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FORMATION AND EVOLUTION OF THE RUSSIAN ARCTIC CRYOLITHOZONE (ON THE SHELF AND THE CONTINENT) IN THE PLEISTOCENE-HOLOCENE

ABSTRACT. The paper addresses age nonuniformity of the permafrost of the Russian Arctic shelf. It has been widely accepted that recent permafrost exists on the present-day shelf, which was formed under subaerial conditions during shelf draining in the Late Pleistocene, was flooded during the subsequent transgression, and now exists as a relic zone. However, there is also modern permafrost forming under submarine conditions. The paper considers the mechanism of its formation. The author suggests a mechanism that involves bottom soil freezing and ice formation due to constant natural transformations in seabed sediments. The proposed mechanism is supported by analyzes of certain sections of the bottom sediments of shelf and of the Pleistocene marine plains (ancient shelves) composed of dislocated sequences with massive ice beds. Analysis of the massive ground ice genesis identified different geological history as well as different transgressive and regressive regime of the Russian Arctic western and eastern sectors. The glacial cover has limited distribution in the Russian North and was absent on the Russian Arctic and Subarctic plains to the East of the Kanin Peninsula.

KEY WORDS: massive ground ice, sheet ice, polygonal wedge ice, subsea permafrost zone, Quaternary history of the Russian Arctic permafrost.

INTRODUCTION

Permafrost (cryolithozone) is one of the elements of the Arctic zone natural

conditions, which substantially complicates shelf development. The present-day Arctic shelf bottom is mostly composed of frozen rocks, often of high ice content (Fig. 1). Most researchers assume that the cryolithozone within the shelf is relic, was formed during the deep Late Valdai (Sartan) sea regression, and was subsequently flooded during the Late Pleistocene – Holocene transgression. During the sea regression, the shelf was drained to a bottom contour of 100–120 m, and thick permafrost was formed within the shelf. At present, the cryolithozone is traced in the Arctic seas as a relic permafrost [Solov'ev, 1981,1988] with a thickness of up to 500 m and greater, for example, in the Laptev and East Siberian seas [Romanovskii et al, 1999, 2011]. However, the main feature of the ground ice distribution is a sharp distinction between the Western and Eastern sectors of the Arctic. In the Eastern sector, polygonal wedge ice is spread almost exclusively, while in the Western sector, the massive ice bed dominates. Genetically, these are essentially different formations, and this fact points to differences in development history of the two parts of the Arctic.

MATERIALS AND DISCUSSION (WESTERN SECTOR OF THE ARCTIC SHELF)

Permafrost with massive ground ice is widely spread in the western sector of the Arctic shelf – the Barents and Kara seas (see Fig. 1). Bottom permafrost was found at depths from 0 to 230 m [Bondarev et al, 2001; Rekant et al, 2005]. Permafrost table can lie at

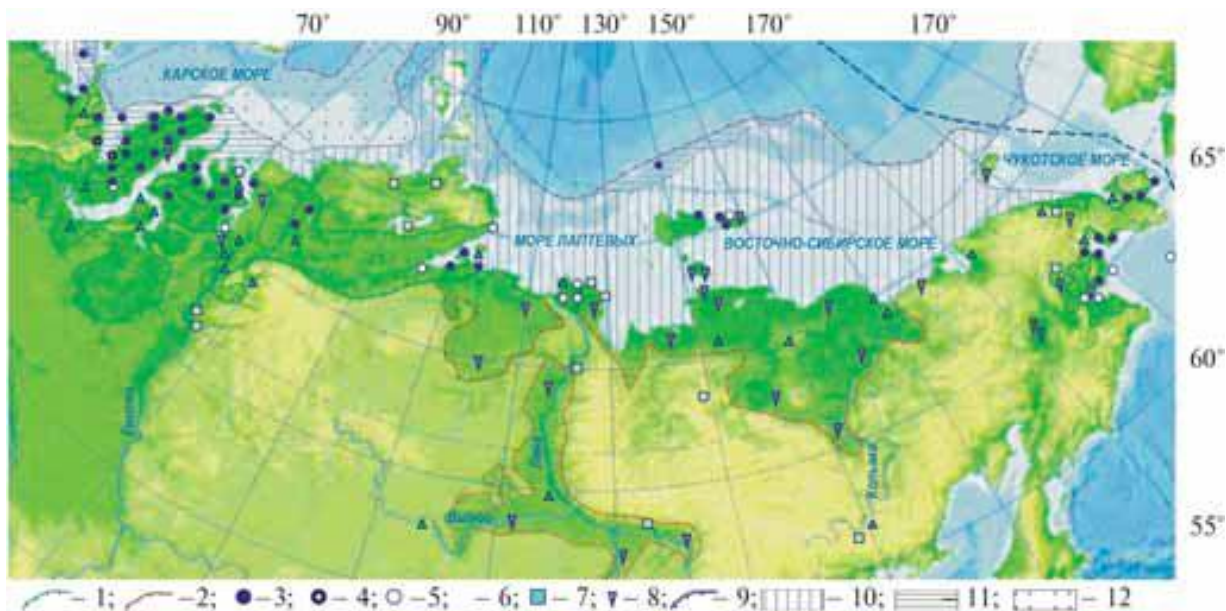


Fig. 1. Map of ground ice. Compiled by N.A. Shpolyanskaya and I.D. Streletskaya; the shelf is shown according to V.A. Solov'ev and S.I. Rokos with some modifications by N.A. Shpolyanskaya.

- 1 – Late Pleistocene marine plain boundaries; 2 – Pleistocene lacustrine-alluvial plains boundaries;
 3–8 – ground ice thick beds; 3 – subsea genesis; 4 – mixed genesis; 5 – coastal-marine genesis; 6 – injected genesis;
 7 – buried (primarily surface); 8 – polygonal wedge ice; 9 – shelf outer boundary; 10–12 – shelf cryolithozone;
 10 – relict permafrost with temperature of 0 to -2°C , up to 200 m or thicker;
 11 – modern permafrost with temperature of 0 to $-1,5^{\circ}\text{C}$, 80–100 m thick;
 12 – cold soils with newly formed permafrost islands with temperature of 0 to $-1,5^{\circ}\text{C}$, 80–100 m thick.

approximately 20–40 m under the sea floor or rise to the bottom surface. Permafrost base sinks to 100 m and lower. In bottom permafrost sediments sections, there is a large amount of ice, sometimes reaching 100% (Fig. 2). Many authors consider the permafrost to be relict here.

