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TEMPERATURE-CONTROLLED FORMATION OF CRYOGENIC STRUCTURE OF PERMAFROST OF THE BOLŠEZEMELSKAYA TUNDRA

Cryogenic structure in Quaternary sediments have attracted attention and stimulated studies because a correct interpretation of the conditions under which it originated may afford wide possibilities of paleogeographic reconstructions. Among the numerous natural factors affecting the course of ice segregation in loose rocks, the temperature regime is one of the first in importance. The modern theory of moisture migration in freezing and frozen grounds and of the formation of ice inclusions suggests the following relation between temperature conditions and character of saturation with ice. The type of cryogenic structure is determined by rate of freezing or by temperature gradient. Formation of structure ice is more likely at low rates of freezing; the probability of development of relatively thicker ice interlayers increases at slower rate of freezing, or lowering of the temperature gradients (Šumskij, 1957; Dostovalov and Kudryavcev, 1967; Popov, 1967).

However, the interrelation between these two significant features and their comparison constitutes a problem that occupies a rather modest place in the modern geocryology for all its abundant materials on both cryogenic structure and particularly the temperature field of frozen rock. The reasons of this state of the question are clear in their general aspects. These are, in the first place, the difficulties of a methodical nature, necessitating operation with such temperature characteristics as have been obtained for the date of investigation. Temperature distribution in frozen rock and the rate of freezing and condensation reflect, in a large measure and especially in the upper part of the section, the effects of recent climate and of the existing landscape components. Whereas cryogenic structure is, as known, a more stable, inert system freguently associated with the natural conditions of previous stages of frozen rock development. Difficulties of another nature come from the complicated organization of investigation of the regime over the temperature field, without which it is impossible to acquire any data on the rates of condensation or on the temperature gradients, while sections and structure are more easily available for study.

One of the first attempts to establish a qualtiative relation between temperature, its changes, and ice content of the grounds under natural conditions was undertaken by Tyutyunov (1951). His experimental observations concerning the role of temperature and moisture regime in the process of freezing of a seasonally thawed layer showed in the section changes in moisture content during the whole period of freezing and alternation of moistured and desiccated seams depending on intensity of freezing. Unfortunately, as Tyutyunov failed to describe the type of the cryogenic structure originated, the possibility is rather reduced of using his results to evaluate the effect of the qualitatively estimated rate of up-freezing upon the formation of a certain cryogenic structure.

Further investigations of these problems dealt mainly with the relation between ice content and mean annual temperatures of frozen rock, as well as mean temperatures over longer periods of time. Comparisons of that type are all the more easier as these characteristics of the temperature field are more readily obtainable as compared with data on temperature gradients. The relation between mean annual temperature and ice content for the Bolšezemelskaya Tundra was traced by Bakulin (1958). Comparing these two phenomena he determined in detail the qualitative data (given in the book on the volume of non-frozen water, cement-ice, and ice content due to ice inclusions), Bakulin reached the conclusion that higher ice content is associated with a more rigorous temperature regime. Like in Tyutyunov's work, iciness is given as a qualitative characteristic of ice content in ground without any analysis of the relationship between temperature and cryogenic structure.

A theoretical evaluation of the possibility of development of a certain type of cryogenic structure depending on the values of annual thermal cycles and mean annual temperatures was ventured by Kudryavcev (1961). Thickness of ice schlieren was regarded as a function of mean annual temperature; hence the formation of certain types of cryogenic structure was considered as being dependent on the temperature-permafrost zonality. According to Kudryavcev, these conceptions are applicable to syngenetic stratified structure developing under conditions of low (below $-3°$) temperatures. These conceptions are however only of a general nature. The paper does not provide any examples of specific links or regional comparisons between various types of a syngenetic structure in different temperature-permafrost zones.

In a somewhat different context the relation between temperature regime and cryogenic structure has been discussed by Žestkova (1966), who holds that ice content in frozen ground horizons at different depths depends on climatic changes over lengthly periods. The existence of horizons with increased ice content (15 to 20%) 30–40 m deep in the series of the epigenetically frozen marine-glacial loam in the Bolšezemelskaya Tundra as described by Žestkova, made her conclude to the existence of a relationship between these horizons and the short-period variations of temperature regime of the surface. Žestkova writes: "... the icy horizons noted in the frozen series at different depths are presumably a result of the short-term climatic changes during individual periods, or individual stages, against the background of long-lasting climatic cycles (for example, those related to interglacials and glaciations 10 to 23 millennia)" (1966, p. 85). This theoretical inference, however, is not supported by an analysis of any specific material. The writer failed to give any regional paleogeographic interpretation of the conditions of development of cryogenic structure in loose deposits in the Bolšezemelskaya Tundra.

An essentially different approach to temperature conditions during formation of cryogenic structure is suggested by Popov (1967). He was the first to study the relation between types of cryogenic structure and values of temperature gradient, which relation is, in fact, the sole characteristic of the temperature field, that determines the physical nature of the ice segregation. According to Popov, the temperature gradient is a criterion for distinguishing between the horizons of active and passive cryolithogenesis. The intensity of ice segregation, the type of cryogenic structure and the dimensions of its components depend on the value of the temperature gradient.

As a result of numerous investigations in the northeastern part of the European North, abundant data on the temperature field and the cryogenic structure of frozen rocks have been accumulated thus permitting to analyse the relation between these characteristics. As noted above, such attempts had been previously made by Bakulin (1958) and Žestkova (1966) whose conclusions were based on data obtained by studying perennially frozen rock in the Bolšezemelskaya Tundra.

