SCIENTIFIC JOURNAL
EARTH'S CRYOSPHERE

Kriosfera Zemli, 2011, vol. XV, No. 4, pp. 8–13 http://www.izdatgeo.ru

EARTH'S CRYOLOGY: ENVIRONMENTAL AND REGIONAL PROBLEMS

GEOCRYOLOGICAL CHRONICLES OF RUSSIA

S.M. Fotiev

Institute of Earth's Cryosphere, Siberian Branch of the Russian Academy of Sciences,
PO box 1230, Tyumen, 625000, Russia; kriozem@gmail.com

The author was the first to compile two geocryological Chronicles: simplified for the Late Cenozoic) (the last 3.5 Ma) and more detailed for the Neopleistocene (Brunhes epoch) (the last 800 kur). These Chronicles comprise the geocryological interpretation of Baikal climatic chronicle, compiled by a large team of scientists, and based on the identified relation of diatom valves and biogenic silica percentage in bottom sediments of the Baikal lake on the severity of natural setting.

LATE CENOZOIC GEOCRYOLOGICAL CHRONICLE (3.1–0.0 Ma)

The Late Cenozoic chronicle is a geocryological interpretation of the Baikal diatomic record[1] [*Karabanov et al., 2000*]. The Late Cenozoic chronicle (Fig. 1) clearly presents time spans within the late Pliocene–Holocene cryogenic period comprized by three cold epochs separated by longterm warm climatic epochs [*Fotiev, 2005*]. The beginning of the Pliocene–Holocene cryogenic period in Siberia coincided with the onset of a stable climate cooling in the Pacific [*Shackleton et al., 1990; Fotiev, 2005*].

First Pliocene cryogenic epoch (3.10–3.08 Ma) was the shortest (only 20 kyr) and the least cold. Judging by the diatom percentage (less than 20 %)[2] it was during that time that the climatic setting provided perennial freezing and formation of permafrost[3]. The mean annual air temperature (T_a) was 3–5 °C lower than that at present. High-temperature (0 to −3 °C), thin (0–100 m), and discontinuous or sporadic permafrost was formed in southern Siberia. At the same time, in northern Siberia, both climate and geocryologic setting were probably severer and permafrost continuous[4].

Between 3.08 and 2.82 Ma BP, when the diatom percentage increased to 48–60 % (with a peak at 2.9 Ma), the climate was relatively warm (Fig. 1). During the 250 kyr time span, permafrost degraded fully or partly in southern Siberia, while no thawing from the surface occurred in the colder northern areas of Siberia. The mean annual ground temperature (T_g) varied considerably though within negative interval.

Second Pliocene cryogenic epoch (2.82–2.47 Ma) was marked by the coldest climate and severest geothermal settingof the entire cryogenic period (Fig. 1) [*Fotiev, 2009*]. There are reasons to hypothesize that the climate became colder as the orogenesis in Eurasia (the Sayan, Altai, Himalaya, Pamir, and Tien Shan mountain systems) impeded the transport of warm and wet air from the Indian ocean. Many scientists agree that the Late Pliocene cooling correlates with the oldest glaciations in the mountains of Transbaikalia, Altai, and Sayan and with the earliest glacial clay deposition marked by iceberg rafting signature in the Baikal lake sediments. Yet, no traces of these old glaciations have been discovered [*Karabanov et al., 2001; Kuz'min et al., 2001a; Williams et al., 2001*].

The second cryogenic epoch, in turn, consisted of nine chrons: five cryochrons and four thermochrons (Fig. 1). The climate of cryochrons was cold or very cold, according to diatom data (0–2 %), with a peak at 2.60–2.63 Ma, and was even colder than during the Sartan time. The mean annual air temperature was 10–15 °C colder than it is now. Permafrost was low-temperature, thick, and continuous, even in southern Siberia, in northern Siberia, continuous permafrost was as cold as −30 °C and thick (>300 m).

