

GEOLOGICAL CRYOGENIC PROCESSES AND FORMATIONS

PARAGENESIS OF DEGRADATION AND AGGRADATION PROCESSES
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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-

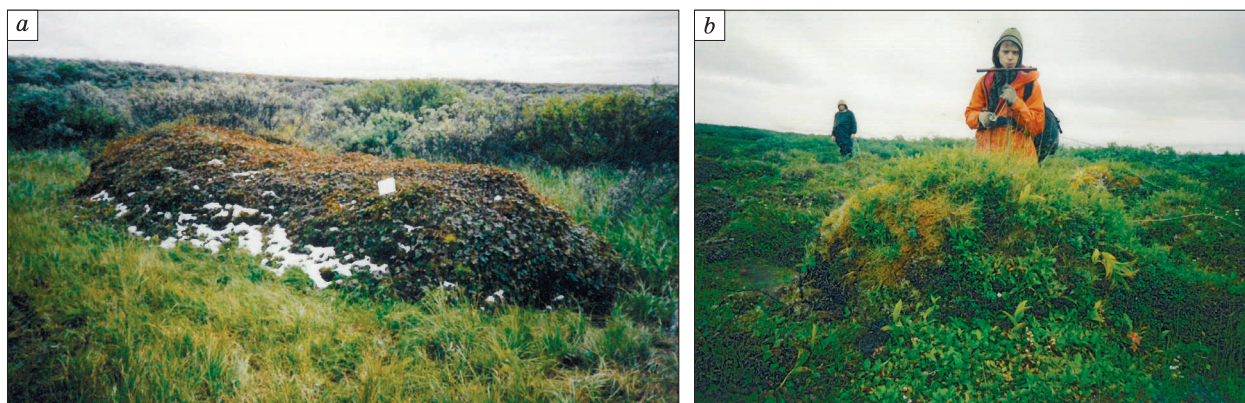


Fig. 1. Heave mounds in the (a) drainage hollow and (b) margin of the drained lake basin, southern part of the Bolshezemelskaya Tundra.

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

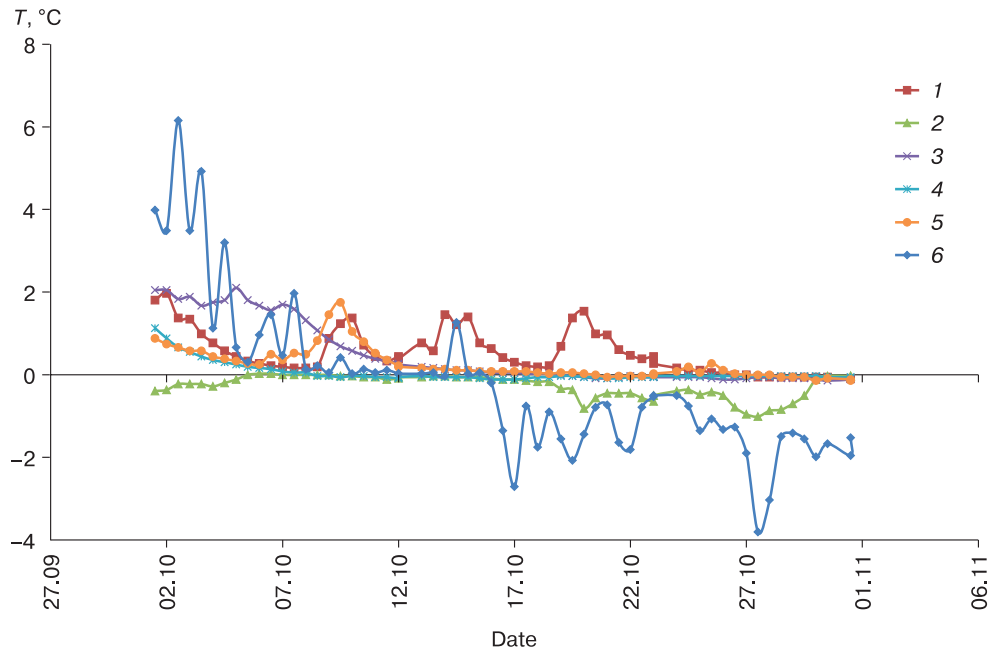


Fig. 2. Ground surface temperature under vegetation cover at 12:00 and 24:00 during the autumn inversion of air temperatures.

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.

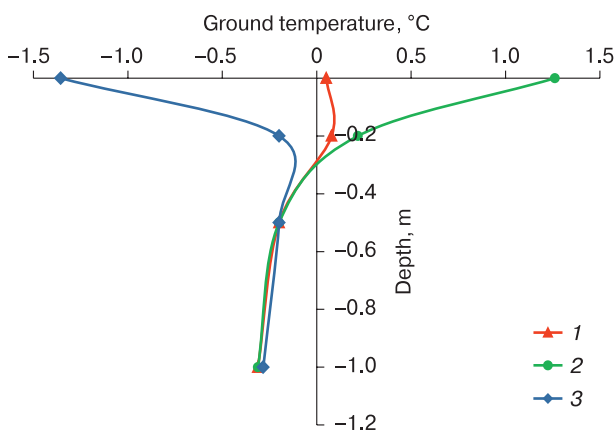


Fig. 3. Fast freezing of seasonally thawed layer at a local site.

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.