Rocks of various origin, such as marine-glacial, bog-lacustrine, alluvial and eluvial, constitute the complex of the Quaternary deposits occurring throughout the eastern part of the Bolšezemelskaya Tundra and reaching thicknesses of 80 to 100 metres. The highest ice content is in bog-lacustrine deposits in which cryogenic structure is traced down the whole thickness of their section. At the same time, owing to their bedding conditions the temperature field is extremely inconstant in extension, with sharp changes in its parametres over short distances. Because of these two features bog-lacustrine deposits provide the most interesting regional example of the effect of temperature regime of frozen rock on cryogenic structure.

Bog-lacustrine deposits crown the section of Quaternary rocks in the depressions of the watershed-divide landscape. They occur, as a rule, on marine-glacial loam being themselves predominantly composed of loam and peat. The contact of the deposits of these genetic types is characterized by highly gradual transition. The uppermost dark-grey or dark-blue, compact, marine-glacial loams of a heavy mechanical composition often enriched with clastic material, pass into lake loam. The latter, of a greenish grey colour, is characterized by a high content of silt (the percentage of silty fraction is 60 to 70%) and a very low one of debris (up to 5%). Its distinctive feature is that it contains plant remnants in the form of stems and pieces of wood with peaty interlayers which are particularly numerous in the upper part of the section where the loam changes into peat. The gradual transition from one lithologic-genetic type into another provides evidence of continuous sedimentation under the prevailing subaqueous conditions, first in sea lagoons and then in relict lakes that shoaled and progressively turned into bogs.

The thickness of lacustrine loam varies widely from 1–2 m to 8–10 m, reaching its maximum in the central parts of the depressions. The thickness of peat is variable, ranging from 0.5–1.5 m to 3–4 m, and still more so is ist distribution over the area. Its horizon does not form a continuous mantle but is dissected into both high and low elements of the polygonal relief. The variability in thickness and occurrence of bog-lacustrine deposits is an indication of different duration of the lake regimes, and of heterogeneity of local conditions in the bogs, probably due to the permafrost polygonal relief.

Lesser thickness of lake deposits corresponds to rapid completion of sedimentation (emptying of lakes). Frost-caused cracking of the bog surfaces resulted in various possibilities of peat accumulation within the polygons and fissures. Further changes in these conditions were introduced by thermokarst. All these factors combined to create a mixed, mosaic-like distribution of peat.

On the whole, high average ice content persisting down to the base of the section and amounting to 20 to 40% is characteristic of the peat-loam lake deposits. In individual horizons it increases up to 60 to 70%. Relatively uniform cryogenic structure corresponds to the equal saturation with ice. Streaky structure predominantes; the massive one, of minor significance is generally found in the sandy gravelly interlayers. In some cases, porphyritic structure is noted in individual horizons. Conspicuous among schlieren structures, are the stratified ones occurring in the majority of the sections (75–80%). Reticulate structure is much rarer. The ice schlieren vary in thickness from 0.2–0.4 cm to 1.5–2 cm and are spaced 1 to 7 cm. According to gradation by thickness of ice schlieren and by their spacing — as suggested by B. I. Vtyurin[1] – fine and medium-sized schlieren, medium-stratified, rarely medium-reticulate structures are the predominant types. In some instances, either thickening of schlieren (spacing reduced up to 0.2–1 cm), or, on the contrary, occurrence of individual thick ice seems (6–10 cm). These are, as a rule, portions near contacts of loam with peat, of sandy-loam and sand seams with loam, seasonally thawing layer with perennially frozen rock, or else horizons of heavily peated loam.

High ice content and its consistent value, the types of structure and the dimensions of its components throughout the whole section of the bog-lacustrine deposits are a characteristic feature of this complex and an indication of syngenetic conditions of freezing (Popov, 1961). It is difficult to picture the formation of a cryogenic section of such type with the freezing of very wet bog-lacustrine deposits after the completion of sedimentation. In that case, considerable enrichment with ice of the upper horizons of the 6 to 10 metre section may certainly have taken place as a result of moisture redistribution with freezing.

Such is, in fact, the character of ice content found in epigenetically frozen marine-glacial loam both on high watershed divides and in places where they underlie lake loam. Large schlieren 10 to 15 cm thick, occassionally up to 20–30 cm, can be found alongside medium-sized schlieren stratified structures in their uppermost part (down a depth of about 10 m). At the same time, average ice content of the deposits decreases down to 10–15% with its very uneven distribution in the section: heavily ice-saturated horizons where ice content reaches 30 to 70% over an extent of 0.3–0.5 m give place to ground interlayers 0.5–1.0 m thick, with massive structure and low ice content (5 to 10%). Bakulin (1958) who studied in detail the ice content of the upper part (down to 15 m) of the section of marine-glacial deposits on high divides where they lie almost directly beneath the surface (under the layer of cover loam) notes a decrease in ice content with depth, which he fixes clearly from a depth of 4 to 5 m (hardly ever more than 5 to 7% at a depth of 13 to 15 m). In some sections, the same kind of regularity was observed by Žestkova (1966) though already as a result of investigations of the entire series of marine-glacial deposits.

Ice veins 1.5 to 2.5 m thick and 0.6 m wide were found in the bog-lacustrine deposits. The veins intersect the peat and the upper part of the loam. In shape they are rather regularly wedge-like having 0.2–0.3 m in width, diminishing in cross-section with transition from peat to loam. They lie at the base of the seasonally thawing layer or near it, at 0.4–0.7 m depths. As a rule, a layer

[1] The gradation is given in "Methods of a complex permafrost-hydrogeological and engineering-geological survey", Moscow University Press, 1970, p. 117.

of clear ice 10 cm thick is found directly above the vein, the layer containing a large number of peat inclusions oriented horizontally and giving the ice a horizontal stratification. Seams of textural milky white ice with peat remnants, 3–6 cm thick, were noted inside the body of the vein at intervals of 0.2 to 0.5 m. These seams differ from the basic mass of ice characterized by vertical striation due to oriented air bubble inclusions, to peat and loam particles, and to minute jointing. Horizontal ice schlieren of equal thickness in the peat and loam correspond to the interlayers of textural ice of the vein. This peculiarity of vein structure and the "cap" of glassy ice above indicate that thickness of the veins increased chiefly due to frontal growth and the veins are therefore of a syngenetic origin (Popov, 1967).