[1] The Baikal continuous climate chronicle of the past 5 Myr has been obtained in the course of exhaustive studies of the lake sediments by a large team involved in the Baikal Drilling Project (BDP) [*Bezrukova et al., 1999; Karabanov et al., 2000, 2001; Williams et al., 2001; Kuz'min et al., 2001a, b; Khursevich et al., 2001; Prokopenko et al., 2001*].

[2] For more details of geocryological interpretation of the Baikal climate chronicle see [*Fotiev, 2005, 2009*].

[3] Permafrost is the rock strata with temperature below 0 °C, and comprises frozen, cryotic and dry frozen deposits.

[4] The latitude zoning of climate parameters held all over the cryogenic period.

Copyright © 2011 S.M. Fotiev, All rights reserved.

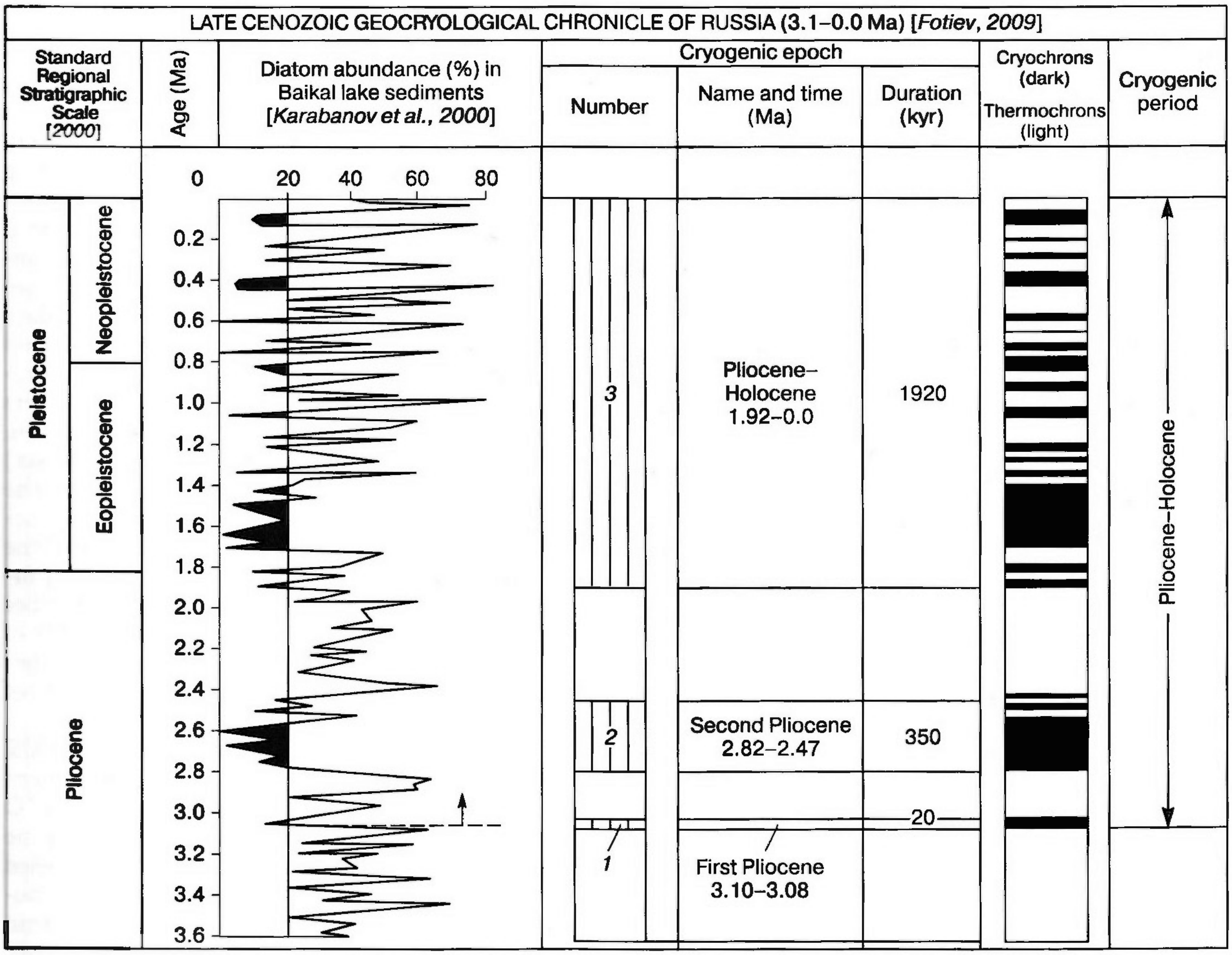

Fig. 1. Late Cenozoic geocryological chronicle of Russia (3.1–0.0 Ma).
Diatom abundances averaged over five data points [*Karabanov et al., 2000*].

The thickness of permafrost exceeded the thickness of freshwater zone in artesian basins and hydrogeological massifs (aquifers). Two layers of frozen and cryotic rocks with cryopegs formed in artesian basins, and layers of frozen rocks and dry permafrost were formed in hydrogeological massifs [*Fotiev, 1978*]. This permafrost structure in the subsurface of the Siberian Platform has persisted till present. Relatively low diatom percentage (19–25 %) indicate quite a cold climate during thermochrons. The permafrost formed during cryochrons never degraded even in southern Siberia where it persisted continuously over 350 kyr (Fig. 1).

In the northern part of permafrost zone[5] T_g though increased significantly during thermochrons yet remained negative, and the permafrost never thawed from the surface. Permafrost degradation from below was negligibly small because of low heat flux and abundant cryopegs within the layer of cryotic rocks. Permafrost in West Siberia apparently degraded from below more rapidly as the heat flux was higher [*Kurchikov, Stavitsky, 1987; Balobaev, 1991*].