However, the very fact of permafrost occurrence at depths from 0 to 230 m indicates its heterogeneous nature. Even if we accept the idea of the Late Valdai sea regression to the isobath of 100–120 m, which led to the shelf soils freezing in subaerial conditions, there are still vast areas with a sea depth from 100–120 to 230 m where permafrost must have formed directly in subsea conditions.

The formation of icy bottom sediments is a known fact, e.g., in the southeaster Barents Sea, southwestern Kara Sea [Mel'nikov and Spesivtsev, 1995; Rokos et al, 2009], and in the eastern Laptev Sea, where frozen marine syngenetic sediments have been encountered [Kassens et al, 2000],

which indicates that the cryolithozone is nonuniform within the Arctic shelf. This zone includes both relict and recent, newly formed, permafrost. Understanding age nonuniformity of the permafrost within the shelf is important for the several reasons.

First, it can help solving many fundamental problems of the thick Pleistocene sediments genesis and Quaternary history of the Russian Arctic regions: certain issues of the Arctic shelf paleogeography can be resolved if we assume that permafrost can form immediately on seabed. Second, by accepting the possibility of formation of the initially submarine cryolithozone, we can explain the origin of certain types of massive ice encountered in sediments of the Pleistocene marine plains (i.e., ancient shelves), and identify the origin of brines at negative temperature (cryopegs) frequently observed in the Arctic regions. There are data on the genetic relation between cryopegs, massive ice beds, and marine host rocks [Streletskaya, Leibman, 2002]. Third, the directional evolution of the shelf

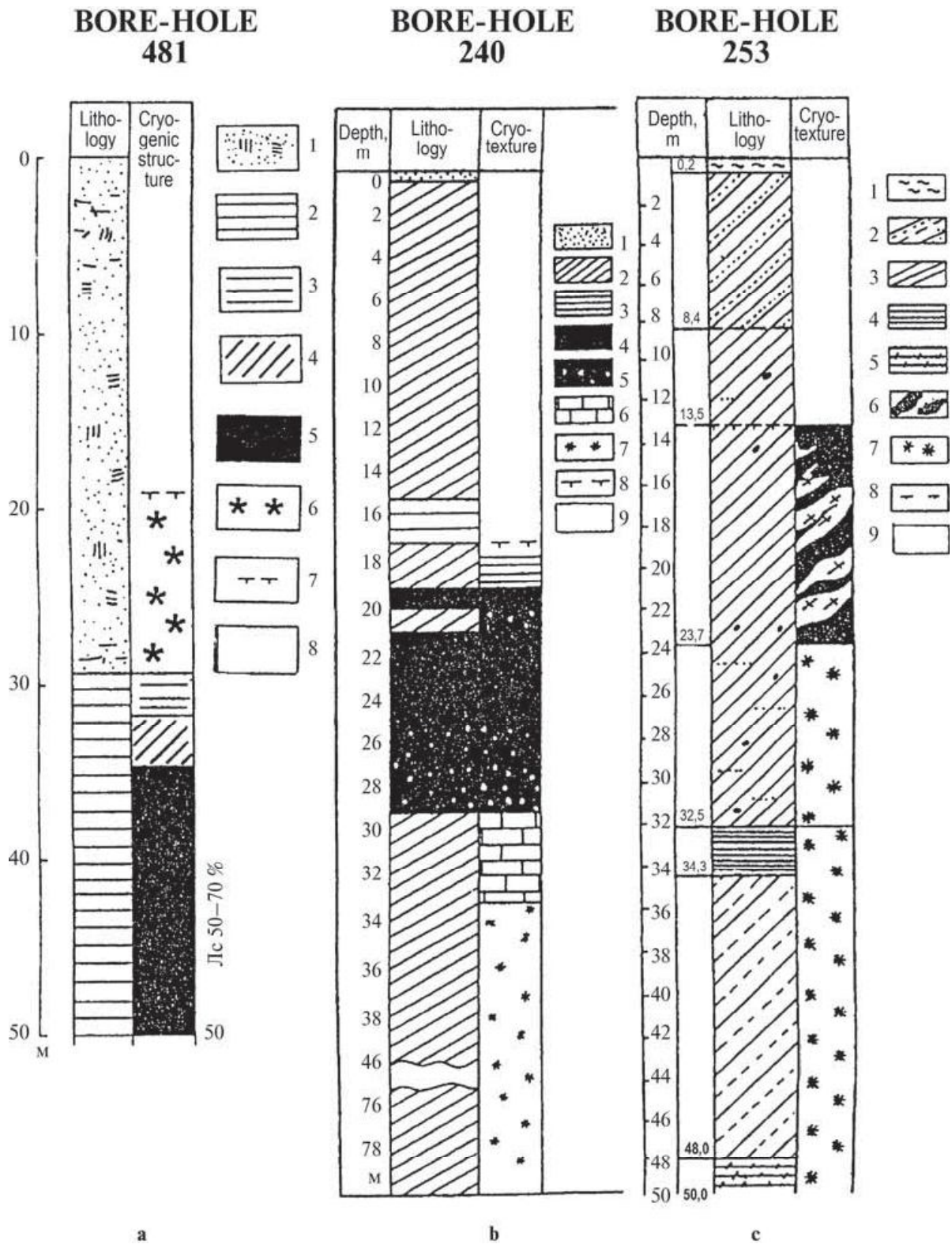


Fig. 2. Bottom sediments section on the Barents-Kara Shelf [Melnikov and Spesivtsev 1995].

a – bore-hole 481 in the Kara Strait area (sea depth is 65 m). 1 – sand with organic material inclusion; 2 – clay; 3–6 cryogenic structure; 3 – horizontally layered; 4 – dislocated subvertical large-schlieren; 5 – ataxitic (sheet ice); 6 – massive; 7 – permafrost table; 8 – negative-temperature deposits.

b – bore-hole 240 in the Baydaratskaya Bay (sea depth is 13–14 m). 1 – sand; 2 – clayey silty loam; 3 – clay; 4–7 cryogenic structure; 4 – sheet ice; 5 – ataxitic; 6 – reticulate; 7 – massive; 8 – permafrost table; 9 – negative-temperature deposits.

c – bore-hole 253 in the Kara Sea on the Rusanovskaya Field (sea depth is 130 m). 1 – silt; 2 – rhythmical interbedding of sand, sandy silty loam and clayey silty loam; 3 – clayey silty loam; 4 – clay; 5 – argillite-like firm clay; 6 – sheet ice; 7 – massive cryogenic structure; 8 – permafrost table; 9 – negative-temperature deposits.

cryolithozone affects the development of the shelf offshore and near-shore parts and, therefore, makes it possible to predict the dynamics of Arctic coasts and assess the degree of their stability. Fourth, the directional development of the shelf permafrost (degradation and growth of relic and recent submarine permafrost, respectively) is responsible for the geoenvironmental conditions within the shelf and, thereby, for the conditions of shelf development.

