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A chain of stamukhas up to 8 meters high formed above the longshore bar in the coastal zone of Sharapov Shar Bay, Kara Sea. March 2023.

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GEOCRYOLOGICAL MONITORING AND FORECAST

DYNAMICS OF PERMAFROST TEMPERATURE REGIME IN DEEP BOREHOLES
IN THE CENTRAL PART OF NORILSK

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In the Arctic zone, the current rise in air temperatures is much faster than in the central part of Russia. This leads to an increase in the temperature of frozen ground of the foundations of buildings and engineering structures, a decrease in their bearing capacity, and an increasing risk of their destruction. At the same time, the temperature regime of frozen ground in the city depends on many factors (microclimatic features, composition and properties of soils, technogenic influence, maintenance practices). In 2022, as part of the program to create an element of a unified system for monitoring the state of permafrost in the municipality of the city of Norilsk and the Taimyr Dolgano-Nenets municipal district, the Polar Division of PJSC MMC Norilsk Nickel began a program for constructing deep boreholes in the territory of the company's responsibility both in the natural environment (as part of the background permafrost monitoring) and in the city. The purpose of this article is to assess the dynamics of the temperature regime of permafrost in the center of Norilsk. Archival materials (since 1959) and the results of geotechnical surveys conducted in Norilsk to study the geological structure, composition, and properties of frozen rocks have been analyzed, and monitoring measurements in a new thermometric borehole drilled to replace the previous one, have begun. The data obtained indicate a significant change in the geocryological conditions in the city. Over the past 65 years (from 1959 to 2024), the ground temperature at a depth of 10 m in the area of the deep thermometric borehole increased by 4.3°C. Significant temperature fluctuations have been recorded down to a depth of 90 meters, and further down the section to a depth of 200 meters the temperature has changed by an average of 0.3°C.

Keywords: frozen ground, permafrost, temperature regime, climate change, Arctic city, Norilsk.

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INTRODUCTION

Currently, there are 1162 settlements with about five million inhabitants in the Arctic permafrost zone. Most settlements are small, with an average population of 622 residents. Eighty-five percent of large settlements are located in the Russian Arctic [Ramage et al., 2021]. Norilsk agglomeration is one of the largest in the Arctic, home to over 183 000 people. The main problems of the city's life support are due to harsh climatic conditions, remote location, and the specific operating conditions of buildings and engineering structures [Shiklomanov et al., 2017].

Geocryological conditions in the Norilsk industrial region are extremely heterogeneous. The thickness of permafrost varies from 5 to 800 m or more, and significant areas (up to 40%) in the intermontane valleys are occupied by taliks, even in the north of the region. Mean annual temperature of the rocks varies in a wide range: from –7 to +2°C [Ershov et al., 1989]. Climate change in the Arctic territories is becoming one of the most important global problems. The warming trend in the Arctic exceeds the global average by more than 2.5 times. Almost all forecasts show that the area of permafrost will decrease [Rosgidro-

met, 2022]. Globally, permafrost temperature has increased by $(0.29 \pm 0.12)^\circ\text{C}$ since 2007 [Biskaborn et al., 2019].

Global warming leads to the thawing of permafrost, a key component of Arctic territories, which poses a significant threat to the integrity of polar infrastructure. Studies have been conducted on climate change and its impact on engineering structures [Grebenets, Sadovsky, 1993; Grebenets, Ukhova, 2008; Hjort et al., 2018, 2022; Streletskiy et al., 2019; Porfirjev et al., 2019; Melnikov et al., 2022; Falaleeva et al., 2023]. However, these estimates are general and are not based on factual material on the geological structure, geocryological conditions, and construction principles, though these characteristics may vary even within the same urban area. In this regard, the most reliable results are those of mathematical modeling performed on the basis of actual measurement data for a specific object [Schneider von Deimling et al., 2021; Hjort et al., 2022]. For example, a significant part of the industrial buildings and engineering structures of the Polar Division of PJSC MMC Norilsk Nickel are located on rocky or thawed soils, so cli-

mate change will not affect their stability. However, historically, the urban area of Norilsk was built using technical solutions for construction on permafrost soils [Grebenets, Sadovsky, 1993; Makarov et al., 2001; Setkov et al., 2005; Shiklomanov et al., 2017].

The construction of multi-story residential and public buildings in the southern part of the city was carried out mainly in the 1940s–1950s. All these buildings were constructed on hard bedrock as the natural foundation. One of the main features of the construction was sufficient deepening of the foundations into the bedrock, because its upper part was strongly destroyed by weathering processes. In such rocky massifs, cracks are sometimes filled with ice inclusions, which, when changing from a frozen to a thawed state, can cause displacement of structures and subsidence of thawing soils in the adjacent territory. The depth of deepening into the bedrock, depending on the thickness of the destroyed zone, ranged from 0.1 to 2.0 m and more [Makarov et al., 2001]. Operating experience shows that if the deepening was done well and below the cracked zone, the building is currently in operation without significant problems. In conditions of deeper bedrock (more than 10 m), concrete columnar foundations with support on coarse-grained soils with low ice content were used in Norilsk. Most of the buildings located in the area of the Nickel Plant (the first metallurgical plant) were built on such soils [Setkov et al., 2005].



Fig. 1. Location of the study area (marked in red).

1 – borehole TN-1.

In the late 1950s, pile foundations proposed by M.V. Kim [Kim, 1964, 1967] began to be used for further construction in the city. The new construction sites were mainly found in the areas of frozen dispersed soils and were characterized by complex geocryological conditions with mean annual permafrost temperatures of -2 to -4°C and the presence of talik zones and ice-rich soils.

Because of these challenging conditions, a 200-m-deep borehole was drilled in the then-undeveloped area in 1959, where temperature measurements were subsequently taken. Recently, these measurements have been resumed after a long pause, which allows us to estimate the temperature dynamics of permafrost soils in the center of Norilsk from 1959 to 2024.

STUDY AREA AND WORK PROCEDURE

The territory of the municipal formation of the city of Norilsk is located north of the Arctic Circle, in the zone of permafrost soils and belongs to the continental part of the Arctic.

This region is characterized by a harsh subarctic climate with the mean annual air temperature (Norilsk station) of -9°C (1934–2023). Winters are long and severe, with the absolute minimum air temperature of -57°C . Summer with above-zero mean daily air temperatures lasts for only four months, and the mean monthly temperature of the warmest month is 18.5°C .

The study area is a built-up area in the central part of Norilsk (address: Leninsky Prospekt, 25A), in the immediate vicinity of the site, there are operating buildings and structures for various purposes (Fig. 1).

In this part of the city, there was a 200-m-deep borehole (D-1284 in Fig. 2) was drilled, and temperature measurements in it began as early as in 1959. However, it was destroyed in the early 2000s. In 2022, as part of the program for monitoring the state of permafrost in Norilsk and Taimyr Dolgano-Nenets

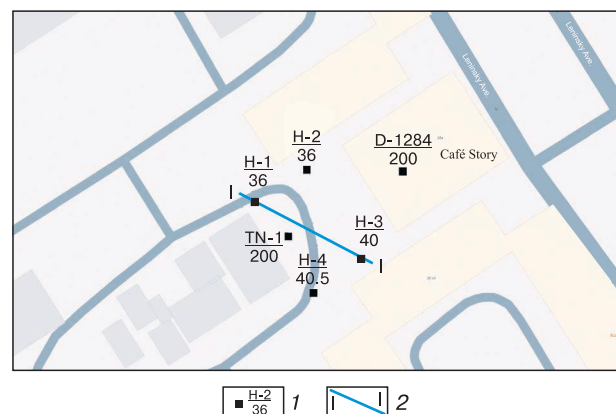


Fig. 2. Location of boreholes in the study area.

1 – borehole number and depth, 2 – line of the engineering-geological section.

municipal district, the Polar Division of PJSC MMC Norilsk Nickel decided to continue monitoring studies of the temperature regime in the city. In this regard, a new 200-m-deep borehole was drilled 30 m from its original location (borehole TN-1 in Fig. 2), and engineering and geological work was carried out to determine the variability of the geological structure and soil properties in the territory (boreholes H-1 to H-4 in Fig. 2). Engineering and geological surveys were carried out by the Norilsk Nickel Technical Services LLC and included the following works:

- Borehole drilling. It was performed using a mechanical core “dry” drilling method with a shortened run and minimum rotation speed of the drill, with full core sampling and careful description of core samples and soils of the mine workings;

- Laboratory determination of the physical thermophysical properties of soils in accordance with current regulatory documents;

- Soil temperature measurements in the borehole conducted in accordance with [GOST (State Standard) 25358-2020, 2020]. The set for field measurement of soil temperature consists of a thermocouple and an ETC-0.1/10 digital electrometer. Temperature measurements using thermometric strings were performed with a spacing of 0.5 m to a depth of 5 m, 1 m (at depths of 5–10 m), and 2 m (down to the borehole bottom). The zero sensor was placed on the ground surface. Before the measurements, the borehole was left to stand for 14 days.

The drilling of the 200-m-deep borehole began in October 2022. The drilling diameter in the loose Quaternary sediments was 122.6 mm; deeper, in the bedrock interval, 96 mm. Geological documentation and photo documentation of the core were carried out during the drilling.

Upon completion of drilling, a casing pipe of 89 mm in diameter was installed and the annulus was cemented; the cement was allowed to harden for 48 h. Then, the borehole was pressured up with compressed air.

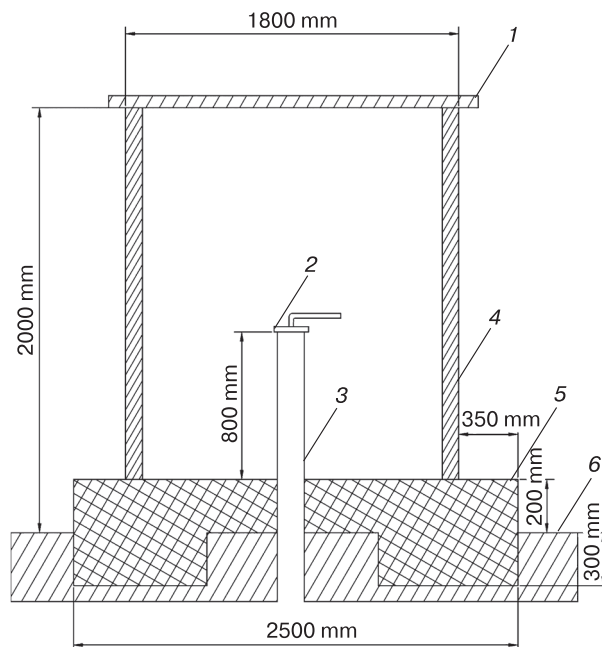


Fig. 3. Diagram of the borehole capping structure.

1 – manhole cover, 2 – device for automatic data transmission, 3 – head, 4 – walls, 5 – concrete slab, 6 – soil.

After drilling and preparing the borehole, a capping structure was installed (Fig. 3) to protect against the negative impact of environmental factors and eliminate the possibility of demolition.

In March 2023, the first soil temperature measurements were taken in the new 200-m-deep borehole; continuous monitoring measurements began in December 2023.

RESULTS

The geological structure of borehole TN-1 is represented by the Lower Triassic intrusive rocks ($\nu\beta T_1$) and Quaternary Ayaklin lacustrine sediments

Table 1. Characteristics of engineering-geological units

EGU no.	Index	Description
1.1	tIV	Filled soil: gravelly soil, frozen, with filler (30.4%) in the form of slightly saline loam, solid in the thawed state, with a medium degree of water saturation
1.2t	tIV	Filled soil: gravelly sand, medium water saturation, medium density, with an admixture of organic matter
1.3t	tIV	Filled soil: fine sand, medium water saturation, medium density
2	IIII–Hak	Silty sand, frozen, icy; in a thawed state – water-saturated, loose
3	IIII–Hak	Sandy loam, frozen, icy; fluid when thawed
4	IIII–Hak	Sandy loam, frozen, low ice content; fluid-plastic when thawed
5	IIII–Hak	Coarse sand, frozen, low ice content; moderately water-saturated in a thawed state, loose
6	IIII–Hak	Loamy sand, frozen, low ice content; plastic in a thawed state
7	IIII–Hak	Medium-grained sand, frozen, low ice content; moderately water-saturated in a thawed state, loose
8	IIII–Hak	Sandy loam with gravel and pebbles, frozen, no visible ice; semi-solid when thawed
9	IIII–Hak	Medium-grained sand, frozen, low ice content; moderately water-saturated in a thawed state; medium density
10	IIII–Hak	Silty sand, frozen, low ice content; water-saturated in a thawed state, medium density
11	$\nu\beta T_1$	Bedrock (gabbro-diorite), very strong, very dense, non-porous, non-softening

(III–Hak). Human-made soils—coarse-grained and sandy—clayey layers – are found at the top of the section. Human-made soils were formed during construction and surface planing by adding imported soil to the design level. At the time of the survey (September 2022), groundwater was not discovered at the work site.

Analysis of field and laboratory data from engineering and geological surveys made it possible to identify 13 engineering-geological units (EGUs) in this territory to a depth of 200 m (Table 1).

The geological section obtained on the basis of engineering and geological surveys is shown in Fig. 4. The difference in the structure of the section in the

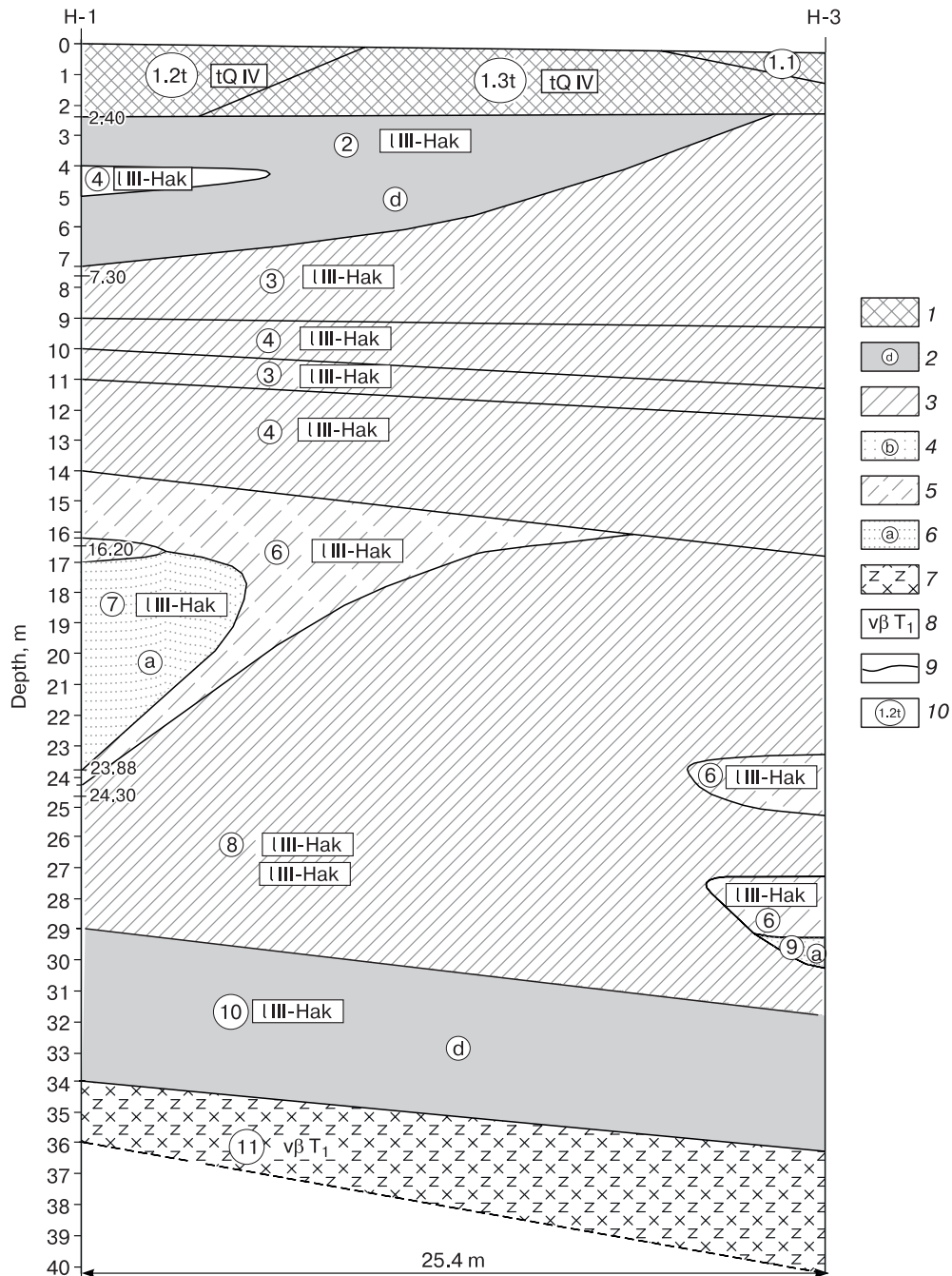


Fig. 4. Geological section along line I–I.

1 – filled soil, 2 – silty sand, 3 – loam, 4 – coarse sand, 5 – loamy sand, 6 – medium-grained sand, 7 – hard rock, 8 – geological index, 9 – EGU boundary, 10 – EGU number designation (description of EGUs is given in Table 1).

Table 2. Calculated characteristics of the soil properties at the site

EGU no.	ρ	ρ_s	w	w_L	w_p	i_{tot}	i_i	D_{sal}	T_{bf}	λ_{th}	λ_f
	g/cm ³		fraction unit					%	°C	$\frac{W}{(m \cdot ^\circ C)}$	
1.1	2.33	2.81	0.06	0.21	0.13	0	0	0.22	—	—	—
1.2t	1.98	2.78	0.14	—	—	—	—	—	—	1.12	—
1.3t	2.02	2.82	0.17	—	—	—	—	—	—	2.07	—
2	1.83	2.81	0.30	—	—	0.46	0	0.08	−0.19	1.65	1.95
3	1.77	2.82	0.36	0.26	0.16	0.4	0.23	0.09	−0.19	1.21	1.6
4	1.87	2.84	0.26	0.27	0.17	0.31	0.1	0.1	−0.2	1.37	1.59
5	1.94	2.83	0.20	—	—	0.36	0	0.08	—	1.81	2.22
6	1.96	2.84	0.19	0.18	0.12	0.26	0.06	0.11	−0.2	1.29	1.54
7	1.89	2.82	0.22	—	—	0.37	0.01	0.07	−0.17	1.71	2.1
8	2.03	2.83	0.14	0.23	0.14	0.15	0	0.09	−0.2	1.25	1.63
9	1.95	2.83	0.17	—	—	0.31	0	0.06	—	1.9	2.46
10	1.94	2.81	0.23	0.00	0.00	0.39	0	0.07	−0.21	2.25	2.25
11	2.9	2.97	0.01	—	—	—	—	—	—	—	—

Note: ρ is the soil density; ρ_s is the soil solid phase density; w is the total soil moisture content for frozen soil, natural soil moisture content for thawed soil; w_L is the upper plasticity limit moisture content; w_p is the lower plasticity limit moisture content; i_{tot} is the total ice content; i_i is the ice content due to visible ice inclusions; D_{sal} is the content of soluble salts; T_{bf} is the soil freezing onset temperature; λ_{th} is the thawed soil thermal conductivity coefficient; and λ_f is the frozen soil thermal conductivity coefficient.

areas of boreholes H-1 and H-3 (spaced apart at 25 m) is clearly seen. In the area of borehole H1, layers EGU 4, EGU 6, and EGU 7 appear, and gabbro-diabase layer (EGU 11) extends down to 200 meters.

Statistical processing of data on soil properties was carried out in accordance with [GOST (State Standard) 20522-2012, 2020]; normative and calculated characteristics of physical and thermophysical properties of soils were determined by laboratory methods (Table 2).

In the upper part of the section (EGU 1.1–EGU IGE 1.3t) to a depth of 1–3 m, there are human-made (technogenic) soils, mainly within the seasonally thawed layer. The thickness of technogenic soils is self-compacted; since the time of construction, it has been caked. Due to the chaotic nature of the fill and the uneven composition of technogenic soils, anisotropy in the physical and thermal properties of these soils is noted.

Differences in the geological structure, insolation patterns, and heat emission from nearby buildings af-

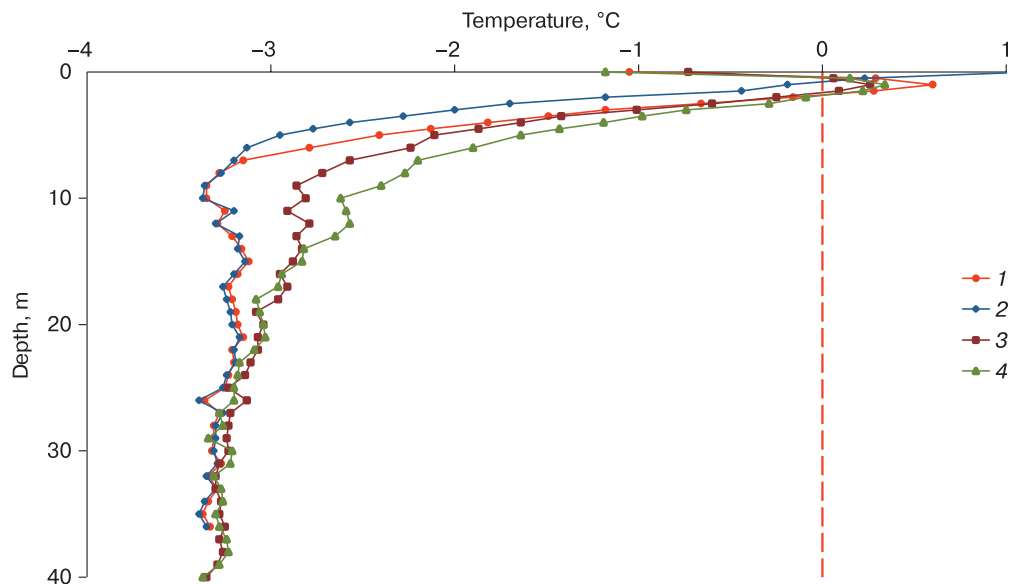


Fig. 5. Soil temperature in the deep well area in September 2022.

1 – borehole H-1; 2 – borehole H-2; 3 – borehole H-3; 4 – borehole H-4.

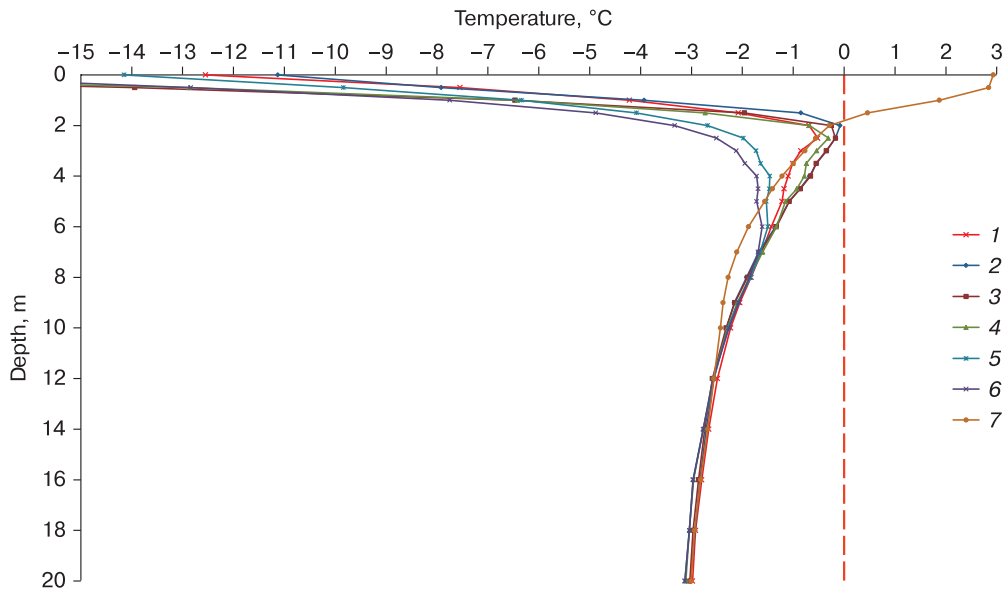


Fig. 6. Soil temperature regime in well TN-1.

1 – Mar. 27, 2023; 2 – Dec. 22, 2023; 3 – Jan. 14, 2024; 4 – Feb. 15, 2024; 5 – Mar. 15, 2024; 6 – Mar. 27, 2024; 7 – Sept. 15, 2024.

fect the temperature regime in the boreholes. At the same time, a significant difference in temperatures is observed for boreholes H-1 and H-2 on one side of the site, and H-3 and H-4 on the opposite side (Fig. 5). At a depth of 10 m, the difference reaches 1°C ; at a depth of 30 m, it virtually disappears, which attests to significant technogenic influence in the upper part of the section. The presence of underground heat-generating collector systems for sanitary and technical utilities in this area may be one of the reasons for such a significant difference in temperature between boreholes H-1, H-2 and H-3, H-4. However, no surveys of the utilities have been conducted.

In March 2023, the first temperature measurements were taken in a new 200-m-deep borehole, and continuous monitoring measurements began in De-

cember. For 1.5 years, temperature fluctuations have been observed to a depth of 12 m; below, the temperature remains virtually unchanged (Fig. 6). Observations will be continued to confirm the stabilization of the temperature regime around the new borehole.

DISCUSSION

The data obtained at the Norilsk weather station indicate that the mean annual air temperature has increased since the mid-1980s (Fig. 7). In the Norilsk industrial region, the rate of this increase reaches $0.7^{\circ}\text{C}/\text{decade}$. In the recent years, abnormally warm annual air temperatures have been recorded, such as -3.78°C (in 2020), against the mean annual air temperature of -8.98°C for the period from 1934 to 2023. The rate of warming varies by season, with the greatest temperature rise in the spring season.

In general, the climate in the study area has become more humid in the past few decades due to an increase in precipitation and changes in its fallout patterns.

Changes in precipitation and air temperature have led to changes in the snow accumulation regime. The formation of the snow cover begins 5–10 days later, and the breakdown of the snow cover takes place about 5 days later.

When analyzing the temperature regime based on stationary observations in urban environments, it is necessary to take into account the urban heat island effect – the difference between air temperatures in the city and in the surrounding area – associated

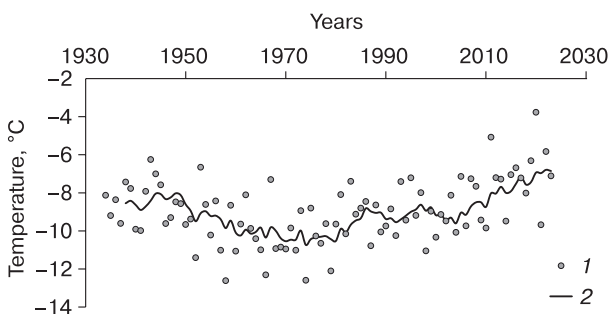


Fig. 7. Change in the mean annual air temperature according to Norilsk weather station data from 1934 to 2023.

1 – actual temperature values; 2 – seven-year averages.

with the local influence of urban development and heat sources in the city. One of the main reasons for the appearance of urban heat island is related to a more efficient absorption and retention of solar heat by the city [Ryu, Baik, 2012]. For polar night conditions in winter in northern cities of Russia, the results of numerical modeling of the urban heat island showed a significant contribution of the anthropogenic factor (due to direct heating) to the formation of this phenomenon [Varentsov *et al.*, 2018].

The paper [Esau *et al.*, 2021] presents an analysis of the characteristics of the urban heat island for settlements in the Arctic with a population of more than three thousand people. For Norilsk, the average intensity of the urban heat island in the recent decades has averaged $+2.7^{\circ}\text{C}$ in summer and $+3.2^{\circ}\text{C}$ in winter. The intensity of the urban heat island in this case is calculated as the difference between the maximum temperature in the city and the average temperature in the region outside the city [Miles, Esau, 2017; Esau *et al.*, 2021]. However, daily values of the heat island intensity can differ greatly from the average values and reach $7\text{--}8^{\circ}\text{C}$ [Varentsov *et al.*, 2014].

To perform calculations, an important task is to assess the impact of the urban heat island on the temperature regime of soils in the city. To assess this factor, a comparison was made between the temperature data from the zero sensor (located at the surface) in borehole TN-1 in the center of Norilsk (measurements are taken three times a day) and the air temperature at the Norilsk weather station (located 2.5 km west of the borehole, measurements are taken every 3 hours) from December 22, 2023 to May 31, 2024 (Fig. 8). As a result, it was found that there is some phase shift in temperature change in the bore-

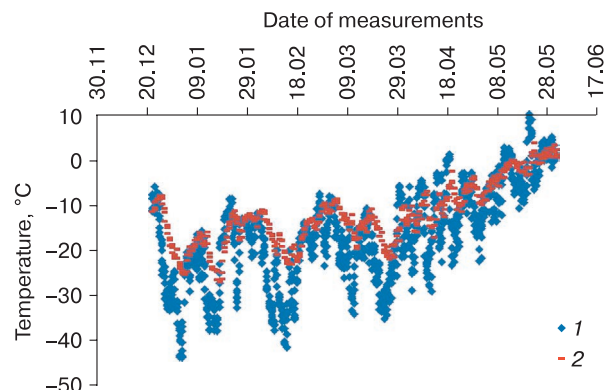


Fig. 8. Air temperature T at the Norilsk weather station and the temperature of the earth's surface in a borehole in the city center.

1 – weather station; 2 – borehole.

hole, and that its amplitude is much smaller, but the values are within the range of air temperature data variability and do not exceed them. The research will be continued, including the use of automatic meteorological stations, to identify differences in the annual cycle in the research area.

Over the past 65 years (from 1959 to 2024), the soil temperature at a depth of 10 m has increased by 4.3°C in the area of deep boreholes. Significant temperature fluctuations have been recorded to a depth of 90 m; deeper, to a depth of 200 m, the temperature has changed by an average of 0.3°C . At the same time, over the period from 1959 to 1985, the soil temperature at a depth of 10 m increased by only 1.9°C (Fig. 9).

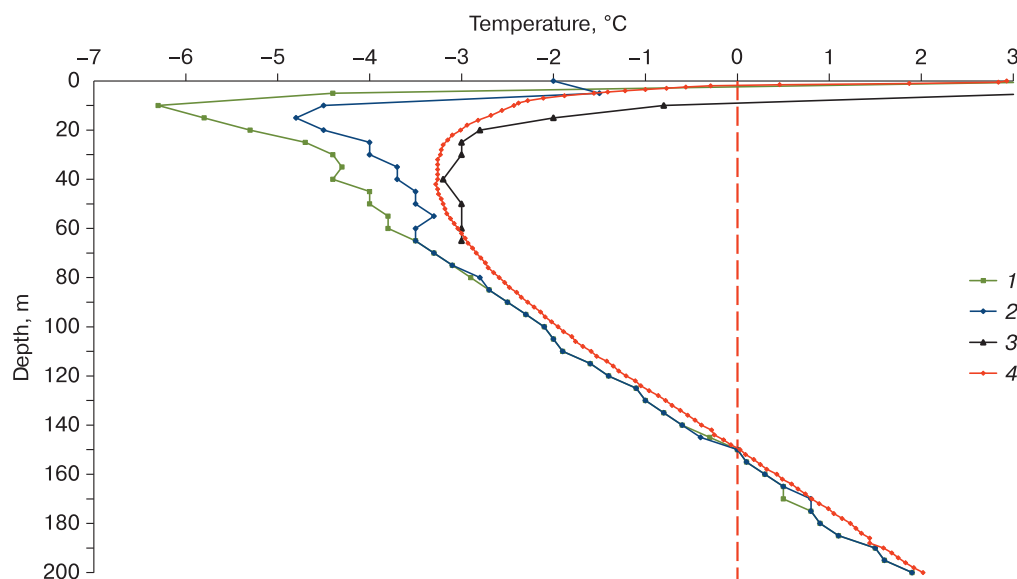


Fig. 9. Dynamics of the temperature regime of soils in the central part of Norilsk.

