are poorly sorted. Maximum clast sizes reach 80 cm in diameter, whereas the average size is about 15–30 cm. In the upper portion of the beds there are traces of subhorizontal macro-layering with varying degrees of sorting. The mineral composition of this deposit includes mostly gneiss, slate, marble, and granitoids (Fig. 10). The fine grained fraction is dominated by hydromica and carbonates.

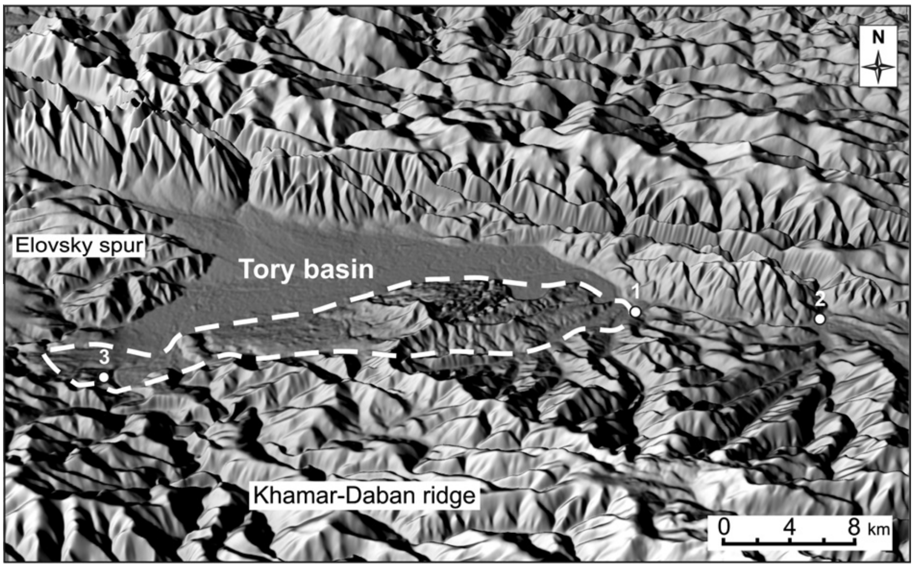

Fig. 9. The eastern part of the Tunka rift basin (based on SRTM digital elevation data). Dashed line indicates the area of inversion (uplifted part of the Tory basin). Numbers correspond to sections: 1 – Tibelti; 2 – Anchuk; 3 – Slavin Yar.

(3) The overlying sand units, predominantly floodplain sediments, are between 15 and 20 m thick. The material is comprised of light-brown, fine-grain sands with interlayers of medium-grain sands with some thin interlayers of pebble gravel. The sequence of horizontal layers alternate with gently sloping wave-like layers and to a lesser degree those which are obliquely laminated. In general, grain-size decreases from the lower layers upwards. The fine grained fraction of these sands, as well as in the underlying sediment, is dominated by hydromica mixed with carbonate. Within the sand unit, at 11 m, 10.5 m, and 8 m below the modern surface, there are three layers each approximately 10–30 cm in thickness consisting of dark grey to black loam and sandy loam with inclusions of discontinuous buried soil horizons containing peat, wood charcoal and terrestrial faunal remains. The layer at 11 m contains Late Pleistocene mammals such as *M. primigenius* Blumenbach, *C. antiquitatis* Blumenbach, *Cervus* sp., and *Capreolus* sp. This layer also preserves a rich variety and abundance of molluscs, including *Aplexa hypnorum* (Linn.), *Galba truncatula* ex. gr. *sibirica* (West.), *Succinea* ex. gr. *putris* (Linn.), *Vallonia costata* (Mull.), *Pupilla muscorum* (Linn.), *Vertigo* cf. *modesta*, *Nesovitrea hammonis* (Strom.), *Bradybaena schrencki* (Midd.), and *Euconulus fulvus* (Mull.) (Shchetnikov