The interval of 2.47–1.92 Ma (the end of Late Pliocene) in southern Siberia was an over 500 kyr spell of moderately warm to warm climate, as indicated by a relatively high (25–65 %) production of diatoms (Fig. 1). The permafrost that formed during the second Pliocene cryogenic epoch fully or almost fully degraded. However, the rather strong and long climate warming did not cause degradation of permafrost from above in the northern areas, though the negative T_g increased markedly.

Third Pliocene–Holocene cryogenic epoch (1.92–0.0 Ma) started 1.92 Ma ago when the diatom percentage in the Baikal lake sediments decreased to 10–12 %, and is ongoing (Fig. 1) [*Fotiev, 2005, 2009*].

[5] Permafrost zone – area extent of the rock strata with temperature below 0 °C.

In the beginning of this cryogenic epoch, there were two short and relatively warm cryochrons during which permafrost formed in southern Siberia.

The interval between 1.75 and 1.40 Ma BP (the first half of Eopleistocene, according to the Russian stratigraphic code) was an epoch of a very cold climate as follows from the Baikal diatomic chronology (Fig. 1)[6]. The epoch is subdivided into seven chrons: four cryochrons and three thermochrons (Fig. 1). During the cryochrons, T_a was 10–15 °C lower than the present one. Low-temperature (up to −10 °C) thick continuous permafrost formed even in the southernmost regions of Siberia. During the extremely cold cryochrons southern limit of permafrost area extended as far as Mongolia and China. Therefore, no zones of discontinuous or sporadic permafrost existed in the southern part of the Siberian Platform. In northern Siberia, T_g decreased to −30 °C or lower, and permafrost thickness likely exceeded 500–700 m. Permafrost degradation during the thermochrons was incomplete even in southernmost Siberia as the climate remained relatively cold (diatoms percentage as low as 12 to 25 %). Therefore, during the first half of Eopleistocene permafrost persisted continuously for at least 350 kyr even in southern Siberia.