1 – July 1959; 2 – December 1985; 3 – July 2006; 4 – September 2024.

At the same time, the geothermal gradient has remained virtually unchanged: 3.7°C per 100 m. It is important to take into account that measurements in the 1960s–1980s were carried out using inertial thermometers, while, at present, semiconductor sensors are used, the accuracy of which is much higher. This may be one of the reasons for the discrepancies in the temperature curves below 90 m, where the influence of the technogenic factor is minimal. In addition, only next year will it be possible to assess whether the new borehole has come to the temperature equilibrium with the surrounding soil after additional heating during drilling, and whether this affects the temperature regime. Measurement data show that, in the upper part of the section, there was a sharp increase in temperature until 2006 (the temperature at a depth of 10 m was –0.8°C), which could be due to the construction of a store nearby, as well as emergency situations in the communications of houses. In the early 1990s, several accidents related to the heat and water supply of the city were recorded, which led to the need to drain water under the buildings. Man-made flooding and salinization of soils (and according to the results of laboratory studies, the salinity of soils in the seasonal thawing layer is three times higher than in the soils located below) led to a widespread increase in the depths of seasonal thawing and changes the soil temperature and hydrogeological conditions. In the period from 1997 to 2002, the largest number of house closures occurred in Norilsk due to irreversible deformation caused by the thawing of frozen soils [Kotov, Avilov, 2023].

It should also be taken into account that in 2023–2024, measurements were carried out in another borehole because it was impossible to restore the initial borehole in the same place. Snow removal was carried out at the location of the borehole (TN-1), which led to a decrease in the soil temperature in the area. Also, the new borehole is now located in a capping structure, so there is no snow cover there, and snow in the Norilsk industrial region has the greatest impact on soil temperature regime [Gribenets, Tolmanov, 2021].

Thus, for Norilsk, as for other cities in the Arctic zone, the main reasons for the increase in soil temperature are improper operation of buildings and structures, failure to take into account geocryological conditions, composition and properties of soils [Kotov, Khilimoniyuk, 2021; Brushkov et al., 2023].

As of September 2024, permafrost thickness at the site has decreased slightly – by 2.3 m over 65 years under the influence of heat flow from the subsoil: in 1959, it was 150 m, and now it is 147.7 m. However, these data are preliminary and may change due to further temperature equilibration in the deep borehole after some warming during drilling.

CONCLUSIONS

As a result of the work, temperature indicators and changes in permafrost thickness, soil composition and properties in the area of deep boreholes in the central part of Norilsk were determined, namely:

- The geological structure of the study area is represented by the Lower Triassic intrusive igneous rocks (vβT₁) and Quaternary Ayaklin lake sediments (III–Hak). Man-made soils represented by coarse-grained and sandy-clayey varieties, lie on top. In this area, 13 engineering-geological units have been identified, and the gabbro-diorite unit has the greatest thickness (more than 150 m). The geological structure on a site measuring 25 by 25 m differs significantly.

- Indicators of physical and thermophysical properties of frozen soils have been calculated for each engineering-geological unit.

- Differences in the geological structure, insolation regime of the territory, and heat emission of nearby buildings affect the temperature regime of the soils. At the same time, there is a significant difference in the temperatures obtained in September 2022 for boreholes H-1 and H-2 located on one side of the site, and H-3 and H-4, on the other side. At a depth of 10 m, the difference reaches 1°C; at a depth of 30 m, it virtually disappears, which indicates a significant technogenic impact in the upper part of this area.

- Over the past 65 years (from 1959 to 2024), there was an increase in soil temperature at a depth of 10 m by 4.3°C in the area of deep boreholes. Significant temperature fluctuations were recorded to a depth of 90 m; below, down to a depth of 200 m, the temperature changed by an average of 0.3°C.

- The geothermal gradient is 3.7°C/100 m.

The obtained data can be used to assess the impact of natural and man-made factors on the temperature regime of soils, to create well-founded forecast models verified on the basis of actual measurements, and to make scientifically sound management decisions on the functioning of urban infrastructure in Norilsk.

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CRYOLITHOGENESIS

ICE COVER IN SHARAPOV SHAR LAGOON (THE KARA SEA) AND ITS IMPACT ON FREEZING OF BOTTOM SEDIMENTS

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The conditions of ice cover formation in the Sharapov Shar Lagoon (Bay) of the Kara Sea and its role in sediment freezing at the fast ice/bottom contact zone are considered. Based on field measurements, remote sensing data, and calculations, characteristic periods of the formation and destruction of the ice cover and its thickness over the past 46 years have been analyzed. It is shown that fast ice in the lagoon is observed for an average of 252 days a year; the ice thickness by the end of winter averages 139 cm and can exceed 160 cm in some winters. The base of sea ice is in contact with the bottom within a significant part of the lagoon: 49% of the area in mild winters and 60% of the area in severe winters. Numerical simulation indicates that at sea depths of more than 0.9 m, a seasonally frozen layer with a thickness of up to 0.8 m for loamy soils and 1.2 m for sandy soils is formed under the fast ice in winter and completely thaws in summer. Calculations have also shown that in the 20th century, there were conditions favorable for the formation of permafrost in the zone where sea ice is frozen to the bottom. The formation of permafrost was possible at depths of up to 90 cm. In the 21st century, rising air temperatures and increasing ice-free periods lead to permafrost degradation. Under modern conditions of global warming, sandy sediments can remain in the frozen state at sea depths of up to 0.7 m; clayey sediments – up to 0.3 m.

Keywords: frozen ground, permafrost zone, sea ice, fast ice, simulation of ground freezing.

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INTRODUCTION

Significant areas of the coasts and shelves of Russia's Arctic seas are occupied by permafrost (including subsea permafrost) which is sensitive to climate changes. The shortening of the ice season and rising air and water temperatures affect the formation and preservation of permafrost and seasonally frozen sediments. Industrial development of the northern territories is impossible without knowledge of the characteristics of the soils composing the bottom and shores of the Arctic seas. Therefore, studies of the relationship between changes in the ice regime of the sea and the formation of permafrost are highly relevant. This issue is addressed in this paper.

The final stage of shelf permafrost degradation began 7–8 kyr ago and continues to this day [Romanovsky *et al.*, 1999]. Over the past 30 years, the subsea permafrost degradation in the Russian Arctic seas has accelerated significantly, reaching 18 cm/yr [Sergienko *et al.*, 2012; Semiletov *et al.*, 2020].

One of the factors of the formation of subsea seasonally frozen ground and permafrost is the ice cover. The presence of permafrost under beaches, coastal marshes, and areas, where sea ice freezes with bottom sediments, has been noted by many researchers [Gri-

goriev, 1987; Zhigarev, 1997], including observations in the area of Cape Kharasavei [Vasiliev *et al.*, 2011, 2017]. In addition, in the area of Cape Burunny (Baidaratskaya Bay of the Kara Sea), frozen sediments with a temperature of –9.9°C were found on an underwater spit (shoal) at the sea depth of 2–3 m; these sediments pinch out at a distance of 2.3 km from the shore [Buldovich, Danilov, 1998]. New formation of permafrost in some shallows of Cape Halkett (Alaska, Beaufort Sea) associated with the development of spit, is also mentioned in [Jones *et al.*, 2009].

According to simulation [Buldovich, Danilov, 1998], in the shallow waters of Baidaratskaya Bay, subsea permafrost with temperatures of –8 to –10°C is formed down to sea depths of 2–3 m, while at depths from 2–3 to 6–8 m, thawed or seasonally frozen sediments are common. Climate warming observed since the end of the 20th century and particularly pronounced in the Arctic [Ogorodov *et al.*, 2022] will obviously exert a negative impact on subsea permafrost, including its complete thawing.

Sharapov Shar Bay is located in the western part of the Kara Sea (Fig. 1). The predominant depth of the water area is 2–3 m. From the west, the bay is

protected from the direct impact of waves and ice by the aggradational Sharapovy Koshki Islands forming lagoon shoals. The eastern shore of the bay is characterized by thermoabrasion and abrasion-thermodenu- dation scarps in frozen sediments and a predominant- ly sandy beach. The bottom and shores of the bay are composed of fine-grained sands, silts, and clays [Mo- tychko *et al.*, 2013].

The height of large waves in the bay does not ex- ceed 110 cm for storms recurring once in 50 years. Tidal fluctuations are weak: the spring tide height is 23 cm, and tidal current velocities are about 5 cm/s. Total currents during storms do not exceed 50 cm/s [Makarov, Vyaly, 2020]. The ice cover is formed under calm conditions, when movement of young ice forms is insignificant, and the processes of ice hummocking and grounded hummock formation have a local char- acter. As a consequence, the bay is almost completely covered with smooth first-year ice [Dumanskaya, 2014].

Most of the bay is located within the Kruzen- shtern field, one of the key oil and gas development sites in the Arctic. The arrangement of wells on arti- ficial islands is being discussed within the framework of development projects [Makarov, Vyaly, 2020; Solo- martin, Kozlov, 2021]. To assess the stability of hydrau- lic structures, it is important to know the character- istics of bottom sediments and their state (frozen, unfrozen cryotic, thawed).

The purpose of this work is to evaluate the role of fast ice in the formation of frozen ground in the shelf zone using Sharapov Shar Bay as an example. No studies focused on the relationship between climate change, the timing and thickness of sea ice, and the formation of seasonally frozen sediments and perma- frost have been conducted until now.

MATERIALS AND METHODS FOR MODELING ICE COVER AND FREEZING OF BOTTOM SEDIMENTS

The timing of *the ice regime phases* (stable ice formation, fast ice formation, breakup of fast ice, com- plete clearing of the water area from ice) in Sharapov Shar Bay was determined for the period of 2000– 2023 based on the analysis of satellite images [https://worldview.earthdata.nasa.gov]:

Terra (MODIS) (October–November and May– July 2000–2023);

Sentinel-1a (October–November, April, June 2014–2023).

The thickness of sea ice in the bay was calcu- lated on the basis of sums of frost degree days (SFDDs) according to data from the Marre-Sale weather station (Roshydromet network) for the pe- riod from 1977 to 2023, excluding the 1978/79, 1979/80, 1980/81 and 2001/02 winter seasons [https://rp5.ru, http://aisori-m.meteo.ru].

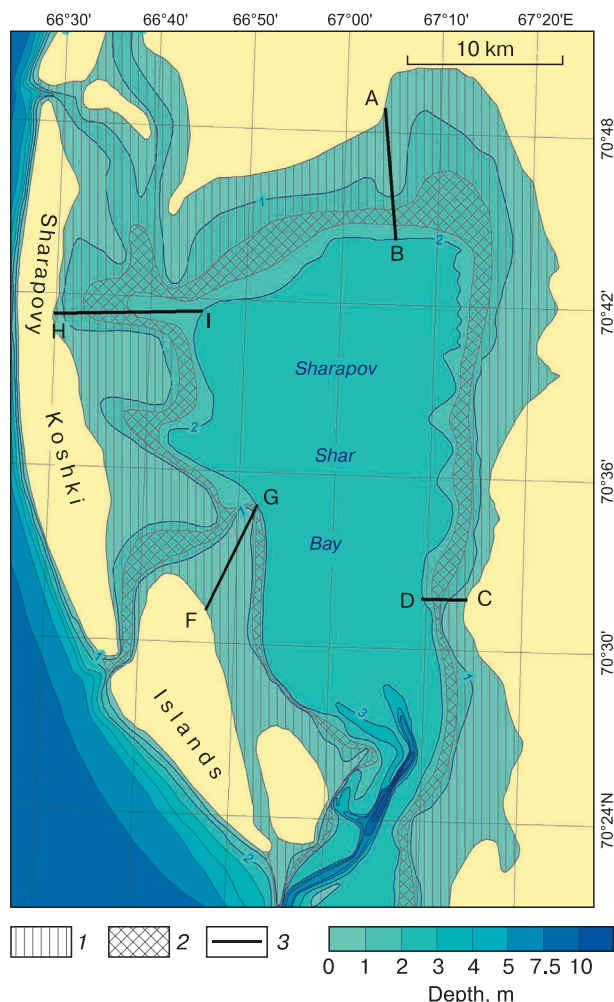


Fig. 1. Bathymetric map of Sharapov Shar Bay.

Hatching shows the boundaries of the maximum contact area of sea ice and seabed established on the basis of the calculated sea ice thickness for the conditions of (1) mild and (2) severe win- ter; lines (3) indicate the location of modeling profiles.

The SFDD was converted to ice thickness using the formula of N.N. Zubov [1956], which showed an adequate correlation with the in-situ data obtained during drilling in the period of the maximum ice cover development (March–April) in 2011–2023 (Fig. 2).

A bathymetric map of Sharapov Shar Bay was constructed to estimate the *position of the bounda- ry of the fast ice–sea bottom contact area* (Fig. 1). The digital model of the seabed relief was obtained from the General Bathymetric Chart of the Oceans (GEBCO) data [https://www.gebco.net], marine navigation maps, and local surveys. All depths were converted to the 1977 Baltic Height System (BS-77) and then referenced to the mean annual sea level (MASL). According to the Kharasavei weather sta- tion (1977–1987; after 1987, this station was abo- lished), the MASL in the bay is 63 cm (BS-77).

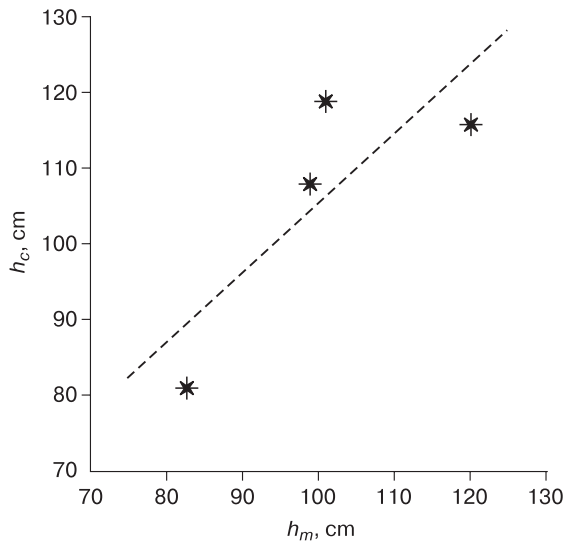


Fig. 2. Measured (h_m) versus calculated (h_c) ice thickness.

Numerical simulation of bottom sediments freezing under sea ice was carried out in the QFrost program [Ershov, 1999]. This work does not consider the history of the formation of the Sharapov Shar Bay sediments, their cryogenic structure, and the development of permafrost in the Late Pleistocene and Holocene, unlike other studies [Romanovsky, Tumskoy, 2011; Kotlyakov et al., 2015; Tumskoy, Kuznetsova, 2022]. Heat fluxes from the Earth's interior that are usually used in modeling shelf permafrost over long periods [Romanovsky et al., 1999, 2003] were not taken into account because of the much smaller modeled depths.

The thermal properties of the modeled sediments given in Table 1 [Aleksyutina, Motenko, 2017] are typical of the bottom sediments with the high water content (30–50%) and skeletal density (1.3–1.6 g/cm³). Because of the lack of actual data, a single-layer sediment section was considered. Bottom sediments of the Kara Sea shelf are saline, and their freezing temperature was taken equal to -1.5°C [Vasiliev et al., 2016]. The initial temperature distribution was derived from observation data at the Marre-Sale station [Dubrovin et al., 2015]. The temperature at the depth of zero annual amplitudes was -1.4°C . The calculations were carried out in two stages: first, the modeling was carried out for the last 30 years of the 20th

Table 1. Thermal properties of sediments used in modeling

Sedi- ment	λ_t	λ_f	C_t	C_f	Q_{phase} , MJ/m ³
	W/(m·K)		J/(m ³ ·K)		
Sandy	1.8	2.0	2800	2000	152.4
Loamy	1.45	1.55	2865	2250	289.3

Note: λ_t , λ_f – coefficients of thermal conductivity of sediment in the thawed and frozen states, respectively; C_t , C_f – volumetric heat capacity of sediment in the thawed and frozen states, respectively; Q_{phase} – heat of phase transition.

century, assuming that the sediments were in the unfrozen cryotic state at the time of the beginning of the calculations; the obtained results were then used as the initial conditions for modeling the state of the sediments in the 21st century. Due to the shallowness of the bay and the fact that the ice cover settles on the bottom, it was decided to calculate one-dimensional problems with varying boundary conditions at the surface.

To model the temperature regime of bottom sediments in the 20th century, mean monthly data on air temperature, snow cover depth, and sea ice thickness for the period of 1970–1999 were used (Table 2). Proskuryakov's formula [Osokin et al., 2017] was used to calculate the thermal conductivity of the snow; for simplification, a fixed value of snow density (250 kg/m³) was used. The coefficient of thermal conductivity of sea ice used in the calculations was constant and equal to 2.1 W/(m·K).

Ice cover in Sharapov Shar Bay develops gradually; fast ice is formed near the shore and grows during the entire cold season. Thus, in the winter season, there are some areas in the bay, where the ice lies on the bottom. Further from the shore, at greater depths, there is a layer of water between the ice and the bottom. While modeling, we took into account the presence of water and ice by setting different boundary conditions at the sediment surface. In the calculations, we assumed that fast ice formation starts from the shore. With increasing sea depth, the thickness of the fast ice increases; however, the time of contact with the bottom decreases. Boundary conditions of the third kind, which include temperature and the heat transfer coefficient, were assumed. Air temperature and the heat transfer coefficient were used for each month of each year. The heat transfer coefficient (α , W/(m²·K)) was calculated taking into account

Table 2. Sea ice and snow characteristics in Sharapov Shar Bay in 1970–1999

Characteristic	Month							
	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.	May	June
Sea ice thickness, cm	30	60	87	106	114	124	132	122
Snow depth, cm	4.6	8.9	12.9	21.0	23.0	26.7	22.8	6.5
Snow thermal resistance, (m ² ·K)/W	0.17	0.32	0.47	0.77	0.84	0.98	0.84	0.24

Table 3. Boundary conditions for different sea depths at the initial modeling stage

h	Parameter	Month											
		Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
30–135	T_{air}	–21.5	–22.4	–17.0	–13.6	–5.3	2.1	7.7	6.8	3.3	0	–13.0	–18.1
30	α	1.624	1.100	1.018	0.894	1.022	2.616	∞	∞	∞	∞	3.225	2.139
60	α	1.318	0.951	0.889	0.793	0.892	1.904	∞	∞	∞	∞	∞	1.638
90	α	1.109	0.837	0.789	0.712	0.791	1.497	∞	∞	∞	∞	∞	∞
105	α	∞	0.790	0.747	0.678	0.749	1.352	∞	∞	∞	∞	∞	∞
115	α	∞	∞	0.721	0.657	0.723	1.271	∞	∞	∞	∞	∞	∞
125	α	∞	∞	∞	0.637	0.699	1.198	∞	∞	∞	∞	∞	∞
135	α	∞	∞	∞	∞	0.676	1.133	∞	∞	∞	∞	∞	∞

Note: T_{air} – air temperature, °C; α – heat transfer coefficient, $\text{W}/(\text{m}^2 \cdot \text{K})$; h – sea depth, cm.

the ice thickness (assuming that it is equal to the depth of the bay). The heat transfer coefficient is inversely proportional to the thermal resistance ($\alpha = 1/\Sigma R$, where R is the thermal resistance), which is calculated as the ratio of the layer thickness to its thermal conductivity coefficient ($R = h_{\text{layer}}/\lambda_{\text{layer}}$).

In winter, there is no stratification of water under the ice, and the water temperature is close to the freezing point of -1.8°C . Mean monthly air temperatures (T_{air} , °C) from 1977 to 1999 according to data from the Marre-Sale weather station and heat transfer coefficients (α , $\text{W}/(\text{m}^2 \cdot \text{K})$) for different sea depths used as upper boundary conditions at the initial stage of modeling are given in Table 3. As the modeling was carried out for the first tens of meters of the section, the heat flux from the subsurface, as the lower boundary condition, was equated to $0 \text{ W}/\text{m}^2$.

For the period from 2001 to 2022, mean air temperatures and heat transfer coefficients for each month of each year were used as upper boundary conditions.

MODELING RESULTS AND DISCUSSION

Ice regime. An analysis of remote sensing data for 22 ice seasons in the 21st century demonstrated that, on average, stable ice formation in the bay begins in the second ten-day period of October. Fast ice in Sharapov Shar Bay usually begins to form one or two weeks after the onset of ice formation. Initially, a

200-m-wide stretch of fast ice is formed near the shore. The minimum number of days for the fast ice to form was observed in 2016 – three days; the maximum was 13 days in 2017 and 2021. On average, fast ice with the ice thickness of 10–15 cm is formed in the third ten-day period of October.

By the beginning of May, the ice cover reaches its maximum development. Melting and disintegration of the ice cover occurs under the influence of solar radiation and heat brought by currents and air masses. In early May, air temperature becomes positive and the formation of ice stops. The first signs of ice melting are the appearance of puddles on the ice surface, which gradually expand. Intensive ice melting occurs in June. Breakup of fast ice typically takes place in the third ten-day period of June.

Clearing of the Sharapov Shar Bay water area from ice begins in late June – early July and proceeds rather quickly, in a few days. Local ice hummocks and grounded ice hummocks are destroyed within one or two weeks after the fast ice breaks up. Full clearance of the water area from ice occurs in the first ten-day period of July.

Summary data on the timing of major ice phases in Sharapov Shar Bay are given in Table 4.

Sharapov Shar Bay is characterized by stable fast ice throughout the entire winter, from 217 to 271 days; on average, 252 days per year. Thus, the stationary ice cover is observed for 7.5–9 months per year. Taking into consideration drift ice, the ice sea-

Table 4. Dates of ice phases in Sharapov Shar Bay

Ice phase	Date		
	early	middle	late
Stable ice formation	1 st ten-day period of October (2013, 2014, 2015)	2 nd ten-day period of October	1 st ten-day period of November (2016, 2022)
Fast ice formation	2 nd ten-day period of October (2014, 2015)	3 rd ten-day period of October	1 st ten-day period of November (2016, 2022)
Fast ice breaking in	1 st ten-day period of June (2020)	3 rd ten-day period of June	1 st ten-day period of July (2019)
Complete sea ice clearance	3 rd ten-day period of June (2015, 2016, 2020, 2023)	1 st ten-day period of July	2 nd ten-day period of July (2014, 2017, 2019)

Table 5. Mean values of the sums of frost degree-days (SFDD) for severe, moderate and mild winters according to data from the Marre-Sale station, 1977–2023

Winter type		Number of winters	SFDD, °C
Severest		1	–4556
Severe	without the severest winter	8	–3968
	with the severest winter	9	–4042
Moderate		22	–3278
Mild	without the warmest winter	10	–2626
	with the warmest winter	11	–2588
Warmest		1	–2212

Table 6. Calculated values of sea ice thickness and sea ice – seabed contact area in Sharapov Shar Bay for mild and severe winters

Winer type	Sums of frost degree-days, °C	Ice thickness, cm	Sea ice – bottom contact area	
			km ²	% from the total area of the bay
Mild	–2588	121	582	49
Severe	–4042	157	716	60

son in Sharapov Shar Bay lasts from 232 to 283 days (on average, 263 days).

Ice thickness. According to the results of the SFDD calculations and typification of winter season based on them [Dumanskaya, 2014], for 42 ice seasons, the winter of 1998/99 (–4556°C) was the severest, while the warmest winter was in 2019/20 (–2212°C). Eight severe winters and ten mild winters were recorded. The 21st century accounts for only two severe winters and nine mild winters. Table 5 presents the SFDD typical values for moderate, severe, and mild winters.

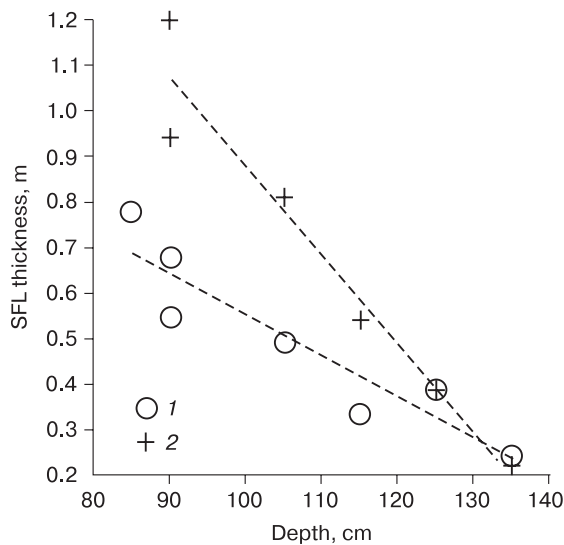


Fig. 3. Estimated thickness of the seasonally frozen layer (SFL) under the fast ice at different water depths for (1) clayey and (2) sandy bottom sediments.

The calculated ice thickness in the coldest season was 167 cm; in the warmest season, 110 cm. The bathymetric map of Sharapov Shar Bay (Fig. 1) shows the boundaries of the ice/seabed contact area for severe and mild winters. The maximum area of the contact in severe winter is 60% of the total bay area; in mild winter, it is about 49% (Table 6).

Results of the sediment freezing simulation.

The formation of a seasonally frozen layer (SFL) in the 20th century was possible at all ice cover thicknesses. The greater the ice thickness, and the later it settles on the seabed, the smaller the SFL thickness. The authors found that, provided fast ice forms in December, sandy bottom sediments do not transfer into the perennially frozen state at the ice thickness of more than 90 cm. For loamy sediments, this threshold is 80 cm because of the higher heat capacity. The SFL thickness varies from 0.2 to 1.2 m for sandy sediments and up to 0.8 m for clayey sediments (Fig. 3).

At the same ice thickness (90 cm), the seasonal freezing depth varies by 20% depending on the time of the ice formation (Fig. 3). At sea depths of up to 0.8–0.9 m, ice forms earlier and fast ice forms by November (ice thickness 30 cm) and December (more than 60 cm).

According to the modeling results, the formation of ice and its freezing with bottom sediments at the beginning of the cold season created favorable conditions for the formation of permafrost in the 20th century. The estimated permafrost thickness for the period of 1970–1999 was up to 3.5–6.6 m (Fig. 4). Sea ice forms throughout the entire winter period, and its thickness increases with increasing depth of the bay. As fast ice begins to form from the shore, some assumptions were made in the calculations. For example, in November, a 30-cm-thick ice layer lies directly

on the bottom at sea depths of less than 30 cm; at greater depths, fast ice is separated from the bottom by a layer of sea water, in our case with a temperature of -1.8°C . The presence or absence of ice contact with the bottom was considered in the model as different

boundary conditions. Table 3 presents averaged air temperatures from 1977 to 1999 and the heat transfer coefficients for different ice thicknesses.

According to previous studies [Vasiliev *et al.*, 2011], permafrost formation on the low shores of

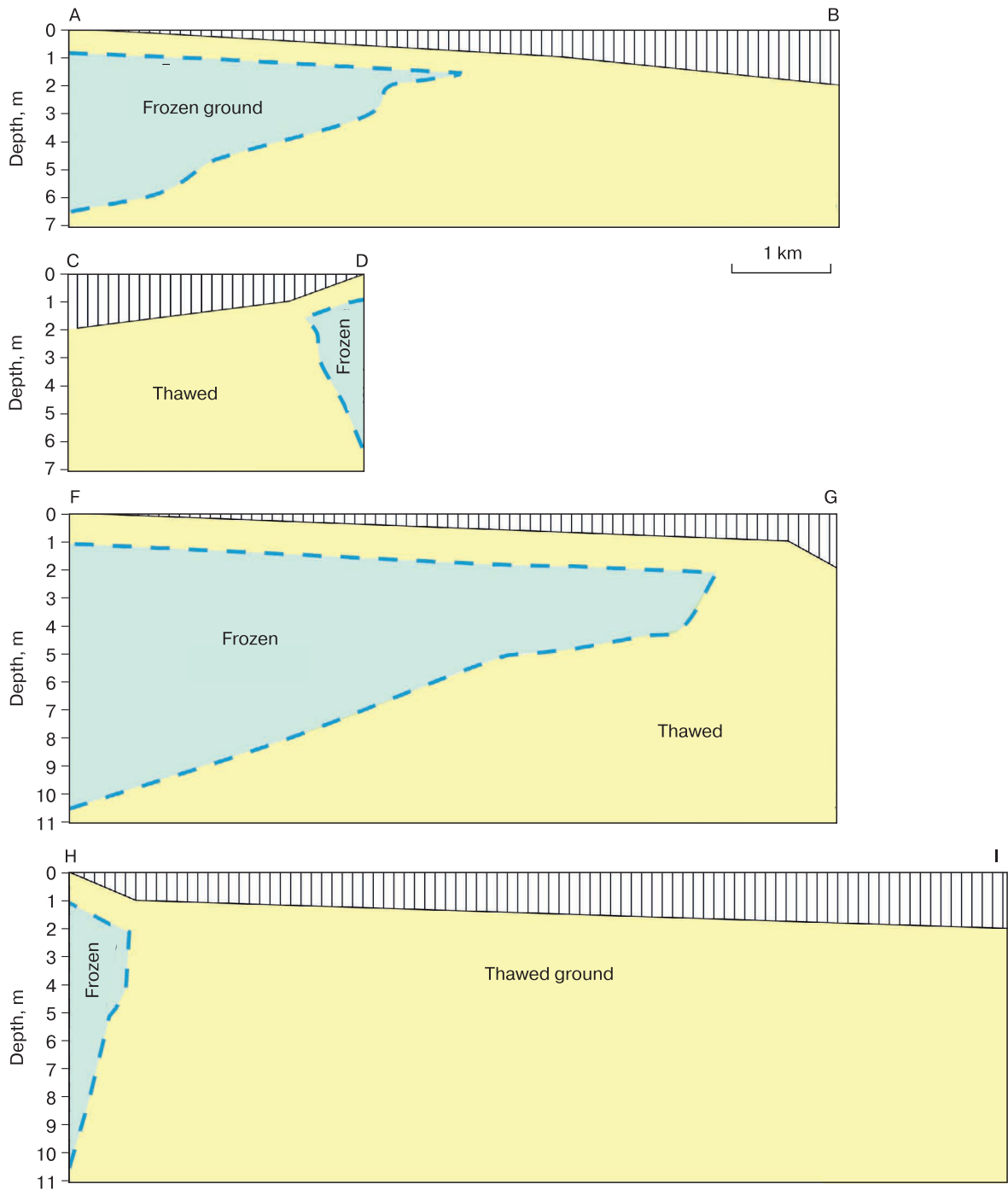


Fig. 4. Schematic sections reflecting the presence of permafrost from modeling data for the last 30 years of the 20th century.

Profiles A–B and C–D are for clayey sediments; sections F–G and H–I are for sandy sediments. The position of the sections is shown in Fig. 1. Hatching indicates sea ice.

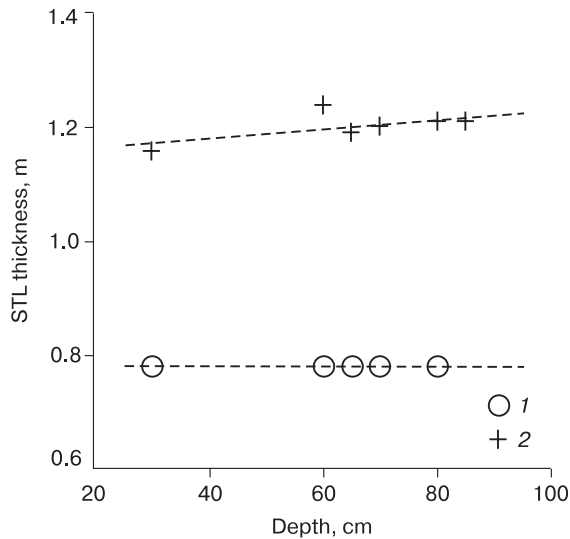


Fig. 5. Calculated thickness of the seasonally thawed layer (STL) in Sharapov Shar Bay at sea depths up to 90 cm for (1) clayey and (2) sandy sediments.