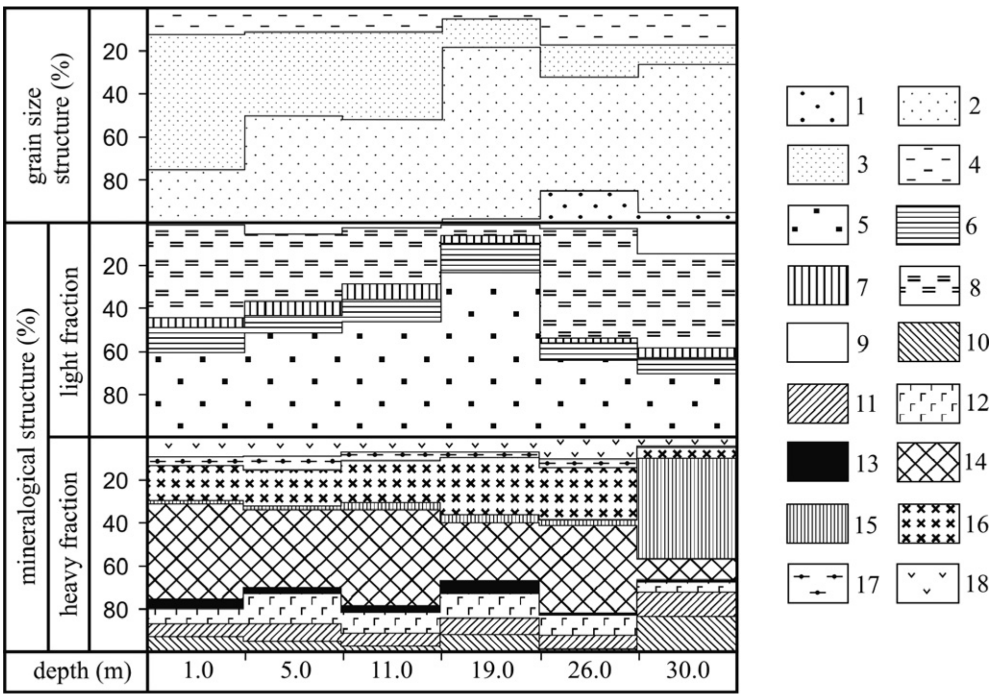

Fig. 10. Lithological diagram of psammite-pelite fractions of deposits at Slavin Yar. Granulometric composition (1–4): 1 – >1 mm; 2 – 0.1–1 mm; 3 – 0.01–0.1 mm; 4 – <0.01 mm. Minerals of light fraction (5–9): 5 – quartz; 6 – plagioclases; 7 – potassium feldspar; 8 – mica; 9 – rock fragments. Minerals of heavy fraction (10–18): 10 – magnetite; 11 – ilmenite; 12 – garnet; 13 – sphene; 14 – amphibole; 15 – pyroxene; 16 – epidote; 17 – apatite; 18 – other minerals.

et al., 2009). All of these species currently inhabit the southwest Cis-Baikal area. Charcoal from the 11 m layer yielded a radiocarbon age of 45 810 ± 4070 ^{14}C yr BP (ИГАН 3133). Bone fragments of *Capreolus* sp. were also found within the layer located at 10.5 m. The layer at 8 m yielded remains of *M. primigenius* Blumenbach, *C. antiquitatis* Blumenbach, and *Cervus* sp. Charcoal from this layer is radiocarbon dated to 37 790 ± 310 ^{14}C yr BP (TO-13278). Limited pollen analyses from the sand series and underlying pebble boulder deposits support vegetation characteristics of the Late Pleistocene (Shchetnikov et al., 2009).

(4) The sands are overlain by a 1.5–2 m-thick sequence of white, carbonate-rich, loessic sandy loam with characteristic columnar structure.

3.2.2. Belyi Yar

The Belyi Yar section (51°43′21″N, 102°40′21″E) is located on the north side of the Irkut River at the foot of the Elovsky inter-basin spur in the eastern part of the Tunka rift basin (Figs. 6 and 7). The section is exposed in two outcrops located approximately 2 km from each other, identified as Belyi Yar I and Belyi Yar II. The sections are divided into two major units: a lower tuff layer and an upper sandy layer. The former is exposed only at Belyi Yar I and consists of green–brown Late Pliocene tufftaceous breccia and tufftaceous sandstone characterised by platy structure, eroded at the upper boundary. The overlying sandy beds are composed of fluvial deposits sub-divided into three layers. The base of the sandy sequence consists of alluvium comprised of brown-ochre to black coarse-grain sands with cross- and wave-like bedding. These sands are well sorted with thin interlayers of sub-angular boulder–pebble gravel beds. Above, there are oxbow sediments consisting of horizontally layered silty dark grey loams and sandy loams which contain thin lenses of peat which alternate with ochre-coloured ferric sands of variable grain size. The uppermost is composed predominately of floodplain sands. These sediments are light-grey, medium-grain sands with small pebble inclusions with horizontal and wave-like bedding. Approximately 15 m and 6 m below the surface at Belyi Yar I and II, respectively, there are two horizons of cryogenic involutions with ice-wedge pseudomorphs present in the upper part of the floodplain sediment. Most likely, the lowermost horizon corresponds to the beginning of the last glacial period, and the upper horizon to the end of the last glacial period (22 000–10 000 yr BP). The upper boundary of the sands has been modified by aeolian processes.