The paper proposes the conceptual author's viewpoint that the Arctic sea bottom sediments can freeze and permafrost can now form within the Arctic shelf. The author suggests a mechanism of the permafrost formation under submarine conditions along with certain criteria for age differentiation of the shelf cryolithozone.

ANALYSIS OF CONDITIONS FOR SYNGENETIC FREEZING OF BOTTOM SEDIMENTS

The possibility of bottom sediment freezing depends on the relationship between the pore water salinity and temperature of sediments [Shpolyanskaya, 1989, 1999]. In this respect, the following data are available.

Temperature. The average annual temperature of the upper horizons of bottom sediments (T_s) is, as a rule, equal to the temperature of the bottom water layer, whose pattern of the spatial variations has been studied in detail. The temperature depends on the Arctic sea depth [Zhigarev, 1997]. Negative temperature is the lowest at zero sea level. With increasing sea depth to 2 m, temperature increases and reaches 0°C at the fast ice boundary, whose maximal thickness equals to sea depth. With further increase in sea depth, temperature continues to increase, becomes positive, and reaches its maximum (2,8°C) at depths of 2–3 m, where ice floats. This is caused by a considerable summer heating of shoals. If sea depth continues to increase, temperature starts decreasing again and crosses 0°C once

again (becomes equal to –0,2°C) at a depth of 7–8 m. Beginning from a sea depth of 16–18 m, bottom water has a stable negative temperature. The temperature decreases to a depth of 30–35 m, where it reaches the minimal values (–1,6 ÷ –1,8°C). Homothermy is observed at depths of 35–40 through 250 m; i.e., the bottom layer temperature remains unchanged at these depths where there are no annual temperature variations. Below these depths, temperature gradually increases to 0,8°C at a depth of 500 m. The described temperature distribution in the bottom water layer indicates that in the Arctic seas, stable negative temperatures are formed at rather large depths of about 40–250 m.

Salinity. The published data on the salt content of bottom sediments make it possible to draw rather definite conclusions on the mineralization character. The salt distribution over the section is not uniform. The mineralization in the surface layer of bottom sediments is usually higher than the salt content of bottom water [Horn, 1972; Shishkina, 1972]. At the same time, pore solution salinity decreases from bottom surface downward. O.V. Shishkina established both these effects when she studied the variations in the chlorine content of pore silt water in different seas. She observed that the chlorine content of the upper layer of sediments was higher than that of the bottom water by 0,6–1,1‰ on the southwestern African shelf, by 0,7–2,3‰ on the Peruvian offshore slope and shelf, by 0,5–0,7‰ on the eastern coast of California, and by 2,12‰ on the Sea of Okhotsk shelf. The chlorine content decreases with increasing depth in many regions: on the offshore slope of the South American continent in the Atlantic, at a large distance from the coast in the northeastern Atlantic at a sea depth of 3000 m (from 19,3‰ in the surface layer to 15,1‰ at a depth of 4,3 m), in the Black (from 12,5 in the surface layer to 9,5 at a depth of 4,5 m and to 4–6‰ at a depth of 8–10 m) and Baltic marginal continental seas, and near the coast of the open Norwegian Sea. The works of I.A. Komarov and D.S.