Western Yamal, including the area of Cape Kharasavei, was observed at sea depths of up to 1.5 m. In the area of Cape Burunny (Baidaratskaya Bay of the Kara Sea), frozen sediments were found on an underwater spit (shoal) at sea depths of 2–3 m; the frozen layer wedged out at a distance of 2.3 km from the shore [Buldovich, Danilov, 1998]. According to the modeling, the new formation of permafrost in

Sharapov Shar Bay is possible at shallower depths because of a higher salinity and lower freezing temperature of saline sediments.

In summer, the thickness of the thawed layer is up to 0.8 m for clayey sediments and 1.2 m for sandy sediments. Note that thawing depth does not depend on the ice thickness (Fig. 5).

Modeling of the second stage showed that permafrost, which could have formed over the last 30 years of the 20th century, begins to thaw from the top in the 21st century with an increase in the seasonal thawing depth (Fig. 6). The thickness of the seasonal thaw layer (STL) is greater in sandy soils than in clayey soils due to differences in their thermophysical properties. Complete degradation of the SFL and merging of the STL with the possible permafrost table was noted for 2022, when the permafrost table for sandy soils was at 407 cm at a water depth of 0.7 m, and for clayey soils at 220 cm at a water depth of 0.3 m.

The rates of permafrost degradation in the 21st century depend on sediment textures. In sandy sediments, permafrost can be preserved at sea depths up to 0.7 m; in clayey sediments, at sea depths up to 0.3 m. Since 2015, the STL thickness has increased significantly in comparison with that in 2001–2014. The differences in the STL and SFL thicknesses between 2006–2009 and 2012–2013 (Figs. 6, 7) indicate that bottom sediments subjected to freezing in winter do not have time to thaw completely in summer. That is, we observe non-stationary conditions, when the STL thickness is less than the SFL thick-

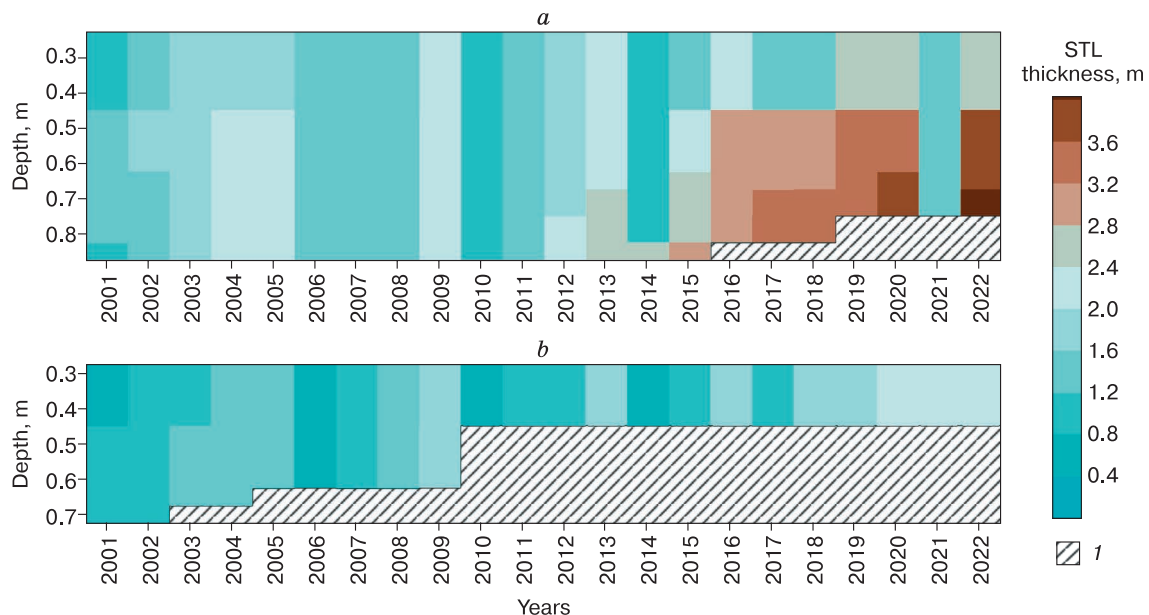


Fig. 6. Changes in the thickness of the seasonally thawed layer (STL) composed of (a) sandy and (b) clayey sediments at different sea depths in 2001–2022.

1 – complete thawing of soils.

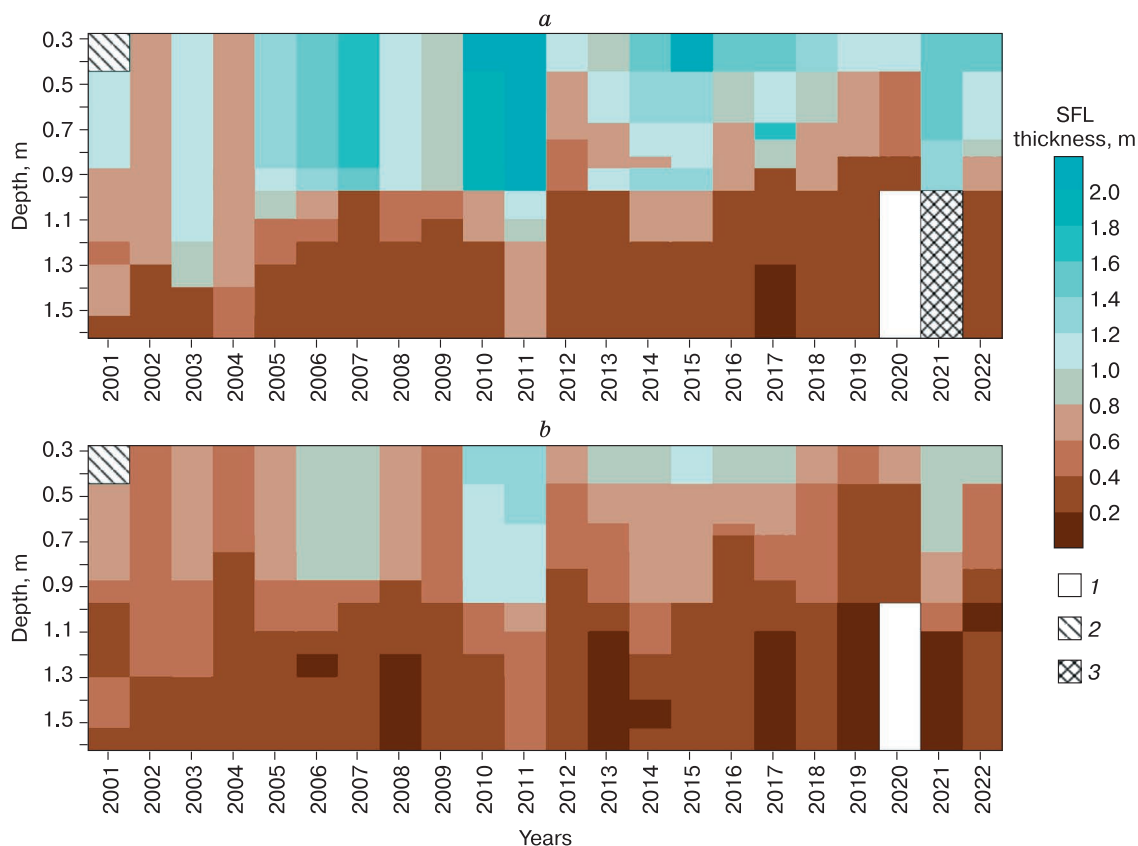


Fig. 7. Changes in the thickness of the seasonally frozen layer (SFL) composed of (a) sandy and (b) clayey sediments at different sea depths in 2001–2022.

1 – SFL is absent, 2 – SFL merged with permafrost, 3 – two-layered structure of the SFL.

ness, and permafrost formation takes place. In the 21st century, these conditions were only observed in 2006–2009 and 2012–2013. In the other periods, the conditions for the permafrost formation did not occur. Over the past decade, the STL thickness has always exceeded the SFL thickness.

M.N. Grigoriev [2006] provides data on the degradation of relict subsea permafrost on the shelf of Eastern Siberia. The degradation rate ranges from 1 to 10–15 cm/yr at contact of permafrost with seawater. New formation of permafrost in subaqueous conditions is possible at sea depths of up to 2.5 m and is observed in shallow-water areas of delta front and shallow bays, where active sedimentation takes place. Modeling indicates that in Sharapov Shar Bay, climatic conditions in the last decades of the 20th century were favorable for the preservation and even local new formation of permafrost. In the 21st century, the probability of permafrost development in Sharapov Shar Bay is absent, and local areas with permafrost that could have formed in the 20th century are degrading.

CONCLUSIONS

A schematic map of the ice cover of Sharapov Shar Bay with indication of the areas, where sea ice is in direct contact with seabed, is presented in this article. It is compiled on the basis of the analysis of long-term remote sensing data, available field observations, and calculated sea ice thicknesses. Modeling of the freezing and thawing of bottom sediments in Sharapov Shar Bay has been performed for a period from 1973 to 2022. The main conclusions of the research can be formulated as follows:

1. The ice season in Sharapov Shar Bay lasts on average from the 2nd ten-day period of October to the 2nd ten-day period of July. The stable fast ice in the bay is preserved for 252 days a year, from the end of October to the end of June.

2. The maximum annual sea ice thickness varies from ~120 cm in mild winters to ~160 cm in severe winters.

3. In winter, ice freezing with the bottom is observed on 50% of the bay water area, resulting in the formation of seasonally frozen sediments.

4. The thickness of the seasonally frozen layer depends on the properties of bottom sediments. At sea depths of more than 80–90 cm, the thickness of the seasonally frozen layer varies in a wide range from 0.2 to 1.2 m for sandy sediments and up to 0.8 m for clayey sediments. The seasonally frozen layer thaws completely during the summer period.

5. The thickness of the bottom sediment layer subjected to thawing in summer does not depend on the ice thickness and is controlled by the properties of the bottom sediment. For sandy sediments, it reaches 120 cm; for loamy sediments, it is about 80 cm.

6. According to modeling data, for the climatic conditions of the period 2001–2022, sediments, in which local development of permafrost took place in the 20th century, could have been preserved in the frozen state at sea depths of up to 0.7 m for sandy bottom sediments and up to 0.3 m for clayey bottom sediments.

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GEOLOGICAL CRYOGENIC PROCESSES AND FORMATIONS

PARAGENESIS OF DEGRADATION AND AGGRADATION PROCESSES
IN FROZEN GROUNDF.M. Rivkin^{1,*}, S.N. Buldovich², D.A. Kaverin³¹ Earth Cryosphere Institute, Tyumen Scientific Centre, Siberian Branch of the Russian Academy of Sciences,
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The regularities of the new formation of frozen ground near the southern boundary of permafrost against the background of a positive climate trend and permafrost degradation are considered. New formation of permafrost against the background of long-term thawing of frozen ground is considered as a paragenetic processes. The spatial heterogeneity of frost heaving and thawing of soils leads to the formation of local irregularities within initially flat hydromorphic areas, where conditions become favorable for the new formation of frozen ground. Even minor altitudinal differentiation of the original surface (0.05–0.20 m) can lead to a radical change in geocryological conditions and formation of fairly low-temperature permafrost. Thus, new formation of frozen ground can take place against the background of general climate warming and permafrost degradation.

Keywords: permafrost, deep position of a permafrost table, frozen ground thaw subsidence, aggradation of frozen ground, perennial ground freezing, frost heaving.

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INTRODUCTION

Permafrost aggradation in talik zones is widespread in high latitudes as well as at near the southern boundary of the permafrost zone. While permafrost aggradation in areas of low mean annual air and, consequently, ground temperatures is rather normal, this phenomenon near the southern boundary of permafrost requires special consideration to understand its spatial and temporal patterns.

Studies on the transformation of geocryological conditions in the upper part of the section near the southern boundary of permafrost indicate that new formation of frozen ground against the background of a positive trend in air temperatures is part of the mechanism of natural permafrost dynamics, including self-preservation of geocryological conditions [Ivanova et al., 2011; Seppälä et al., 2011; Sjöberg et al., 2015; Osadchaya, 2016; Rivkin et al., 2017; Välranta et al., 2021; Rivkin, Buldovich, 2024]. This process significantly influences landscape change patterns and associated paragenetic processes. In particular, large closed taliks and areas with deep permafrost table within wide troughs, drainage hollows, alases (khasyreys), or similar landscapes are subjected to local freezing. Localized thin frozen strata are being formed in the upper part of the section in such areas. In plan, they often appear as mineral-peat mounds of

about 3×3 m in size (Fig. 1a). The thickness of frozen soil in the centers of such mounds 3.0–3.5 m. The layer of annual heat turnover (above the depth of zero annual temperature amplitude) has the same thickness within the mounds. Such mounds with isolated permafrost patches may cover up to 90% of the landscape area (Fig. 1b), which corresponds to the definition of discontinuous permafrost in the surface layer. Thus, a paradoxical geocryological situation appears: the area is classified as thin discontinuous permafrost (up to 90%), whereas the soil remains in the thawed state from a depth of 3.0–3.5 m. In turn, the thawed layer is often underlain by permafrost at depths below 10–12 m. Upon detailed surveys, such areas can be identified as individual geocryological regions.

Though the mechanism of permafrost formation in this case is generally understood, the patterns of this phenomenon appear to be quite ambiguous. Anyway, the process of differentiated permafrost aggradation against the background of a positive temperature trend is associated with local surface heterogeneity: the presence of irregularities (small ridges or mounds) on the initially smooth surface of hydromorphic areas.

The appearance of irregularities on initially flat surfaces significantly alters heat exchange conditions and triggers the mechanism of a local permafrost ag-

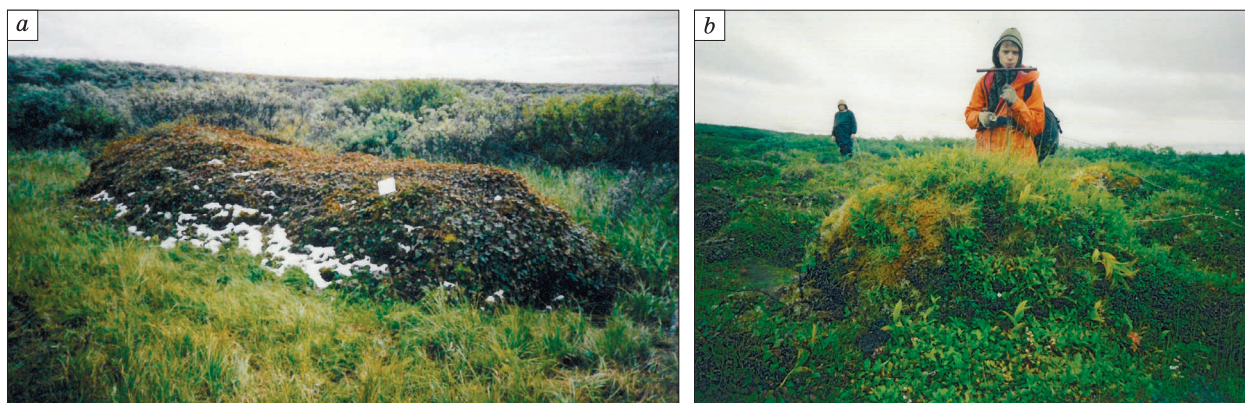


Fig. 1. Heave mounds in the (a) drainage hollow and (b) margin of the drained lake basin, southern part of the Bolshezemelskaya Tundra.

Photo by A.A. Popova.

gradation. The development of cryogenic microtopography on the initially flat smooth surface is usually associated with uneven frost heave. Most intensive heaving on local sites can herewith be triggered by areal inhomogeneity of surface and ground conditions. Some of the combinations of these factors may be random or short-term (seasonal), but subsequently intensify as the permafrost environment evolves.

However, a more quantitatively significant cause of uneven terrain may be associated with uneven thaw settlement in the course of permafrost degradation from the top. In the context of observed climatic changes, these processes are widely developed at the southern periphery of the permafrost zone.

Authors consider several scenarios of permafrost aggradation against the background of a positive temperature trend. First, is permafrost formation triggered by a favorable combination of intra-seasonal factors: air temperature, especially at the beginning of the winter season; snow accumulation regime, particularly the relationship between the onset of low air temperatures and the formation of stable snow cover; soil conditions in the layer of seasonal freezing–thawing; and the occurrence of a combination of favorable freezing and thawing factors over the next few years. Second, is permafrost aggradation against the background of a general lowering of the permafrost table, the appearance of terrain irregularities (due to spatial variations in the ice content within the active layer) on initially smooth hydromorphic surfaces, and the localization of deeper seasonal freezing in these areas. Such microtopographic features create conditions for perennial freezing and permafrost formation.

To study the influence of a positive temperature trend on the geocryological conditions in typical hydromorphic interpolygonal depressions, a series of computational experiments have been conducted. The geocryological conditions typical of waterlogged

depressions, where thawed sediments lie above permafrost at a depth of several meters, were considered. This permafrost is either degrading from the top (residual thaw layer) or has already reached a stationary state (closed talik). As a result of the influence of various local factors, small elevations (up to a few decimeters in height) arise on the initially flat surface, where heat exchange between the soil and the environment differs from that in the surrounding area.

This study focuses on the mechanism of initiation of permafrost aggradation in order to clarify criteria and patterns for stable permafrost formation taking into account the temperature conditions at the surface, the thermophysical properties of vegetation cover, and the cryolithological characteristics of the upper part of the section.

RESULTS AND DISCUSSION

The formation of stable shallow near-surface permafrost, which can occur within a single freeze–thaw cycle, depends on a combination of a set of spatiotemporal factors:

1. Surface conditions (air temperature, snow accumulation regime, microtopography, etc.) that enable rapid ground freezing in localized areas.

2. Soil conditions and pre-winter moisture regime ensuring deep seasonal freezing and the formation of an effective thermal impulse that drives water migration in the freezing soil [*Recommendations...*, 1986; *Orlov et al.*, 1987]. This leads to spatially differentiated water migration towards zones of more intense soil freezing and seasonal frost heave.

3. The interplay of climatic and soil conditions during the seasonal freeze–thaw cycle should ensure the preservation of frozen ground throughout the summer, i.e., winter freezing should not be fully compensated for by summer thawing, which means that

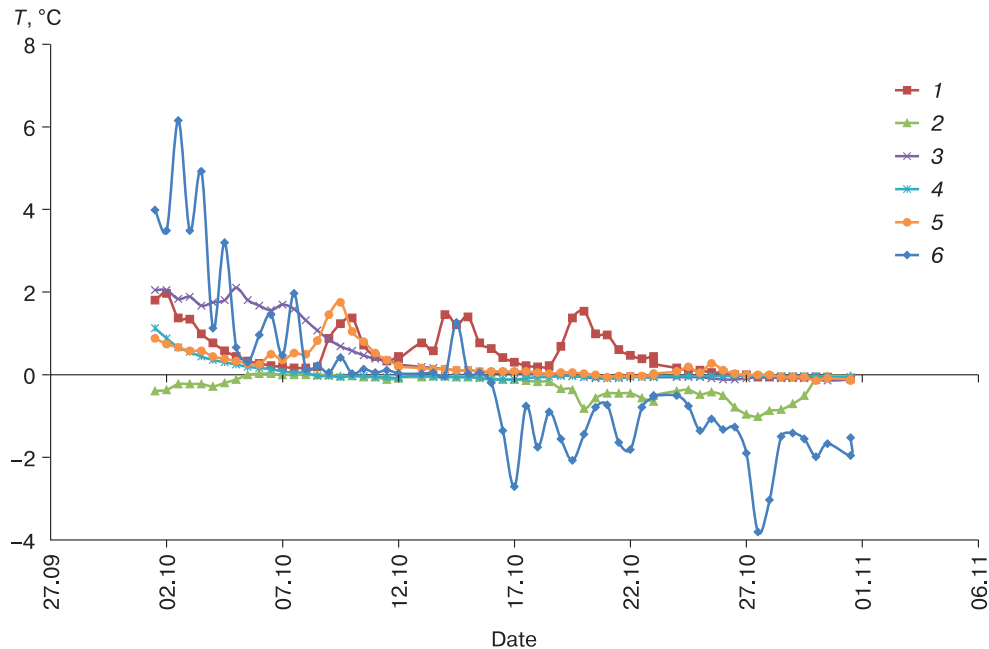


Fig. 2. Ground surface temperature under vegetation cover at 12:00 and 24:00 during the autumn inversion of air temperatures.

1 – 2012, 2 – 2013, 3 – 2014, 4 – 2015, 5 – 2016, 6 – 2019.

the seasonal freezing depth (**SFD**) is greater than the seasonal thawing depth (**STD**) ($SFD > STD$). Additionally, favorable conditions for the growth and/or preservation of frozen ground should persist for several consecutive years. In this way, a stable layer of new permafrost can form. Otherwise, warmer conditions may lead to complete thawing of the newly formed permafrost. However, even in this case, a residual mound will form on the surface, which can become later a site for new permafrost formation under favorable conditions.

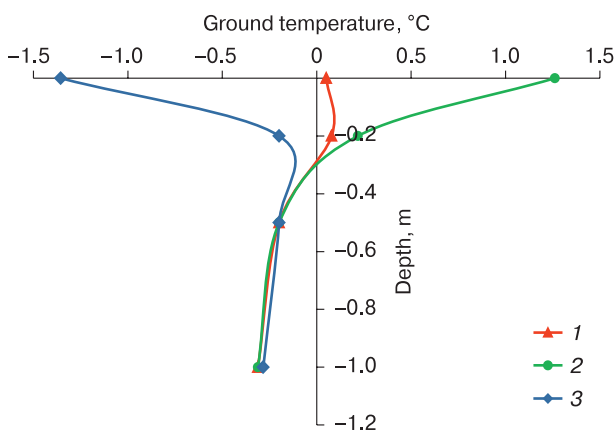


Fig. 3. Fast freezing of seasonally thawed layer at a local site.

1 – Oct. 13, 2019; 2 – Oct. 14, 2019; 3 – Oct. 16, 2019. Measurements at 24:00.

The combination of low air temperatures and spatially variable snow accumulation (including wind-driven redistribution) in early winter creates conditions for faster and deeper freezing in localized areas. Figure 2 shows the results of decade-long monitoring of surface temperatures during the autumn inversion period. Despite the overall warming trend, colder years (e.g., 2019) still provide conditions for deep ground freezing (Fig. 2).

Figure 3 presents a fragment of temperature monitoring data in a peat soil profile, demonstrating how rapidly temperatures can drop within the active soil layer under favorable freezing conditions. Over three days (October 13–16), complete freezing of the seasonally thawed layer occurred in the area with thin snow cover.

Figure 4 illustrates the conceptual model of permafrost aggradation in a hydromorphic thermokarst interpolygonal depression with an initially flat surface. This type of depression is characterized by deep permafrost table (closed talik) and is framed by permafrost (Fig. 4a). The near-surface section consists of thawed peat underlain by thawed loam, with the permafrost table at a depth of 8–9 m. The initially flat and level surface suggests uniform geocryological conditions down to the permafrost table. With the onset of freezing (Fig. 4b), a certain combination of factors is necessary for the formation of stable permafrost: (1) low air temperatures before snow cover formation and (2) uplift of the soil surface (frost heave) as a result of deeper freezing in a local area.

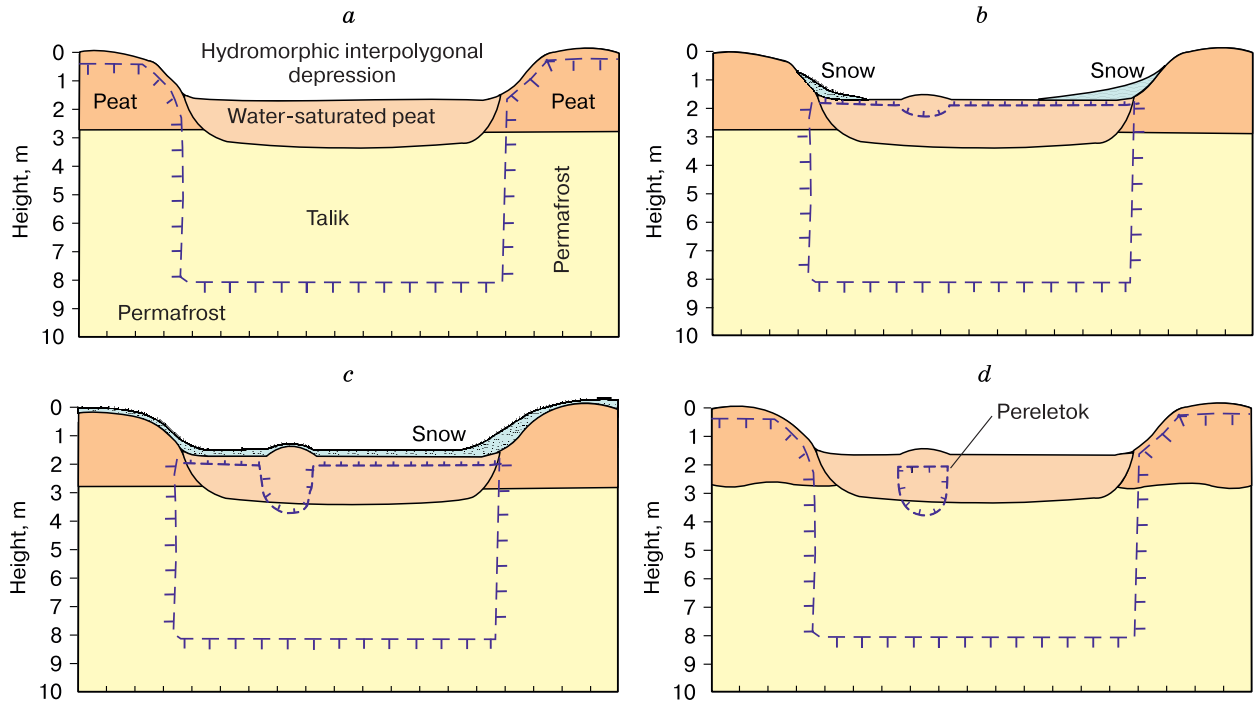


Fig. 4. Principal scheme of permafrost aggradation within the interpolygonal hydromorphic depression (vertical and horizontal scale ratio is 1:1).

a – thermokarst depression, drainage hollow, drained lake basin with a flat surface; *b* – onset of thawing-freezing cycle; the period of a stable transition of air temperature through 0°C and the formation of uneven snow cover; *c* – the end of winter freezing cycle; *d* – the end of annual freeze-thaw cycle; formation of new permafrost (pereletok).

By the end of the winter freeze-thaw cycle (Fig. 4c), deeper freezing occurs in localized zones creating microtopographic irregularities (mounds) on the initially flat surface. The freezing depth at the mounds will exceed that in the adjacent area with a thicker snow cover. Additionally, frost heave will intensify due to water migration toward deeper freezing zones, particularly as a result of freezing of the underlying loam.

The depth of summer thawing (Fig. 4d) remains less than the depth of winter freezing. This cycle ends with the preservation of a frozen soil layer throughout the year, and the mineral-peat mound remains on the initially flat surface. Such mounds favor faster and deeper freezing in subsequent winters.

Figure 5 shows the calculated dependences of frost heave magnitude for loamy substrate underlying peat [Recommendations..., 1986]. The calculations

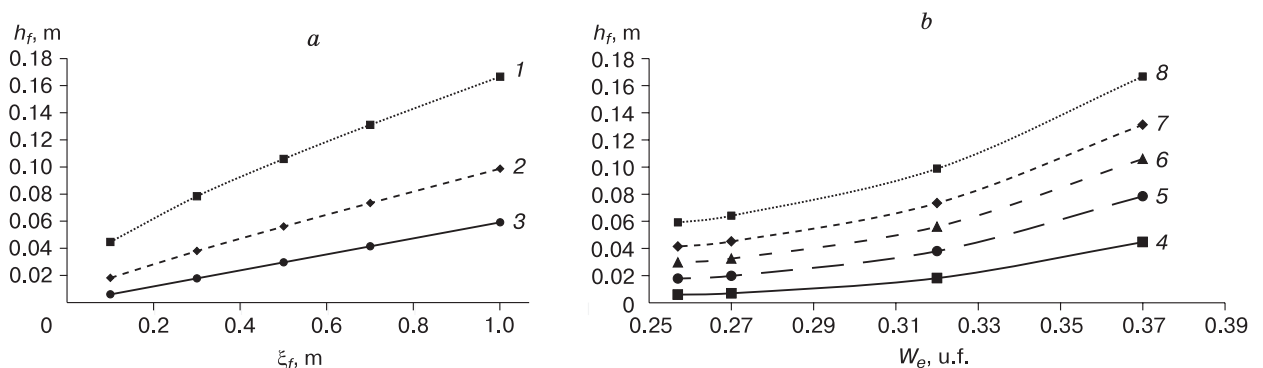


Fig. 5. Calculated dependence of frost heave value (h_f) on (a) freezing depth (ξ_f) and (b) water content of the loam under the peat layer without taking into account peat frost heaving.

1 – $W_e = 0.37$; 2 – $W_e = 0.32$; 3 – $W_e = 0.257$; 4 – $\xi_f = 0.1$ m; 5 – $\xi_f = 0.3$ m; 6 – $\xi_f = 0.5$ m; 7 – $\xi_f = 0.7$ m; 8 – $\xi_f = 1$ m.

were performed for loam with solid phase density of $\rho_s = 2.78 \text{ g/cm}^3$, dry bulk density $\rho_d = 1.8 \text{ g/cm}^3$, and moisture content $W_e = 0.37, 0.32$, and 0.257 . Maximum frost heave occurs at saturation ($W_e = W_{\text{sat}} = 0.37$). The optimal section consists of water-saturated loam overlain by 0.35–0.40 m of peat. Freezing of the loam ensures an increase in frost heave, while the surface peat layer ensures the preservation of the newly formed frozen soil under the active layer localized in the peat.

This situation is typical for hydromorphic depressions, including drainage hollows. If the loam under the peat freezes to a depth of 0.3 m, the frost heave at the surface reaches 0.08–0.10 m (or more, considering migrational ice segregation). Thus, favorable conditions for the initial mound formation are created. Its subsequent development may lead to stable near-surface permafrost aggradation during subsequent years. This explains why aggraded permafrost often occurs along the edges of drained lake basins and interpolygonal depressions, where lithological and surface conditions are optimal. Such permafrost aggradation pro-

cesses occur in the zones of (1) continuous surface permafrost with the residual thaw layer (Fig. 1); (2) continuous permafrost (Fig. 6); and (3) sporadic permafrost (Fig. 7). Photos clearly show the aggradation of permafrost in the localized areas against the background of a generally uniform landscape.

The further development of surface permafrost depends on climatic conditions of subsequent years. A warmer cycle (given the air temperature, snow accumulation regime) may eliminate the newly formed surface permafrost, while colder cycle reinforces it until equilibrium with the modern climate is reached.

An equally significant cryogenic cause of microrelief formation is the uneven surface subsidence resulting from the permafrost thaw. Moreover, in the case of spatially uniform heat exchange between the soil and the enclosing media, the primary driver of differential surface subsidence and, thus, the formation of microtopography is the heterogeneity in the soil properties, particularly, the ice content.

In conditions of uniform lithological properties, uneven subsidence during prolonged thawing can



Fig. 6. Local frost heave mounds within a khasyrey in the northern part of the Bolshezemelskaya Tundra.

Photo by F.M. Rivkin.