Some ideas concerning the paleogeographic conditions at the time of formation of the bog-lacustrine complex are suggested in the works by Popov (1961), Danilov (1962), and Koniščev (1965). The accumulation of these deposits started at the time of the Polar Sea basin regression in the Upper Pleistocene. In the process of regression the lagoon regime changed into a lake regime in the accumulative relief depressions of the continental shelf. As the lakes were being filled with sediments and emptied, they shoaled and turned into bogs where peat accumulation was proceeding. The climatic background at the time of development of the bog-lacustrine deposits, i.e., the beginning of the Upper Pleistocene–the Holocene, was not uniform. As is known during that period recurrent worsening of climate took place at the end and at the beginning of the Upper Pleistocene, which coincided with the Zyryansk and Sartansk glaciations, alternating with warmer periods in the middle of the Upper Pleistocene and in the Holocene (the Karginsk interglacial and the warming up of the climatic optimum).

The beginning of the continental stage in the development of the eastern part of the Bolšezemelskaya Tundra, when it was a shore of the Polar Sea, coincided with the time of the Zyryansk glaciation in the Urals. The climate of that time was characterized by greater frigidity than the present-day climate, and resembled that of the Novaya Zemlya and the Yamal of our days. Under those conditions, intensive up-freezing started not only in the deposits emerged from the sea, that constituted high watershed divides, but also in the bottoms of relatively shallow relict lakes. Up-freezing of bottom deposits down to a depth of tens of metres in shoals is a widely known case on the coast of the existing Arctic seas in Northeastern Eurasia (Grigorev, 1966). Thus sedimentation in lakes was accompanied by simultaneous up-freezing, a fact substantiated by the nature of the cryogenic structure described above.

The temperature rise in the Karginsk period proceeded against the background of increasing continentality due to a considerable retreat of the shoreline northwards. The tundra landscape of high watershed divides changed into forest-tundra landscapes with understocked spruce-birch and pine forests. The frozen rock bedding became considerably deeper, up to 8 m according to Koniščev. A different picture was observed in the depressions of the divide relief. The effect of rising summer temperatures was reduced by the diminished thickness of the snow cover. Under present-day conditions, regions with extra-continental climate (the Transbaikal) and with a snow cover of negligible thickness are known to be characterized by the most rigorous permafrost conditions (Kačurin, 1950) due to the high thermal capacity of the surface water horizon, of the grounds heavily wetted in summer and to their high thermal conductivity when frozen in winter. Therefore the summer warming-up is here hampered and the winter cooling-off proceeds without hindrance. Owing to this, the frozen series in the bog-lacustrine depressions could not only persist, but also increase its thickness as sediments accumulated. According to Koniščev the intensification of solifluction processes on the slopes of the watershed divide hills due to the increased thickness of the seasonally thawing layer accelerated the filling of lakes, their shoaling, and the change of lake sedimentation into bog sedimentation by the end of the Karginsk period.

The next stage was the temperature fall at the end of the Upper Pleistocene, the time of the Sartansk valley-and-mountain glaciation of the Urals. The landscape of the Bolšezemelskaya Tundra was characterized by a common occurrence of *Sphagnum* and *Bryales* bogs. The change of the lake regime into bog regime against the background of an increasingly rigorous climate induced intensive cooling of the upper rock horizons, redoubled by the high thermal conductivity of the heavily icy loam and peat. The result was the frost-caused cracking of rock and the formation of a polygonal pattern with veined ice lattice-work. The slow rate of sedimental accretion, especially of peat under that rigorous climate accounts for the small dimensions of the ice veins.

The temperature rise in the Holocene, just like in the Karginsk time, seems to have proceeded against a background of increased continentality, which phenomenon can according to M. Gričuk and V. Gričuk (1960) – take place at the end of an interglacial epoch. The tundra of high divides gave place to birch forests in the first phase, and to spruce forests mixed with pine and larch in the second (Smirnova, 1959). According to Neištadt's data (1957) forests extended during the climatic optimum as far as the coast of the Barents Sea.

In that time the lowlands were *Sphagnum* and *Bryales* bogs in which intensive peat accumulation took place. The development of a polygonal vein-ice pattern differentiated the microrelief of lowlands thus determining the unevenness of snow accumulation and its appreciable thickness within the depressed portions of the relief. This may have contributed in winter to preservation of the seasonally thawing layer within them as well as a rise of its temperature regime and eventually to its increase in thickness and to development of thermokarst. The latter was intensified by the high ice content of the deposits. Local development of thermokarst caused the differentiation of conditions in permafrost-facies in such depressions in which thermokarst taliks had nearby low-temperature rocks, where sedimentation was accompanied by freezing. The thickness of the Holocene peat varying widely over short distances (from 3 to 0.5 m, even its absence) provides evidence of uneven peat accumulation.