Belyi Yar is one of the most extensively studied Pleistocene sections of the Tunka Cis-Baikal area but the chronology is disputed. It was believed that the lower part of the sandy series contains Early Pleistocene sediment, while its central part is Middle Pleistocene in age (Ravskii et al., 1964). However, new radiocarbon and biostratigraphic data obtained from the sections suggest that the sediment is Late Pleistocene in age (Adamenko et al., 1975; Kul'chitsky et al., 1994). Results from radiocarbon dating of wood charcoal from buried peat in the lower portion of Belyi Yar I yielded an age of 44 200 ± 4500 ^{14}C yr BP (IGAN-3370; Fig. 7).

3.2.3. Tibelti

The Tibelti section (51°45′20″N, 103°15′50″E) is situated along the south bank of the Irkut River within the Tory basin near the mouth of the Bolshaia Tibelti River (Figs. 6, 7 and 9). The exposure reveals two terrace deposits. The sediment of the lower terrace is dominated by cross-bedded sandy channel alluvium, whereas the upper terrace is characterised by horizontally and wave-like layered floodplain sands of various grain sizes. The sands are overlain by a layer of ochre-coloured and poorly sorted boulder–pebble gravels. A similarity between the terrace deposits is the presence of horizons of similar thickness at their base with an abundance of fossilised organic remains. Initially, these thin-layered, fine-grain sands with interlayers of peat and gyttja and numerous inclusions of organic detritus were considered to date to the Early Pleistocene (Ravskii et al., 1964). Subsequent radiocarbon dating indicates that this deposit is Late Pleistocene in age (31 860 ± 370 ^{14}C yr BP [SOAN-1583]; 40 060 ± 820 ^{14}C yr BP [SOAN-1592]) (Logachev, 1981), results generally consistent to that of the Belyi Yar I section.

3.2.4. Anchuk

The Anchuk section (51°45′47″N, 103°25′42″E) is located in the Bistraya basin on the west bank of the Irkut River where a complex of terrace levels occur as the river enters the Zyrkuzun gorge (Figs. 7 and 9). The three terraces above the floodplain (60 m, 40 m and 25 m) are eroded terraces with scattered pebble debris on their surfaces. A lower 14 m terrace is socle-shaped and composed of lithified alluvium, including layers of ochre-coloured fine-pebble and boulder–pebble gravels and sands of various grain sizes with fragments of lignitised wood and interlayers of lignite. The sediments, exposed 7–8 m above the Irkut River level, dip between 25° and 61° toward the west. Within the 14 m terrace, one of the sand layers contains remains of small rodents of the olagurodont-mimomis group (Adamenko et al., 1984). These data allow correlation of this deposit with the Anosov series of the Late Pliocene. Approximately 400 m downstream from here, a landslide has exposed clayey sandstone with lenses of coal and fragments of carbonised wood which directly overlie the crystalline bedrock of the Bistraya basin. V.M. Klimanova (Kul'chitsky, 1985) obtained pollen from here which corresponds to the coal-bearing series which formed during the Late Oligocene and Miocene (Logachev, 1958a).

The Neogene cemented deposits of the 14 m terrace underlie alluvium. The lower part of this alluvium contains a poorly sorted grey boulder–pebble gravel overlain by pebble gravel beds and cross-layering sands of various grain-sizes. The latter yielded a vertebra of *M. primigenius* Blum. dated to 20 390 ± 120 ^{14}C yr BP (OxA-24817). The section is topped by a thin, loamy covering up to 1.5 m thick, which at the base contains ice-wedge pseudomorphs and other evidence of cryoturbation that formed during the last glacial stage of the Late Pleistocene.

3.2.5. Shimki

The Shimki section (51°40′08″N, 101°57′14″E) is situated near the western periphery of the Tunka basin, along the south bank of the Irkut River near Shimki village (Figs. 6 and 7). The exposed section (Shimki I) consists of sandy-clayey deposits approximately 8 m thick, which forms the first terrace above the floodplain. In the lower 0.5 m of the section, gravel and poorly sorted cross-bedded sands dominate. The sands are light brown with ochre-coloured interlayers with inclusions of small pebbles. These are overlaid with relatively thin-layered, dark grey clays with inclusions of wood fragments which alternate with ochre-coloured and horizontally layered fine-grain silty sands. The thickness of this sequence, interpreted to be oxbow sediments, is 1.5–2 m. Above these sediments are floodplain deposits of cryoturbated, relatively homogenous, thin-layered grey sandy loams with interlayers of dark grey loams, sands and peat up to 5 m in thickness. The lower sandy-clay deposits were described by Ravskii et al. (1964) and considered to be Late Pleistocene (Kazan interglacial period; 130 000–110 000 yr BP) in age on the basis of fauna remains. However, radiocarbon dating of a wood fragment from the base of this sequence returned an age of 9260 ± 80 ^{14}C yr BP (TO-10549).