Fig. 7. Remnants of ground-mineral seasonal frost mounds within a waterlogged hydromorphic depression (Kevo area, Finland).

Photo by F.M. Rivkin.

Table 1. Thermophysical properties of the ground

Ground	ρ_d , g/cm ³	w	w_w	λ_{th}	λ_f	C_{th}	C_f	Q_{ph}
		u.f.		W/(m·°C)		kJ/(m ³ ·°C)		kJ/m ³
Loam	1.55	0.28	0.09	1.48	1.70	3310.0	2660.0	100 200.0
Water-saturated peat	0.18	4.88	1.22	0.55	1.30	4000.0	2630.0	220 100.0
Slightly moist peat	0.18	<1.5	–	0.17	0.17	–	–	–

Note: ρ_d is the dry bulk density; w is the total water content; w_w is the content of unfrozen water; λ_{th} and λ_f are thermal conductivities of thawed and frozen soil, respectively; C_{th} and C_f are volumetric heat capacities of thawed and frozen soil, respectively; and Q_{ph} is the volumetric heat of phase transition of water.

arise from variations in external thermal exchange conditions related to local features of the surface vegetation cover or its disturbances.

Crucially, in both cases, the onset of permafrost degradation leads to surface elevation differentiation. In turn, the appearance of terrain surface irregularities causes changes in the thermal exchange conditions on microhighs and in microlows as compared with the initial flat surface. These changes occur despite the small amplitude of heights. The most influential factor is the thinning in snow cover on elevated areas and snow accumulation in microlows due to snow redistribution by wind. Moreover, under snow-drift conditions, the snow cover on the microhighs (mounds) may appear with a significant delay compared to surrounding areas; sometimes, only after air temperatures drop below 0°C (Fig. 1a). In waterlogged mires with peat soils, changes in thermal exchange conditions are related to peat drainage on elevated mounds. Drainage sharply reduces thermal conductivity of peat, effectively creating an insulating layer [Rivkin, Buldovich, 2024].

Both factors significantly lower the mean annual soil temperature on developing mounds: reduced snow cover diminishes its warming effect in winter, while a drier peat layer drastically reduces heat flux into the ground in summer. A feedback loop also exists: the reduction in summer heat absorption due to the insulating effect of dried peat on microhighs is accompanied by the weakening of the warming effect of a thinner snow cover in winter. Moreover, rapid development of new moss vegetation on drained and warmer microhighs exerts an additional cooling effect.

Thus, permafrost degradation in areas with spatial heterogeneity in the ice content or initial surface conditions leads to the formation of microtopographic irregularities. Their amplitude increases proportionally to the deepening of perennial thaw. This leads to differentiation of thermal exchange conditions between soil and adjacent media leading to a counterintuitive paragenetic effect – a significant cooling of elevated loci against the background of a generally positive temperature trend.

To study the influence of surface elevation differences on permafrost aggradation, we conducted a se-

ries of calculations. The objective was to quantitatively assess the impact of low-amplitude surface elevations (hummocks, tussocks) on permafrost conditions. We employed a simplified geological section typical for waterlogged areas in northern tundra: a 0.4-m-thick peat layer overlying loam, with all layers being water saturated. Table 1 presents the thermophysical properties of the sediments. We used recent (5-year average) climate characteristics from the Vorkuta region as input climate data. Mean monthly air temperatures are provided in Table 2.

The snow depth at the end of winter was considered to be $h_{sn} = 0.5$ m with snow density $\rho_{sn} = 0.28$ g/cm³. Our calculations used the winter-average snow depth – following a near-parabolic snow accumulation pattern during winter, the mean cold-season value equals 2/3 of the maximum, or $h_{sn} = 0.33$ m. The thermal conductivity of snow calculated using Proskuryakov's formula was $\lambda_{sn} = 0.31$ W/(m·°C), giving the initial average thermal resistance of snow $R_{sn} = 1.06$ (m²·°C)/W.

Changes in the mean annual ground temperature and seasonal freeze-thaw depths within the formed elevations may result from several key factors. First, snow cover depth in areas with significant snowdrift transport is reduced on elevated elements of the microtopography, which decreases the insulating effect of the snow and lowers ground temperatures. Second, drainage of the upper peat layer in elevated areas drastically changes the thermophysical properties of peat with reduced moisture (Table 1), which creates

Table 2. Mean monthly air temperatures recorded at the Vorkuta weather station (averaged through 2018–2023 [Climate..., 2024])

Month	t_i , °C	Month	t_i , °C
Jan.	–18.2	July	14.1
Feb.	–15.0	Aug.	10.9
Mar.	–10.6	Sept.	–18.2
Apr.	–5.4	Oct.	–15.0
May	–0.1	Nov.	–10.6
June	8.6	Dec.	–5.4

Note: t_i – mean monthly air temperature. $\Omega_s = 28\,300^\circ\text{C}\cdot\text{h}$, $\Omega_w = -54\,480^\circ\text{C}\cdot\text{h}$ – sums of summer and winter temperatures, respectively.

an additional insulating surface layer with substantial thermal resistance. Therefore, our model treats the drained peat section specifically as such an insulating layer.

To evaluate the influence of these two factors on the mean annual ground temperature and seasonal thawing–freezing depths (STD–SFD), we solved a series of steady-state one-dimensional problems across a spectrum of surface elevation changes and corresponding landcover property variations [Buldovich, 2016]. This formulation accounts neither for climate change dynamics nor for the growth rate of these elevations. We assume that, prior to surface differentiation, the temperature regime at the ground surface combined with snow accumulation patterns maintained spatially uniform positive mean annual ground temperatures constant over time. The formation of peat mounds and their partial drainage were treated as instantaneous events. Subsequently, depending on the mound height (and corresponding thickness of drained peat layer and snow cover), we calculated equilibrium mean annual ground temperatures and STD/SFD. The temperature value and its sign indicate both the influence of the mound height on the ground temperature regime and the subsequent direction of permafrost development.

Numerical modeling data show that the reorganization of the geocryological system after changes in heat exchange processes at the surface proceeds very rapidly – when new conditions lead to stable negative mean annual temperatures in initially thawed areas, permafrost aggradation typically begins within the first one to two years, rarely after the changes, rarely longer. Furthermore, the decrease in the mean annual temperature at the base of the active layer after changes in heat exchange conditions at the surface reaches equilibrium values with a delay, often significant, depending on the downward propagation rate of the freezing front. Thus, immediately after the mound formation, rapid ground cooling begins within its boundaries, which at sufficient elevation leads to sub-zero temperatures, replacement of seasonal freezing with seasonal thawing, and initiation of permafrost aggradation.

The analytical calculation method was selected because it enables separate quantitative assessment of all active factors influencing ground temperatures. Certainly, the one-dimensional scheme substantially simplifies reality by neglecting lateral heat exchange between mounds and surrounding ground. However, the analysis specifically aimed to evaluate the impact of changes in the heat exchange factor within the mounds themselves. Calculations employed an analytical methodology for determining geocryological characteristics in two-layer active layer systems, developed at the Department of Geocryology, Moscow State University [Buldovich, 2016] with verification

through numerical computer modeling using the WARM program [Khrustalev et al., 1994].

The problem was solved using two scenarios. In the first scenario, snow cover variations were excluded from calculations to evaluate the sole effect of partial drainage of the peat layer. We assumed uniform snow depth and thermal resistance within both the mound and the surrounding terrain. The thickness of the drained peat layer was set equal to the height of the mound and increased proportionally with the mound growth.

In the second scenario, we considered the wind-driven snow redistribution, when the snow surface topography becomes independent of the underlying topography (which is characteristic of real-world conditions). In this case, the snow depth decreases within elevated areas, while the thickness of the drained peat layer correspondingly increases.

Our calculations revealed exceptional sensitivity of geocryological conditions to these two factors demonstrating the influence of the altitudinal differentiation of the surface, even when it is as small as a few centimeters.

Let us first examine scenario I, where the snow cover characteristics remain constant and independent of the surface elevation. The sole modification associated with surface uplift is an increase in the thickness of drained peat (with a corresponding decrease in the thickness of the underlying water-saturated peat layer). The calculation results are presented in Table 3 and Fig. 8. The initial conditions with no elevations were as follows: mean annual ground temperature $t_{\xi} = +0.65^{\circ}\text{C}$, SFD $\xi_f = 0.87$ m. A noticeable lowering of the mean annual ground temperature and an increase in the SFD occur as the elevation increases. Even at the mound's height of only 0.075 m ($h_b = 0.075$), t_{ξ} turns to negative with SFD $\xi_f = 0.7$ m. Further, at the mound's height of 0.20 m, t_{ξ} lowers down to -0.89°C with STD $\xi_{th} = 0.36$ m. Thus, the formation of a mound of 0.2 m in height ($h_b = 0.2$) leads to a decrease in the ground temperature by 1.5°C , which initiates ground freezing and the development of isolated permafrost.

The data in Table 3 demonstrate that the primary cooling factor is the reduction in the warming effect of the snow (Δt_{sn}) from $+5.0^{\circ}\text{C}$ to $+2.4^{\circ}\text{C}$ (with a difference reaching 2.6°C). As the snow's thermal resistance (R_{sn}) remains constant, this reduction clearly results from decreased annual heat flux through the snowpack during winter. This decrease combines with the insulating effect of the drained peat layer, which restricts heat penetration into the ground during summer. As Table 3 shows, the heat flux B nearly halved at the mound height of only 0.20 m.

Notably, this cooling trend is partially offset by a crucial factor in the ground's annual heat exchange:

Table 3. Calculated geocryological characteristics according to two scenarios

No.	h_b , m	R_{dp} , ($m^2 \cdot ^\circ C$)/W	H_{wp} , m	R_{sn} , ($m^2 \cdot ^\circ C$)/W	$H_{CFL/CTL}$, m	t_{ξ} , $^\circ C$	B , MJ/ m^3	Δt_{sn} , $^\circ C$	Δt_{λ} , $^\circ C$
Scenario I									
1	0	0	0.40	1.06	0.87	0.65	149.05	5.01	-1.37
2	0.025	0.15	0.375	1.06	0.81	0.39	135.32	4.55	-1.17
3	0.05	0.295	0.35	1.06	0.77	0.17	124.00	4.17	-1.01
4	0.075	0.44	0.325	1.06	0.71	-0.01	115.11	3.87	-0.86
5	0.10	0.59	0.30	1.06	0.61	-0.21	103.48	3.48	-0.70
6	0.15	0.88	0.25	1.06	0.46	-0.58	85.50	2.87	-0.46
7	0.20	1.18	0.20	1.06	0.36	-0.89	71.58	2.41	-0.29
Scenario II									
1	0	0	0.40	1.06	0.87	0.65	149.05	5.01	-1.37
2	0.025	0.15	0.375	0.977	0.92	0.11	140.90	4.37	-1.27
3	0.05	0.295	0.35	0.893	0.78	-0.34	130.91	3.71	-1.06
4	0.075	0.44	0.325	0.81	0.65	-0.80	118.33	3.04	-0.85
5	0.10	0.59	0.30	0.727	0.53	-1.20	106.91	2.46	-0.68
6	0.15	0.88	0.25	0.56	0.35	-1.84	88.88	1.58	-0.43
7	0.20	1.18	0.20	0.393	0.21	-2.31	74.53	0.93	-0.25

Note: Scenario I – variable thickness of drained peat vs constant snow cover depth; Scenario II – variable thickness of drained peat and variable snow cover depth. Designations: h_b – mound height, R_{dp} – thermal resistance of dry peat, H_{wp} – thickness of water-saturated peat, R_{sn} – thermal resistance of snow, $H_{CFL/CTL}$ – thickness of seasonally frozen / seasonally thawed layer, t_{ξ} – mean annual ground temperature, B – annual heat turnover of the ground, Δt_{sn} – warming effect of snow, and Δt_{λ} – thermal offset.

the thermophysical temperature shift (Δt_{λ}) within the seasonally frozen part of the section.

This negative temperature shift (thermal offset) reaches its maximum absolute values specifically in water-saturated biogenic deposits, the reason why peatlands are considered “cold” grounds. In the considered scenario, the thermal offset initially shows a substantial value of $\Delta t_{\lambda} = -1.37^\circ C$, but begins to decrease rapidly with the mound growth accompanied by an increase in the mean annual ground temperature. At a $h_b = 0.20$ m, it reduces to just $-0.29^\circ C$, meaning it has decreased (in absolute value) by more than four times. Two primary mechanisms drive these changes: (1) the annual temperature amplitude sharply decreases beneath the drained peat layer controlling thermal offset Δt_{λ} and (2) the thickness of the water-saturated peat layer (where thermal conductivity differences between thawed and frozen states are most pronounced) gradually decreases with mound growth, which reduces the absolute Δt_{λ} value. Consequently, the rise in ground temperature from the reduced thermal offset occurs much slower than the reduction in the warming effect of the snow. This imbalance ultimately leads to significant ground cooling and permafrost aggradation.

In the second computational scenario (scenario II), we account for the thermal effects of peat drainage and the corresponding reduction in the snow depth over local topographic elevations (Ta-

ble 3; Fig. 8, curve 2). While the overall pattern of changes remains qualitatively similar to that in scenario I, our calculations reveal significant quantitative differences. Thus, the mean annual ground temperature t_{ξ} crosses into negative values at minimal mound height $h_b = 0.03$ m and further declined to $-2.31^\circ C$ at $h_b = 0.20$ m, while maintaining $+0.65^\circ C$ in adjacent flat terrain. In general, all previously identified (Scenario I) factors retained the same trend but were more contrasting.

Our analysis demonstrates that even minimal surface elevation variations (0.05–0.20 m) can trig-

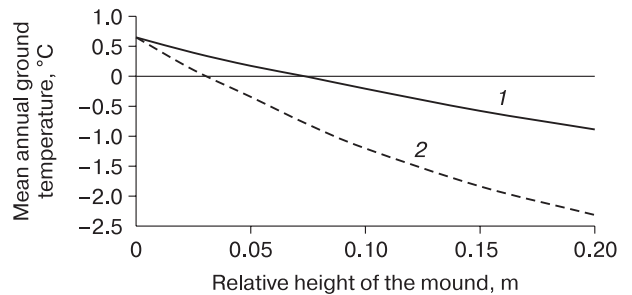


Fig. 8. Change in the mean annual ground temperature (at the base of the active layer) as a function of mound height.

1 – Scenario I with variable thickness of the drained peat layer and constant snow cover; 2 – Scenario II with changes in both dry peat thickness and snow cover depth.

ger dramatic permafrost modifications, leading to the formation of isolated permafrost patches with rather low temperatures. Remarkably, this localized permafrost aggradation occurs despite overarching climate warming trends.

It appears that the most likely mechanism of the new formation of permafrost is associated with local migrational frost heave. Water migration takes place from adjacent areas, with the flow moving through thawed fine-grained mineral sediments underlying peat layers. For directional water movement into the zone of developing uplift, the presence of some negative temperature anomaly is required. This anomaly, as shown above, may result from the appearance of a small (first centimeters) uplift that forms somewhat randomly due to local snow scouring or during an exceptionally cold year, etc.

In this regard, an important result of the calculations performed is that with increasing height of the mound, a unidirectional rapid decrease in the active layer thickness takes place. This means that the migrational ice accumulated during the cold season near the seasonal thawing front will no longer be able to melt completely during the following summer, and the process of ice accumulation and preservation (along with the accompanying mound growth) acquires a stable multi-year character. To this, a migrational frost heave may also be added due to water inflow from surrounding thawed deposits as noted earlier (Fig. 5). Thus, local intensification of frost heave processes may also be a consequence of uneven permafrost degradation.

Another cause of elevation differentiation in originally flat surfaces may be the process of permafrost degradation occurring under modern climate change conditions. The deepening of the deep permafrost table in areas of discontinuous permafrost or the expansion and deepening of closed taliks may be accompanied by uneven surface subsidence. Of course, the amplitude of the resulting irregularities is small, but the calculations performed suggest that it may be sufficient for permafrost aggradation in certain areas. Thus, a paragenesis of permafrost aggradation and degradation processes is realized, as noted above.

While surface factors during the formation of cryogenic microrelief act to amplify differences in thermal regimes and permafrost conditions between areas, lateral thermal interactions serve to smooth the influence of contrasting upper boundary conditions on the spatial configuration of permafrost. In this case, the relative extent of cold and warm zones and their temperature differences play a significant role. The permafrost beneath peat mounds will form isolated patches, whose thickness depends on both the areal dimensions of the elevations and the mean annual ground temperatures in adjacent areas (Fig. 1).

It should be emphasized that, in thermodynamic terms, we consider the final stage of thawing under

newly established conditions – that is, in areas, where the permafrost table has already become deepened through various processes. This complex permafrost configuration nevertheless represents a fully equilibrium (stationary) state achieved precisely through spatial thermal interactions between thawed and frozen zones. Naturally, with continued climate warming, further changes will occur ultimately leading to complete thawing of the newly formed permafrost.

CONCLUSIONS

The paragenesis of permafrost thawing and aggradation against the background of positive climate trend is a natural phenomenon.

Local permafrost aggradation is invariably linked to the development of surface irregularities on an initially flat terrain. The appearance of initial mounds may result from either (1) differential frost heave across the area or (2) uneven surface subsidence caused by permafrost thawing from the top. The first mechanism occurs as a consequence of seasonal frost heave and may transform into perennial freezing under favorable surface and subsurface conditions. The second mechanism is a direct result of long-term spatially variable surface subsidence due to permafrost degradation.

Elevated areas undergo modified heat exchange conditions between the ground and the adjacent media. This leads to sufficient cooling of soils on elevated elements of the microtopography to initiate perennial freezing and permafrost aggradation even under overall climatic warming. Moreover, it can be argued that this warming trend itself and the accompanying permafrost degradation are the driving forces of new permafrost formation.

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SURFACE AND GROUND WATERS
IN TERRESTRIAL PERMAFROST REGION

**THE ROLE OF SUPRAPERMAFROST SUBAERIAL TALIK AQUIFERS
IN SUPPLYING WATER TO SMALL TRIBUTARIES OF THE LENA RIVER ACCORDING
TO FIELD OBSERVATIONS**

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The article provides an assessment of the distribution of suprapermafrost subaerial aquifer taliks and their influence on the processes of runoff formation of two small rivers, Shestakovka and Markhinka, whose catchments are similar in area and are located near the city of Yakutsk in the continuous permafrost zone. The talik area fraction under pine forests in the Shestakovka River basin reaches 23%; in the entire basin, 9.5%. Under pine forests of the Markhinka River basin, taliks are not developed. The decisive role of suprapermafrost subaerial aquiferous taliks in the formation of low-flow river flow in summer and autumn, as well as in the dynamics of autumn freezing of watercourses, is shown. During the summer low-flow period, the Shestakovka River flow is maintained due to the discharge of groundwater from taliks and the deep active layer in pine forests. During the period of autumn–winter freezing, channel and slope aufeis are formed due to the cryogenic expulsion of groundwater from taliks and the deep active layer upon its freezing from the surface. These aufeis serve as an additional source of river water during spring floods. In the absence of taliks in the Markhinka River basin, the flow usually stops at the end of the spring flood; in summer, the river is a chain of hydraulically unconnected expansions; in autumn, it freezes and has no discharge.

Keywords: *permafrost, river flow, groundwater, ground penetrating radar, low-flow period, water chemistry, talik coverage, Central Yakutia.*

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INTRODUCTION

Permafrost has a significant influence on the processes of river runoff formation, which makes these processes in the permafrost zone fundamentally different from those in the regions with seasonally frozen ground [Piguzova, 1973; Sokolov, 1989; Arzhakova, 2001; Woo, 2012; Bring et al., 2016; Bröder et al., 2020]. In the areas of continuous permafrost, where the cryogenic aquiclude lies everywhere and close to the surface, a set of specific hydrological and hydrogeological processes and phenomena are observed: (1) limited interaction of surface and groundwater [Woo, 2012], (2) seasonal and interannual dynamics of the cryogenic aquiclude (permafrost table) [Shepelev, 2011], (3) seasonal and multiyear cryogenic redistribution of river runoff [Alekseev et al., 2011; Lebedeva et al., 2017], (4) aufeis formation and aufeis-derived runoff [Sokolov, 1975; Makarieva et al., 2022], (5) cryogenic floods, (6) in situ condensation of soil water [Bantsekina, Mikhailov, 2009], (7) winter freezing of the rivers to the bottom [Arzhakova,

2001], and (8) recharge of surface waters and groundwater by ground ice melt in its seasonal and multi-year dynamics [Woo, 2012].

Under continuous permafrost, groundwater contribution to small and medium-size rivers is negligible [Arzhakova, 2001]. With the permafrost aquiclude close to the surface and complete freezing of the seasonally thawed layer (**STL**) in winter, small rivers only function during the warm season, i.e., on average, from the end of April to the end of October. Complete freezing of the STL is not characteristic of all river basins in Central Yakutia. In some river basins, there are supra- and interpermafrost lenses of thawed water-saturated sediments – taliks – have been discovered. They do not freeze in winter and, together with the groundwater of the STL, act as sources of river water. The hydrological regime of such rivers differs from that of the rivers without taliks in their catchments. The depth of the permafrost table is not uniform even within a single small catch-

ment area and strongly depends on landscape conditions [Lebedeva *et al.*, 2023a]. For example, on the left bank of the Lena River, in the sandy sediments of the accumulative-denudation plain, in areas mostly covered by monodominant pine forest, suprapermafrost subaerial aquiferous taliks of up to 20 m in thickness are widespread and contain significant reserves of fresh groundwater [Boitsov, 2002; Lebedeva *et al.*, 2019, 2023b]. Taliks are confined to medium- or coarse-grained sands [Lebedeva *et al.*, 2023a]. They are absent under larch forests and peat bogs, where the STL thickness reaches only 0.5–1.0 m.

The aim of this study is to assess the distribution of suprapermafrost subaerial (open) taliks in the catchments of two small tributaries of the Lena River and to identify their role in the formation of the water level and hydrochemical and temperature regimes of the river runoff under conditions of continuous permafrost. This work is based on the results of geophysical surveys and parallel hydrological and hydrochemical observations in the two catchments in 2023.

STUDY OBJECTS

Field studies were conducted in catchments of the Shestakovka (170 km²) and Markhinka (227 km²) rivers – left tributaries of the Lena River located 25–30 km apart from one another in the northeastern part of the Prilensky Plateau (Central Yakutia) (Fig. 1, Table 1).

The climate of the study area is sharply continental, with severe long winters and short hot summers. The mean annual air temperature for the period 1951–2020 at the nearest Yakutsk weather station is –9.2°C, the mean monthly temperature in January is –39.6°C; in July, +19.2°C. The mean annual precipitation is 240 mm, of which 68% falls as rain and 32% as snow.

Geomorphologically, both catchments are located on an accumulative-denudation plain dissected by drainage valleys [Ivanov, 1984]. In both catchments, mixed larch–birch forests grow in ravines, hollows, and on level surfaces, while pine forests occupy better drained slopes and elevated watersheds. Swampy areas occur along streams and around lakes (Table 1). Google Earth satellite images and field descriptions were used to construct landscape maps of the two catchments (Fig. 1).

In the 1970s–1980s and in 2015–2023 researchers from the Melnikov Permafrost Institute conducted observations of the elements of the near-surface water cycle and discovered suprapermafrost subaerial aquiferous taliks in some areas under pine forests [Boitsov, 1985, 2002; Lebedeva *et al.*, 2019, 2023a,b; Shamov *et al.*, 2023]. However, their spatial distribution remained unclear. The coverage of pine forests in the Markhinka River catchment is smaller than that in the Shestakovka River catchment (Table 1), which suggests a smaller proportion of the area occupied by taliks.

The channels of both rivers have a distinct structure consisting of alternating lake-like widenings (beaded-channel pools) and narrows [Tarbeeva *et al.*, 2019]. In the Shestakovka River channel, widenings are developed in the upper and middle reaches; in the Markhinka River channel, they occur throughout the entire channel.

The upper part of the geological section of the Shestakovka River catchment to a depth of 30 m consists of loose sandy sediments with rare interlayers of clay [Boitsov, 2002]. In the bottoms of the valleys and ravines, wet sands (water content 15–25%) are covered by an organic layer; sphagnum peat bogs are often formed on the river terrace at the foot of watershed slopes.

The thickness of continuous permafrost in the area under consideration reaches 400–500 m. The temperature of permafrost at a depth of 100 m varies in the range of –0.6 to –1.0°C [Boitsov, 1985]; and at the depth of zero annual amplitudes, depending on the landscape, it varies from positive (close to 0°C) values to –2...–3°C and lower. The thickness of the STL varies from 4 m on gentle slopes and watersheds under with pine forest to about 0.5 m in swamps and peat sphagnum peat bogs [Varlamov *et al.*, 2023].

STUDY METHODS

At the end of August 2017, in order to detect and map talik zones, geophysical surveys using ground-penetrating radar (**GPR**) were carried out on 15 survey profiles under pine forests in various geomorphic positions within the Shestakovka River catchment; each profile was about 300 m or more in length [Lebedeva, 2018]. As a rule, the profiles were oriented across the slope. Four profiles were laid on the south-facing slopes; five, on the north-facing slopes; four, on the northwest-facing slopes; one profile was laid on the west-facing slope; and one, on the southeast-facing slope. Slope gradients ranged from 1° to 10°. The height of the pine stands varied from 8 to 20 m, and the canopy density ranged from 25 to 95%. The ground cover was represented by lichens and bearberry alternating with areas covered by needle litter. Traces of forest fires were observed on six profiles, and traces of former logging activities were noted on three profiles.

In August 2023, analogous geophysical surveys were carried out on twelve 300-m-long profiles within the Markhinka River catchment. Slope gradients ranged from 1° to 19°. Three profiles were laid on the west-facing slopes; three, on the east-facing slopes; five, on the south-facing slopes; one profile was laid on the north-facing slope. All twelve profiles were under pine forests with the canopy density of 25–65%. The ground cover consisted of bearberry, lingonberry, and lichens alternating with patches covered by needle litter. In general, landscape conditions were simi-

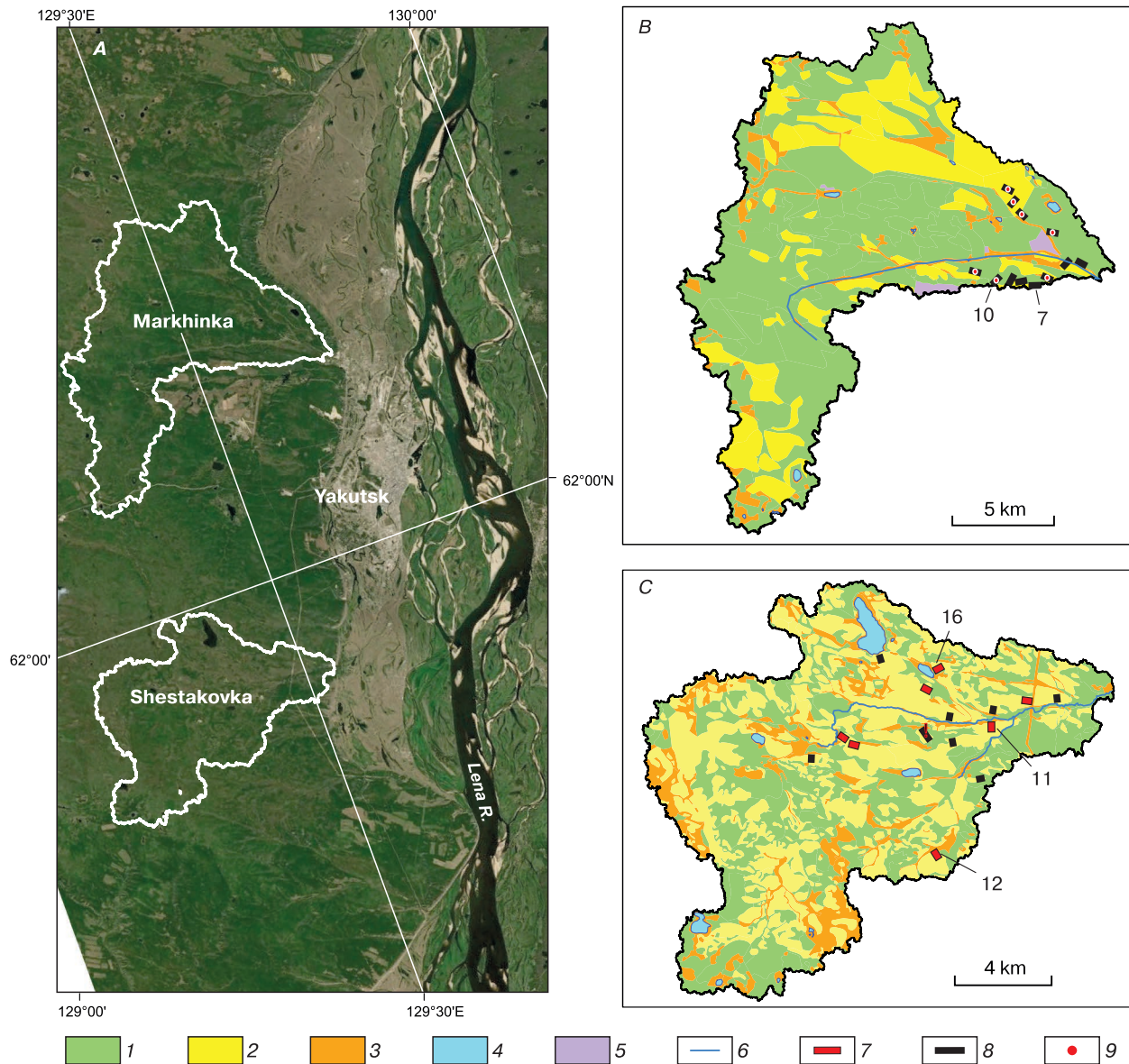


Fig. 1. Location of catchments on the satellite image (A) and schematic landscape maps of the Markhinka River (B) and Shestakovka River (C) catchments.

1 – dwarf shrub and dwarf shrub–moss larch forests; 2 – dwarf shrub and dwarf shrub–moss pine forests; 3 – bogs and swampy areas; 4 – lakes; 5 – anthropogenically altered areas; 6 – rivers; 7 – GPR profiles (11, 12, 16), where taliks were detected; 8 – GPR profiles (7, 10), where taliks were not detected, 9 – control boreholes on the profiles in the Markhinka River catchment.

Table 1. Characteristics of catchments of the Shestakovka and Markhinka rivers

River	Area, km ²	Height, m a.s.l.			Slope, degrees			Dwarf shrub and dwarf shrub–moss larch stands, %	Dwarf shrub and dwarf shrub–moss pine stands, %	Swamps, %
		average	min	max	average	min	max			
Shestakovka, Kamyrdagystakh gauging station	170	233	99	304	7.2	0	66	42	41	15
Markhinka, temporary gauging station	227	203	82	280	7.5	0	50	66	26	7

lar to those in the Shestakovka River catchment, where taliks were found.

In the GPR surveys, an AB-150 antenna unit with the center frequency of 150 MHz was used (OKO-3 GPR series produced by Logis LLC). Garmin GPSmap 64S navigator was used for georeferencing. To measure the distance traveled, a DPI odometer attached to the antenna unit and calibrated in the field before each survey was used. Recording was conducted in continuous profiling mode; trace time (sweep) was 400 ns, and signal accumulation was 16. Each radargram has two scales: the left scale is used to determine the time of arrival of reflected signals to the receiver, and the right scale denotes depth in meters calculated for a dielectric permittivity equal to 6. During data processing and interpretation, this value was corrected according to the drilling data, and the presence of local heterogeneities on the radargram. Almost all radargrams contain interference signals from double reflection from the STL boundary. The bottom of taliks on radargrams is detected in the form of high-amplitude reflected GPR signals – axes of in-phase pulses of the reflected wave. The maximum values of the amplitudes of reflected signals are determined in the lower part of taliks, near the permafrost table, where thawed rocks are fully water-saturated. Despite the attenuation of signals on the radargrams obtained in the Markhinka River catchment, the probing depth was at least 25 m when exponential amplification of signals was applied. For processing and interpretation of GPR data the GeoScan32 software was used.