The second half of the Eopleistocene and in Neopleistocene was a period of rather variable climate in southern Siberia judging by dramatic and essential variations in diatom percentages between 0 to 80 % (Fig. 1). The climatic and geocryological setting during cryochrons were almost as cold as those in the first half of Eopleistocene, while climate of thermochrons wes much warmer. During extremely warm thermochrons when the diatom production increased to 80 % (peak at 1.01 Ma BP), T_a was 3–4 °C warmer than now, and permafrost in southern areas degraded rapidly from both above and below (Fig. 1). In the northern Siberia with its severe climate T_g remained very low during both cryochrons and thermochrons varying within the negative interval and aggrading from below.

NEOPLEISTOCENE (BRUNHES EPOCH) GEOCRYOLOGICAL CHRONICLE (0.8–0.0 Ma)

The geocryological chronicle is based on the interpretation of the Baikal biogenic silica ($SiO_{2\,biog}$) record (Fig. 2) [*Prokopenko et al., 2001; Fotiev, 2005, 2009*]. The Neopleistocene geocryological chronicle thirty nine well pronounced chrons are subdivided: 20 cryochrons and 19 thermochrons (Fig. 2). The cryochrons (with negligibly low $SiO_{2\,biog}$) look like U-shaped troughs in the $SiO_{2\,biog}$ chronicle, with flat or inclined floors, while the thermochrons (more than 80 % $SiO_{2\,biog}$) appear as prominent sharp peaks [*Karabanov et al., 2001; Prokopenko et al., 2001*]. This pattern is a clear evidence of a steadiness of severe climate in cryochrons and abrupt dramatic warming during thermochrons.

Compiling the geocryological chronicle required the use of three chronologies. The 39 chrons distinguished by the author in the Neopleistocene of Siberia (20 cryochrons and 19 thermochrons) correlate with 19 marine isotope stages (MIS) with 10 cold and 9 warm stages of the oxygen isotope ($\delta^{18}O$) stratigraphy (Fig. 2) [*Shackleton et al., 1990*]. The time limits of some chrons perfectly match the MIS boundaries, but some isotope stages comprise several (up to six) distinct cryochrons and thermochrons[7], which the $\delta^{18}O$ stratigraphy actually misses. The generally accepted Standard Regional Stratigraphic Scale of the Quaternary of the West Siberian Plain [*2000*] includes 14 horizons (epochs), i.e. less than the number of MIS. In order to cancel this disparity, some MIS of different ages were in some models brought together [*Fotiev, 2009*], which distorted the age bounds of some intervals.

The model suggested by the author [*Fotiev, 2005, 2009*], is based on the correlation of the northern West Siberian glacials and interglacials with the $\delta^{18}O$ scale after [*Karabanov et al., 2001*]; the gaps in the latter chronology (MIS 9, 10, 12, 13, 19) are labeled with conventional names. Cryochrons and thermochrons are distinguished [*Fotiev, 2005, 2009*] within the age limits of each MIS [*Shackleton et al., 1990*], and correlated with the respective intervals of the regional stratigraphy (Fig. 2).

The time between 800 and 426 kyr BP (Early Neopleistocene) in Siberia included 17 chrons (9 cryochrons and 8 thermochrons) corresponding to eight (four cold and four warm) marine isotope stages. The Middle Neopleistocene interval 426–127 kyr BP comprise 18 chrons (9 cryochrons and 9 thermochrons) correlated with six (3 cold and 3 warm) stages of the marine $\delta^{18}O$ stratigraphy. The Late Neopleistocene from 127 to 11 kyr BP is divided into four chrons (2 cryochrons and 2 thermochrons) that correspond to four (two cold and two warm) MIS. The West Siberian Karginian warm cycle (MIS-3) does not show up in the Baikal biogenic silica record[8] (Fig. 2), and that was the reason why MIS-4, MIS-3, and MIS-2 were joined into a single Zyryanka–Sartan cryochron.