In localities where perennially frozen rock was lying near the surface, the thick moss cover and high moisture content (200 to 500 per cent) of peat in the seasonally thawing layer provided for their being safely isolated from thawing and warming up. The present-day analogues of the Holocene peat-bogs in the Bolšezemelskaya Tundra strongly suggest the possibility of permafrost rock existing under such conditions. In the zone of the present forest-tundra in the northeast of the European part of the Soviet Union, 200 to 250 kms southwards of the southern boundary of the Bolšezemelskaya Tundra, frozen rock development is associated with localities of bog-lacustrine deposits, and the rock lies at a depth of 0.5 to 0.7 m (just like in the Bolšezemelskaya Tundra), temperature is $-1°$ to $-1.5°C$, and the surrounding high divides are built up of thawed rock occupying 75 per cent of the area (Ivanova, 1964).

The time following the climatic optimum was characterized by worsening climate, forest retreat southwards and the forest landscapes being replaced by tundra landscapes with frozen rock lying not deep on hills and ridges. Conditions similar to the recent ones were setting in and peat accumulation was being completed. At the same time, differentiation of permafrost conditions intensified by the further development of thermokarst due to thick accumulations of snow in the ever deepening inter-polygons of the polygonal relief.

Thus, at the continental stage of development of the Bolšezemelskaya Tundra, the frozen rocks of the watershed divide depressions and basins had, as a result of the specificity of their landscape conditions, a certain thermal inertia in respect to the climatic changes, and were characterized by such basic permafrost indices (including temperature features) as were stable in their general aspects, or at least sligthly variable, if at all. That was facilitated at different stages by water (lakes, swamps) and by the presence of moisture-saturated moss and peat covers which provided safe isolation against intensive warming-up of the grounds by the rise of summer temperatures due to increasing continentality. On the other hand, the increased thickness of the snow cover modified the severity of the cooling stages. Hence the observational data relative to the present-day temperature regime of frozen bog-lacustrine deposits and of the upper part of the underlying marine-glacial loam might prove helpful in interpreting the conditions under which their structure originated.

Permafrost conditions in general, and the temperature field of the frozen rocks in the Bolšezemelskaya Tundra in particular, vary widely throughout the area. This is especially true of the divide depressions within which permafrost contrasts reach their maximum. The reason for this is in the development of a legible polygonal pattern with the polygon centres mounting 4–5 m above the interpolygon troughs. Polygonal relief introduces sharp difference in the conditions of snow accumulation, namely from 0.2 to 0.35 m on the polygons up to 2–2.5 m in the interpolygons. The latter, in particular are swamped and their centres are frequently occupied by lakes. Therefore, perennially frozen rocks within them lie at depths of 10 to 30 m being occasionally altogether absent whereas on the polygon blocks they are observed at depths of 0.3 to 0.6 m in the peat and 0.8–1.0 m in the loam. Consequently ground temperatures also display a wide range of fluctuations. In the pseudo-taliks of the interpolygons, the mean annual temperature in the base of the layer of annual variations is of 0.3 to 0.5°, and that of the frozen rocks is not lower than —0.2° to —0.3°. Some interpolygons are found to be insufficiently deepened as compared with the polygons in which the thickness of snow cover is relatively small (ca 0.5—0.7 m); as a result, they are not yet affected by thermokarst processes, and the frozen rocks within have rather low temperatures of —1.5° to —1.7°.

Likewise a wide temperature range is observed within the polygons from the lowest values in the east of the Bolšezemelskaya Tundra, which are —2.5° to —3°, up to —1° to —1.5°. Alternation of one type of temperature field with another is most frequent, mamely at distances of 15–20 to 100–200 m, according to the dimensions of the polygonal relief component. This permits to use the observational data, collected in relatively small areas that represent the entire regional range of fluctuations of the temperature field.

Comparison of the temperature regime in the layer of annual variations in polygons and in interpolygons (Tab. I, Fig. 1: E) shows that within the most homogeneous lithologic composition (the section predominantly loamy, Fig. 1: A) the temperature regime of the grounds experiences abrupt changes. Thawed grounds turn into frozen grounds over a distance of about 360 metres, and the mean annual temperature at the base of the annual fluctuation layer varies 0.3—0.6° and —2.2 to —2.5°. The formation of taliks in the interpolygons is, as noted, due to the warming effect of the thick snow cover