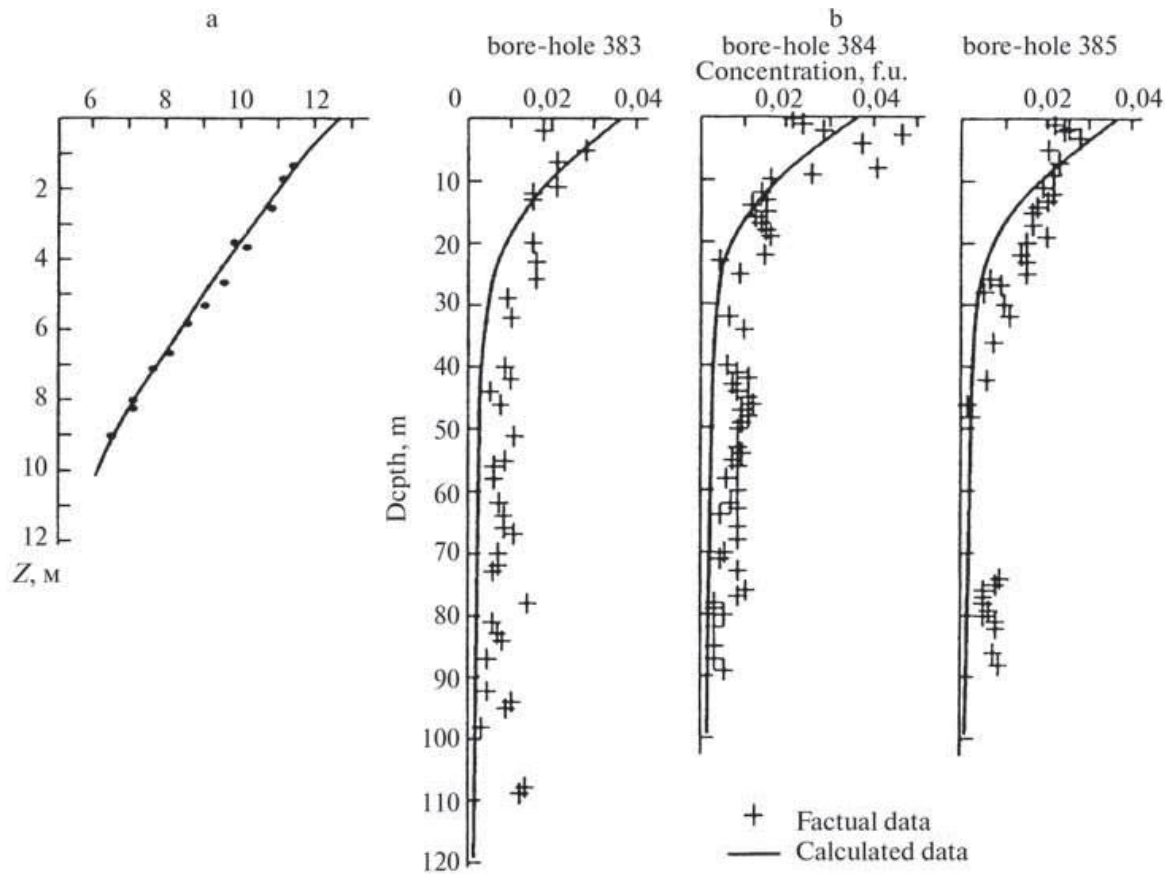


Fig. 3. Vertical variations in the pore solution salinity in bottom sediments:
a – Black Sea [Shishkina, 1972], b – Barents Sea [Komarov and Lukovkin, 2001]

Lukovkin [2001] and A.N. Khimenkov and A.V. Brushkov [2003] also report that the pore water concentration in bottom sediments decreases down the section (Fig. 3). These data indicate that the directional variation in the bottom sediment salinity with changing depth is a universal phenomenon.

The cause of such salt distribution has not yet been determined, although this cause has been discussed in literature. R. Horn [1972] considered that ion exchange at the “seabed – seawater” interface can cause a decrease in the salt content. During such exchange, positive absorption and, consequently, an increase in the ion concentration in the pore solution can be observed in some cases, and negative absorption and a decrease in the ion concentration can be observed in other cases, e.g., in montmorillonite, which is the known mineral component of marine sediments.

However, R. Horn considers that this disagrees with the fact that the salinity in the

upper layers of silt water remains higher than the seawater salinity. R. Horn also presents viewpoints of other researchers, who assume that sediment salinity decreases with depth as a result of sediment consolidation. However, R. Horn presents results of the experiment indicating that pressure does not affect the composition of water in clay. Moreover, ion diffusion forces should have equalized any concentration gradients since bottom sediments become less compact toward their surface. R. Horn did not find a satisfactory explanation of the existent salinity distribution in marine sediments.

Several researchers, O.V. Shishkina [1972] and other, assumed that decrease in salinity with increasing depth reflects a certain earlier geological stage, when glaciers thawed and seawater became less saline since fresh thawed water was discharged into sea. However, it is difficult to agree with this assumption for the following reasons. First, salinity also decreases at low latitudes in open sea, where glacial water could hardly

pronouncedly affect seawater mineralization. Second, fresh water is always less dense than saline water and usually floats over the sea surface in the form of lenses. A fresh-water lens thickness of even the largest world rivers, such as Orinoco and Amazon, is not more than 10–15 m, although these rivers flow into the open sea over a distance of several hundred kilometers [Fedorov, 1981]. In the not very deep Gulf of Ob, the fresh water of the Ob River is underlain by the heavier saline water of the Kara Sea [Zhigarev, 1997]. This phenomenon can be explained by differences in the fresh and seawater densities. Thus, the maximal fresh water ($S = 2\text{‰}$) density at a temperature of 4°C is 1.00163 g/cm^3 , whereas the seawater ($S = 35\text{‰}$) density at a temperature of -1°C is 1.028126 g/cm^3 . Therefore, seawater freshening in the epoch of deglaciation hardly affected bottom water and could not influence mineralization of bottom sediments. This could take place only in isolated cases in the near-shore shallow-water areas of the shelf. Third, diffusion, which would inevitably have originated during the concentration gradient formation after deglaciation, should have equalized salinity of sediments during geological epochs, no matter how slow this diffusion may be. A stable mineralization gradient should be maintained due to some permanent process.

It seems that the process of thermal diffusion can explain the cause of the considered salinity distribution in bottom sediments. The analyzed medium (bottom sediments) is simultaneously characterized by gradients of concentration and temperature related to the heat flux within the Earth. In this case, the process of thermal diffusion, which means that mass is transferred under the action of temperature gradient, proceeds in addition to the diffusion process. The flux of ions directed oppositely to temperature gradient originates in bottom sediments. This process is described by the formula [De-Groot, Mazur, 1964]

$$I = -D'\rho C(1 - c)\text{grad}T - \rho D\text{grad}C, \quad (1)$$

where I is the concentration flux, D' is the thermal diffusion coefficient, ρ is the density of a medium, C is the solution concentration, T is the temperature, and D is the diffusion coefficient.

The first term in the right-hand side of the equation describes the process of thermal diffusion, i.e., a salt flux caused by temperature gradient and directed oppositely to this gradient (from lower soil layers toward the soil surface). The second term in the right-hand side of the equation describes the flux directed oppositely to the direct diffusion, which originates due to concentration gradient and tends to balance the flux of thermal diffusion.

The condition of equilibrium of these oppositely directed salt concentration fluxes, usually dependent on ordinary and thermal diffusion, is specified by the formula

$$\text{grad}C = -\frac{D'C(1-C)}{D}\text{grad}T. \quad (2)$$

The salt content of pore water in bottom sediments will decrease until the stationary gradient, corresponding to the condition (2), is established. Since “the marine sediments – seawater system” is open owing to permanent accumulation of sediments, such equilibrium is not formed, and the salt flux from bottom to top is constantly maintained.

Based on these concepts, another phenomenon related to the above-mentioned (increased salinity in the bottom soil upper layer as compared to bottom water salinity, a satisfactory explanation for which is still absent) can also be explained. As was noted above, the intensity of diffusion related only to the concentration gradient is insignificant, especially at low temperatures. At the same time, the process of thermal diffusion that can result in a pronounced ion flux terminates at the “bottom soil – water” interface since the temperature gradient disappears in the bottom water layer. Under these conditions, salts migrating from below

are adsorbed and are reliably retained in the upper layer of bottom soils.

Thus, bottom sediments can freeze at a certain depth below the seabed since the salinity of pore water in bottom sediments decreases down the section.