Low $SiO_{2\,biog}$ contents in the Baikal lake sediments indicate several coldest and longest cryochons

[6] Extremely low diatom percentages of 0–3 % (the lowest at 1.63 Ma) indicate a climate and geothermal setting colder than during the Sartan time.

[7] These climate oscillations are most often attributed to orbital forcing [*Kuz'min et al., 2001b*].

[8] According to some models [*Baulin et al., 1981*], the zone of permafrost degradation from the surface extended in southern West Siberia as far as 65–66° N during the optimum of the Karginian warm stage.

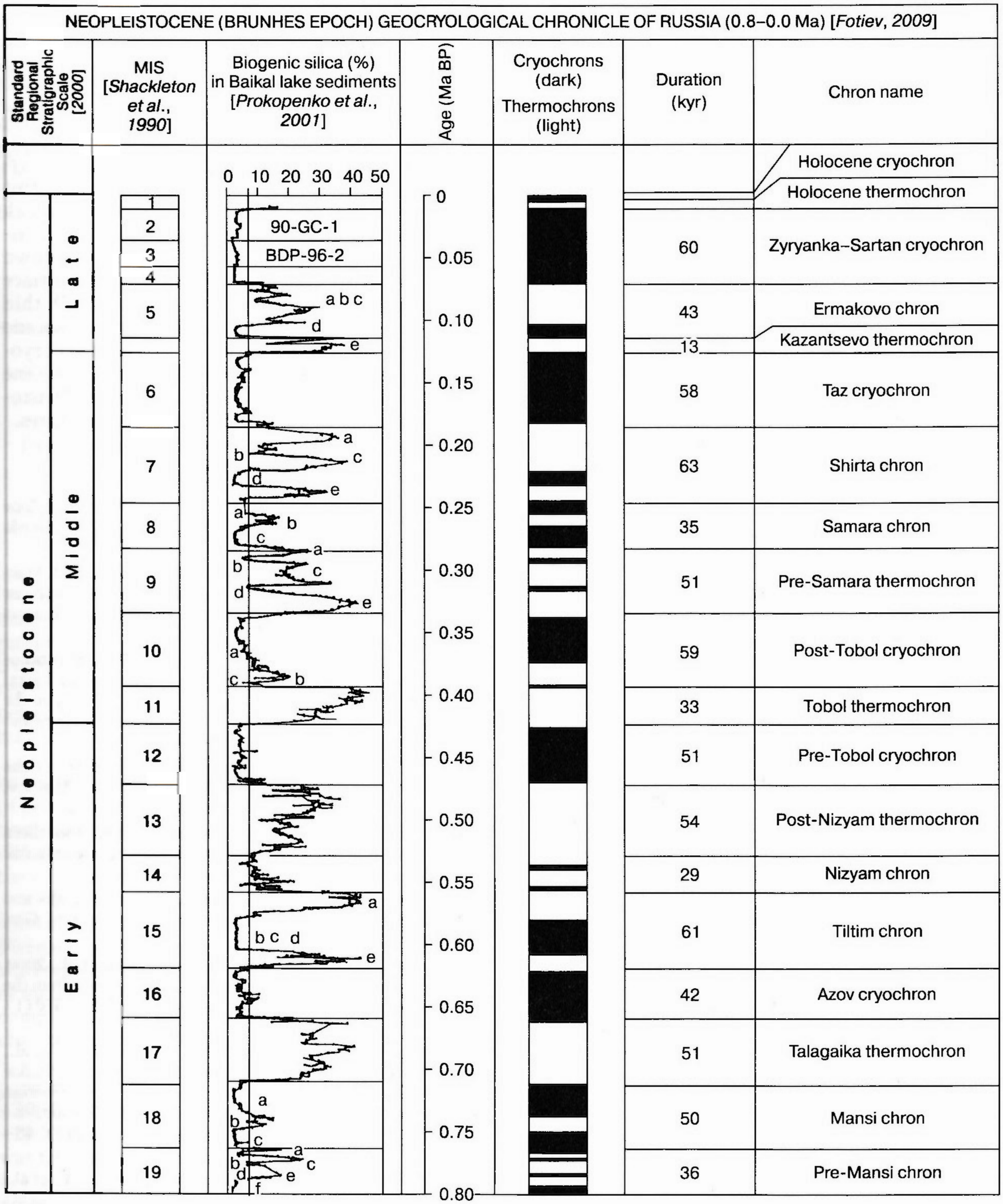

Fig. 2. Neopleistocene (Brunhes epoch) geocryological chronicle of Russia (0.8–0.0 Ma).

in Neopleistocene, namely (Fig. 2), Asov (MIS-16) and pre-Tobol (MIS-12) in the early Neopleistocene, post-Tobol (MIS-10a), Samara (MIS-8c), and Taz (MIS-6) in the Middle Neopleistocene, and the Zyryanka-Sartan (MIS-4, MIS-3, MIS-2) in the Late Neopleistocene. The climatic and geocryological setting during the extremely cold Neopleistocene cryochrons were as follows [*Fotiev, 2009*]: (1) T_a throughout the permafrost area was 8–15 °C lower than that at present; (2) long-lasting, low-temperature (–3 to –25... –30 °C), thick (from 300 to 700–1500 m) and continuous permafrost formed over the entire permafrost area; (3) zones of high-temperature (0 to –3 °C), thin (0 to 150 m) and discontinuous (including sporadic) permafrost were outside the region in West Siberia and were absent from East Siberia as the southern permafrost extent reached the territory of Mongolia and China; (4) the subaerial permafrost area was of the largest extent due to perennial frozen ground in the emerged shelf and southern margins of permafrost area. It extended from 47 to 78° N (over 3300 km) along the 70° E meridian, and from 52 to 79.5° N or over 3000 km along 105° E (excluding the southern mountains and the northern islands).