In September 2023, verification drilling down to the permafrost table was performed at seven out of twelve geophysical profiles in the Markhinka River catchment, using a hand-held rig. At each drilling point, a brief landscape description was made, and a thawed soil sample was taken from the base of the STL to determine the gravimetric water content and particle-size distribution of the sediments.

To compare the hydrological regime of the studied rivers during the warm season of 2023, detailed parallel observations were carried out at the Kamyrdagystakh stream gauging station on the Shestakovka River and a temporary stream gauging station on the Markhinka River. These observations included daily measurements of water levels and discharges during the high-water period and less frequent measurements during the summer–autumn low-flow period; of river, atmospheric, and groundwater to determine the chemical composition; and visual observations of the river conditions. A total of 62 water samples were collected. Water discharges were measured using the area-velocity method with a Potok-1 flow velocity meter. Water levels and temperatures in the rivers were recorded using Hobo U20-L04 digital loggers with a 3-h time resolution. In addition, during discharge measurements, the water level was determined

using stakes preinstalled in the channel. Water conductivity and temperature during discharge measurements were recorded using a HydroMaster-200 thermal conductivity meter manufactured by HM Digital, Republic of Korea.

Camera traps KeepGuard KG762 installed before the snowmelt season near both gauging stations recorded the situation on both rivers every 3 h during the spring flood period and every 6 h during the low-flow period.

In the geochemical laboratory of the Melnikov Permafrost Institute, the content of macrocomponents in the water samples was determined using titrimetry and capillary electrophoresis methods.

Boundaries of the catchments were determined automatically using QGIS software and the Arctic-DEM surface model with a resolution of 10 m.

Daily weather records for 2023 at the Yakutsk weather station located 10–15 km from the stream gauging stations were obtained from [<http://meteo.ru/>]. As the Yakutsk weather station is located in the river valley, and interfluvial positions of the catchments are 100 m higher, observation data from the Yakutsk weather station are only conditionally representative of the catchments.

RESEARCH RESULTS

Structure of the upper section of permafrost in the Shestakovka River catchment. On 7 out of 15 geophysical survey profiles, areas identified as water-bearing taliks can be traced at distances of tens to hundreds of meters and have a thickness varying from 1 to 15 m. The radargrams of corresponding parts of the profiles have a length of 120 m (profiles 12 and 16) and 260 m (profile 11) and are shown in Fig. 2 at a scale of (length : depth) = 1 : 3. The location of the profiles is shown in Fig. 1.

The taliks identified at the seven profiles are generally located on slopes facing watercourse valleys, lake basins, and swamps. Profile 16 is located to the east of Malaya Chabyda Lake. Three shallow (2.0 to 3.5 m) taliks of about 10 m in length can be identified on its radargram (Fig. 2a). Profile 12 is oriented from southeast to northwest and is directed towards the swamp. At the end of the profile, a talik of 40–45 m in length and 10.5 m in thickness is found. Profile 11 is located on the right slope of the Shestakovka River valley and is oriented from north to south. Four taliks with a thickness from 3 to 13 m have been identified on it. The configuration of one of the taliks is rather complex because of the uneven depth of the permafrost table.

At the end of August, the thaw depth was similar across all the presented profiles ranging from 2.4 to 3.7 m (average 2.9 m) on profile 12, from 2.8 to 3.6 m (average 3.1 m) on profile 16, and from 2.3 to 3.1 m (average 2.9 m) on profile 11.

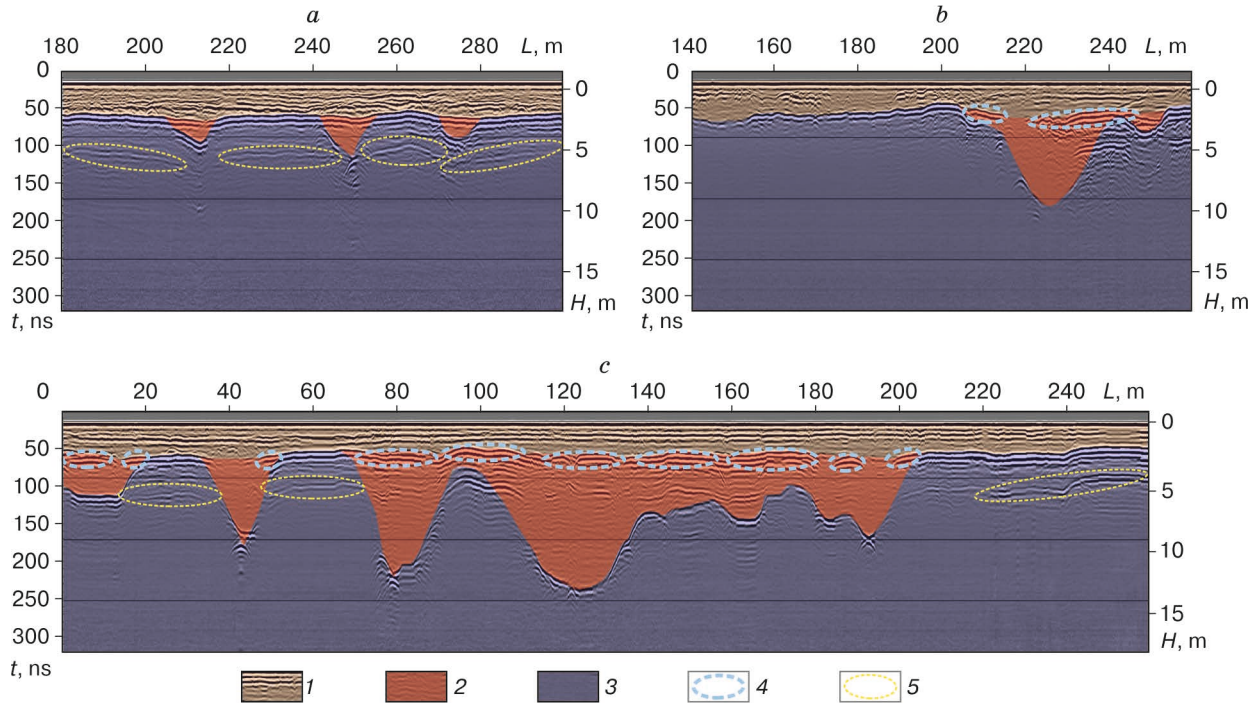


Fig. 2. Sections of interpreted GPR profiles 16 (a), 12 (b) and 11 (c) with the presence of taliks on radar-grams.

1 – seasonally thawed layer on August 29, 2017; 2 – taliks; 3 – perennially frozen sediment; 4 – lenses of seasonally frozen layer; 5 – double reflection from the boundary of seasonally thawed layer; t – time sweep of the radargram, ns; L – profile length, m; H – radargram depth, m at $\epsilon = 6$ (ϵ is dielectric permittivity).

Taking into account the fact that the thawing depth does not reach its maximum at the end of August, frozen interlayers remain in the upper part of the talik and disappear in September.

Previously, it was shown on the example of one key site that the talik is extended downslope towards the Shestakovka River valley [Lebedeva et al., 2019, 2023b]. The revealed configuration of taliks on profiles across the slope indicates that taliks are often oriented downslope and have the shape of nonfreezing ‘channels’ in the thickness of frozen rocks; they act as groundwater filtration pathways to rivers, lakes, and other depressions in the relief.

The total length of the 15 profiles surveyed in the Shestakovka River catchment reached 5953 m, of which the presence of suprapermfrost taliks was recorded in sections with a total length of 1361 m. It was impossible to identify the association of taliks with slopes of a particular steepness or aspect, forests with a particular canopy density, or the dominance of a particular type of the ground cover. As all the 15 profiles in the Shestakovka River catchment were established in relatively homogeneous landscape conditions (dwarf shrub and dwarf shrub–moss pine forests), the talik coverage in these landscape units is estimated at about 23%. For other landscape units,

taliks are not typical because of thin STL, which freezes completely in winter [Varlamov et al., 2023; Lebedeva et al., 2023a]. Taking into account the total area of the above landscapes (Table 1), the talik coverage in the entire Shestakovka River catchment is 9.5%.

Structure of the upper section of permafrost in the Markhinka River catchment. The distribution of pine forests indicates the possible development of suprapermfrost subaerial taliks similar to those in the Shestakovka River catchment. However, geophysical surveys in the Markhinka River catchment did not reveal taliks on any of the 12 profiles down to a depth of 25 m. Radargrams of two profiles – 7 (284 m) and 10 (290 m) – are shown in Fig. 3; their location is shown in Fig. 1. According to the geophysical data, the STL thickness varies between 2.3–3.1 m (average 2.9 m) along the entire length of profile 7 and between 2.2–3.2 m (average 3.0 m) along profile 10.

Verification drilling along the geophysical sections revealed that the permafrost table under pine forests in the Markhinka River catchment lies at depths from 1.6 to 2.6 m, which is close to GPR profiling data (from 1.5 to 3.2 m). This confirms the efficiency of the GPR method to determine the boundary between thawed and frozen layers in the near-surface horizons composed of sands and loamy sands.

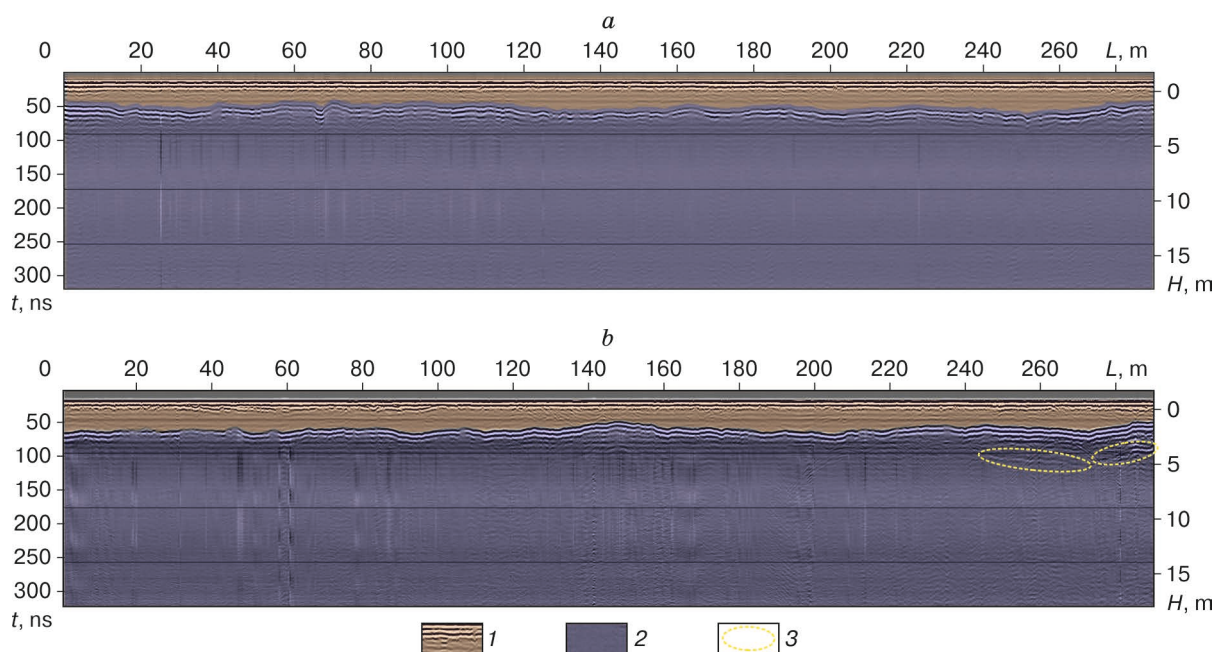


Fig. 3. Sections of interpreted GPR profiles 7 (a) and 10 (b).

1 – seasonally thawed layer on August 25, 2023; 2 – permafrost; 3 – double reflection from the boundary of the seasonally thawed layer.

The drilled sections are mainly composed of fine sand with interlayers of sandy loam, loam, and medium-grained sand. In two boreholes, at depths of 1.8 and 0.9 m, a water-saturated sand layer of up to 0.1 m in thickness has been found. This layer overlies interlayers of thawed loam above the frozen aquiclude. The weight moisture content in the soils at the foot of the STL varies from 9 to 24%. Six out of seven soil samples taken at the bottom of the STL during drilling represent fine-grained sands with a predominance of particles 0.10–0.25 mm (46–71%).

According to the results of geophysical surveys, the depths of seasonal thawing in catchments of the Shestakovka and Markhinka rivers are close under similar landscape conditions, but taliks were only identified in the Shestakovka River catchment. The geophysical and drilling works in the catchments allow us to conclude that there are virtually no suprapерmafrost subaerial taliks under pine forests in the Markhinka River catchment. The attenuation of GPR signals with depth in this catchment is more pronounced than that in the Shestakovka River catchment, which may be related to the presence of sandy loamy and loamy interlayers in sandy sediments.

Water and temperature regime of the Markhinka and Shestakovka rivers in 2023. On the Shestakovka River, the runoff started on May 2, two days earlier, and the spring flood level reached its peak on May 8, a day earlier than that on the Markhinka River (Fig. 4, Figs. 5a and 5b).

After the beginning of runoff, the water level in the Shestakovka River fluctuated insignificantly (within 30 cm) until May 5. On May 5, against the background of significant and rapid warming from 12:00 to 20:00, it rose by 60 cm. On the Markhinka River, the corresponding rise was only 14 cm.

In accordance with the dynamics of water levels, the discharge of the Shestakovka River increased from 0.875 m³/s (May 5) to 5.17 m³/s (May 6), and that of the Markhinka River – from 0.056 m³/s to 0.33 m³/s.

Subsequently, the water level and discharge of the Shestakovka River demonstrated a distinct diurnal variation driven by fluctuations in air temperature, which determined the intensity of snowmelt in the catchment. The most intense increase in the discharge occurred in the afternoon; during nighttime cold spells, especially frosts, the inflow of meltwater from the catchment decreased, the river water level stabilized and then slightly dropped until the beginning of air warming in the morning. On the Markhinka River, diurnal fluctuations in the runoff were less pronounced and smoothed.

Already in the first days of the runoff, ice near the stream gauging station at the Markhinka River completely melted, and river water began to warm up in the daytime, while the water temperature of the Shestakovka River fluctuated around zero degrees Celsius until May 13, and ice remained in the channel. The inflection points of both hydrographs during the spring flood period (Fig. 4) correspond to the re-

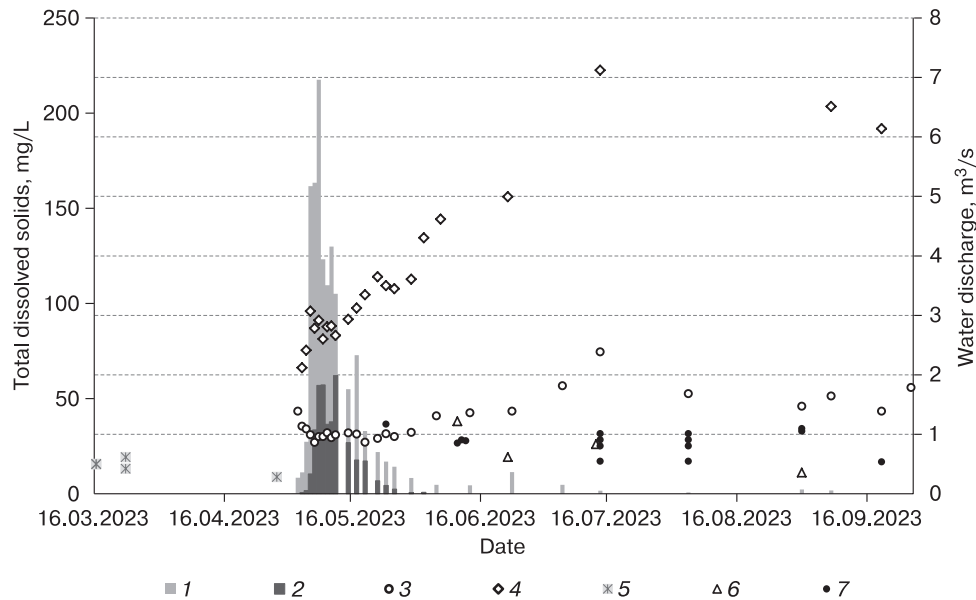


Fig. 4. Dynamics of water discharge measured on the (1) Shestakovka and (2) Markhinka rivers and total dissolved solids in water of the (3) Shestakovka and (4) Markhinka rivers, (5) snow, (6) rainwater, and (7) groundwater sampled in the catchments of these rivers during the warm season of 2023.

turn of cold weather and the decrease in snowmelt runoff intensity in the river valleys.

The discharge of the Shestakovka River exceeded that of the Markhinka River by several times (by an order of magnitude or even more during the spring flood decline) (Fig. 4).

The Shestakovka River is characterized by a wider range of fluctuations in the water level and discharge, as well as by a more diffuse relationship between them during the rising and falling floods (hysteresis): at equal water levels, the discharge during the falling flood is up to three times higher than that during the rising flood (Fig. 6). The ambiguity in the relationship between water discharge and water level in the Markhinka River is less pronounced, allowing the curve to be approximated by a single nonlinear function. Deviations of the empirical points from the approximating curves correspond to periods of sharp changes in air temperature and corresponding changes in the river's supply of meltwater.

The discharge of the Markhinka River ceased on June 6. In the area of the stream gauging station, the river represented a series of water-filled pools, with no flow between them. The water level in the pools gradually dropped, and the tracks of dry land between them became overgrown. Several rain-induced rises in the water level failed to restore the flow. The Shestakovka River's flow continued throughout the warm season of 2023.

During the period of the main spring flood from May 7 to May 26, the daily runoff depth from the Shestakovka River catchment was 2.3–7.6 times

greater than that from the Markhinka River catchment. During the spring flood's peak, this ratio reached 21. The total runoff depth from the Shestakovka River catchment from May 2 to June 5 was 27.7 mm, while that from the Markhinka River catchment was 6.6 mm (4.2 times less).

Water temperature of the Markhinka River was higher than that of the Shestakovka River throughout the low-flow period by 4–12°C. The difference in water temperatures between the two rivers also had a diurnal variation with maximum values at 16–17 hours.

Ice regime of the Markhinka and Shestakovka Rivers in 2023. The onset of ice formation on both rivers was observed on September 21 amid overnight subzero air temperatures (Figs. 5c and 5d). In the absence of runoff, ice growth on the Markhinka River was controlled only by air temperature fluctuations. Stable ice cover formed on October 14. The water level remained virtually unchanged during the ice formation period. No aufeis formation was recorded (Fig. 5e).

Ice formation on the Shestakovka River was complicated by the presence of runoff: the areas of flowing water froze later than stagnant areas, and ice drift and destruction of young shore ice were frequently observed during sharp rises in the water level associated with a narrowing of the flow cross-section. Therefore, this river froze later than the Markhinka River despite the same temperatures. Thus, a warming at the end of the first ten-day period of October caused a massive ice drift that completely destroyed

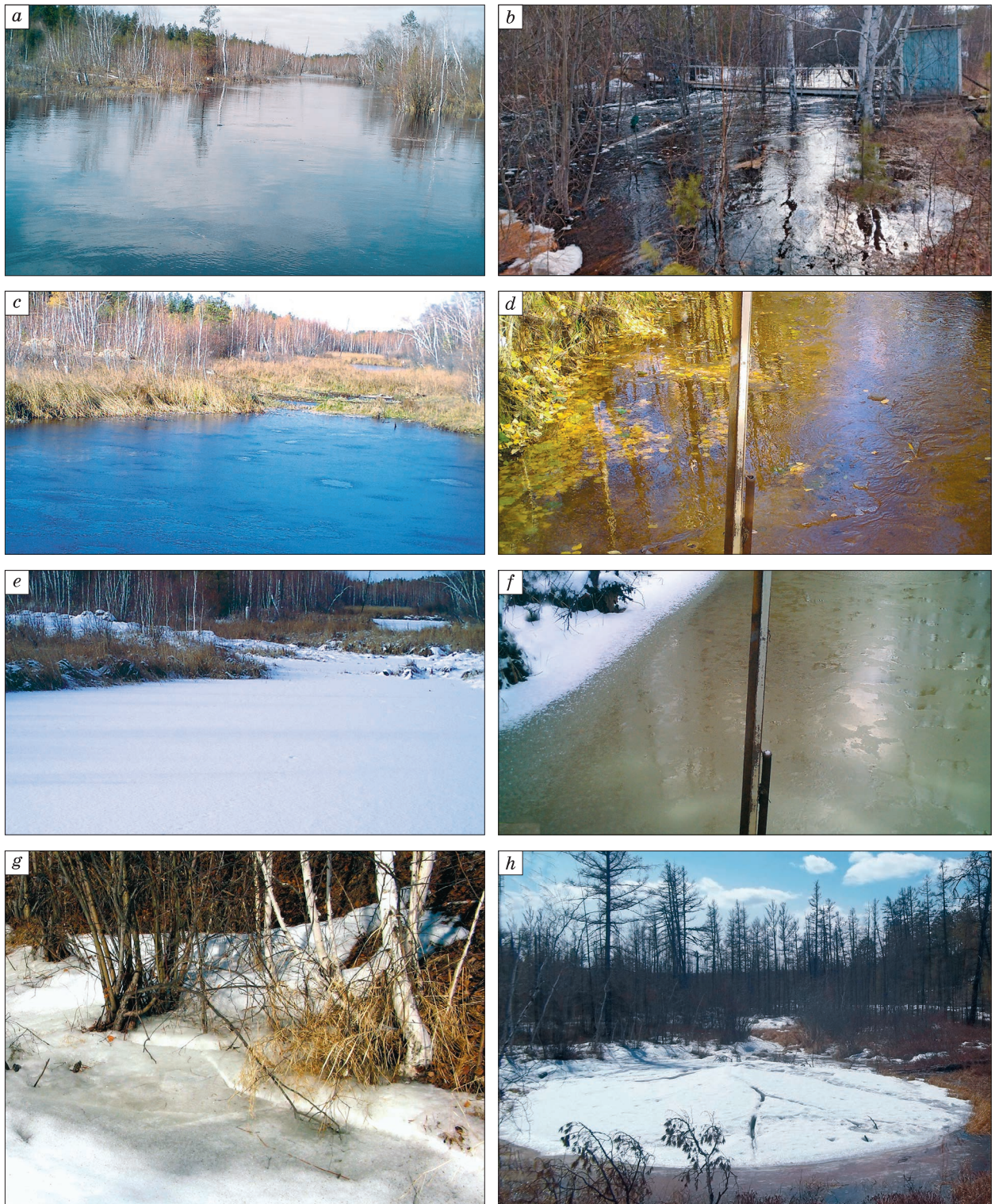


Fig. 5. Hydrological phenomena on the studied rivers in different seasons of 2023.

a – spring flood on the Markhinka River, May 8; *b* – spring flood on the Shestakovka River, May 8; *c* – first ice formation and the absence of runoff of the Markhinka River, September 21; *d* – first ice formation and runoff on the Shestakovka River, September 21; *e* – established ice cover on the Markhinka River, October 25; *f* – water outflow in the channel of the Shestakovka River, October 25; *g* – ice subsidence at the water edge of the Markhinka River, April; *h* – cryogenic ice heave on the Shestakovka River, April.

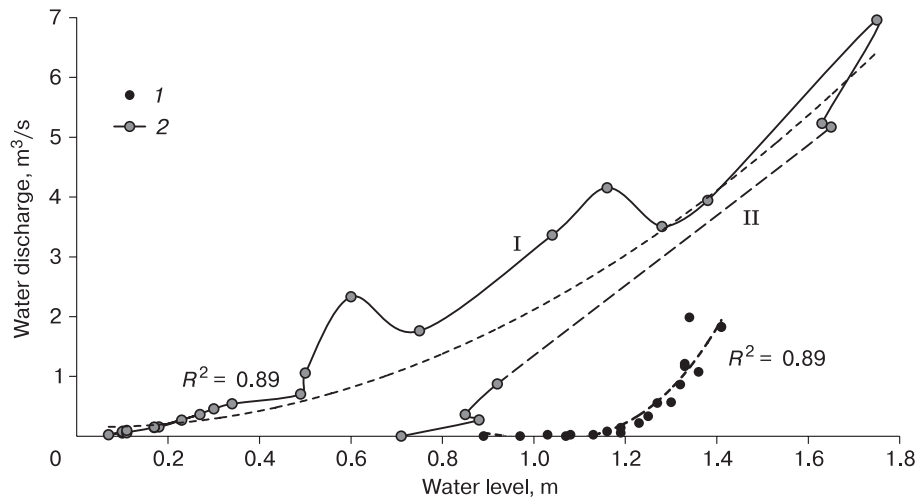


Fig. 6. Relationship between water level and discharge during the spring flood period on the Markhinka (1) and Shestakovka (2) rivers, May 2023.

For the Shestakovka River: I – spring flood decline phase, II – spring flood rise phase.

the previously established thin ice cover and caused an ice jam with a significant rise in water level. A few days later, with a steady decrease in air temperature, ice began to form again in the area of the gauging station. At the same time, the water level rose again due to a narrowing of the flow cross-section.

The thaw on October 22–25 led to the formation of a polynya (Fig. 5f). On October 30, as the air temperature dropped, channel aufeis began to form. November frosts caused regular water expulsion onto the ice surface and the aufeis growth.

Chemical composition and dynamics of total dissolved solids (TDS) in the Markhinka and Shestakovka rivers in 2023. In terms of chemical composition, the water in the Shestakovka and Markhinka rivers is ultra-fresh, of bicarbonate magnesium-calcium and calcium-magnesium compositions, less often mixed in cations with a predominance of calcium (35–43%-eq., Fig. 7). The water reaction is slightly acidic at the beginning of the spring flood (pH 5.4–6.3) and neutral at the end (pH 6.7–7.0). Five snow samples collected in March–April 2023 belong to ultra-fresh bicarbonate calcium waters with a TDS of 8–20 mg/L and neutral reaction (Fig. 4). The TDS of rainwater in June–July is slightly higher, up to 38 mg/L. Talik groundwater in the Shestakovka River basin in summer 2023 is also ultra-fresh, with a TDS of 17–32 mg/L (Fig. 4) and pH 6.5–6.9.

The seasonal dynamics of water TDS in the two rivers differs (Fig. 4). During the spring flood period, it changes in the opposite directions. In the Shestakovka River, water reached its maximum (43.4 mg/L) at the beginning of the runoff and then decreased to 27 mg/L on May 7, on the eve of the spring flood peak. Intensive melting of snow and aufeis formed on

the slopes of the river valley in the winter contributed to a decrease in TDS (27–32 mg/L) of the river water until the end of May. Desalination of the river water was mainly due to a decrease in the concentration of sulfate and chloride anions and calcium and sodium cations (Fig. 7). Only in June, against the background of an increase in evaporation, a significant decrease in water discharge and the transition of the river to a low-flow regime, water TDS gradually increased up to 75 mg/L in mid-July. This increase was associated with an increase in the concentration of all macrocomponents, except for potassium, the concentration of which remained almost constant throughout the summer (Fig. 7).

The minimum value of water TDS in the Markhinka River (66 mg/L, Fig. 4) was observed at the beginning of the spring flood flow on May 4, which was due to the inflow of ultrafresh meltwater from the river catchment into the channel. As the spring flood level rose, stagnant water from pools enriched in dissolved substances was involved in the flow, and water TDS of the Markhinka River increased to 96 mg/L on May 6 and stabilized at values of the level of 81–91 mg/L during the passage of maximum spring flood discharge in May 7–15. During the spring flood decline after May 15 and up to the complete cessation of flow in early June, water TDS increased from 91 to 110 mg/L. In June and July, the concentration of dissolved solids in the water of pools increased to 222 mg/L (Fig. 4). The increase in water TDS was mainly due to bicarbonate, calcium, and magnesium ions (Fig. 7). The concentration of sulfate and chloride ions changed insignificantly. Changes in the concentration of sodium and potassium ions in the waters of both rivers are noteworthy. From the begin-

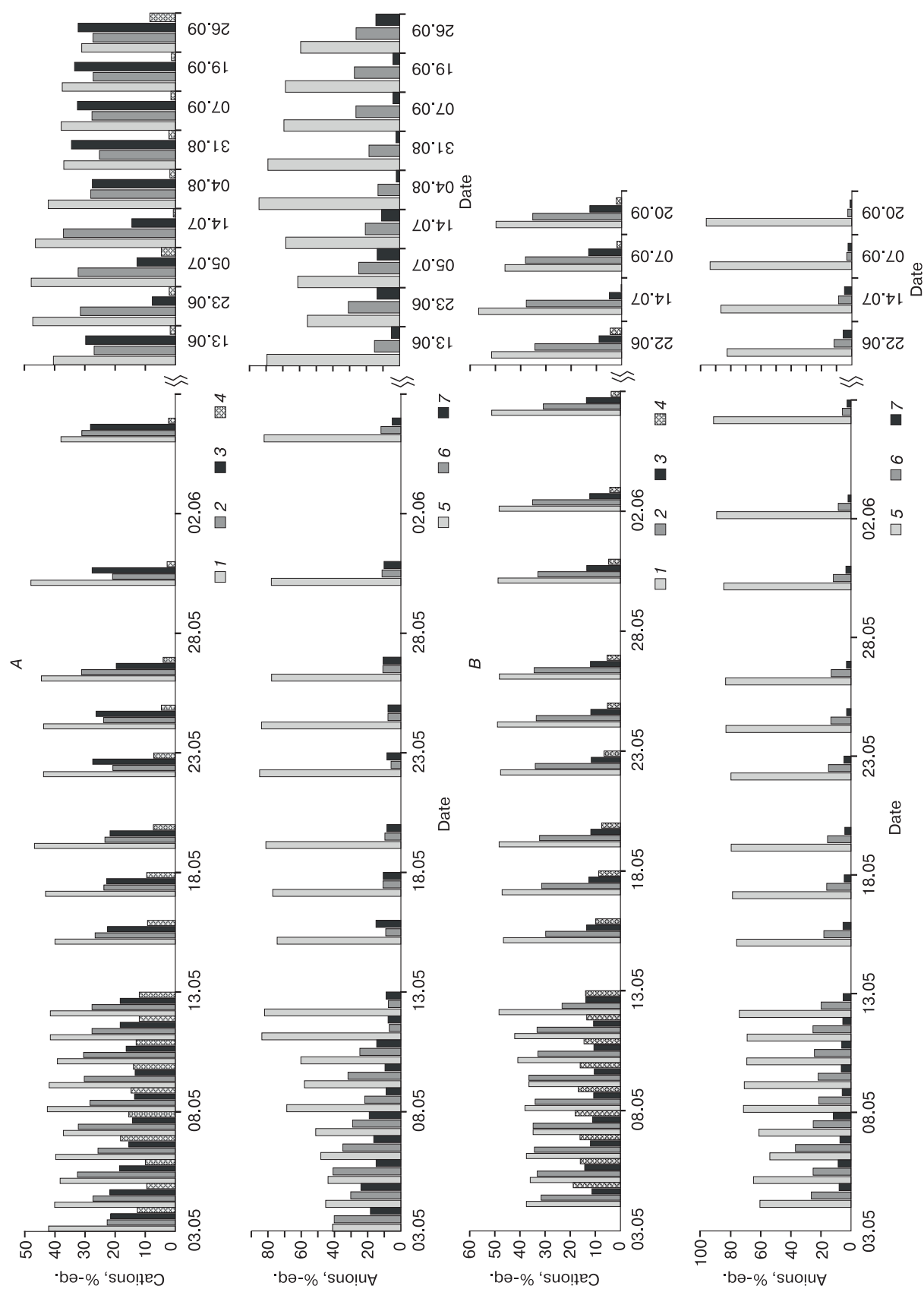


Fig. 7. Changes in cation and anion composition of water in the Shestakovka (A) and Markhinka (B) rivers.
1 – Ca^{2+} , 2 – Mg^{2+} , 3 – Na^+ , 4 – K^+ , 5 – HCO_3^- , 6 – SO_4^{2-} , 7 – Cl^- .

ning of the spring flood until May 12, the shares of these ions were comparable (12–20%-eq. each). Then, until mid-June, against the background of increased concentration of calcium and magnesium ions, the concentration of potassium ion in the water of the Shestakovka and Markhinka rivers decreased to 2 and 4%-eq., respectively (Fig. 7).