This may be supported by a simple calculation based on the published data. The freezing temperature of seawater with a salinity of 35‰ is $-1,91^{\circ}\text{C}$. According to the factual data, the natural water temperature in the bottom layer of the Arctic Ocean is $-1,8^{\circ}\text{C}$; therefore, this water never freezes. The salinity of bottom sediments in the upper section is even higher; therefore, these sediments are also cooled, rather than frozen, to a certain depth. However, the salinity pronouncedly decreases with increasing depth, and temperature at a certain depth becomes sufficient for sediments to freeze. The combined variations in the bottom soil temperature (with $\text{grad } t = 0,04^{\circ}/\text{m}$) and salinity (according to [Shishkina, 1972]) plotted versus depth (Fig. 4) indicate that such conditions can originate at depths from 4,5 to 10–11 m.

As sediments accumulate, the depth interval where the condition of pore water freezing is satisfied shifts upward synchronously with the upward motion of the seabed surface. The frozen sequence syngenetically grows in the same direction, from bottom to top. It is known that dewatering of accumulating sediments is a very slow process. Therefore, slightly lithified and very wet bottom soils change into a layered ice-soil sequence.

As the frozen sequence grows upward, the temperature of the lower frozen layer increases in accordance with the temperature gradient. Therefore, this growth is finite and continues until the temperature of the growing sequence reaches 0°C . Figure 4 indicates that the frozen sequence thickness can be not more than 40–50 m at the above temperature gradient and salinity since this sequence will thaw from below according

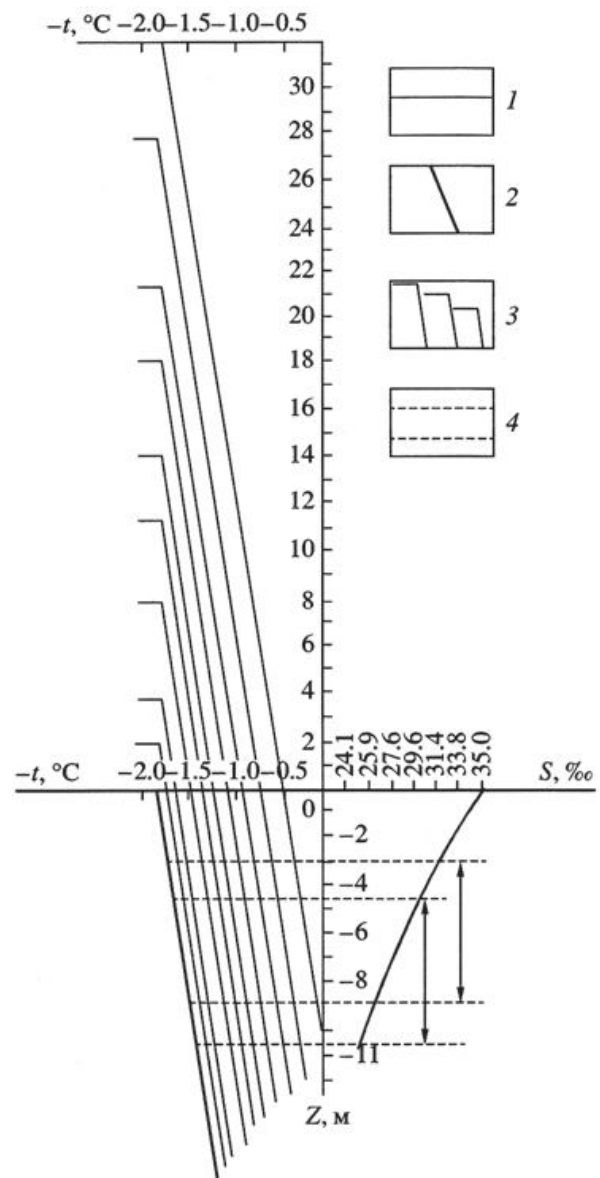


Fig. 4. Joint graph of vertical variations (Z_M) in temperature ($T^{\circ}\text{C}$) and salinity ($S\text{‰}$) of bottom sediments in the Arctic Shelf.

1 – seabed surface; 2 – recent temperature of bottom soils (vertical $\text{grad } T = 0,04^{\circ}/\text{m}$); 3 – variation in bottom soil temperature with accumulation of sediments and upward motion of the seabed surface; 4 – depth interval with freezing bottom sediments. This interval also moves upward with accumulation of sediments

to a change of negative temperature into positive ones. At other temperature gradient and salinity, thickness of deposits to 100 m can remain in a frozen condition.

It is necessary to consider one more problem. Heat released during pore water freezing should propagate from the freezing front into bottom water layers. Nevertheless, the temperature of this water does not increase in this case.

According to the observations, this temperature is always constant and the lowest at these depths because vertical density mixing constantly takes place in seawater. Since seawater density increases with increasing salinity and with decreasing temperature at close salinity values, bottom water moves upward and becomes warmer due to the influx of phase transition heat and is replaced by colder water coming to the seabed surface. As a result, a stable density and temperature stratification rather rapidly recover in water.

WATER CRYSTALLIZATION IN BOTTOM MARINE SEDIMENTS

When submarine freezing is considered, two main questions arise: 1 – how pore water crystallizes and ice body is formed in bottom sediments; 2 – how salt ions migrate and are distributed during silt water crystallization.

Freezing of bottom sediments can also be characterized as freezing of completely saturated soils and crystallization of highly mineralized water. In this case, according to V.I. Golubev's [2000] experiments, water distributed between mineral particles combines the surface of all particles into one system; therefore, emerging ice crystals rapidly grow and form basal ice. Since bottom soils freeze in a certain depth interval (see above), the freezing front is replaced here by the freezing zone, i.e., the zone with negative temperatures. Ice crystals can originate simultaneously at many points of this zone, where the only necessary condition is satisfied: pore solution salinity should be such that the solution freezing temperature would correspond to this temperature. Growth of crystals is hindered by adjacent crystals and soil particles. Because of mineral particles, growth of ice crystals becomes slower and continues mostly due to the growth of bases along the surface of mineral particles in sedimentation layers, where dissolved gas and salt inclusions are adsorbed during the growth of crystals. Therefore, the emerged ice streaks always inherit the form of even

folded soil beds. In the process of growth of ice crystals, crystal faces force out ions and molecules of dissolved salts, which results in the formation of the zone of increased solution concentration near crystals and its growth with time. Two processes proceed in this case. Growth of ice crystals slows down due to increase in salt concentration in the fluid boundary layer. On the other hand, diffusion of salts from the boundary layer into the liquid medium becomes more intense due to the increased difference between salt concentration in the boundary layer and liquid medium. Some time after the growth onset, a dynamic equilibrium is established between the processes of increase in the concentration in the solution boundary layer due to growth of crystals and decrease in the concentration due to salt diffusion. Under constant thermodynamic conditions, crystals continue growing at a certain steady rate. As a result, the two-phase system is formed. This system includes fresh ice crystals and concentrated saline water uniformly distributed in bands between the crystals. At contacts between the ice crystals, saline water is pressed out of these bands and is uniformly adsorbed on the surface of minerals in the ice-soil sequence. The ice remains fresh.