According to relatively high contents of $SiO_{2\,biog}$ in the sediments of Lake Baikal, the warmest and longest were (Fig. 2) the Talagaika (MIS-17) and post-Nizyam (MIS-13) thermochrons in the Early Neopleistocene, the Tobol (MIS-11) and Shirta (MIS-7) thermochrons in the Middle Neopleistocene, and the Kazan (MIS-5e) in the Late Neopleistocene (127–114 kyr BP). The extremely warm thermochrons of Neopleistocene were characterized by the following climate and permafrost-related features [*Fotiev, 2009*]: (1) T_a was 2–4 °C higher than now and dramatically (10–15 °C) higher than during the previous cryochrons; (2) permafrost degraded completely in the southern and partly in the central part of permafrost area; (3) permafrost did not degrade from the surface in the northern permafrost area where the climate was still cold, though T_g increased considerably remaining however below zero; (4) permafrost degraded partly from the surface only in submerged limnic environments; (5) the N–S extent of permafrost decreased as a result of marine transgression and complete permafrost degradation in the southern regions.

The Holocene comprises two chrons (Fig. 2) subdivided according to the specific features of cryogenic metamorphism. T_a during the thermochron from 11.0 kyr BP to the end of t he thermal optimum[9] rose 10–15 °C higher than in the minimum of the Sartan thermal minimum almost all over the permafrost area. Nevertheless, active permafrost degradation (full or partial), both from above and from below, was limited to the plains of southern Siberia. In the northern part of permafrost area, T_g was rising essentially but remained negative, partial degradation of permafrost was restricted to areas underneath numerous thermokarst lakes. During the cryochron from the end of the thermal maximum to present, T_g became 3–5 °C lower than in the thermal optimum on all over the permafrost area; high-temperature, thin, and discontinuous or sporadic permafrost formed in central and southern regions of Siberia.