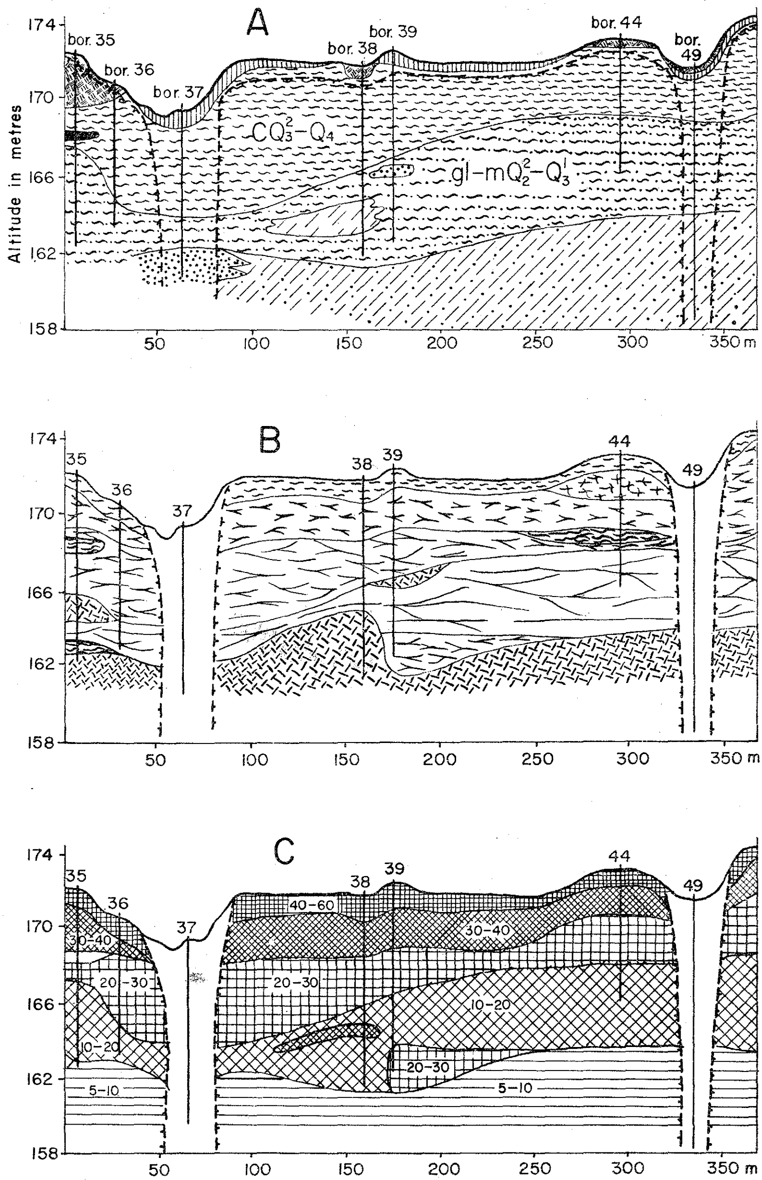
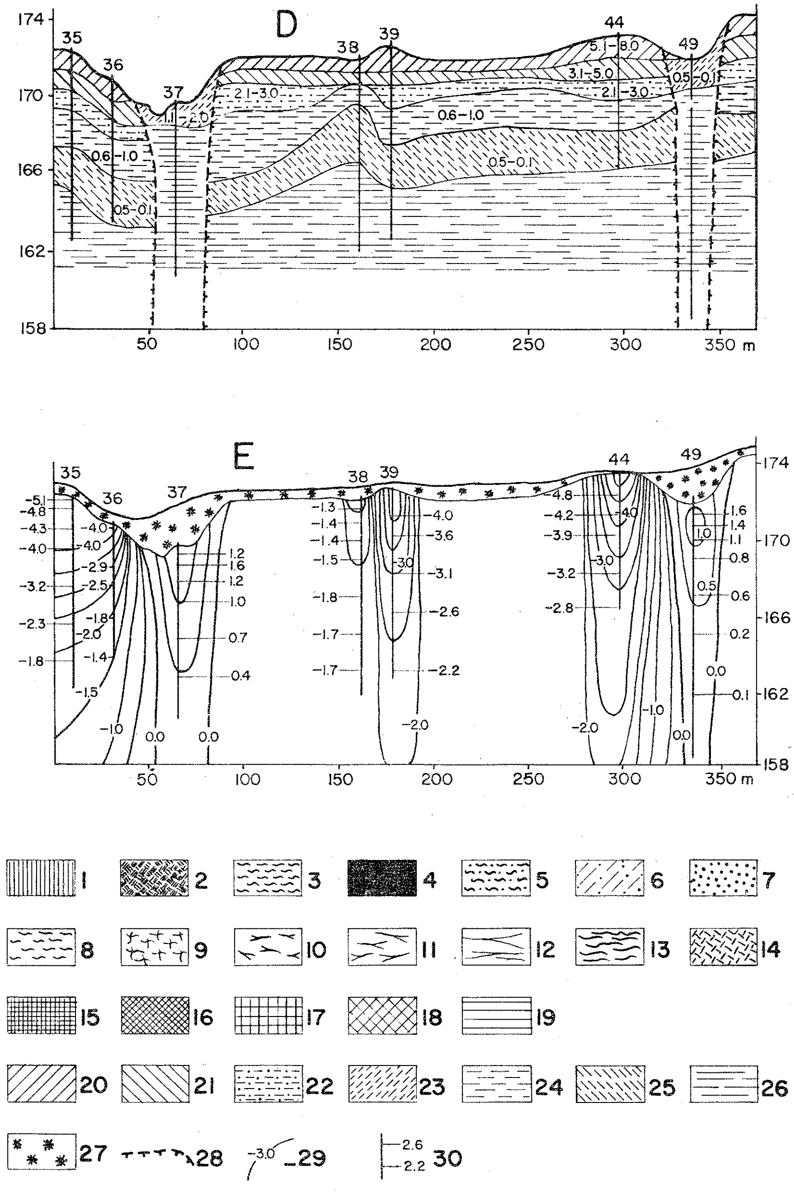


Fig. 1

A. Lithologic-genetical types of rock; 1. cover loam; 2. peat; 3. lacustrine loam; 4. lacustrine sandy loam; 5. marine-glacial loam; 6. marine-glacial sandy loam; 7. marine-glacial sand

B. Cryogenic structure: 8. fine-schliern densely stratified; 9. fine-schlieren densely reticulate; 10. fine- and medium-schlieren densely stratified; 11. medium-schlieren medium-stratified; 12. medium-schlieren remotely stratified; 13. coarse-schlieren remonte- and medium- stratified; 14. massive structure

C. Ice-content: 15. 40–60%; 16. 30–40%; 17. 20–30%; 18. 10–20%; 19. 5–10%



D. Temperature gradient: 20. 5.1°/m–8.0°/m; 21. 3.1°/m–5.0°/m; 22. 2.1°/m–3.0°/m; 23. 1.1°/m–2,0°/m 24. 0.6°/m–1.0°/m; 25. 0.5°/m–0.1°/m; 26. below 0.1°/m

E. Thermoisopleths of mean annual temparature: 27. snow cover; 28. frozen rock bounds; 29. thermoisopleths of mean annual temperature; 30. mean annual temperatures according to the data of regime observations in boring holes

(2–2.4 m), which completely isolates the interpolygonal grounds from the winter cooling.