Salt distribution in frozen bottom sediments is inevitably closely related to facial conditions. Under the conditions of a deeper sea, when the finest (muddy) sediments with a high skeleton specific surface and, thereby, with a high surface energy are formed and freeze, salts pressed out during pore solution crystallization are completely adsorbed by the surface of mineral particles. Therefore, free pore water concentration always corresponds to a given freezing temperature, and ice crystals originate and grow without interruptions. Under the conditions of a shallower sea, when mostly sandy-silty sediments with usually low total surface energy are formed, uniform distribution of saline pore water over the surface of mineral particles is not observed. Pore solution concentration increases with increasing ice crystals, and unfrozen zones

with a very high solution concentration are gradually formed since salts remain dissolved. Ice formation stops. In sediments that are newly accumulated in the upper section, ice is not formed until the process of thermal diffusion brings salinity to a certain value. In this case, water begins to freeze and ice starts forming. Ice crystals continue growing until the concentration of the pressed out solution reaches a critical value. The unfrozen zone with a very high solution concentration originates again. Cryopeg lenses in freezing bottom sediments are formed in this way.

The considered dependence of salt ion distribution on facial conditions is also discussed in some publications. For example, when considering the processes of absorption at the “water – seabed” boundary, R. Horn [1972] notes that absorption becomes more intense with decreasing grain size of absorbing particles. Therefore, deep-water fine-grained sediments are enriched in trace elements in contrast to coarser sediments accumulated in shallower water.

The shelf cryolithozone analysis shows that the structure features of ice-rich bottom sediments of the Barents-Kara Shelf, for example, at the Kara Strait or on Rusanovskaya Field [Bondarev, 2001; Rokos, 2009] comply

with the initially subsea freezing mechanism described.

ANALYSIS OF CERTAIN FROZEN SECTIONS WITH MASSIVE ICE BEDS WITHIN MARINE PLEISTOCENE PLAINS (ANCIENT SHELVES)

The Pleistocene marine plains of northern Eurasia (Western sector: the Western Siberia and North-East of European Russia) – ancient shelves, whose sections show evidence of processes that proceeded under submarine conditions – can be considered an analog of the recent Arctic shelf. The study of the dislocated frozen sections with massive ice beds, which are widespread within such plains, indicates that the above schematic formation of the shelf (submarine) cryolithozone is rather clearly defined in these sections (for example, Fig. 5).

The structure of such a sequence indicates the formation of the permafrost on the recent shelf. The occurrence character and structure of similar ice beds (studied by the author in northern Western Siberia) reflect the sedimentation type of the entire sequence structure and indicate that these beds were formed under subaqueous conditions. As a rule, this section represents alternation of ice beds of thickness 10–15



Fig. 5. Massive ground ice bed. Western Yamal Peninsula. Photo by G. A. Rzanitsin

cm and soil interlayers of thickness less than 1 cm. Soil interlayers are not monolithic and also include ice microstreaks extended along bedding, parallel to one another. The layers are deformed and show a complex pattern: parallel layers or layers interlacing in a complicated manner, sometimes horizontal, in other cases arched or of a complex shape, located closely to one another, at a distance of several centimeters. A similar ice-soil marine sequence with highly coordinated layers, present even when the configuration is very complex, could form only under submarine conditions during syngenetic freezing of accumulating bottom sediments.



Fig. 6. Ground ice in the Barents-Kara Shelf dislocated bottom sediments [Melnikov and Spesivtsev 1995, Bondarev et al. 2001].

The hypothesis on submarine cryolithogenesis and related growth of dislocated ice beds was first suggested by Popov [1984]. The idea is based on the fact that sedimentation is primarily accompanied by underwater landslide processes, which results in plicative dislocations in bottom sediments. Such dislocations originate in sediments with excess moisture content and almost noncompacted sediments showing properties of floating and thixotropic soils. As a result of changing seabed slope during movement of such sediments, the entire sequence becomes shrunken, and not only regular but also reverse and drag, gentle and steep, folds of various dimensions are formed here. Plicative dislocations are traced over a distance of several tens and hundreds of meters along the strike and several hundred meters in vertical. In frozen sequences, these dislocations are accompanied by a regular ice distribution along sedimentation folds and massive ice beds are integrated with hosting dislocated sediments (Fig. 6). Such relationship between dislocations and ice indicates that sediment deformation and ice formation are syngenetic.

At present, many researchers accept that massive ice beds were often formed

immediately under submarine conditions, for example, [Khimenkov and Brushkov, 2003].

This mechanism is supported by numerous facts. For example, in the Baydaratskaya Bay area, in 42 drill sections, 12 cryopeg interbeds were found [Mel'nikov and Spesivtsev, 1995]. There is only one reason for that – interruptions in the ice formation during silty-sandy-loam and silty sediments freezing.

The above mechanism of formation of the submarine cryolithtozone is clearly defined in the section of massive ice beds in central Yamal, studied by Streletsкая and Leibman [2002]. The geological section from this work (Fig. 7) shows the sequence of marine Kazantsev sediments with a thick massive ice bed and cryopeg lenses. Ice is represented by alternation of ice and soil laminae of the sedimentation type. The sequence is generally salinized. However, Fig. 7 clearly demonstrates that cryopegs are not present in the clay part of the section and are frequently occur in the sand section. It should be noted that Streletsкая and Leibman differently explain origination of cryopegs; however, the discussion of differences