The boundary between the Northern and Southern geocryological zones is quite prominent at present and coincides with the link up of permafrost layers of Holocene and Pleistocene ages. The permafrost is low-temperature (–2 to –13 °C or lower), thick (300 to 1500 m), continuous, predominantly of Pleistocene age north of this boundary, sinking down southward of this boundary, replaced in the surface layer by mainly high-temperature (0 to –2 °C), thin (0 to 150 m), discontinuous, permafrost of Holocene age. In the northern part of the Southern geocryological zone, within the artesian basins the Holocene permafrost is separated from the deep-seated Pleistocene permafrost by a layer of positive temperatures.

References

Balobaev V.T., 1991. The Thermal Field of the Permafrost Zone in the North Asian Lithosphere [in Russian]. Nauka, Novosibirsk, 193 pp.

Baulin V.V., Chekhovsky A.L., Sukhodolsky S.E., 1981. Main evolution stages of permafrost in Russian European Northeast and in West Siberia, in: The History of Permafrost in Eurasia [in Russian], Nauka, Moscow, pp. 41–60.

Bezrukova E.V., Kulagina N.V., Letunova P.P., Shestakova O.N., 1999. Climate and vegetation changes in the Baikal region for the last 5 Ma, according to palynological data on the Baikal sediments. Geologiya i Geofizika (Russian Geology and Geophysics), 40 (5), 739–749 (722–731).

Fotiev S.M., 1978. The Hydrogeothermal features of the permafrost area of the USSR [in Russian]. Nauka, Moscow, 236 pp.

Fotiev S.M., 2005. Cryochrons and thermochrons of southern Siberia for the past 5 Ma: Paleogeographic implications of Baikal lake sediment data. Kriosfera Zemli, IX (1), 13–27.

Fotiev S.M., 2009. Cryogenic Metamorphism of Rocks and Groundwater: Conditions and Consequences [in Russian]. Geo, Novosibirsk, 279 pp.

Karabanov E.B., Kuz'min M.I., Williams D.F., et al., 2000. Late Cenozoic global cooling in Central Asia: evidence from the sedimentary chronicle of Lake Baikal. Dokl. Earth Sci., 370 (1), 61–66.

Karabanov E.B., Prokopenko A.A., Williams D.F., et al., 2001. Glacial and interglacial periods of Siberia: the Lake Baikal paleoclimatic chronicle and correlation with West Siberian stratigraphic scheme (the Brunhes Chron). Geologiya i Geofizika (Russian Geology and Geophysics), 42 (1), 48–63(41–54).

Khursevich G.K., Karabanov E.B., Prokopenko A.A., et al., 2001. Detailed diatom biostratigraphy of Baikal sediments during the Brunhes chron and climatic factors of species formation. Geologiya i Geofizika (Russian Geology and Geophysics), 42 (1), 108–129 (97–116).

[9] The ages and durations of the Holocene climate optimum varied over the permafrost area.

Kurchikov A.R., Stavitsky B.P., 1987. Heat Flow in Petroleum Provinces of West Siberia [in Russian]. Nedra, Moscow, 134 pp.

Kuz'min M.I., Karabanov E.B., Kawai T., et al., 2001a. Deep drilling on Lake Baikal: main results. Geologiya i Geofizika (Russian Geology and Geophysics), 42 (1), 8–34 (3–28).

Kuz'min M.I., Karabanov E.B., Prokopenko A.A., et al., 2001b. Late Cenozoic rhythms and climate change in Asia, from Baikal Deep Drilling data, in: Global Change [in Russian]. Izd. SO RAN, Filial "Geo", Novosibirsk, pp. 146–159.

Prokopenko A.A., Karabanov E.V., Williams D.F., et al., 2001. Biogenic silica chronicle of the Lake Baikal response to the climatic forcing during the Brunhes. Quatern. Res., 55, 123–132.

Shackleton N.J., Berger A., Peltier W.R., 1990. An alternative astronomical calibration of the Lower Pleistocene time scale based on ODP site 677. Trans. Roy. Soc., Edinburgh, Earth Sciences, 81, 251–261.