In winter, water was only preserved in distinct widenings of the river. By the beginning of the spring flood, the TDS of water under the ice cover in pools varied within 84–220 mg/L in the Shestakovka River and reached 426 mg/L in the Markhinka River.

DISCUSSION

Differences in the structure of the upper section of permafrost in the catchments of the Shestakovka and Markhinka rivers. The presence of sub-aerial aquiferous closed taliks in the areas of pine forests in the Shestakovka River catchment has been confirmed by drilling of 20-m-deep boreholes at two key sites in the 1980s and in the recent 8 years [Boitsov, 1985; Boitsov, Lebedeva, 1989; Lebedeva et al., 2019]. The taliks are confined to medium- to coarse-grained sandy sediment; less often, to fine-grained sands; interlayers of loam or clay were not detected in the boreholes drilled in the areas of talik development [Lebedeva et al., 2023a]. The widespread occurrence of closed taliks in the Shestakovka River basin is confirmed by the presence of numerous slope and channel aufeis. The two largest aufeis studied have a length of more than 80 m. Their occurrence and growth are associated with the discharge of groundwater from taliks and deep STL, which intensifies in winter due to the development of cryogenic pressure.

Despite similar landscape conditions, taliks have not been detected along the profiles studied in the Markhinka River catchment. We hypothesize that this is due to the somewhat heavier texture of the sediments, as indicated by borehole data from seven profiles. However, more data on grain-size distribution are needed to confirm this hypothesis.

According to the composition of dissolved ions, the Markhinka River water is bicarbonate water with magnesium-calcium or calcium-magnesium cations, whereas the Shestakovka River water has a more complex cationic composition containing up to 34%-eq. of sodium (Fig. 7). According to the specificity of the cationic composition and TDS of the river water, it can be assumed that its chemical composition in the Shestakovka catchment is formed due to water interaction with well-permeable aluminosilicate rocks, while sandy sediments in STL of the Markhinka River catchment also contain calcium carbonates precipitating from suprapermafrost groundwater during freezing and partly dissolving under the impact of infiltrating atmospheric precipitation.

Changes in the concentration of potassium ions in waters of the Shestakovka and Markhinka rivers in summer correlate with the observed differences in the texture of the sediments. Potassium is a highly biophilic element and belongs to the active water migrants. The decrease in the potassium concentration in the water of the Shestakovka River with sandy sediments in the aeration zone is more pronounced than that in the Markhinka River, where the upper layers of Quaternary sediments in the catchment contain more silt particles.

The area of taliks in the Shestakovka River basin reaches 9.5%. This is a high value for Central Yakutia, and it exceeds previously reported value of 4% [Anisimova, 2003]. This suggests the existence of significant reserves of shallow talik groundwater in some areas of pine forests and their participation in the formation of the discharge of small rivers in the area of continuous permafrost, though the signs of landscape indication of aquiferous closed taliks within pine forests have not yet been identified.

The relationship between the features of spring flood development and the structure of the channel and the upper part of the permafrost. Channel widenings and beaded-channel pools are developed only in the upper and middle reaches of the Shestakovka River [Tarbeeva et al., 2019] and throughout the entire channel of the Markhinka River. On average, pools on the Markhinka River are larger (in the area of the gauging station, their width is 12–20 m, length 35–50 m, and depth 3.0–3.5 m) than on the Shestakovka River (7–12, 12–18, and 1.5–3.5 m, respectively) [Tarbeeva et al., 2019]. On the Markhinka River, traces of the ice surface subsidence during river freezing (Fig. 5g) were found; in contrast, the pools of the Shestakovka River often have convex ice surface (Fig. 5h). The concave shape of the ice in the Markhinka River pools indicates insignificant inflow of suprapermafrost groundwater into the channel or its absence after the onset of ice formation. At the same time, pools of the Shestakovka River continue to be fed by water from the freezing STL and taliks, and during further winter freezing, cryogenic pressure develops in taliks and in unfrozen lenses of water in pools of the channel, which leads to some heaving of the ice cover in pools of the Shestakovka River.

The earlier onset dates and higher runoff layer depth in the Shestakovka River may be related to lower water retention capacities of the channel and the entire catchment. The smaller size and number of pools in the Shestakovka River channel and the presence of ice mounds in them (Fig. 5h) reduce the overall capacity of the channel network, while the concave shape of the ice cover in the large and numerous pools of the Markhinka River (Fig. 5g) implies a larger storage of water in the channel until overflow and runoff onset, and also contributes to the reten-

tion of part of the runoff. In addition, groundwater from the taliks discharges onto the surface and forms streams in spring; together with melting channel and slope aufeis, which lasts 2–3 weeks longer than snowmelt, it contributes to a higher discharge of the Shestakovka River in comparison with the Markhinka River.

Hysteresis of the relationship between discharge and water level of the Shestakovka River during the spring flood (Fig. 6) is caused by a general decrease in the ice surface and water surface elevation due to melting of more than 0.5-m-thick ice layer on the river bottom. In the Markhinka River channel, the ice near the gauging station is less thick and melts much earlier, which explains a simpler relationship between river discharge and water level. Aufeis in the Shestakovka River channel is formed in November–December due to supraperafrost groundwater from taliks and deep STL, the discharge of which is intensified during seasonal freezing of the ground. In addition to direct contribution to the river runoff, channel aufeis maintains the temperature of the river water close to 0°C until their complete destruction, and also causes a complex relationship between the water level and discharge of the Shestakovka River. On the Markhinka River, channel aufeis does not form, water begins to warm up earlier, and the relationship between water level and discharge is closer.

The increase in TDS of water of the Markhinka River during the spring flood in 2023, on the third day of runoff, may be due to melting of pools, the water of which has the maximum concentration of dissolved substances before the spring flood. The accumulation of chemical elements in the water of pools is due to evaporation during the previous summer period; it continues in winter, when the water freezes under conditions of impeded water exchange. The cryogenic metamorphism of the water of pools is evidenced by an increase in the concentration of bicarbonates of magnesium and sodium. The opening and “inclusion” of pools with their mineralized and oxygen-deficient waters in the already developing process of spring runoff formation was also observed earlier on the Shestakovka River [Shamov *et al.*, 2023]. However, the volume and number of pools in the Shestakovka River are smaller in comparison with the Markhinka River, so they have no significant influence on TDS of river water.

The relationship between the features of the hydrological regime of rivers during the summer low-flow period and during the period of autumn freezing with the spread of taliks. The hypothesis about the predominant role of talik groundwater and deep STL is confirmed by the cessation of flow of the Markhinka River in the summer low-flow period; there are no taliks in the pine forests of the Markhinka River, while the area of taliks in the catchment of the Shestakovka River reaches 9.5%, and the

river discharge is continuous until complete freezing in November–December.

During the summer–autumn low-flow period, the Shestakovka River maintains its flow due to groundwater supply from taliks and from deep STL of sand massifs in pine forests, and partly due to the inflow of water from melting aufeis. Some of them are found in summer under the moss layer in the area of waterlogged footslopes. During the whole summer–autumn low-flow period, the water temperature of the Shestakovka River is significantly lower than that of the Markhinka River, which can also be explained by the inflow of cold water from taliks and STL into the channel of the Shestakovka River, as well as by the favorable warming conditions in the Markhinka channel pools.

The differences between the two rivers during the freezing period are also related to the presence of groundwater supply to the Shestakovka River from talik aquifers. The Markhinka River freezes in the absence of runoff and limited groundwater supply, while the Shestakovka River flow continues even at subzero temperatures and is accompanied by the inflow of significant volumes of water into the channel, water level difference due to cryogenic narrowing of the flow cross-section, the formation of channel aufeis, and, later, as the river freezes and a seasonally frozen layer is formed above the talik zones, cryogenic pressure develops, and numerous and extensive slope aufeis are formed.

CONCLUSIONS

The study examines the role of supraperafrost subaerial aquifers in the formation of the runoff of two small tributaries of the Lena River in its middle reaches – the Shestakovka River and the Markhinka River.

In the Shestakovka River catchment, in sandy sediments covered by pine forests, supraperafrost subaerial aquiferous taliks of up to 15 m in thickness and hundreds of meters in length are formed, which is confirmed by ground-penetrating radar profiling and drilling. The area of taliks is estimated at 23% under pine forests and 9.5% of the entire catchment. No taliks were found in the Markhinka River catchment under similar landscape conditions, which we hypothesize is due to a heavier texture of sediments in the upper part of the section.

The runoff of the Shestakovka River continued throughout the warm period, in contrast to the Markhinka River, where runoff stopped on June 6.

The prolonged freezing of the Shestakovka River until mid-November was complicated by the presence of runoff, which led to regular ice drifting and the formation of channel aufeis due to expulsion of groundwater from taliks. The Markhinka River froze in mid-October in the absence of runoff, and no aufeis formed in its valley.

The temperature and TDS of water in the Shestakovka River were lower than those in the Markhinka River due to the influx of cold and ultrafresh water from melting aufeis and from suprapermafrost groundwater from taliks. Water of the Markhinka River had higher TDS and temperature, especially in the summer low-flow season, due to intensive evaporation and lack of groundwater supply.

The hydrological differences between the two rivers are explained both by the talik area fraction and the associated aufeis and by the structure of the channel. A large number of channel pools along the entire course of the Markhinka River retain part of the runoff, while slope and channel aufeis in the Shestakovka valley provide an additional source of water for the river.

The study highlights the significant role of suprapermafrost aquiferous taliks in shaping the hydrological regime of small rivers under conditions of continuous permafrost.

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GASES AND GAS HYDRATES IN THE EARTH'S CRYOSPHERE

CARBON DIOXIDE RELEASE FROM MELTED GROUND ICE SAMPLES
UNDER CONDITIONS OF AN INCUBATION EXPERIMENTP.B. Semenov^{1,*}, A.O. Kil¹, E.V. Shatrova^{1,2}, N.G. Belova^{3,4}, O.V. Levochkina⁴, M.O. Leibman⁴,
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This work aims to study the quantitative transformation of dissolved organic matter into CO₂ in ground ice samples collected from typical exposures of the Russian Arctic under conditions of an incubation experiment. The CO₂ emission during incubation has two peaks, the first being associated with the utilization of organic carbon dissolved in ground ice; and the second, with the mobilization of organic matter from the solid fraction of melted ground ice. Melted ice samples are characterized by high values of biodegradable dissolved organic carbon (>65%) and a significant positive correlation ($R^2 = 0.82$) between the values of cumulative CO₂ emission and biodegradable dissolved organic carbon. Our data indicate a high potential of ground ice as a source of labile carbon in the ecosystem under thermodenudation.

Keywords: ground ice, dissolved organic matter, carbon dioxide, incubation experiments, biodegradable dissolved organic carbon.

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INTRODUCTION

The state of permafrost in a changing climate is of great scientific and practical interest. According to some researchers, the amount of carbon stored in the permafrost may exceed its content in the modern atmosphere by two times or more [Schuur *et al.*, 2008]. The observed and predicted global warming contributes to the permafrost thawing and, as a consequence, to the release of carbon from buried organic matter (OM) into the modern carbon cycle [McCalley *et al.*, 2014; Dean *et al.*, 2018]. The involvement of relict OM during the degradation of permafrost into the modern carbon cycling leads to the emission of additional amounts of carbon-containing gases – methane (CH₄) and carbon dioxide (CO₂), which can be a trigger for the so-called positive feedback contributing to the acceleration of climate warming [Schuur *et al.*, 2015; Heinz, Zak, 2018; McGuire *et al.*, 2018; Natali *et al.*, 2019].

Massive ground ice, including wedge ice and ice beds, are important components of the permafrost areas in the Russian Arctic in terms of the carbon cycle.

Ice beds and host sediments in the Yamal Peninsula are marked by relatively high methane concentrations that correlate with the organic carbon content in the sediments [Streletskaia *et al.*, 2018; Semenov *et al.*, 2020, 2024; Zadorozhnaya *et al.*, 2022]. Massive ground ice serves as a source of labile dissolved organic matter (DOM) in modern permafrost ecosystems. Surface watercourses formed as a result of ground ice melting transport DOM across landscapes to the base level of erosion. This is accompanied by biogenic mineralization of DOM to CO₂ under aerobic conditions and to CH₄ under anaerobic conditions [Vonk *et al.*, 2013]. All of the above argues for close attention to ground ice as a source of organic matter and greenhouse gases (GHG).

To study GHG emissions, methodological approaches are used that involve in situ gas flux measurements [Kraev *et al.*, 2017; Oblogov *et al.*, 2020; Zadorozhnaya *et al.*, 2022] or experimental incubation with periodic measurements [Vonk *et al.*, 2013; Knoblauch *et al.*, 2018; Shirokova *et al.*, 2019; Prijac *et al.*

al., 2022]. The advantage of the latter is the possibility to monitor carbon transformation in controlled/closed conditions of an incubated vessel. Some studies focus on measuring the loss of organic carbon due to its biogenic mineralization during the incubation and on determining biodegradable dissolved organic carbon (BDOC) [Vonk et al., 2013, 2015; Shirokova et al., 2019]. Other studies are focused on the dynamics of GHG emissions and fluxes over time [Brouchkov, Fukuda, 2002; Tanski et al., 2019; Alekseeva et al., 2021]. It is also necessary to distinguish between methodological approaches that involve measuring the dynamics of DOM and DOC in the filtrates of thawed permafrost samples [Vonk et al., 2013, 2015; Shirokova et al., 2019; Prijac et al., 2022], and approaches that involve the analysis of the dynamics of total and particulate organic matter (POM) and the corresponding carbon fractions: total organic carbon (C_{org} , TOC) and particulate organic carbon (POC) [Knoblauch et al., 2013, 2018; Shakil et al., 2022]. Experiments show that significant and stable generation of CH_4 by methanogenic archaea in anaerobically incubated thawed permafrost samples begins after 3–4 years of zero activity (lag phase), which is probably due to the adaptation of microbiota. Therefore, for experiments to determine methane production in permafrost samples to be reliable, they must be long-term [Knoblauch et al., 2013, 2018]. The fastest and most intense process of biogeochemical OM mobilization is aerobic respiration of bacteria, accompanied by CO_2 emission, which can be demonstrated by short-term incubation experiments [Vonk et al., 2013, 2015; Tanski et al., 2019; Alekseeva et al., 2021]. DOM undoubtedly includes the most labile fraction of OM, primarily spent on biogenic generation of CO_2 , but the POM pool is much larger in terms of total C_{org} content [Shakil et al., 2020]. In this paper, we analyze the dynamics of CO_2 production in filtered and intact samples (initially undisturbed monoliths of ground ice after melting), including three ice bed samples from the Yamal Peninsula thermocirque exposures and two wedge ice samples (from the Yamal Peninsula and from the vicinity of Norilsk city).

The aim of the work is to study, in the course of experimental incubation, some patterns of DOM transformation into GHG in ground ice samples during their melting. The main objectives are to (1) analyze the temporal dynamics of CO_2 emission in ground ice melt samples, (2) quantitatively assess the contribution of solid/suspended matter to CO_2 emission, (3) determine the amount of BDOC and the effect of suspended/solid matter on DOC mineralization under experimental conditions. The idea of this work is not to simulate natural conditions of seasonal melting of ground ice but to demonstrate the potential of ground ice to release labile carbon and generate GHG. In this regard, conditions are created for ex-

perimental incubation that favor biogenic generation of GHG in the aerobic regime in all samples.

STUDY AREA AND GENERAL CHARACTERISTICS OF THE SAMPLES

Five samples of ground ice monoliths were collected in typical exposures of Central Yamal (at the Vaskiny Dachi (VD) Research station) and the environs of Norilsk (N) city (Industrial District, the Daldykan Stream) (Fig. 1). The general description and initial basic geochemical characteristics of the samples are given in Table 1. Figure 2 shows the studied geocryological sections indicating the sampling location of each monolith used in the incubation experiments.

The monolith of massive ice bed (VD1ib) was sampled from the scarp of the thermocirque TC-2A (“Vaskiny Dachi” Research Station), approximately 1.9 m below the upper edge of the exposure represented by alternating layers of “black” ice and “white” transparent (glassy) ice with inclusions of sand and dark gray sandy loam. Gas bubbles in the ice monolith were round in shape and up to 2 mm in diameter. The upper edge of the thermocirque had the absolute height of 21 m a.s.l. with its base at 13.7 m a.s.l. (Figs. 1b, 2a, and 2b).

Monolith VD2ib was sampled from the scarp of thermocirque TC-5 (“Vaskiny Dachi” Research Station) from the ice exposure approximately 3 m below the upper edge, at the interface of gray loam and transparent glassy ice. This thermocirque was located on a slope and has its upper edge at 25 m a.s.l. and its base at 3 m a.s.l. (Figs. 1b, 2a, and 2b).

Monolith VD3ib was sampled from an ice layer exposed in the eastern scarp of the TC-5 thermocirque. The ice lens was located 120 cm below the upper edge of the thermocirque and had a thickness of about 40 cm and visible width about 1.5 m. The ice was clean, transparent, with numerous bubbles (Figs. 1b, 2a, and 2b).

Ice wedge sample VDiw was taken in a pit dug in a polygonal peat bog (“Vaskiny Dachi” Research Station). The head of the Holocene ice wedge was exposed at a depth of about 0.5 m in an interpolygonal trough. The ice was almost transparent with characteristic vertical stratification. The monolith was collected to the right of the ice wedge axis (Figs. 1b, 2a, and 2b).

Ice wedge sample Niw was taken a Holocene peat bog on the right bank of the Daldykan Stream (Norilsk Industrial Region), 200 m from the channel. The peat bog extended from north to south for 150 m and had a width of about 50 m. The polygonal relief was at the degradation stage, and the difference in height between the centers of polygons and interpolygonal troughs could be up to 2 m. The diameter of the polygons varied within 8–13 m (up to 17 m) (Figs. 1c, 2a, and 2b).

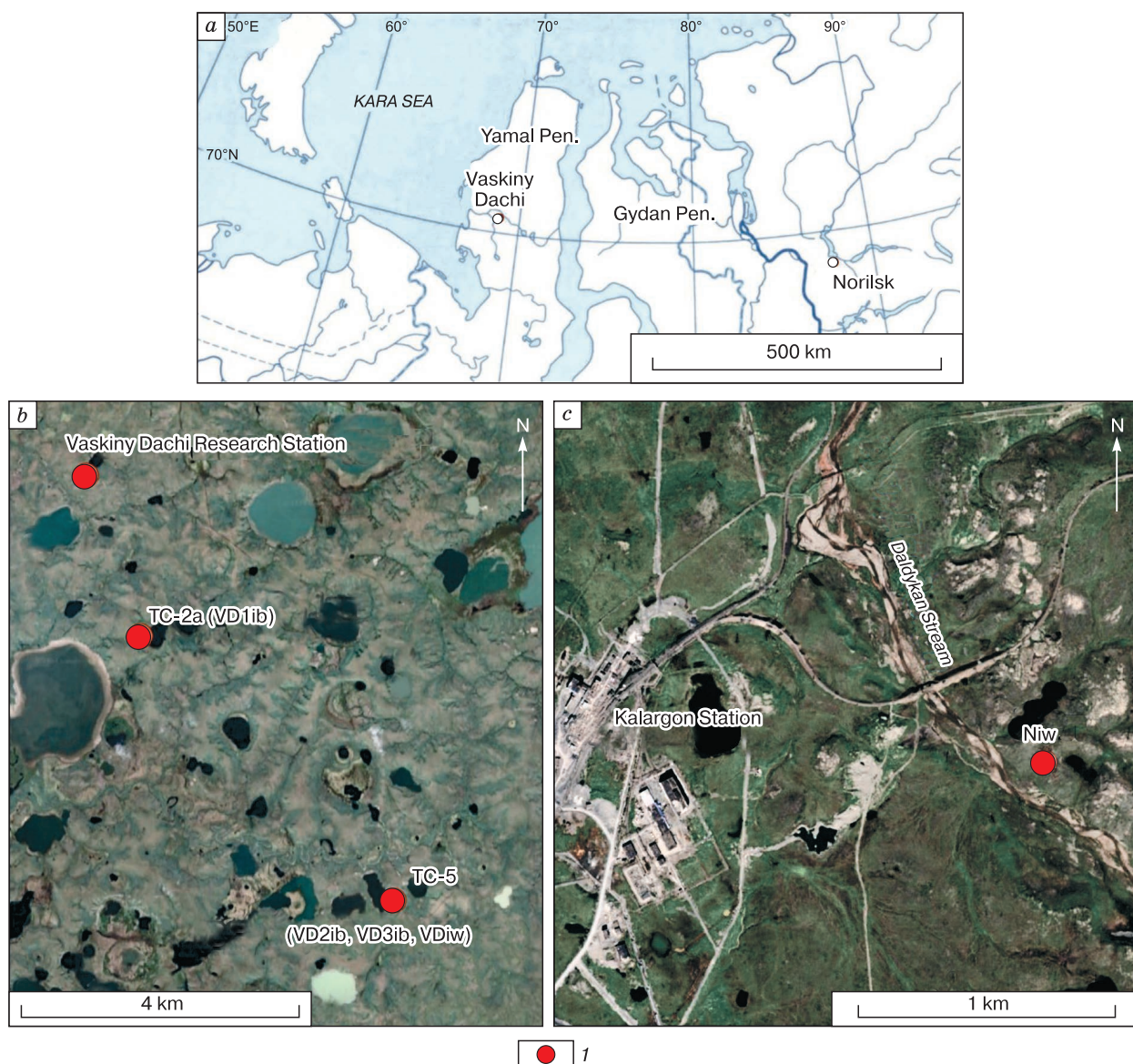


Fig. 1. An overview map of the ground ice sampling (a); a detailed map of sampling location (VD1ib, VD2ib, VD3ib, VDiw) in Central Yamal (b); a detailed map of sampling location (Niw) in Norilsk Industrial Region, the Daldykan Stream (c).

1 – sampling sites; TC – thermocircue, ib – ice beds, iw – ice wedges, VD – Vaskiny Dachi, N – Norilsk.

Table 1. General characteristics and initial geochemical parameters of ice samples

Sample ID	Ice type	Sampling site	Lithogenic fraction, wt. %	TDS, mg/L	pH	DOC, mg/L	CH ₄ , ppmV	CO ₂ , ppmV
VD1ib	Ice bed	Yamal Peninsula, "Vaskiny Dachi" Research Station	11.33	74.58	7.82	42.66	14 802.73	2435.95
VD2ib	Ice bed		41.89	397.01	7.08	126.53	3485.78	2030.94
VD3ib	Ice bed		3.61	40.57	7.20	7.14	1019.20	2138.37
VDiw	Ice wedge		2.95	20.90	6.19	62.34	418.15	34 998.70
Niw	Ice wedge	Norilsk, Daldykan Stream	0.01	9.27	6.62	5.35	3.90	1334.59

Note: TDS – Total Dissolved Solids in water; pH – hydrogen index, DOC – dissolved organic carbon, CH₄ – methane, CO₂ – carbon dioxide, ppmV – parts per million by volume; Sample ID: ib – ice bed; iw – ice wedge, VD – Vaskiny Dachi, N – Norilsk.



Fig. 2. Cryolithological sections with monolith sampling locations (a); ground ice monoliths used for incubation experiments (b); vials with ground ice melts (c).

1 – sampling points; filtered variants of melts are marked with “f”; ib – ice beds, iw – ice wedges, VD – Vaskiny Dachy, N – Norilsk.

MATERIALS AND METHODS

To study the initial gas composition of the samples, ground ice monoliths were placed in 75-mL glass vials. Vials were pre-washed with 1 N HCl and deionized water, then dried and weighed. Each sample placed in the vial was weighed to obtain the sample mass. The space above the sample (headspace) was washed with pure argon and hermetically sealed. The sealed vials were shaken using an orbital shaker (Heidolph Unimax 2010) for 3 h at maximum rotation speed (3500 rpm). The content of solid material (lithogenic fraction in % of the initial sample mass), gas composition (CH_4 , CO_2 , N_2), and DOC in ice

melts were measured according to the methods described in previous publications [Semenov *et al.*, 2020, 2024; Pismeniuk *et al.*, 2023].

For the experimental studies, ice monoliths were melted at $+4^\circ\text{C}$ for 24–48 hours. The resulting initial melts of each of the five samples were used to prepare two experimental variants (subsamples): intact (unfiltered) and filtered ($0.45\ \mu\text{m}$). Comparison of the intact and filtered variants allowed for the selective determination of CO_2 and biodegradable DOC associated with the dissolved and suspended (lithogenic) fractions of the ground ice samples, respectively. The filtered samples were marked as “f” (fVD1ib, etc.). Due to the limited amount of material (ground ice

monoliths), only three out of five samples were duplicated (VD2ib, VD3ib, Niw). Overall, 16 samples were subjected to the incubation. Figure 2c shows photographs of incubation variants of ground ice melts.

Suspensions (intact melts) and melt filtrates of ground ice samples were poured into pre-weighed and washed with 1 N HCl and deionized water vials; approximately 100 mL of material was placed in each vial according to the scale marked on the vial. The vials with samples were weighed and the mass of the samples was recorded and subsequently used for quantitative calculations. The main study was based on the experimental incubation of samples under aerobic conditions at room temperature, tracking the dynamics of CO₂, and the amount of BDOC at the beginning and at the end of the incubation in each variant [Vonk *et al.*, 2013, 2015; Shirokova *et al.*, 2019; Tanski *et al.*, 2019; Alekseeva *et al.*, 2021]. To equalize the conditions of “microbiological colonization” of the incubated 16 samples the inoculation was used. Filtrate of an aqueous suspension of cultural soil (0.75 µm, Watman GF-F) was used as an inoculum injected into each sample (100 µL). This method allows introducing a sufficient amount of active aerobic microbiota into each incubated sample [Vonk *et al.*, 2013].

The duration of the incubation experiment was 61 days. During the incubation, periodic sampling was performed (in addition to the initial point) on days 6, 15, 20, 26, 38, 50, and 61. The concentration of CO₂ and other gases was measured after shaking the incubation flasks for 3 hours on an orbital shaker (Heidolph Unimax 2010). The procedure for collecting gas and filtrate samples, as well as measuring gas concentrations (CO₂, etc.) and DOC, are similar to those described above for determining the initial geochemical characteristics of ground ice samples. For each measurement, the current mass of the sample and the volume of gas (headspace) were taken into account. Gas concentrations were calculated using the ideal gas equation, Henry's Law, and Bunsen coefficients [Yamamoto *et al.*, 1976; Ioffe, Vitenberg, 1984]. The CO₂ generation (emission) during the experiment was expressed as the mass of carbon CO₂ per mass of melt (CO₂-C, µg/g), as well as the mass of carbon per mass of melt per day of the experiment (CO₂-C, (µg/g)/day) [Knoblauch *et al.*, 2013]. Changes in the CO₂ concentration over time were plotted as two curves: a dynamic curve, where the amount of CO₂ released during each interval of the incubation was visualized, and a cumulative curve, where the portions of CO₂ released during each interval were summed up. Thus, the end point of the cumulative curve characterized the maximum amount of CO₂ released during the experiment. The BDOC content was determined by the difference in its concentrations at the beginning and end of the incubation test [Vonk *et al.*, 2013].

RESULTS

Initial geochemical characteristics of the samples

The work involved three samples of ice beds (ib) and two samples of ice wedges (iw). Sample VD2ib had the highest content of lithogenic fraction reaching 41.89 wt.% (Table 1). The lowest content of lithogenic fraction was in the ice wedge sample Niw collected in the vicinity of Norilsk. Both ice wedge samples (VDiw and Niw) had pH values <7 (up to 6.19), which was probably due to their localization within the polygonal peatlands. The highest pH value (7.82) was recorded in sample VD1ib. The maximum value of total mineralization (TDS) was found in sample VD2ib 397.01 mg/L, while the minimum (9.266 mg/L) was measured in sample Niw. Sample VD2ib was enriched in dissolved salts by more than 5 times compared to ice bed samples VD1ib and VD3ib, which was probably due to a higher content of the lithogenic fraction captured by ice from the enclosing sediments of marine origin.

The maximum DOC content (126.53 mg/L) in sample VD2ib was 23.6 times higher than the minimum DOC content (5.35 mg/L) in sample Niw. The sample from the Holocene ice wedge VDiw was remarkable for the high DOC content (62.34 mg/L), and an abnormally high CO₂ concentration (34 598 ppmV) in the gas phase of the melt. In all samples, CO₂ concentrations were many times higher than the modern atmospheric background values, indicating the enrichment of the source water in carbon dioxide. More or less similar values of CO₂ concentration were determined in ice bed samples: 2435.95 ppmV (VD1ib), 2030.94 (VD2ib), and 2138.37 ppmV (VD3ib). The maximum concentration of methane (CH₄) measured in the VD1ib sample – 14 802.72 ppmV – was many times higher than all the other values measured in this experiment. The minimum concentration of CH₄ (3.90 ppmV), close to the equilibrium atmospheric concentration, was characteristic of the Niw sample. In general, ice wedge samples were characterized by lower methane concentrations than ice bed samples, but the range of variation between them exceeded the difference between the lowest concentration of CH₄ in ice beds (1019.20 ppmV, sample VD3ib) and the highest concentration of methane in ice wedges (418.14 ppmV, sample VDiw). The studied collection of the ground ice samples is sufficiently diverse in the distribution of key parameters to represent material for experimental studies.