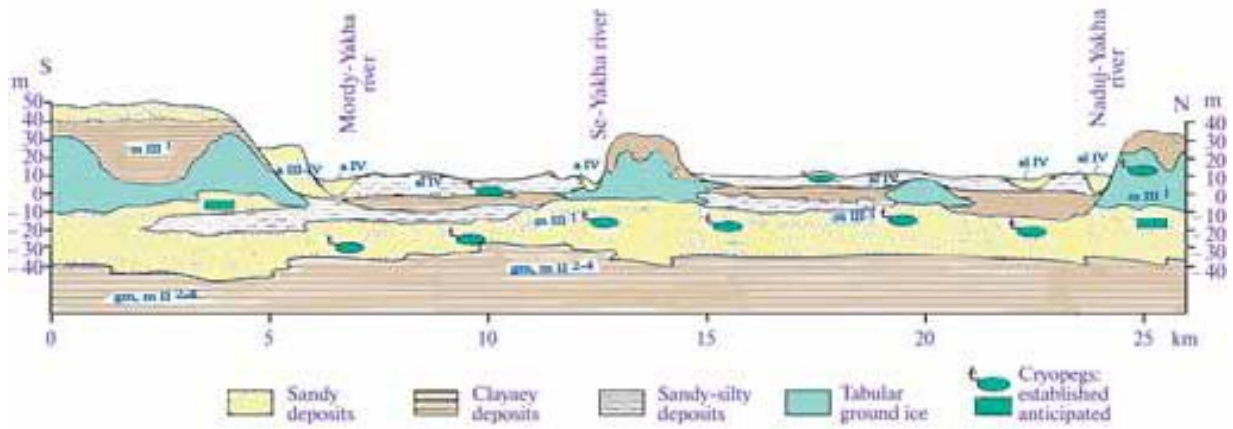


Fig. 7. Schematic geological transect with tabular ground massive ice and cryopeg lenses in Quaternary deposits compiled for Central Yamal, across Mordy-Yakha, Se-Yakha, Naduj-Yakha rivers [Streletsкая and Leibman, 2002]. The uniform ice body cutting by river valleys is visible.

in opinions with these researchers is the subject of an independent paper. Here, we use the factual data, which, in our opinion, confirm the proposed mechanism of pore water crystallization in marine sediments.

Features of the proposed mechanism are also traced in other sections of the Pleistocene marine plains. Thus, a distinct change in the cryogenic structure following a change in lithology is observed in the ice bed section in the Tadibeyakha River basin on western Gydan Peninsula [Shpolyanskaya, 1999] (Fig. 8).

Ice and soil laminas regularly alternate in the clay section (Fig. 8a), which indicates that ice was continuously accumulating in a growing frozen layer. In the sandy-loam (Fig. 8b), thicker and pronouncedly dislocated ice beds are separated by soil layers with a massive structure, which indicates that ice accumulation terminated due to increased pore solution concentrations.

The data of Streletsкая and Leibman [2002] demonstrate that ice and hosting sediment mineralization distribution along the section confirms the proposed mechanism of submarine cryolithogenesis. Thus, the total mineralization of pore solutions in clays near the ice bed is 20 542 mg/l, whereas the mineralization of ice interlayers in the same clay decreases to 189 mg/l, and the

mineralization of the massive ice bed in most of its body is not higher than 78 mg/l. This results from salt pressing out of the zone between crystals during growth of ice crystals and from salt absorption by clay particles of rocks containing ice. The mineralization of solution in sand (cryopeg lenses) reaches 58 507 mg/l, whereas the mineralization of pore solution in sand outside these lenses is only 2407 mg/l. It is clear that, despite of a high salinity of clay, cryopegs are absent in these sediments. All salt ions are retained by the surface of soil mineral particles, whereas pore solutions in sand are desalinated because salt ions pressed out of pores by growing ice crystals accumulate in closed zones, forming cryopeg lenses, rather than are absorbed by sand particles.

A change in the salt ion composition along the section also does not contradict the ice bed formation under submarine conditions. According to O.V. Shishkina [1972], the chemistry of seawater in muddy sediments substantially changes. Sulfates are reduced in bottom sediments of northern and inner seas, as a rule, enriched in organic matter; as a result, sulfate concentration in silt water decreases and is followed by a replacement of SO_4^- ion by CO_3 and HCO_3 ions. Silt water is transformed into chloride-alkaline water of low sulfate content. Chloride-sodium-calcium water, the composition of which differs from the seawater composition,