On the north bank of the Malaia Taitorka River, opposite Shimki village, another section of this terrace is exposed (Shimki II), which has been described by Ufimtsev et al. (2002). Its composition is very similar to the exposure described above, with a reported age of 11 180 ± 70 ^{14}C yr BP (GIN-8091) from peat found 5.8 m below the surface (Fig. 7).

3.2.6. Kyren

The Kyren section (51°40′33″N, 102°10′02″E) is located in a quarry exposure approximately 30 m thick connected to the gently sloping, irregular plain along the base of the Khamar Daban Range (Figs. 6 and 7). In general, the section is homogeneous with medium- and fine-grained sands. Most layers are 1–1.5 m thick with some thin, horizontal wave-like and parallel floodplain stratification a few centimeters thick. In the upper part of the section, these horizons alternate with cross-bedded sandy layers of similar thickness, inclined to 20–25°. The sands indicate a channel facies with rare wedge-shaped forms. In some cases, the layers are crumpled into "turbulent" textures and contain concentrations of olive-coloured rolled layers of mud up to 5 cm in size.

The Kyren section contains numerous well preserved shells of relatively cold-adapted molluscs, including *Aplexa hypnorum* (L.), *Succinella* ex gr. *oblonga* (Drap.), *Succinea putris* (L.), *Vallonia tenuilabris* (Al. Br.), *Pupilla* ex gr. *muscorum* (L.), *P.* tx gr. *sterrii* (Voith), *Vertigo* (I.) *modesta*, *V. alpestris* and *Columella columella* (G. Martens) (Ufimtsev et al., 2002). All of these species are presently found in Cis-Baikal. A thermoluminescence date obtained from sands at 11 m below surface yielded an age of 31 500 ± 2300 yr BP (Ignatova et al., 1999).

3.2.7. Turan

The Turan section (51°38′38″N, 101°43′56″E) is situated on the elevated eastern edge of a deformed terrace of the Tury basin which borders an inter-basin spur, rising to 100 m (Figs. 6 and 7). This part of the basin belongs morphologically to the Khamar Daban Range.

Exposed in a road side quarry, the thickness of unconsolidated deposits in the Turan section is approximately 25–30 m. In the upper part of the section, the ochre coloured cross-bedded gravel-bearing fluvial sands alternate with relatively thin layers of poorly sorted pebble beds. These are overlain by a 1 m thick dark grey loam and up to 2 m thick pale sandy loam on top. A thermoluminescence date of 58 000 ± 10 000 yr BP was obtained on sands from 6.9 m below the surface. In the lower part of the sequence the section contains pebble beds, sand lenses and interlayers up to 1 m thick. Pebble layer thicknesses are up to 3–5 m. A thermoluminescence date from a depth of 25.0 m below surface yielded an age of 76 000 ± 9000 yr BP.

4. Summary

4.1. Cenozoic history of the Baikal rift zone

The Baikal rift zone began to form over 50 million years ago. According to N.A. Logachev (1978) the sedimentary infill of the rift basin can be sub-divided into two components based on age, deposit type, and structure. The lower part is Eocene–Early Pliocene in age, with deposits of fluvial sandstone, lacustrine siltstone and mudstone, and rare beds of palustrine brown coal, diatomite and marl near the periphery of the depressions. These sediments accumulated under subtropical (Eocene) to moderately warm (Oligocene, Miocene) climatic conditions. During the Eocene and Miocene, sedimentary basins were wider than in later periods. Plateau-like structures bordering the basins were dissected by rivers. At the margins of the basins, sharp unconformities divide lower and upper strata indicating increased tectonic activity and renewed erosion during the Middle Pliocene. At this time the boundaries of the basins reached their present position and Lake Baikal formed as a deep-water reservoir. The infill of the upper part of the rift basin is Middle Pliocene–Quaternary in age, dominated by fluvial coarse-grained sediments developed primarily at the margins of sedimentary basins where fragmented sandy-gravel, pebble and boulder deposits occur. Crustal movements along the Baikal rift are still taking place today.