An analysis of the mean annual temperatures of the polygons shows that their values are not entirely determined by the conditions of heat exchange at the surface. Three polygons taken for comparison (Borings 35, 39, 44) had equal thicknesses of the snow cover amounting to 0.2 or 0.35 m and a similar type of vegetation cover namely hillocky-spotty dwarf-shrub and wild-rosemary tundra. However, most diversified is the composition of the upper part of their section: Boring 35 shows a peat polygon whose upper 2.9 m consist of peat; in Boring 44 peat is 0.3 m in thickness (the polygon has a mixed peat-mineral substratum); peat is absent from the mineral polygon of Boring 39, and the cover loam extends to a depth of 0.75 m. Downward, the polygons are built of lake loam down to depths of 4.8 m (Boring 35), 3.8 m (Boring 44) and 5.5 m (Boring 39) respecively; farther downward, the lake loam passes into marine-glacial loam. Except for the half-metre interlayers of lacustrine sandy loam (Boring 35: 4–4.5 m) and the sand including pebbles in the upper part of the section of the marine-glacial series (Boring 39: 6–6.5 m), the uniformity of the section, loamy in mechanical composition, is entirely undisturbed. With such a lithology, the peat polygons should be naturally expected to exhibit the minimal temperatures, and the maximal ones in the mineral polygons. In reality however the lowest mean annual temperature, amounting to $-2.5°$ at the base of the annual variation layer, is observed in Boring 44 (peat-mineral polygon); it is $-2.2°$ in the mineral polygonal grounds, and $-1.8°$ in the peat polygon. And observations of the regime show that down to a depth of 5 m, formation of the temperature field proceeds in conformity with heat exchange at the surface and with the thermal-physical properties of the rocks. The lowest winter and summer temperatures and the minimum thickness of the summer thawing layer (0.5 m) correspond to the most "cold-loving" type of ground, which is peat whereas they increase successively in the peat-mineral and mineral polygons (Tab. I).

This regularity is disturbed at a depth of 7 m where the distribution of mean annual temperature is as follows: $-2.3°$ in the peat polygon, $-2,8°$ in the peat-mineral one, and $-2.5°$ in the mineral polygons. This phenomenon is related to lateral heat flows the action of which was noted by numerous investigators (Redozubov, 1946; Mukhin, 1964). The polygons under consideration are located at various distances in the region of a thermokarst lake occupying the central part of a watershed-divide depression. The peat polygon is a lake shore, and Boring 35 is drilled at a distance of 17 m from the water edge. Therefore heat exchange in the grounds of the upper part of the section is determined by superficial and lithologic conditions while

Table 1

Some characteristics of frozen rocks

Boring	Elements of polygonal relief	Height of the polygon (m)	Thickness of snow cover (m)	Depth of seasonal thawing (m)	Depth of annual oscillations of temperature (m)	Mean annual temperature (°C)									Mean summer temperature (June to August)							Mean winter temperature (December to March)						
						0.0 m	0.5 m	1.0 m	2.0 m	3.0 m	5.0 m	7.0 m	9.0 m	At the base of annual oscillations	0.0 m	0.5 m	1.0 m	2.0 m	3.0 m	5.0 m	7.0 m	0.0 m	0.5 m	1.0 m	2.0 m	3.0 m	5.0 m	7.0 m
35	Peat polygon	2.7	0.2–0.3	0.5	10–11	–5.4	–5.1	–4.8	–4,3	–4.0	–3.2	–2.3	–1.8	–1.8	0.4	–0.4	–1.3	–3.2	–4.2	–3.8	–2.8	–13.8	–11.6	–9.2	–5.7	–4.0	–2.6	–1.8
36	Slope of the same polygon		0.2	0.6	9–10	–4.4	–4.0	–3.5	–2.9	–2.5	–1.8	–1.4	–1.1	–1.0	0.9	–0.3	–1.3	–2.4	–2.4	–2.5	–1.6	–11.0	–8.6	–6.3	–3.7	–2.4	–1.4	–1.2
39	Mineral polygon	0.7	0.2–0.25	1.0	11	–4.5	–4.5	–4.2	–3.8	–3.4	–3.0	–2.5	–2.3	–2.2	2.9	–0.5	–1.8	–2.8	–3.7	–3.8	–3.2	–12.3	–9.6	–7.5	–4.6	–3.5	–2.5	–2.1
38	Interpolygon		0.6	1.0	8	–0.8	–1.3	–1.4	–1.4	–1.5	–1.8	–1.7	–1.7	–1.7	2.1	–0.1	–1.0	–1.6	–1.9	–2.0	–1.7	–3.7	–2.4	–1.6	–1.0	–1.2	–1.6	–1.6
44	Peat-mineral polygon	1.5	0.35	0.65	11	–5.2	–5.0	–4.8	–4.2	–3.9	–3.2	–2.8	–2.5	–2.5	1.8	–0.3	–1.2	–2.7	–3.5	–3.8	–3.3	–14.4	–10.9	–8.9	–5.9	–4.1	–2.6	–2.3

lower downwards the warming-up effect of the lake raises at a depth of 7 m the mean annual temperature by at least 0.5° as compared with the portions where lateral heat flows are absent or insignificant. The mineral and peat--mineral polygons (Boring 39 and 44) are situated at 80 and 215 m from the lake respectively. Interpolygons built of frozen ground of a rather low temperature, −1.7° are situated near the hole of the mineral polygon (Boring 38) while the peat-mineral polygon adjoins an interpolygonal depression where occurs very deep talik – unbottomed by boring (Boring 49). However, it is just in the latter case that the lowest temperatures of polygonal grounds (Fig. 1: E) are observed, a fact providing evidence of considerable differences in the intensity of lateral heat flows depending on the type of taliks. Even in cases of well-defined but dry interpolygons, their warming-up influence upon the low horizons of rocks is insignificant, though the lateral effect of the under-lake taliks should not be disregarded. It extends within a radius of up to several tens of metres.