Greenhouse Gas Generation under Experimental Incubation

The results of gas composition measurements during experimental aerobic incubation of ground ice melts demonstrate that CO₂ generation obviously differs in various ground ice samples, as well as in in-

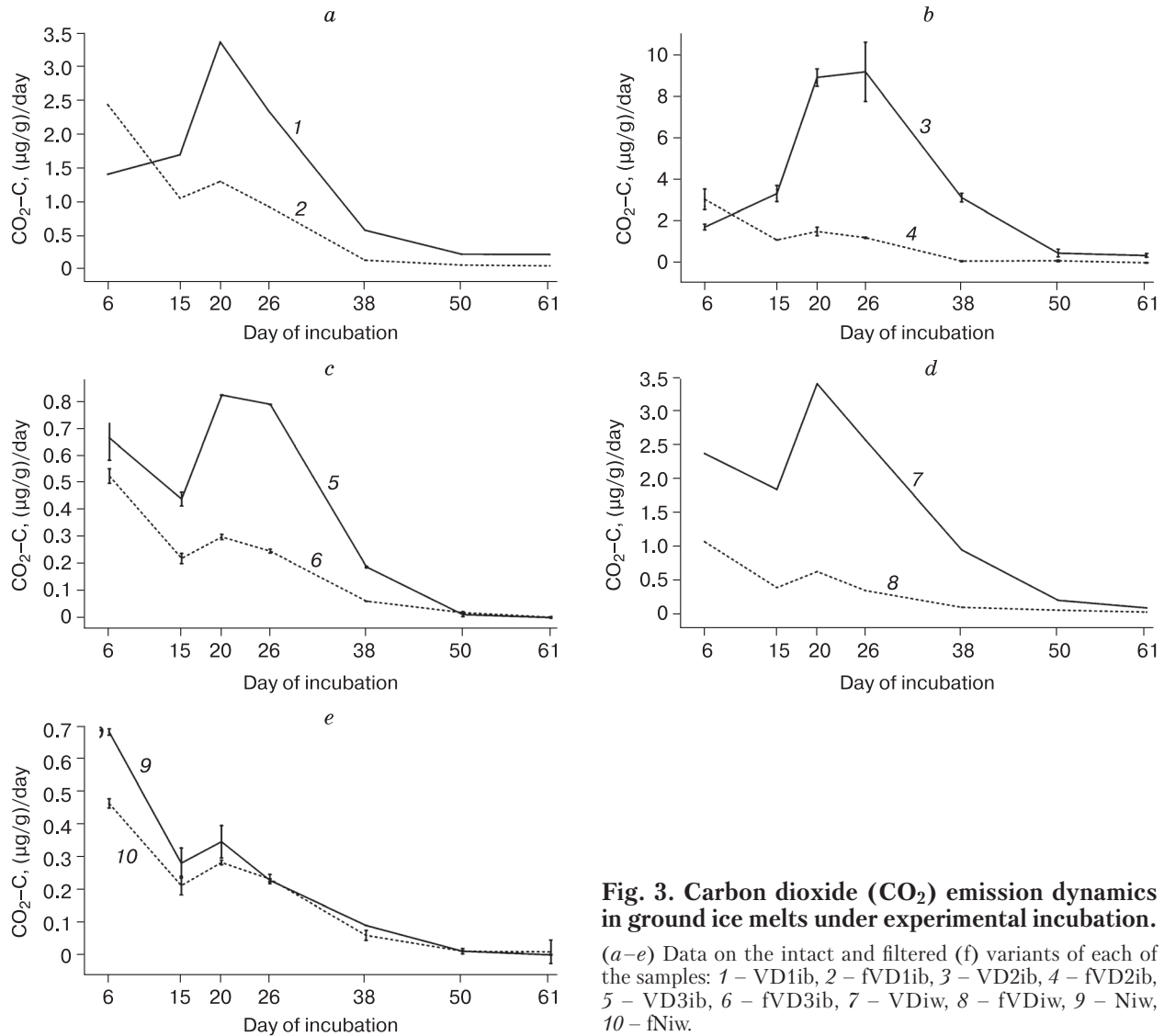


Fig. 3. Carbon dioxide (CO₂) emission dynamics in ground ice melts under experimental incubation.

(a–e) Data on the intact and filtered (f) variants of each of the samples: 1 – VD1ib, 2 – fVD1ib, 3 – VD2ib, 4 – fVD2ib, 5 – VD3ib, 6 – fVD3ib, 7 – VDiw, 8 – fVDiw, 9 – Niw, 10 – fNiw.

tact and filtered variants of each of the studied samples. No reliable increase in methane concentration in the gas phase was recorded in any of the samples during the experimental incubation of ground ice melts. In this regard, no data on methane emission are provided in the text or in the illustrations.

Diagrams illustrating CO₂ emission in ground ice samples are shown in Figs. 3 and 4. The plots indicate the dynamics of CO₂ emission for each of the two variants (intact and filtered). The maximum value of cumulative CO₂ emission (Figs. 4c and 4d) for the entire incubation period (61 days) was 173.40 CO₂-C, μg/g in the intact ice bed sample VD2ib from the central part of Yamal, which had the highest content of the solid fraction among the analyzed ice samples (41.8 wt.%). The minimum value of cumulative CO₂ emission (10.63 CO₂-C, μg/g) was found in the fil-

tered melt of sample Niw prepared from a transparent ice wedge with an extremely low content of the lithogenic fraction (~0.01 wt.%). Thus, the measured minimum and maximum values of CO₂ cumulative emission differ by more than 17 times. The curves of the CO₂ generation dynamics in intact samples are characterized by a sharp rise during 15–26 days of incubation, when the maximum values of the average daily CO₂ emission are observed (up to 9.160 CO₂-C, μg/(g·day) in the VD2ib sample). The standard deviation (SD) values between the duplicate samples of CO₂ emission measurements, plotted on the graphs as error bars for the VD2ib, VD3ib and Niw samples (Figs. 3b, 3c, 3e; Figs. 4a–4d), are notably higher in the intact variants than in the filtered ones, which is especially noticeable for VD2ib sample with the highest content of the lithogenic fraction.

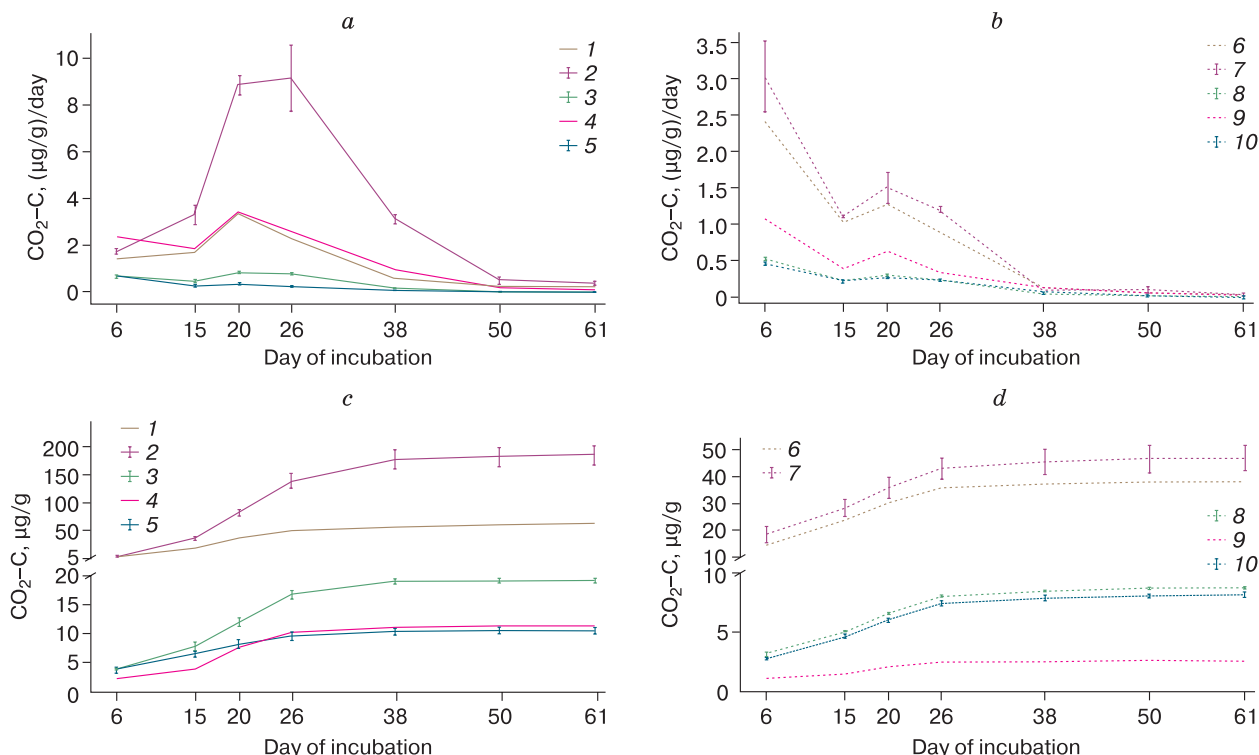


Fig. 4. Carbon dioxide (CO₂) emission from the ground ice melts under experimental incubation.

(a, b) – emission dynamics curves; (c, d) – cumulative emission curves. 1–5: intact (continuous lines, 1 – VD1ib, 2 – VD2ib, 3 – VD3ib, 4 – VDiW, 5 – Niw), 6–10: filtered (dotted lines, 6 – fVD1ib, 7 – fVD2ib, 8 – fVD3ib, 9 – fVDiW, 10 – fNiw).

Biodegradable dissolved organic carbon (BDOC) under experimental incubation conditions

BDOC values (calculated as a difference in DOC concentrations at the initial and final points of incubation) vary from 83.39 mg/L (ice bed sample VD2ib with the highest solid fraction content and cumulative CO₂ emission value for the entire incubation period) to 0.27 mg/L in the filtrate of transparent melt (Niw) with the lowest values of the above parameters. The DOC data under experimental conditions are summarized in Table 2. In general, for the studied samples, BDOC demonstrates higher values in the intact variants. The ratios of BDOC concentrations in intact and filtered samples are (in descending order): 13.2 (VD2ib), 2.23 (VD1ib), 1.814 (Niw), 1.55 (VD3ib). Sample VDiW is an exception: it demonstrates a higher BDOC value in the filtered variant (2.98 mg/L versus 1.30 mg/L in the intact sample).

DISCUSSION

The work involves a collection of five samples, four of which were obtained in the area of “Vaskiny Dachi” Research Station. Ground ice exposures and permafrost thawing conditions in this location have been a subject of numerous studies [Dvornikov *et al.*,

2017; Khomutov *et al.*, 2017; Leibman, Khomutov, 2019; Savvichev *et al.*, 2021].

The curves of the CO₂ emission dynamics in the incubated samples generally form different patterns for the intact and filtered variants. The former are characterized by peak values of the average daily CO₂ emission in the period from the 15th to the 38th day of incubation with a maximum in the range of 26th–28th day. The latter are characterized by the highest intensity in the period from the 6th to the 15th day with a maximum at the earliest stage of incubation (the 6th day). The temporal dynamics of CO₂ generation undoubtedly reflects the mobilization of labile OM into the local carbon cycle. Probably, the first stage of peak CO₂ emission is associated with the use of the most labile fraction of ground ice DOC, which already resides in the solution and is directly accessible to microorganisms. Intensive CO₂ generation during the first days of incubation was previously noted for various frozen rocks enriched in labile OM [Lee *et al.*, 2012; Knoblauch *et al.*, 2013]. The next peak of CO₂ emission (15th–38th day), most pronounced in intact samples, is apparently due to the supply of an additional portion of organic carbon from the solid component of ice melts. This portion plays a significant quantitative role in the resulting (cumulative) CO₂

Table 2. Concentrations of dissolved organic carbon (DOC) and biodegradable dissolved organic carbon (BDOC) in an incubation experiment

Sample	DOC _{init}	DOC _{fin}	BDOC	
	mg/L		mg/L	%
VD1ib	42.66	25.28	17.38	40.74
fVD1ib		34.88	7.78	18.24
VD2ib (1)	126.53	43.14	83.39	65.91
VD2ib (2)		48.22	78.32	61.90
VD2ib av			80.85	63.90
VD2ib, SD			3.59	2.84
fVD2ib (1)	126.53	119.47	7.06	5.58
fVD2ib (2)		121.35	5.19	4.10
fVD2ib av			6.12	4.84
fVD2ib, SD			1.33	1.05
VD3ib (1)	7.14	4.74	2.40	33.61
VD3ib (2)		4.88	2.26	31.65
VD3ib av			2.33	32.63
VD3ib, SD			0.10	1.39
fVD3ib (1)	7.14	5.51	1.63	22.83
fVD3ib (2)		5.78	1.36	19.05
fVD3ib av			1.50	20.94
fVD3ib, SD			0.19	2.67
VDiw	62.34	61.04	1.30	2.09
fVDiw		59.36	2.98	4.78
Niw (1)	5.35	4.75	0.60	11.21
Niw (2)		4.97	0.38	7.10
Niw av			0.49	9.16
VDiw, SD			0.16	2.91
fNiw (1)	5.35	5.04	0.31	5.79
fNiw (2)		5.12	0.23	4.30
fNiw av			0.27	5.05
fNiw, SD			0.06	1.06

Note: DOC – dissolved organic carbon, BDOC – biodegradable dissolved organic carbon, DOC_{init} and DOC_{fin} – DOC values before and after incubation, respectively; av – average value; SD – standard deviation; f – filtered variants; ib – ice beds; iw – ice wedges, VD – Vaskiny Dachy, N – Norilsk.

emission from ice melts. The curves of CO₂ dynamics emission in the intact and filtered variants of the Niw sample, which contains almost no solid inclusions, practically coincide, illustrating the minor role of suspended matter in the mineralization of OM to CO₂ in this sample. The mobilization of OM from the solid phase can be associated with desorption from the organomineral matrix and with the dissolution/destruction of detrital forms of OM in the sample [Groeneveld *et al.*, 2020]. Suspended material of the incubated mixture also contributes to the fixation and development of microbial biomass [Battin *et al.*, 2008; Shakil *et al.*, 2020]. The above-mentioned processes are certainly activated by ice melting and high temperatures in the incubated melts (+22°C) during the incubation period. It should be noted that in this

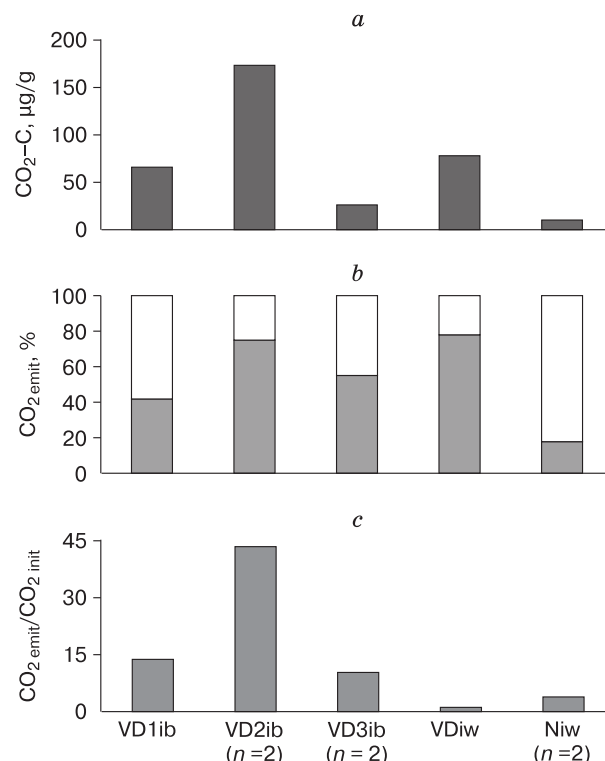


Fig. 5. Comparison of cumulative CO₂ emission values in the ground ice melt as a result of 61-day incubation (a); contribution of dissolved and suspended OM (associated with the solid phase of melts) to CO₂ emission (b), the ratio of cumulative CO₂ emission to the initial CO₂ concentration in the ground ice sample (c).

For duplicate incubations ($n = 2$) averaged values (VD2ib, VD3ib, Niw) are shown.

work, the authors did not aim to reproduce the natural conditions of ground ice melting, but focused on comparing data on different samples placed in equal conditions of a short-term incubation experiment.

It follows from Fig. 5a that the highest values of cumulative emission are characteristic of two ground ice samples, particularly the ice bed sample VD2ib (173.40 CO₂-C, μg/g) and the ice wedge sample VDiw (78.035 CO₂-C, μg/g), which generally illustrates quite comparable potential for CO₂ generation out of ground ice of different origin. It is curious that the above-mentioned samples (VD2ib and VDiw), collected from the thermocirques of central Yamal, are characterized by similar values of the suspended organic matter contribution (75.02 and 77.95%, respectively) to the cumulative CO₂ emission with a more than 20-fold difference in the solid fraction content of the incubated melts (41.88 and 2.95% by weight, respectively). This indicates the critical importance of the solid fraction of ground ice as a source of biolabile DOM mineralized by aerobic microbiota

with the release of CO₂, and also confirms the special role of ice wedges in the local carbon cycle during ground ice melting in thermocirques. Similar patterns were noted in a number of studies describing the process of OM mobilization from suspended material of frozen rocks of thermokarst landscapes [Tanski et al., 2017; Shakil et al., 2020]. In addition, the literature describes the stimulating effect of ice wedge melts on the mineralization of DOM in aqueous extracts of the ice complex under experimental incubation conditions (the Kolyma River bank, Duvanny Yar) due to the co-metabolism known as the priming effect [Vonk et al., 2013; Textor et al., 2019]. A comparison of the initial CO₂ concentrations in the ice bed melts and the concentrations of CO₂ released as a result of the labile OM biogenic mineralization (CO₂ emissions during incubation) shows that these parameters are apparently not related. This is illustrated by the plot in Fig. 5c, where the above-mentioned variables ratio (CO₂ emission/CO₂ initial) shows a significant scattering that does not follow any pattern. Thus, the initial CO₂ concentration in the ice bed sample does not predict the potential for CO₂ generation from the enclosed OM.

Based on the temporal dynamics of the ground ice OM biogenic mineralization during the incubation, we suggest the following phases of CO₂ emission: the first phase (6th to 15th day of incubation) is associated with mobilization of DOM in the melt; the second one, with the release of DOM from the suspended matter (15th to 38th day); then, a decrease in the CO₂ emission is observed in all incubated samples (38th to 61st day). A sharp fall in the activity of aerobic respiration (the main biochemical process providing mineralization of OM to CO₂) on the 28th to 40th day of incubation may be associated with the depletion of labile OM or bioavailable forms of nitrogen, limiting the intensity of biochemical processes in the obvious absence of oxygen starvation (the O₂ concentration in the head space of the samples did not fall below 16%). Probably, a similar pattern of CO₂ emission with two peaks caused by the sequential utilization of DOM and POM, respectively, is also a characteristic of natural conditions of thermodenudation within ground ice exposures.

A correct comparison of the data on cumulative CO₂ emission from ground ice samples with similar data available in the literature on frozen rocks is not entirely feasible, since in the latter cases the values are recalculated per dry sample weight [Knoblauch et al., 2013; Tanski et al., 2019]. In the present work, the amount of CO₂ released during incubation is recalculated per mass of melted ground ice samples containing both water and solid fraction. However, even very rough comparisons taking into account the above-mentioned discrepancies indicate significantly lower CO₂ emission by ground ice melts in relation to bot-

tom sediments or frozen rock samples [Knoblauch et al., 2013; Tanski et al., 2019; Alekseeva et al., 2021].

Incubation experiments demonstrate a significant loss of labile (biodegradable) DOC (BDOC) during ice melt incubation. The maximum BDOC recorded is 65.91% over 61 days of incubation in the sample VD2ib, which is also notable for the maximum initial DOC concentration (126.53 mg/L) and maximum solid fraction content of 41.88% by weight. Sample VD2ib is also characterized by an extremely high total dissolved solids (TDS) value of 397.01 mg/L, which may indirectly indicate the contribution of marine OM to the labile carbon pool that provides the maximum CO₂ emission during the experiment. The mentioned BDOC value in the VD2ib sample (65.91% over 61 days) is comparable with the values (about 34% over 14 days of incubation) reported for surface water samples in the Pleistocene ice complex zone sampled on the Kolyma River bank [Vonk et al., 2013, 2015]. Special attention should be paid to the higher BDOC values in the intact variants compared to the filtered ones, which are typical for all samples except for VDiw (Fig. 6). Thus, the ratios of the BDOC values in the intact variants to the filtered ones are in descending order: 13.2 (VD2ib), 2.23 (VD1ib), 1.81 (Niw), 1.56 (VD3ib). This feature contradicts our initial expectations, according to which a more intensive DOC consumption should be observed in filtered variants, where DOC is the only source of energy for aerobic chemoorganotrophic microbiota, whereas in intact melts there is an additional source of OM associated with the solid phase (SOM). Based on this data, we cannot exclude that more intensive DOC consumption and higher values of CO₂ emission in the intact samples are caused by the activation of biogeochemical processes

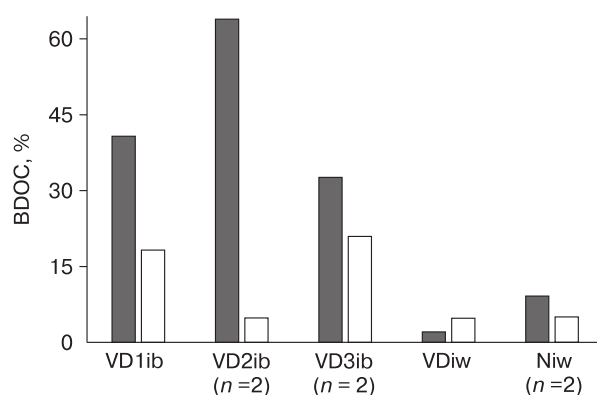


Fig. 6. Comparison of biodegradable dissolved organic carbon (BDOC) concentrations in the ground ice melt as a result of 61-day incubation.

Dark filled columns – intact variants, white filled columns – filtered variants. For duplicate incubations ($n = 2$) average values are shown (VD2ib, VD3ib, Niw).

BDOC – biodegradable dissolved organic carbon.

in the presence of solid material rather than by the release of the additional portion of DOC from the SOM pool. However, in this work, involving a rather narrow set of both experimental variants and variables, we prefer a more straightforward hypothesis that the labile DOM is released from the solid phase of melts (due to desorption and dissolution), which is based mainly on the comparison of the cumulative CO₂ emission values in the intact and filtered melts. The sample of ice bed VD2ib is remarkable for the greatest influence of the solid fraction on the mineralization of DOM, and the sample Niw is notable for the opposite trend, where a large value of BDOC is noted for the filtered variant. Further detailed studies should bring us closer to understanding the obtained BDOC distribution patterns.

Figure 7 shows a graph of the relationship between the values of cumulative CO₂ emission and concentrations of BDOC in ground ice melts indicating a significant positive correlation ($R^2 = 0.82$, $p = 0.003$). Both axes of the plot have a logarithmic dimension for a clearer presentation of the low values. A previously mentioned feature is evident here: for the majority of samples there is a more significant decrease of DOM in the intact variants compared to the filtered ones, which is associated with a more intense CO₂ emission. The only exception is the ice wedge sample VDiw collected at the “Vaskiny Dachi” Research Station: its intact variant shows disproportionately high values of CO₂ emission (78.035 CO₂-C, µg/g) in combination with relatively low values of BDOC. Formally, this can be explained by the compensation of the BDOC loss due to extraction from the solid phase. Taking into account the low content of the solid fraction in the VDiw melt (2.95%

by mass), we assume the presence of weakly transformed detrital forms of OM which replenish the labile DOM through dissolving in the aqueous phase of the melt. We should also consider the massive ice bed sample VD3ib with higher ratio of CO₂ emission to the BDOC content than of the rest of the melts, except for VDiw. This fact is difficult to interpret unequivocally at this stage of research, but it definitely characterizes the specifics of the composition of DOM utilized by aerobic chemoorganotrophic microbiota during incubation. Apparently, the correspondence of the BDOC values to cumulative CO₂ emission is a promising tool for the analysis of experimental data, as well as the differentiation of ground ice samples in relation to the local carbon cycle. Further studies with ground ice monoliths at the level of the dissolved and suspended organic matter chemical composition, accompanied by CO₂ and BDOC measurements, will allow us to clarify the patterns obtained in this work and improve the diagnostic approach based on the experimental incubation results.

The results of this study indicate a sufficiently high lability of the ground ice DOM, as well as complex and multifaceted interactions between DOM and SOM, which do not fit into mechanistic concepts.

CONCLUSIONS

The studies have shown that the dynamics of CO₂ generation reflects two temporal stages of labile OM mobilization into the local carbon cycle, where the first stage is associated with the use of exclusively ground ice DOM, and the second is due to the mobilization of labile OM from the solid fraction of ground ice melts. The maximum measured value of cumulative CO₂ emission is 17 times higher than the minimum, indicating high variability of the CO₂ emission potential of melting ground ice.

For ground ice samples relatively enriched in the solid fraction, the significance of suspended organic matter (associated with the solid fraction of melts) has been demonstrated as a source of biochemically labile DOM readily mineralized by aerobic microbiota with the release of CO₂. At the same time, the relative contribution of solid material is different, which indicates the influence of its composition on the CO₂ emission potential. Incubation experiments show a drastic decrease in biodegradable DOM, up to 65.91% in 61 days indicating high lability of the ground ice organic matter. A significant positive correlation ($R^2 = 0.82$) was found between the values of cumulative CO₂ emission and the values of BDOC in ground ice melts. At the same time, higher values of BDOC in intact variants compared to filtered ones (where DOM serves as the only carbon source for the generated CO₂) are characteristic of most ground ice samples.

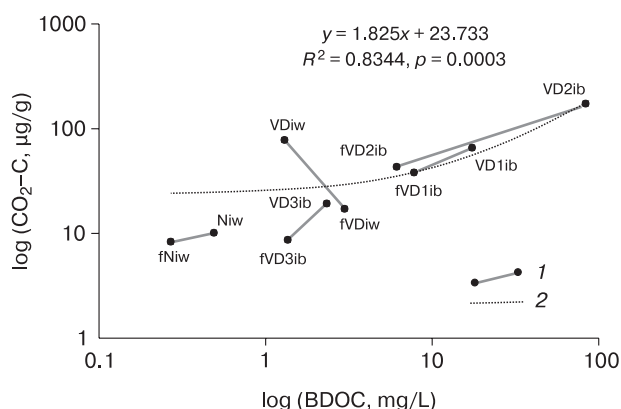


Fig. 7. Correlation between the content of biodegradable dissolved organic carbon (BDOC) and cumulative CO₂ emission measured in the aerobic incubation experiment (61 days).

1 – the lines connecting points for filtered (f) and intact variants of corresponding ground ice samples; 2 – regression trendline. For other designations, see Figs. 1 and 5.

The results of this study indicate a sufficiently high lability of the ground ice DOM, as well as complex and multifaceted interactions between DOM and SOM, which do not fit into mechanistic concepts.

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PERMAFROST ENGINEERING

FORECAST OF PILE FOUNDATION STABILITY BASED ON THE RESULTS OF TEMPERATURE OBSERVATIONS IN A NEARBY BOREHOLE

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The possibilities of using the results of permafrost temperature measurements for a short-term forecast of permafrost temperature regime by the extrapolation method and assessment of the stability of the pile foundation on the basis of this forecast are considered. The soil temperature has been monitored over the entire depth of the pile foundation, including one temperature sensor located under the lower end of the pile. The processing of the results of temperature measurements has been carried out according to the outlined methodology using three computer programs, links to which on the Internet are given in this article.

Keywords: *temperatures observations, permafrost soil, forecast, stability, processing technique.*

Recommended citation: Khrustalev L.N., Emelyanova L.V., 2025. Forecast of pile foundation stability based on the results of temperature observations in a nearby borehole. *Earth's Cryosphere* XXIX (1), 57–64.

INTRODUCTION

The stability of foundations on permafrost is known to be assessed at the design stage based on the permafrost temperature forecast, which is carried out using an analytical or numerical solution to the problem of thermal conductivity with phase transitions of water in the soil (Stefan's problem). The accuracy of solving such a problem depends on the accuracy of setting boundary conditions, which include the thermal impact of the environment and constructions on the foundation soil. Unfortunately, there are many factors in analytical and numerical solutions that are not adequately taken into account, such as the thermal resistances of snow and vegetation and the thermal effect of groundwater. This leads to design errors that can be detected using geotechnical monitoring. One of the main tasks of geotechnical monitoring is to monitor the evolution of the temperature field of the permafrost at the base of constructions. To prevent accidents, it is necessary to know the predicted values of the permafrost soil temperature, which can be obtained by extrapolating the past and present temperature conditions of the permafrost soil to the future. According to the authors, this approach is more efficient than calibrating mathematical models for their use for predictive purposes. As is known, the Stefan problem has no inverse solution, i.e., its solution does not allow us to obtain initial and boundary conditions. Therefore, a direct solution is used for calibration, in which initial and boundary conditions are intuitively selected in order to obtain an adequate result.

Thus, the first problem considered in this article is the development of a sufficiently accurate method for extrapolating permafrost temperature fields over time, that is, to find an adequate extrapolation function and establish the optimal lead (forecast) period.

The second problem is the mathematical processing of the predicted permafrost temperature field in order to obtain calculated parameters for assessing the stability of the pile foundation. It includes calculating the average temperature of the permafrost near the side surface of the pile and under its lower end, as well as the thickness of the thawed soil layer above the permafrost and the seasonal freezing layer.

Finally, the third problem is to calculate the stability of the pile foundation as a result of the impact of the payload and frost heave forces. Safety factors characterizing (a) the impact of the payload (the ratio of the pile's bearing capacity to the payload) and (b) frost heave forces (the ratio of the forces preventing the pile from being heaved out to the frost heave forces) are also calculated.

Extrapolation of temperature observations over time

Predictive calculations of permafrost temperature fields based on the actual ground temperature constitute a problem of extrapolating a discrete set of measurement moments over time. In this case, the inverse Stefan problem arises, which, as noted above, is mathematically incorrect and does not have an unambiguous solution. The way out of this situation is a compromise approach, in which the physical nature of the problem serves as the basis for finding a class of

support functions that make it possible to find the best extrapolation function by relatively simple calculations. To implement it, the authors used an approximate method for calculating temperature fields in the foundations of constructions, called the equivalent temperature method [Khrustalev, 1971]. The idea of the method is to reduce Stefan's three-dimensional problem to a set of one-dimensional problems. Omitting the details, which can be found in the cited monograph, let us write the dependence of temperature (T) on time (τ):

$$T = cE(\tau) + d, \quad (1)$$

$$E(\tau) = 1 - \operatorname{erf}\left(\frac{y}{2\sqrt{a\tau}}\right), \quad (2)$$

where c , d are unknown parameters associated with the values of the equivalent and initial temperature, °C; y is the depth of the calculation point, m; a is the thermal diffusivity of the soil, m²/°C; and $\operatorname{erf}(u)$ is the integral of probabilities, a tabulated function, dimensionless.

For unknown parameters, there are soil temperature values measured at many points at certain points in time and forming a system of equations:

$$c_i E_{ij} + d_i = T_{ij}, \quad i = 1, 2, \dots, n, \quad j = 1, 2, \dots, m, \quad (3)$$

where n is the number of ground temperature measurement points in the borehole; m is the number of ground temperature measurements over time, separated by a period equal to a multiple of the number of years; E_{ij} is the value of function (2) at the point y_i at the τ_j .

Measured soil temperature values (in the area of thermal influence of the construction) are distorted by measurement errors and human-made and natural random deviations. Therefore, for $m > 2$, the system of equations (3) is inconsistent and only yields approximate solutions. However, if we take into account the two most recent temperature measurements for unknown parameters c and d , this system of equations has a rigorous analytical solution, namely:

$$\begin{cases} c_i E_{i,m-1} + d_i = T_{i,m-1} \\ c_i E_{i,m} + d_i = T_{i,m} \end{cases}, \quad (4)$$

$$c_i = (T_{i,j} - T_{i,m-1}) / (E_{i,j} - E_{i,m-1}); \quad (5)$$

$$d_i = T_{i,m} - c_i E_{i,m}. \quad (6)$$

Substituting (5) and (6) into (1), we obtain the desired extrapolation function:

$$T_{i,k} = c_i E_{i,k} + d_i, \quad k > m. \quad (7)$$

Calculation using formula (7) can be performed automatically. For this purpose, the authors have developed a computer program (macro A, see Supplementary materials).

Determination of design parameters for assessing the stability of a pile foundation

The predicted values of the soil temperature in the control borehole are determined as follows.

First, the number of the sensor (measured from the soil surface), where a positive temperature is detected is set to $n1$. Then the number of the last sensor located above sensor $n1$ and having a negative temperature is set to $n2$. Then, two important parameters are obtained by interpolation – the depth of soil thawing H at the time of the soil temperature forecast and the depth of seasonal soil freezing D_f :

$$H = Y_{n1+1} - (Y_{n1+1} - Y_{n1}) \cdot (T_{n1} - T_{bf}) / (T_{n1} - T_{n1+1}),$$

$$D_f = Y_{n2} + (Y_{n2+1} - Y_{n2}) \cdot (T_{bf} - T_{n2}) / (T_{n2+1} - T_{n2}),$$

where Y_i is the depth of the i -th temperature sensor, m; T_i is the soil temperature indicated by the i -th temperature sensor; T_{bf} is the freezing–thawing temperature of the soil, °C.