4.2. Tunka rift during the Oligocene–Middle Pliocene

The Tunka rift forms part of the Baikal rift zone. The tectonic history of the Tunka rift is one of periodic change forming a variety of basins and uplifts. During the Oligocene–Middle Pliocene, relatively weak tectonic activity occurred with trough-like down warping of basins and the accumulation of more than 1700 m of carbon-rich fine-grained sediment, formed under humid and warm climatic conditions. The development of rich thermophilic broadleaf vegetation with an admixture of subtropical forms (e.g. magnolia, myrtle, and laurel) developed near the edges of the basins where with time coal developed. The central parts of the basins accumulated sands, silts, clays, and diatomaceous algae in lacustrine–marshy environments. At this time, Lake Baikal was connected to the Tunka rift basin. Relatively strong tectonic activity between the Miocene and Pliocene initiated widespread fluvial channel and floodplain sedimentation resulting in relatively thick beds of coarse alluvium.

4.3. Tunka rift during the Late Pliocene–Quaternary

During this period, deposition of coarse material exceeded 1000 m in some locations and basins acquired their present morphologies. Late Pliocene volcanic cinder cones and lava flows have been identified. Non-depositional unconformities mark the top of the Tertiary sediments. For the most part, the Late Pliocene climate was milder than today. Early and Middle Pleistocene deposits are poorly exposed being overlain by Late Pleistocene sediments. During the middle of the Late Pleistocene colder and drier climatic conditions prevailed, coinciding with the formation of mountain-taiga and trough-steppe landscapes. The sediments are comprised largely of alluvial, proluvial, volcaniclastic, glaciofluvial and glacial deposits and to a lesser extent, lacustrine–marshy deposits. Late Pleistocene fauna and flora remains have aided in establishing a chronological framework. Paleosols, glacial sediments and cryogenic material indicate that at times the climate was cool or cold.

Late Pleistocene deposits in the Tunka rift basins can be characterised by three types. The first, and most extensive, are surficial material deposited along the lower parts of the basins and in piedmont areas. The characteristic deposit is coarse proluvium such as talus cones near the north side of the rift and development of alluvial fans and deltas in outlet areas of the valleys of the northern slope of the Khamar Daban Range. During part of the Late Pleistocene glaciers occupied a limited area and therefore, parts of the piedmont slope contain glacial and glaciofluvial deposits.

The second type of sediments in the Tunka rift are terrace deposits within inter-basin spurs, including the lower parts of the Mondin and Bistraya basins. The end of the Middle Pleistocene and first half of the Late Pleistocene was a period of accumulation of coarse alluvium and glaciofluvial deposits. Accumulation resulted in the infilling of deep (deeper than today) valleys incised into bedrock. This was followed by more uplift and incision during the second half of the Late Pleistocene and Holocene.

The third type of sediment is formed in basin bottoms which are subjected to wind erosion and redeposition of sandy and silty material. This is in part caused by the geographic position of the Tunka basin which provides a wide and extensive (over 200 km) wind corridor where aeolian processes and deposition could take place. These processes were accompanied by the formation of dunes, deflation basins and the creation of aeolian microrelief (Akulov and Rubtsova, 2011). Aeolian deposits formed in two geomorphological contexts: (1) as mantle-shaped sandy covers which overlay the gently rolling windward western slopes of inter-basin spurs, and (2) as accumulation of loess-like loam and sandy loam on the inclined plains on the slopes of the Khamar Daban Range. The surface deposits of the Tunka rift contain an admixture of pelite-aleurite material, which gives them a loessic character. The maximum period of aeolian sedimentation in the Tunka rift basin occurred at the end of the Late Pleistocene–early Holocene, and at the present time large parts of the Tunka valley is subject to wind erosion.

Acknowledgments

This work was made possible through the support of the Russian Fund of Fundamental Research (11-05-00666, 11-05-00677, 12-05-00544), the Grant Council of the President of the Russian Federation (MK-2557.2010.5), the Baikal Archaeology Project and Social Sciences and Humanities Research Council of Canada, Natural Sciences and Engineering Research Council of Canada, and the Research-Educational Center 'Baikal'. We thank Professor G.F. Ufimtsev (The Institute of the Earth's Crust, Russian Academy of Sciences) for productive discussions and assistance with this paper. We also thank A.V. Sizov, I.V. Shibanova and I.M. Mashchuk who assisted with fieldwork at some sites. Thanks also to Adrian Lister and Tony Stewart who included several of the fauna specimens recovered from our work in their on-going study of Pleistocene megafauna extinctions in Eurasia. We also extend our thanks to Andriy Chernevych who translated parts of an early version of this paper from Russian into English.