The distribution of mean annual temperatures in the section of the polygons reveals their clearly displayed gradient character (Fig. 1: E, Tab. I). They vary from −5.0 to −4.0° at the base of the seasonally thawing layer to −1.4 to −2.5° at the depth of annual zero amplitudes and the range of their fluctuations within the interval of 0.5 to 5.0 m is from 12° to 17° up to 2.6 to 3.0°. With such an amplitude of temperature changes, even within its negative values, a great amount of heat is spent on phasal transitions due to the non-frozen water which fact preconditions a gradient distribution of temperature. According to B a k u l i n's data (1958), the amount of non--frozen water in loam is of 10 to 15% at a temperature of ca −2°.

A different character of temperature distribution in the section is observed in the interpolygon constitued by frozen grounds (Boring 38). With a rather low mean annual temperature at the base of the annual variation layer (−1.7°), nearly the same as in the peat polygon (−1,8°), a degradation dependence of the temperature curve is observed in it (Fig. 1: E), as noted for "run-off bands" by R e d o z u b o v (1946). Such a distribution is in conformity with the warming-up effect of the 0.6 m thick snow cover, the significance of which is decisive in the Bolšezemelskaya Tundra in respect to formation of an active layer either merging or not with perennially frozen rock. In the interpolygon in which Boring 38 is drilled, the warming-up effect of snow is attenuated by the 0.6 m thick peat horizon.

An analysis of the course of mean annual temperatures shows that the diversified, often variously oriented influence of heat exchange factors at the surface of the thermophysical properties of the grounds themselves, and of the lateral heat flows creates a most variegated picture in this respect (Fig. 1: E). The effect of the "cold" factors is counteracted by the "warm" ones, and *vice versa*, for example, the cooling effect of the thick peat horizon (2.9 m) and the warming influence of the under-lake talik (Boring 35), or of the warming through snow and cooling through peat (Boring 38).

Sharp variations of the mean annual temperature values occur at short distances both over the area and in the section. At the same time, an analysis of the ice content shows that its character is much more persistent (Fig. 1: B, C). This is evidenced both by its qualitative content and by the types of cryogenic structure. Volumetric ice content due to ice inclusions below the base of the seasonally thawing layer and down to the base of the lake loam averages 25–40%, with gravimetric moisture of the loam and peat being 30–40% and 550–650% respectively. Ice content in the marine-glacial loam is reduced to an average of 10–15%, but contrary to the lacustrine loam, its value is rather inconsistent in the section and rises in individual horizons up to 20–40%. As noted above stratified structure is typical of the entire series of bog-lacustrine deposits. Schlieren from 0.3–0.5 cm to 1.5–2.0 cm thick are set 1 to 7 cm apart occasionally constituting a network (Boring 44). A certain predominance of stratified structure with fine-schlieren densely spaced (0.3–0.5 cm) is noted in the upper part of the section down to a depth of about 3 m in the peat and loam. A break in structural uniformity is due to inclusions of coarser material in the loam. In Boring 35, three thick ice schlieren (10–15 cm) were found at depths of 3.55 to 4.0 m above the sandy--loam interlayer.

In the marine-glacial loam 1.5–2.0 cm thick schlieren become more widely spaced owing to thickening of mineral interlayers which are 10–20 cm thick. Horizons of stratified structure alte nate moreover with portions of massive structure. In some cases, thick schlieren (3–10 cm) and remotely stratified structure (Boring 44) are noted near the contact with lake loam.

Comparison of mean annual temperatures and ice content of rock shows that the latter (its qualitative evaluation and the structure) is characterized by consistency and homogeneity both within the peat and loam polygons and in the interpolygons, while mean annual temperatures change most abruptly in both absolute value and in trend of formation (Fig. 1: B, C, E). This leads to the conclusion that the values of mean annual temperature and those of ice content must be most cautiously correlated. In the case in question there is no relation whatever and the association of higher ice content with lower temperatures (Bakulin, 1958) is by no means confirmed.

In the paleogeographic context, the trend of changes in the temperature field seems to have been similar to that of the modern one. The behaviour of the lake loam base and the distribution of peat show (Fig. 1: A) that the sites of Borings 35 and 36 had always been subjected to considerable warming-up effect of lake water (it neighbours upon the maximum thickness of lake loam, Boring 37). The distribution of mean annual temperatures at the stage of lake shoaling, with the initiation of a polygonal relief must have been all the more similar to that of to-day. The appreciable thickness of peat in Boring 35 is no secondary phenomenon but appears due to the local conditions of peat accumulation. Hence, the distribution of mean annual temperatures similar to that existing at present (at least in the tendencies of its development, if not in absolute values), did not determine the peculiarities of ice formation in the deposits of the bog-lacustrine complex during the process of its syngenetic up-freezing.

Data collected by observations relating to the temperature regime permitted to establish a relation between winter values of temperature gradients and cryogenic structure. A regular accretion of their values is observed both in time and in the section from December till March, with either shifting of their maximum values from the near-surface horizons downward or their simultaneous equalization to a depth of 2 m. Deeper downwards they successively decrease (Tab. II). In polygons, maximum temperature gradients are noted down to a depth of 0.5 m (in the layer of seasonal thawing) and are 5.5 to 8.0°/m in January and February. The limit of gradient formation exceeding 1°/m is at 5 m, i.e., more or less at the base of the bog-lacustrine deposits; their values do not exceed 0.5°/m to 0.8°/m in the interval of 5 to 7 m, they are no more than 0.2°/m deeper downward in the annual cycle, and are absent in winter. On the whole, temperature gradients, just like temperatures, diminish (fade away) and lag with depth.