The Standard Regional Stratigraphic Scale of the Quaternary of the West Siberian Plain [in Russian]. SNIIGGiMS, Novosibirsk, 2000, 64 pp.

Williams D.F., Karabanov E.B., Prokopenko A.A., et al., 2001. The Late Pliocene–Pleistocene sedimentary chronicle of Lake Baikal: benchmarks for paleoclimatic and biostratigraphic reconstructions. Geologiya i Geofizika (Russian Geology and Geophysics), 42 (1), 35–47 (29–40).

Received
5 March 2011

Kriosfera Zemli, 2011, vol. XV, No. 4, pp. 13–16 http://www.izdatgeo.ru

PERMAFROST RESPONSE TO CLIMATE WARMING

V.N. Konishchev

Lomonosov Moscow State University, Department of Geography, 1, Leninskie Gory, Moscow, 119991, Russia; vkonish@mail.ru

The response of permafrost to global Late Pleistocene–Holocene warming has been investigated in the ice complex of Yakutia. Thermokarst erosion of the complex was found out to depend on the properties of landscape and its components that change as a consequence of warming. In addition to the degradation tendency, aggradational stabilization in ice complex remnants may occur in certain conditions as formation of a protective layer.

Any external effect on permafrost, including climate, is never direct, unlike on a glacier surface, but is rather mediated by the overlying vegetation, soil, active layer, i.e., by various landscape components. The resulting positive and negative feedbacks control the permafrost responses which differ in intensity and may show unexpected trends.

Changes in surface conditions attendant with warming or cooling may change the evolution trend of permafrost and cause its aggradation or degradation. They may act either together with or against the climate trend and, correspondingly, amplify or damp the climate effect [*Koreisha et al., 1997*].

An illustrative example in this respect may come from dynamics of the ice complex over the latest Pleistocene–Holocene interval. An ice complex (IC) is a complex of syngenetically frozen sediments, tens of meters thick, which are thermally unstable. Such complexes were deposited in harsh Late Pleistocene climates between 50–40 and 12–11 kyr BP.

At that time, the mean annual temperature of permafrost was as low as –25...–28 °C or locally –30 °C in northern Yakutia and no warmer than –10 °C in central Yakutia, while the air temperature was naturally still lower [*Konischev, 1997*]. The map in Fig. 1 shows the extent of the Yakutian ice complex except shelf where it was destroyed by the Holocene transgression of the Arctic seas.

The dramatic climate change at the Last Glacial–Holocene boundary caused permafrost warming and gave rise to thermokarst erosion of the ice complex, which has been the main evolution process in its terrain-forming deposits through the past 13–12 kyr.

The basic landforms in the present North and Central Yakutian coastal lowland include, besides river valleys, the so-called *yedomas* (IC remnants) and *alas* depressions (result of subsidence of thawed permafrost) that reach surface areas of tens of square kilometers. Alases occupy up to 75 % of the northern Yakutian coastal plains [*Lomachenkov et al., 1965; Bosikov, 1978*] and up to 50 % in central Yakutia [*Ivanov, 1981*].

Although being approximate, these estimates indicate that the ground surface composed of IC deposits is more strongly affected by thermokarst processes in northern Yakutia than in its central part. The reason is that alases in both northern and central areas of Yakutia result from water ingression (showing up as the number and sizes of lakes) rather than from climate warming per se or from other temperature agents.

The glacial deposition completed in the end of the last cryochron (Last Glacial), and the top surfaces of its remnants (yedomas) generally have experienced no denudation since then. This inference follows from many radiocarbon dates showing latest Pleistocene ages of the youngest IC deposits [*Ivanov, 1981; Kaplina, 1981*]. Therefore, through the Holocene the IC deposits have been subject to different alteration trends [*Shur, 1988*].

Copyright © 2011 V.N. Konishchev, All rights reserved.