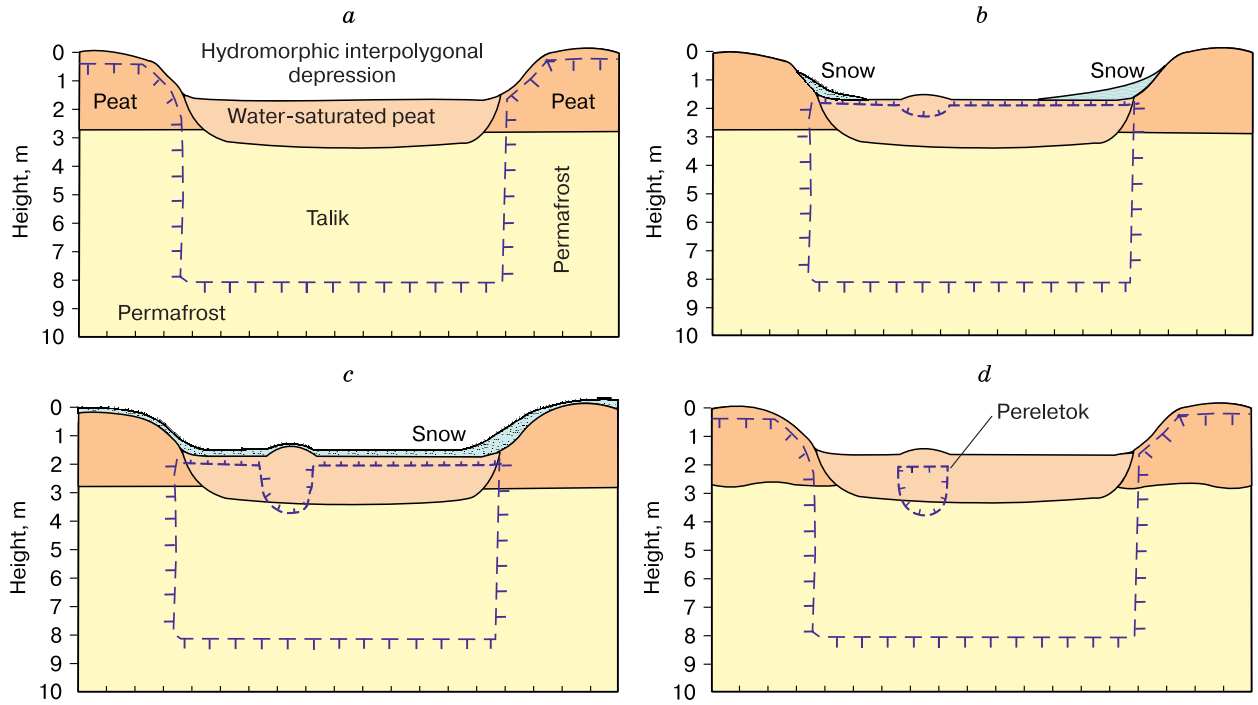


Fig. 4. Principal scheme of permafrost aggradation within the interpolygonal hydromorphic depression (vertical and horizontal scale ratio is 1:1).

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

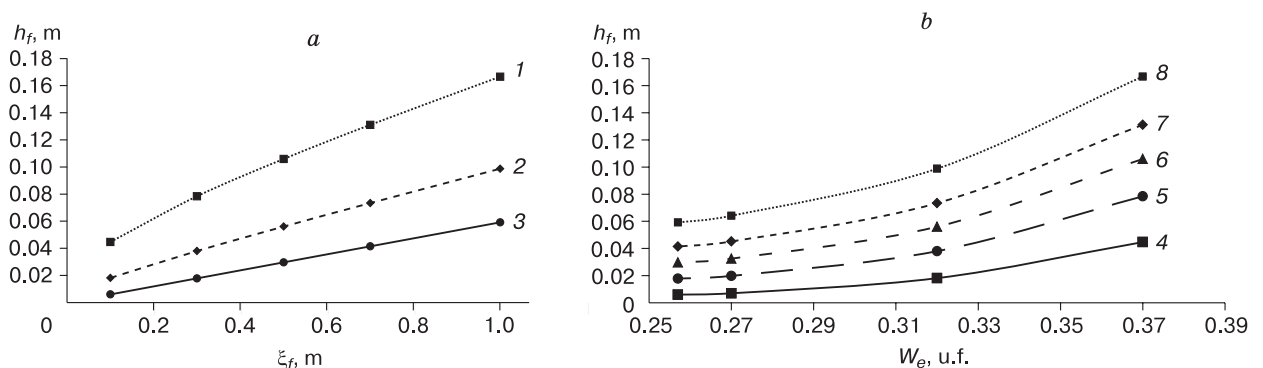


Fig. 5. Calculated dependence of frost heave value (h_f) on (a) freezing depth (ξ_f) and (b) water content of the loam under the peat layer without taking into account peat frost heaving.

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 snowdrift 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-

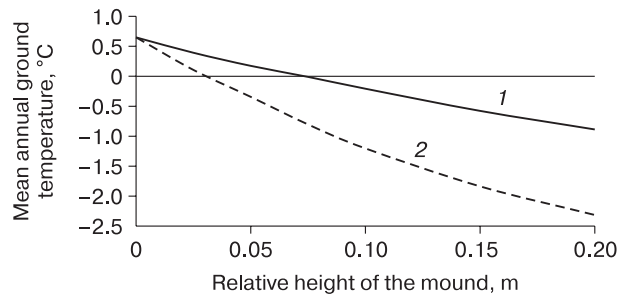


Fig. 8. Change in the mean annual ground temperature (at the base of the active layer) as a function of mound height.

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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