The values of maximum winter gradients within polygons of various types are rather similar at corresponding depths: 6.8 to 8.0°/m down to 0.5 m; 5.2 to 6.0°/m at 0.5 m to 1.0 m; 3.7 to 4.8°/m at 1 to 2 m; 2.1 to 3.2°/m at 2 to 3 m; and 1.2 to 1.6°/m at 3 to 5 m. The distribution of temperature gradients over the area corresponds to the character of distribution of ice content and cryogenic structure. The boundaries of these three characteristics of the frozen series are subject to a tendency determining the formation of temperature gradients, i.e., to the conditions of heat exchange in the winter, whereas the thermoisopleths of mean annual temperatures reflect the annual complex of factors affecting the temperature field (Fig. 1: B, C, E, D).

As shown above, the specific conditions of landscapes of low watershed divides against the background of all the changes that occurred in the climate of the Bolšezemelskaya Tundra during the Upper Pleistocene and the Holocene, contributed to preservation, within their bounds, of the most stable microclimate, especially in winter, and consequently also of the temperature field in these regions. Proceeding from this fact the present-day temperature gradients may be extrapolated under certain conditions for the time of de-

Table II

Winter temperature gradients in the layer of annual temperature variations

Depth in metres	Temperature gradient (°C/m)																			
	December					January					February					March				
	Boring					Boring					Boring					Boring				
	35	36	39	44	38	35	36	39	44	38	35	36	39	44	38	35	36	39	44	38
0.0–0.5	3.6	3.2	4.8	2.8	1.4	6.8	6.8	7.4	5.4	1.2	5.0	5.2	5.8	8.0	5.2	2.0	4.2	3.6	3.4	3.2
0.5–1.0	4.0	3.6	3.4	3.8	0.0	6.0	5.0	3.8	3.6	0.2	6.0	5.2	6.0	5.4	3.0	3.8	5.0	4.2	3.6	2.8
1.0–2.0	2.0	0.9	1.2	2.0	–0.3	3.0	2.3	2.5	2.6	–0.3	4.8	3.7	3.8	3.9	0.6	4.2	3.3	4.1	3.4	2.2
2.0–3.0	0.3	0.0	–0.5	0.4	–0.4	1.0	0.8	0.9	1.4	–0.4	2.4	2.2	2.0	2.3	–0.2	3.2	2.0	2.1	2.7	0.8
3.0–5.0	0.0	0.0	–0.15	0.0	–0.8	0.2	0.2	0.15	0.4	–0.2	1.0	0.6	0.65	1.05	–0.2	1.6	1.2	1.3	1.5	0.0
5.0–7.0	0.0	0.0	0.0	0.0	–0.2	0.2	0.1	–0.05	0.0	0.0	0.4	0.0	0.2	0.2	0.0	0.8	0.5	0.55	0.55	0.0

velopment of the cryogenic structure in the bog-lacustrine deposits. It should be noted that fine-schlieren, densely stratified or reticulate structure and volumetric ice content of 50 to 60% correspond to the maximum gradients of 7 to 8°/m in the seasonally thawing layer. Therefore it might be assumed that the formation of medium-stratified structure with mixed medium- and fine-schlieren and of 30–40% ice content proceeded with gradients of 5 to 6°/m, which is on the whole consistent with the gradation suggested by Popov (1967), according to which fine-reticulate and densely stratified structure originates within limits of 1 to 8–10°/m. The predominance of fine-schlieren structure in the upper horizons can be attributed to the relative increase of cont nentality in the Holocene.

Coarse-schlieren structure develops, according to Popov, with gradients of 1 to 0.2–0.3°/m. Such structure, unrelated to lithology, was found in the upper part of the marine-glacial deposits underlying the lake loam (Boring 44). The date of its formation has been attributed to the initial stage of existence of relict lakes under conditions of a rigorous moderately continental climate during the Zyryansk glaciation. Considering the warming-up effect of lake water and the sufficiently thick snow cover (coastal situation of the region), it might be assumed that the temperature gradient in the layer of passive cryodiagenesis (summits of the marine-glacial deposits) did not exceed 1°/m, and ice schlieren of 10 to 30 cm originated under its effect. The presence of fine-schlieren structure lower down the section can be explained by a decrease in moisture content of the diagenetically compacted loam, increased by redistribution of moisture and its upward shifting with the development of thick schlieren. Therefore the finer structure of the epigenetic series is probably due to deficiency in moisture that began with its redistribution in the process of freezing, or otherwise primarily conditioned – rather than to an increase in gradient values.

Investigations of the role of temperature conditions in the formation of cryogenic structure demonstrate the existence of a direct connection between the latter and temperature gradients that are among the most important factors determining the physical essence of the process of moisture redistribution with freezing. Winter temperature gradients which reflect the conditions of freezing and condensation determine the temperature conditions in the time of development of cryogenic structure in a less ambiguous manner than do mean annual temperatures. The latter characterize the heat exchange standard in the annual cycle and are formed under the action of a much larger number of factors including those operating as well in summer. In an analysis of permafrost facies the features of the temperature field during the winter season should be therefore taken into account, together with other factors determining the formation of cryogenic structure (Katasonov, 1965).

In permafrost studies, no matter how detailed any under-estimation of these factors may lead to erroneous extrapolations concerning the formation of course of mean annual temperature for cryogenic structure and for values of temperatures as related to ice content in frozen rock.

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