Next, the average temperature of the permafrost (T_e) is calculated using data from sensors located in the depth range from H to L (depth of immersion of the pile in the soil, m), and the temperature near the lower end of the pile (T_z) at the time of the forecast is calculated by interpolation:

$$T_z = (L - Y_{n1}) \cdot (T_{n1+1} - T_{n1}) / (Y_{n1+1} - Y_{n1}) + T_{n1}.$$

The calculation of these parameters can be performed in a computer program developed by the authors (macro B, Supplementary materials).

Calculation of the stability of the pile foundation based on the predicted temperature values

Calculation of the stability of a pile foundation includes checking two limiting conditions – bearing capacity and heaving.

Let's consider them sequentially.

Verification of the bearing capacity limits. In this case, monitoring the stability of the pile foundation is reduced to checking the value of the bearing capacity safety factor (k_u), which should always be greater than one during the foundation operation period, that is:

$$k_u = F_u / (F_v + F_g) > 1,$$

where F_u is the bearing capacity of the pile, kN; F_v is the payload on the pile, kN; F_g is the friction force of the thawed soil against the side surface of the pile, kN. In addition:

$$F_u = F_{u,T} + F_{u,b},$$

$$F_{u,T} = \gamma_c R S_p,$$

$$F_{u,b} = \gamma_c \gamma_{ca} R_{af} u_p (L - H),$$

$$F_g = 0.8 \gamma_{cf} f_c u_p (H - D_f),$$

where $F_{u,T}$ and $F_{u,b}$ are the bearing capacities of the lower end and the lateral surface of the pile, respectively, kN; γ_c is the coefficient of working conditions taken according to Table 7.2 [SP 25.13330.2020], dimensionless; γ_{ca} is the coefficient taking into account the pile material, dimensionless and taken equal to 1 for concrete, 0.7 for steel, and 0.9 for wood; R is the design pressure on the frozen soil under the lower end of the pile, kPa; s_p is the area of the lower end of the pile, m²; R_{af} is the design resistance of the frozen soil or soil mortar to shear on the lateral freezing surface within the part of the pile located in the frozen soil, kPa; u_p is the perimeter of the pile, m; L is the depth of pile penetration into the ground, m; and H is the thickness of the thawed layer above the permafrost table, m. Parameters R and R_{af} are determined according to the data given in Tables B1 and B3 of Supplementary materials [SP 25.13330.2020] depending on the calculated values of the soil temperature near the lower end of the pile T_z and its lateral surface T_e ; γ_{cf} is the coefficient of soil working conditions on the lateral surface of the pile determined according to Table 7.4 [SP 24.13330.2021] depending on the method of the pile driving, dimensionless; f_c is the calculated resistance of thawed soil along the lateral surface of the pile taken according to Table 7.3 [SP 24.13330.2021], kPa; D_f is the depth of seasonal soil freezing, m.

Verification of the heaving limit conditions. In this case, stability monitoring of the pile foundation is reduced to checking the value of the heaving safety factor k_f :

$$k_f = F_y / F_f > 1,$$

$$F_y = F_v + F_g + F_{u,b},$$

$$F_f = \gamma_{ca} \gamma_c \tau_{fh} u_p D_f,$$

where F_y and F_f are the force keeping the pile from heaving (temperature-dependent) and the heaving force, respectively, kN; τ_{fh} is the specific tangential heaving force taken according to Table 7.8 [SP 25.13330.2020], kPa.

A computer program (macro C, see Supplementary materials) was compiled to calculate the stability forecast of the pile foundation in automatic mode.

It should be noted that the described method, like most of the methods for analytical calculation of temperature fields and thawing “bowls” at the base of constructions, assumes that the soil is homogeneous. In the case of layered soil, averaged soil characteristics are used in the calculation. Averaging is performed within the averaging depth using formula for the soil thermal conductivity and formula for averaging the remaining characteristics of the frozen soil. The depth of the location of the last calculated tem-

perature sensor in the control borehole is taken as the averaging depth:

$$\lambda = \sum_{i=1}^n h_i / \sum_{i=1}^n \frac{1}{\lambda_i} \cdot h_i,$$

$$A = \sum_{i=1}^n A_i \cdot h_i / \sum_{i=1}^n h_i,$$

where λ_i , A_i are the numerical values of the soil characteristics of the i -th layer; h_i is the thickness of the i -th layer; n is the number of i -th layers within the depth of averaging.

Obtaining data for verification of the extrapolation method and compilation of an example calculation

Comparison of the forecast results with the data from field observations is replaced by a conditional comparison with the data of mathematical modeling of the permafrost temperature field at the base of the support of an aboveground gas pipeline located at a distance of 2 m from the building, since modeling allows us to obtain a fairly accurate picture of the soil temperature field under the specific conditions.

Mathematical modeling based on the solution of Stefan's problem was implemented in the WARM program [Emelyanov et al., 1994]. The algorithm of this program includes the finite difference method using the balance method and the enthalpy formulation of the problem [Ershov, 1999].

Problem statement. A two-dimensional model area of 24 × 24 m divided into blocks in horizontal and vertical directions with a spacing of 1 m is considered. On the right side of the model area (0 to 6 m), there is a building (half of a building). A 10-m-long support of the aboveground gas pipeline is found 8 m from the left boundary, and a 10.5-m-deep tem-

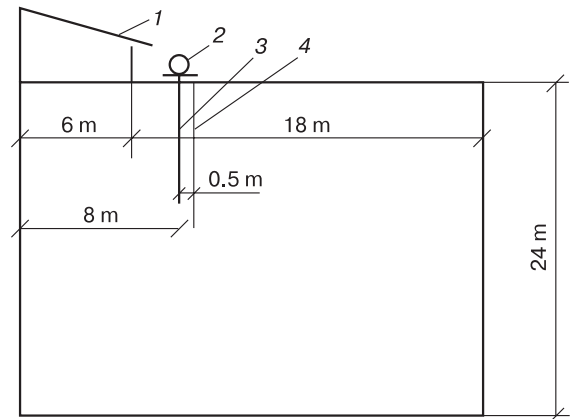


Fig. 1. Calculation scheme of the model (two-dimensional model).

1 – building (half), 2 – aboveground gas pipeline, 3 – aboveground gas pipeline support, 4 – temperature borehole. The model area is split into horizontal and vertical blocks with a spacing of 1 m.

Table 1. Boundary condition of the 3rd type outside the building

Month	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
T_{air}	-42.0	-35.6	-22.0	-6.8	6.2	15.6	18.8	14.9	6.1	-7.9	-28.2	-39.5
α	0.66	0.66	0.62	1.14	3.63	3.60	3.51	3.44	3.37	2.13	0.94	0.81

Note: T_{air} – atmospheric air temperature, °C; α – coefficient of heat exchange between the ground surface and the atmosphere, $\text{W}/(\text{m}^2 \cdot ^\circ\text{C})$.

perature borehole is 8.5 m away from the left side of the model area and 0.5 m from the support of the gas pipeline with a depth of 10.5 m. The design of the computational model area is shown in Fig. 1.

Boundary conditions. Boundary conditions of the second type with zero heat flow were set at the lower and lateral boundaries of the model area. At the upper boundary, first-type (under the building) and third-type (outside the building's contour) boundary conditions were set. Thus, a constant temperature of 20°C was set at the boundary under the building (from 0 to 6 m). Outside the building (from 6 to 24 m), the temperature of atmospheric air and the heat exchange coefficient between the surface and atmospheric air were set (Table 1).

The air temperature shown in Table 1 was adjusted for global warming from the start of the modeling in the form of a linear temperature trend equal to 0.05°C/yr.

Initial conditions. The initial constant temperature of permafrost equal to -3.0°C was taken for the entire model area.

Soil characteristics. The soil was a sandy loam with constant properties throughout the calculation area. Thermal conductivity was 1.28 $\text{W}/(\text{m} \cdot ^\circ\text{C})$ for the thawed soil and 1.65 $\text{W}/(\text{m} \cdot ^\circ\text{C})$ for the frozen soil; heat capacity was 599 $\text{W} \cdot \text{h}/(\text{m}^3 \cdot ^\circ\text{C})$ and 458 $\text{W} \cdot \text{h}/(\text{m}^3 \cdot ^\circ\text{C})$, respectively. The latent heat of phase transition was 25 250 $\text{W} \cdot \text{h}/\text{m}^3$, and the temperature of the beginning of the phase transition of water was -0.1°C.

Gas pipeline support. The support consisted of one steel pile with an outer diameter of 0.3 m. The pile was driven into the permafrost soil to a depth of 10 m. The vertical load on the top end of the pile was 200 kN.

Temperature control borehole. The temperature borehole was located 2.5 m from the building and 0.5 m from the gas pipeline support. Temperature sensors were installed at depths from 0.5 to 10.5 m with 1-m spacings.

The simulation results – temperature values at the loci of temperature measurement sensors (8.5 m from the left edge of the model area to a depth of 10.5 m) after 10, 11, ... 15 years from the start of the simulation are given in Table 2.

Verification of the extrapolation method for temperature forecasting

Data on model soil temperatures calculated by the extrapolation method with a help of macro A (see Supplementary materials) and by the numerical modeling based on the results of field observations and the WARM program [Emelyanov *et al.*, 1994] are given in Table 3. From the presented material, it is seen that the forecast error (according to macro A) within a 1-year-lead time does not exceed 0.2°C. However, with an increase in the lead time, it increases almost linearly. Thus, at the lead time of 2 years, the error exceeds 0.3°C and at the lead time of 4 years, it approaches 0.6°C. Therefore, for geotechnical monitoring, a 1-year-long lead time is preferable, which allows for a satisfactory match between mathematical modeling data (WARM program) and the extrapolation forecast method (macro A, see Supplementary materials).

Example of a pile foundation stability prediction

A stability forecast was made for the gas pipeline support (Fig. 1). To reduce the length of the paper, the forecast was first made for only one year (1992). The forecast for the subsequent years was carried out similarly.

The initial parameters were the data specified in the formulation of the temperature field modeling problem, as well as the temperature fields themselves obtained during numerical modeling using the WARM program (Table 2). The calculation was performed sequentially using macros A, B, and C (see Supplementary materials). The input parameters of the subsequent macro were the output parameters of the previous macro. The results of predicting the stability of the pile foundation are given in Table 4.

From the analysis of the data in Table 4, it follows that the pile support in all months of 1992 has safety factors greater than one and, therefore, remains stable. Moreover, the observer comes to this conclusion a year earlier, i.e. in 1991. This is very important, because had the outcome been different, a year remains in reserve to correct the situation (e.g., via artificial cooling of the support base).

Table 2. **Results of numerical simulation of the soil temperature regime at the loci of temperature sensors using the WARM program**

Y, m	Ground temperature by month, °C											
	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
1	2	3	4	5	6	7	8	9	10	11	12	13
<i>1990</i>												
0.5	-9.03	-9.72	-8.05	-6.1	-2.33	-0.1	2.58	10.31	8.66	4.23	-0.1	-0.1
1.5	-0.1	-0.1	-0.1	-3.1	-1.76	-0.28	-0.1	-0.1	3.17	3.36	1.44	0.84
2.5	0.66	0.43	0.38	-0.1	-0.1	-0.1	-0.1	-0.1	-0.04	2.24	1.6	1.15
3.5	0.62	0.47	0.42	-0.02	-0.1	-0.1	-0.1	0.34	0.51	1.22	1.09	1.05
4.5	-0.10	-0.1	-0.1	-0.1	-0.1	-0.1	-0.1	-0.1	-0.1	-0.1	-0.1	0.63
5.5	-0.10	-0.1	-0.1	-0.1	-0.1	-0.24	-0.29	-0.21	-0.12	-0.1	-0.1	-0.1
6.5	-0.32	-0.31	-0.3	-0.34	-0.42	-0.49	-0.53	-0.48	-0.41	-0.37	-0.34	-0.32
7.5	-0.51	-0.49	-0.47	-0.5	-0.58	-0.65	-0.69	-0.67	-0.62	-0.58	-0.54	-0.51
8.5	-0.69	-0.66	-0.64	-0.64	-0.69	-0.76	-0.81	-0.81	-0.78	-0.74	-0.71	-0.67
9.5	-0.86	-0.83	-0.8	-0.79	-0.81	-0.86	-0.91	-0.93	-0.91	-0.88	-0.85	-0.82
10.5	-1.01	-0.98	-0.95	-0.93	-0.94	-0.97	-1.0	-1.02	-1.02	-1.01	-0.98	-0.95
<i>1991</i>												
0.5	-8.84	-9.72	-7.99	-5.93	-2.24	-0.10	2.89	10.37	8.91	4.37	-0.10	-0.10
1.5	-0.1	-0.10	-0.10	-2.91	-1.68	-0.27	-0.10	-0.10	3.92	3.68	1.57	0.90
2.5	0.77	0.52	0.46	-0.10	-0.10	-0.10	-0.10	-0.10	2.00	2.51	1.81	1.26
3.5	0.88	0.70	0.64	0.45	0.0	-0.01	0.21	0.54	1.23	1.65	1.49	1.18
4.5	0.58	0.50	0.46	0.34	0.06	-0.03	0.13	0.36	0.66	0.89	0.88	0.77
5.5	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10
6.5	-0.27	-0.14	-0.11	-0.13	-0.20	-0.27	-0.27	-0.23	-0.17	-0.12	-0.10	-0.10
7.5	-0.47	-0.36	-0.32	-0.32	-0.38	-0.44	-0.46	-0.44	-0.40	-0.36	-0.33	-0.31
8.5	-0.64	-0.56	-0.52	-0.49	-0.52	-0.58	-0.61	-0.61	-0.58	-0.55	-0.52	-0.49
9.5	-0.79	-0.74	-0.70	-0.66	-0.67	-0.70	-0.74	-0.75	-0.74	-0.71	-0.68	-0.66
10.5	-0.92	-0.89	-0.85	-0.81	-0.80	-0.82	-0.85	-0.86	-0.86	-0.85	-0.82	-0.8
<i>1992</i>												
0.5	-8.64	-9.72	-7.87	-5.87	-2.17	-0.10	3.20	10.43	9.14	4.50	-0.10	-0.10
1.5	-0.10	-0.10	-0.10	-2.83	-1.61	-0.24	-0.10	-0.10	4.32	3.93	1.64	0.94
2.5	0.83	0.55	0.48	-0.10	-0.10	-0.10	-0.10	0.98	2.30	2.65	1.91	1.33
3.5	0.97	0.77	0.69	0.50	0.45	0.22	0.31	0.88	1.45	1.81	1.63	1.32
4.5	0.67	0.57	0.52	0.46	0.36	0.19	0.23	0.56	0.94	1.17	1.18	1.050
5.5	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	0.21	0.45	0.59	0.63	0.60
6.5	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10
7.5	-0.30	-0.29	-0.28	-0.29	-0.32	-0.36	-0.37	-0.37	-0.34	-0.32	-0.30	-0.29
8.5	-0.47	-0.45	-0.43	-0.43	-0.46	-0.50	-0.53	-0.54	-0.52	-0.49	-0.47	-0.45
9.5	-0.63	-0.60	-0.58	-0.57	-0.58	-0.62	-0.65	-0.66	-0.65	-0.63	-0.61	-0.59
10.5	-0.78	-0.74	-0.72	-0.70	-0.70	-0.72	-0.74	-0.76	-0.76	-0.75	-0.73	-0.71
<i>1993</i>												
0.5	-8.50	-9.71	-7.86	-5.77	-2.07	-0.10	3.49	10.5	9.34	4.63	-0.10	-0.10
1.5	-0.10	-0.10	-0.10	-2.76	-1.54	-0.22	-0.10	-0.10	4.61	4.14	1.69	0.96
2.5	0.89	0.60	0.53	-0.10	-0.10	-0.10	0.57	1.15	2.51	2.76	1.97	1.37
3.5	1.10	0.87	0.79	0.60	0.53	0.51	0.70	1.11	1.59	1.89	1.70	1.37
4.5	0.93	0.81	0.74	0.67	0.60	0.57	0.61	0.82	1.04	1.25	1.24	1.11
5.5	0.55	0.49	0.45	0.42	0.33	0.29	0.31	0.41	0.54	0.66	0.69	0.66
6.5	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.1	-0.10	-0.10	-0.10	-0.10
7.5	-0.28	-0.27	-0.26	-0.27	-0.30	-0.33	-0.35	-0.25	-0.20	-0.16	-0.13	-0.11
8.5	-0.43	-0.41	-0.40	-0.40	-0.42	-0.47	-0.49	-0.43	-0.40	-0.36	-0.33	-0.30
9.5	-0.57	-0.55	-0.53	-0.52	-0.53	-0.57	-0.59	-0.58	-0.55	-0.53	-0.50	-0.47
10.5	-0.69	-0.67	-0.66	-0.64	-0.64	-0.66	-0.68	-0.69	-0.67	-0.66	-0.63	-0.61

End of Table 2

1	2	3	4	5	6	7	8	9	10	11	12	13
1994												
0.5	-8.38	-9.71	-7.80	-5.69	-1.98	-0.10	3.73	10.55	9.44	4.71	-0.10	-0.10
1.5	-0.10	-0.10	-0.10	-2.71	-1.43	-0.21	-0.10	-0.10	4.75	4.24	1.72	0.98
2.5	0.92	0.62	0.55	-0.10	-0.10	-0.09	0.70	1.19	2.62	2.84	2.02	1.41
3.5	1.14	0.91	0.82	0.63	0.56	0.55	0.81	1.17	1.68	1.98	1.78	1.44
4.5	0.98	0.86	0.79	0.72	0.66	0.63	0.70	0.88	1.15	1.36	1.37	1.23
5.5	0.60	0.55	0.51	0.48	0.45	0.38	0.39	0.49	0.74	0.88	0.93	0.89
6.5	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	0.02	0.34	0.42	0.47	0.48
7.5	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.14	-0.10	-0.10	-0.10	-0.10
8.5	-0.28	-0.27	-0.26	-0.26	-0.27	-0.29	-0.31	-0.33	-0.30	-0.29	-0.28	-0.27
9.5	-0.43	-0.42	-0.40	-0.39	-0.40	-0.42	-0.44	-0.46	-0.45	-0.44	-0.42	-0.41
10.5	-0.57	-0.55	-0.54	-0.52	-0.52	-0.53	-0.55	-0.57	-0.57	-0.56	-0.55	-0.53
1995												
0.5	-8.27	-9.70	-7.77	-5.62	-1.94	-0.10	3.98	10.6	9.53	4.79	-0.10	-0.1
1.5	-0.10	-0.10	-0.10	-2.65	-1.37	-0.12	-0.10	-0.10	4.86	4.34	1.76	1.00
2.5	0.96	0.65	0.57	-0.10	-0.10	0.31	0.76	1.22	2.71	2.90	2.06	1.44
3.5	1.21	0.98	0.88	0.70	0.61	0.68	0.90	1.24	1.76	2.05	1.84	1.49
4.5	1.10	0.97	0.89	0.82	0.75	0.74	0.82	1.00	1.23	1.43	1.43	1.29
5.5	0.83	0.77	0.72	0.68	0.63	0.61	0.62	0.70	0.81	0.93	0.98	0.94
6.5	0.46	0.43	0.41	0.39	0.34	0.30	0.30	0.34	0.40	0.47	0.51	0.51
7.5	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10	-0.10
8.5	-0.26	-0.25	-0.24	-0.24	-0.25	-0.28	-0.29	-0.29	-0.29	-0.28	-0.27	-0.26
9.5	-0.39	-0.37	-0.36	-0.36	-0.37	-0.39	-0.41	-0.42	-0.41	-0.4	-0.39	-0.38
10.5	-0.51	-0.49	-0.48	-0.47	-0.47	-0.48	-0.50	-0.51	-0.51	-0.51	-0.50	-0.49

Note: Y is the depth of the temperature sensor, m.

Table 3. Error of forecasting the ground temperature by the extrapolation method

Lead time	1 year			2 years			3 years			4 years		
	October 1, 1992			October 1, 1993			October 1, 1994			October 1, 1995		
Y, m	T_{mod}	T_{pr}	$T_{\text{mod}} - T_{\text{pr}}$	T_{mod}	T_{pr}	$T_{\text{mod}} - T_{\text{pr}}$	T_{mod}	T_{pr}	$T_{\text{mod}} - T_{\text{pr}}$	T_{mod}	T_{pr}	$T_{\text{mod}} - T_{\text{pr}}$
0.5	4.50	4.49	0.01	4.63	4.60	0.03	4.71	4.70	0.01	4.79	4.79	0.00
1.5	3.93	3.96	-0.03	4.14	4.21	-0.07	4.24	4.44	-0.20	4.34	4.64	-0.30
2.5	2.65	2.75	-0.10	2.76	2.96	-0.20	2.84	3.15	-0.31	2.90	3.32	-0.42
3.5	1.81	2.03	-0.22	1.89	2.37	-0.48	1.98	2.67	-0.69	2.05	2.95	-0.90
4.5	1.17	1.76	-0.59	1.25	2.54	-1.29	1.36	3.24	-1.88	1.43	3.87	-2.44
5.5	0.59	-0.10	0.69	0.66	-0.10	0.76	0.88	-0.10	0.98	0.93	-0.10	1.03
6.5	-0.10	0.10	-0.20	-0.10	0.30	-0.40	0.42	0.47	-0.05	0.47	0.63	-0.16
7.5	-0.32	-0.17	-0.15	-0.16	0.01	-0.17	-0.10	0.16	-0.26	-0.10	0.30	-0.40
8.5	-0.49	-0.38	-0.11	-0.36	-0.23	-0.13	-0.29	-0.10	-0.19	-0.28	0.02	-0.30
9.5	-0.63	-0.56	-0.07	-0.53	-0.43	-0.10	-0.44	-0.31	-0.13	-0.40	-0.20	-0.20
10.5	-0.75	-0.71	-0.04	-0.66	-0.58	-0.08	-0.56	-0.47	-0.09	-0.51	-0.37	-0.14
Average	-	-	0.20	-	-	0.34	-	-	0.44	-	-	0.57

Note: Y is the depth of the temperature sensor, m; T_{mod} and T_{pr} are model ground temperature based on field observations (WARM) and predicted ground temperature (macro A, Supplementary materials), respectively, °C; $T_{\text{mod}} - T_{\text{pr}}$ is the error of the prediction, °C (absolute units).

Table 4. The results of forecasting the stability of the pile foundation

January						July					
F_v	200	F_u	276.07	K_u	1.38	F_v	200	F_u	340.68	K_u	1.70
F_y	326.78	F_f	145.14	K_f	2.25	F_y	326.78	F_f	0	K_f	∞
February						August					
F_v	200	F_u	285.54	K_u	1.43	F_v	200	F_u	354.28	K_u	1.77
F_y	326.78	F_f	145.14	K_f	2.25	F_y	326.78	F_f	0	K_f	∞
March						September					
F_v	200	F_u	287.31	K_u	1.44	F_v	200	F_u	346.65	K_u	1.73
F_y	326.78	F_f	145.14	K_f	2.25	F_y	326.78	F_f	0	K_f	∞
April						October					
F_v	200	F_u	298.46	K_u	1.49	F_v	200	F_u	336.94	K_u	1.68
F_y	326.78	F_f	178.13	K_f	1.83	F_y	326.78	F_f	0	K_f	∞
May						November					
F_v	200	F_u	308.20	K_u	1.54	F_v	200	F_u	272.89	K_u	1.36
F_y	326.78	F_f	178.13	K_f	1.83	F_y	326.78	F_f	85.77	K_f	3.81
June						December					
F_v	200	F_u	329.11	K_u	1.65	F_v	200	F_u	264.88	K_u	1.32
F_y	326.78	F_f	85.77	K_f	3.81	F_y	326.78	F_f	85.77	K_f	3.81

Note: F_u , F_y , F_f are the bearing capacity of the pile, the force holding the pile from being heaved out, and the heave force, respectively, kN; F_v is the payload on the top of the pile, kN; and K_u , K_f are the coefficients of the reserve in terms of bearing capacity and frost heave, respectively, dimensionless.

CONCLUSIONS

The methodology for processing permafrost temperature monitoring data described in this paper is a rapid method that does not require a large amount of initial data, which is difficult to obtain using software packages for solving the Stefan problem with their calibration based on field observations. In this regard, the proposed method is of great importance for the operational control of permafrost temperature conditions. It enables early warning of structure failures and the adoption of management decisions to prevent them. This suggests the need to incorporate the proposed rapid method into geotechnical monitoring in the permafrost zone.

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SUPPLEMENTARY MATERIALS

The computer programs (macros) offered to the reader's attention are written in Visual Basic Application language using the Excel 2007 platform (VBA Excel). To view them in the Excel worksheet corresponding to the selected macro, click the "Developer" button on the command line and then the "Visual Basic" button. The programs are ready for practical use, for which the user must perform two very simple operations: fill in the yellow field on the Excel worksheet with input (source) data and click the "Run" button at the top of the worksheet. The calculation result will be displayed in the gray field on this worksheet.

Access to the programs is provided via the following links:
macro A – <https://disk.yandex.ru/i/pE3IAHgw31G5Lw>
macro B – <https://disk.yandex.ru/i/W0r5Z2i9guBQLA>
macro C – <https://disk.yandex.ru/i/Au9FRimdtks9uQ>

CHRONICLE

ON THE 90th ANNIVERSARY OF IGOR VLADIMIROVICH KLIMOVSKY

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On August 19, 2024, Igor Vladimirovich Klimovsky, a famous Russian permafrost scientist, leading researcher at the Melnikov Permafrost Institute, Siberian Branch of the Russian Academy of Sciences, a Candidate of Geographical Sciences, and author of major monographs and maps on permafrost, as well as historical and biographical books about permafrost researchers, celebrated his 90th birthday.

Keywords: *permafrost, geocryology, natural processes and phenomena, expeditionary research, permafrost researcher.*

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On August 19, 2024 Igor Vladimirovich Klimovsky, a Candidate of Geographical Sciences and a leading researcher at the Melnikov Permafrost Institute, Siberian Branch of the Russian Academy of Sciences, celebrated his 90th birthday. We warmly and heartily congratulate our dear colleague on his anniversary and wish him health, optimism, and success in his endeavors.

I.V. Klimovsky (Fig. 1) was born on August 19, 1934 in Yakutsk. In 1953, he graduated from the city's secondary school No. 9 with a silver medal and entered the Faculty of Geography of Lomonosov Moscow State University, where he specialized in the Geography of Northern and Polar Countries (Fig. 2). After graduating from one of Russia's leading universities, I.V. Klimovsky returned to Yakutsk. Since 1958, only one organization is mentioned in his employment book – the Permafrost Institute of the Russian Academy of Sciences, where he has been working for 66 years.

His entire subsequent creative life was inextricably linked with Yakutia and geocryological science. This remarkable devotion to the institute was undoubtedly rooted in his love for geocryology and the romance of scientific inquiry. During the first thirty years of his work at the institute, Igor Vladimirovich went on expeditions almost every year. The romance of challenging fieldwork captivated him, built his character, and, most importantly, shaped his understanding and vision of natural processes and phenomena, which could not be acquired in the quiet of his office. The geography of his field research is very impressive: the glaciers of the Suntar-Khayata Ridge, the Yakutian diamond-bearing province, Kotelný Island in the Laptev Sea, northern Transbaikalia, the

Western Sayan Mountains, the Cheremukhovskiy coal basin, the Baikal-Amur Mainline zone, the Buor-dakh mountain-glacial massif, and elsewhere... Based on the results of his field studies I.V. Klimovsky wrote dozens of scientific and technical reports and published more than 110 scientific papers and monographs, such as *Permafrost of the Stanovoy Highlands and the Vitim Plateau* [Nekrasov, 1967], *The Last Glaciation and Permafrost in the Southern Verkhoyansk*

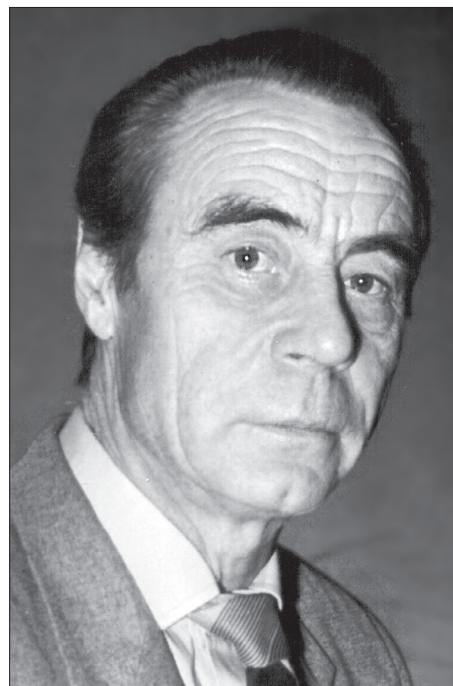


Fig. 1. Klimovsky Igor Vladimirovich.



Fig. 2. Group of students of the Department of Geography of Northern and Polar Countries, Faculty of Geography, Lomonosov Moscow State University.

In the first row, on the left is I.V. Klimovsky (Moscow, 1956).

Range [Nekrasov et al., 1973], *Permafrost Landscapes of Yakutia* [Fedorov et al., 1989], and *Permafrost of the Yakutian Diamondiferous Province* [Klimovsky, Gotovtsev, 1994].

With his active participation, the following fundamental collective cartographic and monographic scientific generalizations were compiled and published: *Permafrost along the BAM (Baikal-Amur Mainline)* [Nekrasov, Klimovsky, 1978], *Catalogue of Glaciers of the USSR* [Nekrasov, Sheinkman, 1981], *Permafrost Landscape Map of the Yakut ASSR* [Fedorov et al., 1991], and *Academic Geocryology in Yakutia* [Kamensky, 1997].

In recent years, unable to take part in expeditionary work for health reasons, he pays great attention to studying the history of geocryological science, writes essays, brochures, and books about prominent permafrost researchers. In this field, no less important for the development of any science, Igor Vladimirovich has proven himself a thorough, objective, and scrupulous researcher. Thus, in 2008, his book about the founder of the Yakutian geocryological scientific school, Academician P.I. Melnikov was published [Klimovsky, 2008] (Fig. 3). The book received many enthusiastic reviews from fellow permafrost scholars. The author deeply analyzed a huge factual material collected in the archives of Yakutsk, Moscow, St. Petersburg, Novosibirsk, and Tyumen.

I.V. Klimovsky devoted considerable attention to teaching and training future researchers. For 25 years, without interruption from work at the institute, he trained scientific personnel, lectured on general permafrost science to students of the Biological, Geographical, and Geological Exploration faculties of



Fig. 3. Delegates of the VIII Geographical Congress of the USSR (Kiev, 1986).

From left to right: Ph.D. S.E. Mostakhov, Academician P.I. Melnikov, and Ph.D. I.V. Klimovsky.

Yakutsk State University (YaSU), conducted student practices, and supervised a number of theses. Being the chairman of the State Examination Commission of the Faculty of Biology and Geography of YaSU, he received the academic title of associate professor in 1986. Many of his students became prominent scientists and specialists. Some of them are candidates of science and are now successfully working in the field of permafrost studies.

For his achievements in the development of scientific research, scientific-organizational, pedagogical, public, and outreach activities I.V. Klimovsky was awarded three medals, as well as the badges "Honored Veteran of the Siberian Branch of the USSR Academy of Sciences", "Veteran of Labor", etc. In 1985, he was awarded the honorary title "Honored Worker of the National Economy of the Republic of Sakha (Yakutia)", and in 2000, the Russian Geographical Society issued him a diploma for outstanding achievements in the field of geography.

It is worth noting that even the most dedicated scientists are not immune to the passage of time. We

would like to wish Igor Vladimirovich strength and perseverance in overcoming this inevitable challenge and many more years of active service to Russian permafrost science.

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