References

Adamenko, O.M., Belova, V.A., Popova, S.M., Adamenko, R.S., Ponomareva, E.A., Efimova, L.I., 1975. Biostratigraphy of upper Pleistocene sediments of the Tunka depression. Russian Geology and Geophysics 6, 78–85 (in Russian).

Adamenko, O.M., Adamenko, R.S., Belova, V.A., Grachev, A.F., Lopatin, D.B., 1984. The age of molassa sequences of the Baikal rift zone based on small mammal fauna. In: Shatsky, S.B., (Ed.), Environment and Life on the Boundary of Cenozoic Epochs in Siberia and the Far East, Nauka, Novosibirsk, pp. 189–193 (in Russian).

Akulov, N.I., Rubtsova, M.N., 2011. Aeolian deposits of rift zones. Quaternary International 234 (1–2), 190–201.

Arzhannikova, A.V., Arzhannikov, S.G., Semenov, R.M., Chipizubov, A.V., 2005. Morphotectonics and Late Pleistocene–Holocene deformations in the Tunka system of basins (Baikal rift, Siberia). Annals of Geomorphology (Zeitschrift für Geomorphologie) 49 (4), 485–494.

Chipizubov, A.V., Smekalin, O.P., Semenov, R.M., 2003. Ancient seismic dislocations and related ancient earthquakes in the Tunka fault zone. Geology and Geophysics 44 (6), 587–602 (in Russian).

Delvaux, D., Moyes, R., Stapel, G., Petit, C., Levi, K., Miroshnichenko, A., Ruzhich, V., San'kov, V., 1997. Paleostress reconstructions and geodynamics of the Baikal region, Central Asia, Part 2, Cenozoic rifting. Tectonophysics 282, 1–38.

Derbyshire, E., 1983. On the morphology, sediments, and origin of the Loess Plateau of central China. In: Gardner, R., Scoging, H. (Eds.), Mega-geomorphology, Conference of the British Geomorphological Research Group. Oxford University Press, Oxford, pp. 172–194.

Firsov, L.V., Panychev, V.A., Orlova, L.A., 1985. The catalogue of radiocarbon dates. IGG SB AS USSR, Novosibirsk (in Russian).

Florensov, N.A., 1960. Mesozoic and Cenozoic Depressions in Cis-Baikal. AN SSSR, Moscow (in Russian).

Florensov, N.A., (Ed.), 1964. Geology of the USSR. The Buryatian ASSR. Part 1, vol. XXXV. Nedra, Moscow (in Russian).

Florensov, N.A., 1969. Rifts of the Baikal mountain region. Tectonophysics 8, 443–456.

Hase, Y., Krivonogov, S.K., Iwauchi, A., 2003. Geomorphological development of the Tunka depression in the Baikal rift zone in Siberia, Russia. In: Kashiwaya, K. (Ed.), Long Continental Records from Lake Baikal. Springer-Verlag, Tokyo, pp. 61–71.

Ignatova, N.V., Kulagina, N.V., Mashchuk, V.A., Shibanova, I.V., Ufimtsev, G.F., Perevalov, A.V., Rezanova, V.P., Vogt, TH., 1999. Soft sediments of the Tunka rift, Eastern Siberia. In: Rifting in Intracontinental Setting: Baikal Rift System and Other Continental Rifts – Abstract Book of Third Annual Meeting of the Project IGCP400 Geodynamics of Continental Rifting, Tervuren, Irkutsk, pp. 73–74.

Kashik, S.A., Mazilov, V.N., 1994. Main stages and palaeogeography of Cenozoic sedimentation in the Baikal rift system (Eastern Siberia). Bulletin Centres Recherche-Exploration-Production Elf-Aquitaine 18 (2), 453–462.

Krivonogov, S.K., 1995. Late Pleistocene sedimentation in the Tunka, south Baikal, Russia, compared with the northern territories. In: Abstracts of the XIV INQUA Congress, Berlin, p. 149.

Kuimova, L.N., Sherstyankin, P.P., 2003. Hydrophysical processes in Lake Baikal at the transition from subtropical to temperate climate during the last 5 Ma. Doklady Earth Sciences 390, 588–592.

Kul'chitsky, A.A., 1985. On Cenozoic deposits of the Bystraya basin in Cis-Baikal. In: Logachev, N.A. (Ed.), Issues of Geology and Paleogeography of Siberia and the Far East: Collection of Scholarly Works, Irkutsk University, Irkutsk, pp. 85–89 (in Russian).

Kul'chitsky, A.A., Osadchy, S.S., Misharina, V.A., Popova, S.M., Chernyaeva, G.P., Fileva, T.S., Orlova, L.A., Krivonogov, S.K., 1994. Results of study of the Tunka basin sandy deposits (Belyi Yar I and Belyi Yar II sections), in Baikal and its mountainous surrounding. In: Abstracts of the Irkutsk Geomorphological Workshop, October 1994, Irkutsk University, Irkutsk, pp. 100–103 (in Russian).

Logachev, N.A., 1958a. Cenozoic continental deposits of the Baikal type of depressions. Izvestia Akademiy Nauk SSSR, Geologicheskaya Seria 4, 18–21 (in Russian).

Logachev, N.A., 1958b. On the genesis of Quaternary sands of Cis-Baikal. Geology and Geophysics 1, 84–95.

Logachev, N.A. (Ed.) (1974). Uplands of Cis-Baikal and Trans-Baikal. The History of the Surface Topography of Siberia and the Far East. Nauka, Moscow (in Russian).

Logachev, N.A. (Ed.), 1981. Baikal Region. Guidebook A-13, C-13. XI INQUA Congress, Moscow.

Logachev, N.A., 1984. Neotectonics. In: Logachev, N.A. (Ed.), Geology and Seismicity of the BAM Zone (From Baikal to Tunda), Nauka, Novosibirsk, pp. 55–91 (in Russian).

Logachev, N.A., 1993. History and geodynamics of Lake Baikal in the context of the Eastern Siberia rift system: a review. Bulletin Centres Recherche-Exploration-Production Elf-Aquitaine 17 (2), 353–370.

Logachev, N.A., 2003. History and geodynamic of the Baikal Rift. Russian Geology and Geophysics 44 (5), 391–406.

Logachev, N.A., Mohr, P. (Eds.), 1978. Geodynamics of the Baikal rift zone. Tectonophysics 45, 1–105.

Logachev, N.A., Zorin, Y.A., 1987. Evidence and causes of the two-stage development of the Baikal rift. Tectonophysics 143, 225–234.

Logachev, N.A., Zorin, Y.A., 1992. Baikal rift zone: structure and geodynamics. Tectonophysics 208, 273–286.

Lvov, A.S., 1924. From the Geological Past of the Middle Part of Irkut Valley, from Tunka Trough to Zyrkun Ridge. Reports of the Russian Geographic Society vol. 46, No. 3, pp. 1–84 (in Russian).

Martinson, G.G., 1948. Fossil sponges of the Tunka basin in Cis-Baikal. Reports of the USSR Academy of Sciences 2(5), 49–53 (in Russian).

Mats, V.D., 1993. The structure and development of the Baikal rift depression. Earth – Science Reviews 34, 81–118.

Mazilov, V.N., Lomonosova, T.K., Klimanova, V.M., Kashaeva, G.M., Ershov, R.I., Semjonova, G.O., 1972. Lithology of Teritary Deposits in Depressions: South-west of the Baikal Rift Zone. Nauka, Moscow (in Russian).

Mazilov, V.N., Kashik, A.A., Lomonosova, T.K., 1993. Oligocene deposits of the Tunka basin. Geology and Geophysics 8, 81–87 (in Russian).

McCalpin, J.P., Khromovskikh, V.S., 1995. Holocene paleoseismicity of the Tunka fault, Baikal rift, Russia. Tectonics 202, 504–685.

Parfeevets, A.V., San'kov, V.A., 2006. Geodynamic conditions of evolution of the Tunka branch in the Baikal rift system. Geotectonics 5, 377–388.

Pavlov, S.F., Kashik, S.A., Lomonosova, T.K., 1976. Cenozoic Deflation Crusts and Sediment Formations of Western Cis-Baikal Region. Nauka, Novosibirsk (in Russian).

Popova, S.M., Mats, V.D., Chernyaeva, G.P., Shimaraeva, M.K., Kulchitskiy, A.A., Vorobova, G.A., Klimanova, V.M., Kononov, E.E., 1989. Paleoliminological Reconstructions. Baikal Rift Zone. Nauka, Novosibirsk (in Russian).

Ravskii, E.J., Alexandrova, L.P., Vangeigem, E.A., Gerbova, V.G., Golubeva, L.V., 1964. Anthropogene Deposits of Southern East Siberia. Nauka, Moscow (in Russian).

Shchetnikov, A.A., 1999. Structure and morphodynamics of the edges of the Tunka basins. Geography and Natural Resources 4, 75–82 (in Russian).

Shchetnikov, A.A., 2008. The manifestation of a global mechanism of orogenesis in the Baikal rift zone (using the Tunka rift as an example). Geography and Natural Resources 29, 226–229.

Shchetnikov, A.A., Ufimtsev, G.F., 2004. Relief Structure and Recent Tectonics of the Tunka Rift (southwestern Lake Baikal region). Nauchnyi Mir, Moscow (in Russian).

Shchetnikov, A.A., Ufimtsev, G.F., Skovitina, T.M., 1997. Antecedent valleys of the southern Cis-Baikal region. Geography and Natural Resources 4, 86–95 (in Russian).

Shchetnikov, A.A., Filinov, I.A., Shibanova, I.V., Mashchuk, I.M., Sizov, A.V., 2006. New section of upper Cenozoic deposits "Slavin Yar" in Tunkinsy rift valley. In: Abstracts of the INQUA Symposium "Stratigraphy, Paleontology and Paleoenvironment of Pliocene–Pleistocene of Transbaikalia and Interregional Correlations", BSC SB RAS, Ulan-Ude, p. 83.

Shchetnikov, A.A., Filinov, I.A., Shibanova, I.V., Mashchuk, I.M., Sizov, A.V., 2009. New section of upper Cenozoic deposits "Slavin Yar" in the Tunka rift valley. Stratigraphy and Geological Correlation 17 (4), 114–119.

Sherman, S.I., Medvedev, M.E., Ruzhich, V.V., Kiselev, A.I., Shmotov, A.P., 1973. Tectonics and Volcanism in the South-west of the Baikal Rift Zone. Nauka, Novosibirsk (in Russian).

Shibanova, I.V., 1996. New data on fresh-water mollusks of the Tunka basin loose sediments. In: Sklyarov, E.V., (Ed.), The Russian Foundation for Basic Research in the Siberian region (Earth's crust and mantle), vol. 1, IZK SO RAN, Irkutsk, pp. 138–139 (in Russian).

Trofimov, A.G., Kulagina, N.V., Popova, S.M., Shibanova, I.V., 1995. New data on the Pleistocene series of the Tunka basins. In: Sklyarov, E.V. (Ed.), The Russian

Foundation for Basic Research in the Siberian Region (Earth's crust and mantle), vol. 1, IZK SO RAN, Irkutsk, pp. 50–51 (in Russian).
Ufimtsev, G.F., Shchetnikov, A.A., 2002. The Tunka rift as a special element among continental rifts. In: Abstracts of the Conference "Tectonics and Geophysics of the Lithosphere", GEOS, vol. 2, Moscow, pp. 254–256 (in Russian).
Ufimtsev, G.F., Perevalov, A.V., Rezanova, V.P., Shchetnikov, A.A., 1999. Subsiding row of volcanoes of the Tunka basin. Izvestiya Vuzov 4, 158–160 (in Russian).
Ufimtsev, G.F., Shibanova, I.V., Kulagina, N.V., Mashchuk, I.M., Perevalov, A.V., Rezanova, V.P., Vogt, T., Ignatova, N.V., Misharina, V.A., 2002. Upper Pleistocene and Holocene deposits of the Tunka rift. Stratigraphy and Geological Correlation 10 (3), 90–99.
Ufimtsev, G.F., Perevalov, A.V., Rezanova, V.P., Kulagina, N.V., Mashchuk, I.M., Shchetnikov, A.A., Rezanov, I.N., Shibanova, I.V., 2003. Radiothermoluminescence dating of Quaternary sediments of the Tunka rift. Russian Geology and Geophysics 44 (3), 226–232.
Ufimtsev, G.F., Shchetnikov, A.A., Filinov, I.A., 2004a. River terraces of the Tunka rift valley. Geomorphology 1, 113–122 (in Russian).
Ufimtsev, G.F., Shchetnikov, A.A., Filinov, I.A., 2004b. Late Quaternary lithogenesis in the Tunka rift. In: Abstracts of the 2nd Siberian International Conference of Young Scholars in Earth Sciences, Novosibirsk State University, Novosibirsk, pp. 173–175 (in Russian).
Ufimtsev, G.F., Shchetnikov, A.A., Filinov, I.A., 2009. Inversions in the contemporary geodynamics of the Baikal rift zone. Geology and Geophysics 50 (7), 798–810.
Vogt, H., Vogt, T., 2007. Morphotectonic evolution of two depressions at the southern border of the Baikal rift system. Geomorphology 86 (3–4), 480–495.