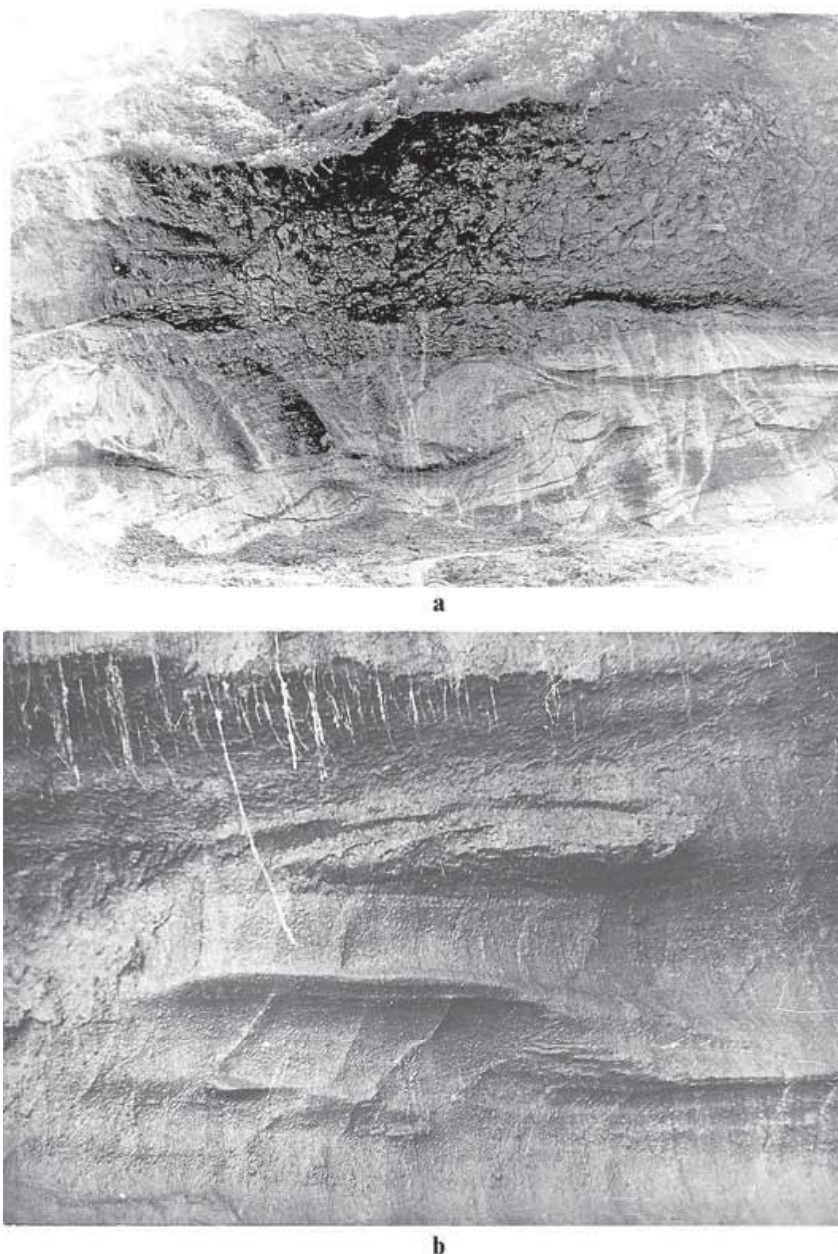


Fig. 8. Dislocated massive ice bed of the subsea genesis in the “Tadibeyakha” section (Western Gydan). Photo by N.A. Shpolyanskaya

A change in the cryogenic structure evidently follows a change in the lithology. a – clay section: uniform alternation of ice and soil laminas over the entire section indicates that ice continuously accumulates in a growing layer of frozen sediments. b – silty section: thicker and pronouncedly dislocated ice layers are separated by layers of massive soil, which indicates that ice accumulation was terminated due to increased concentration of pore solutions.

is formed during diffusion that causes exchange processes. Precisely this is seen in the ion structure of the considered section. The amount of SO_4^{2-} ions sharply decreases from seawater to pore solutions of clay and sand. The amount of HCO_3^- increases in the same direction: from seawater to pore solutions. The role of Na^+ in the ion composition structure of pore solutions in clay increases compared to seawater. This corresponds to the transformation of the water ion structure in bottom sediments

described by O.V. Shishkina. In this case, the marine spectrum of ions is clearly defined in pore water of clays, ice interlayers, main beds of massive ice, and cryopegs.

One more issue should be considered. The presence of salt ions (which, seemingly, precipitated during ice formation) in cryopegs can also be explained. It is known [Doronin and Kheisin, 1975] that joint presence of different ions in solution decreases salt precipitation temperature.

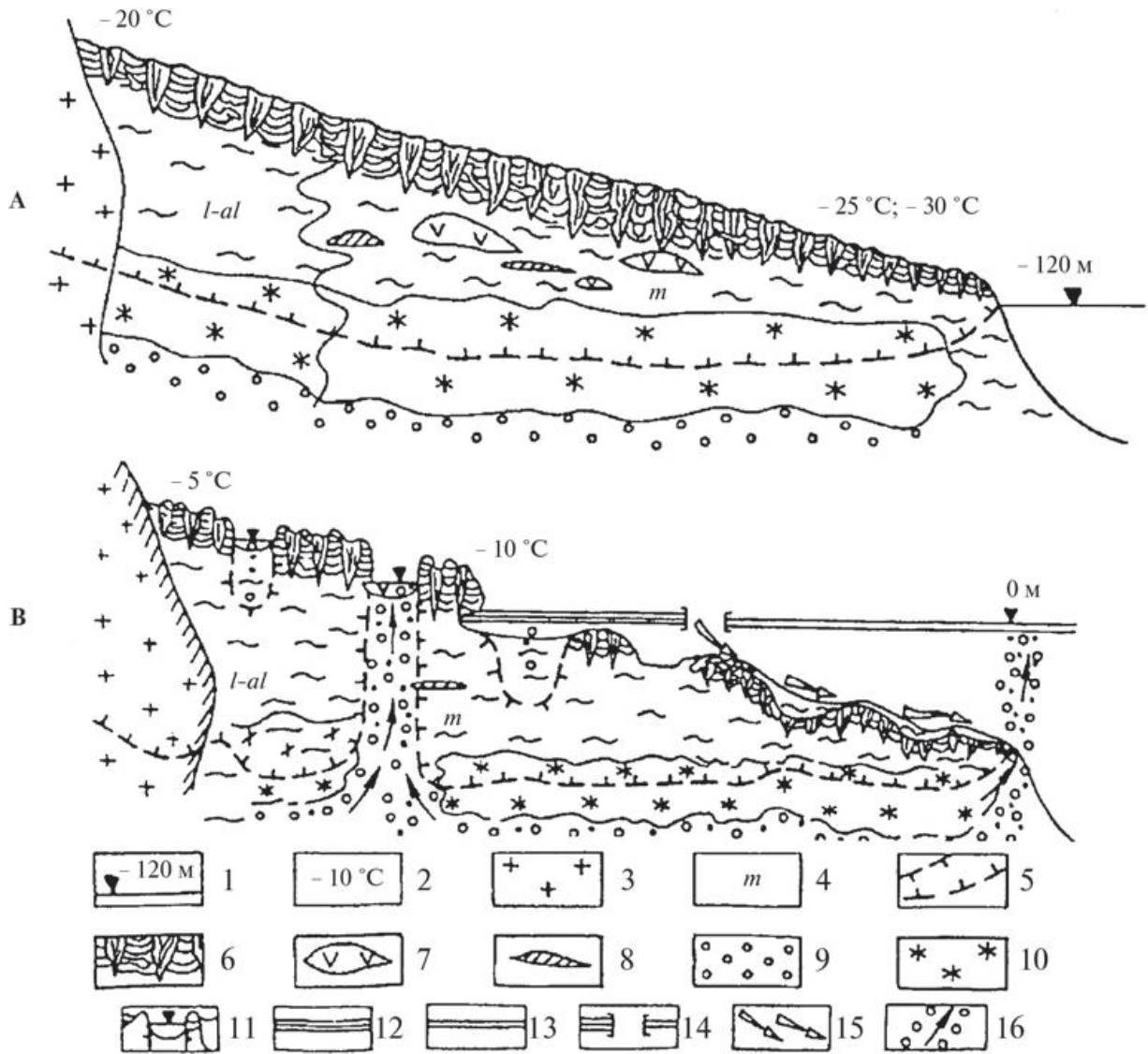


Fig. 9. The Laptev Sea shelf structure [Romanovskii et al, 1997].

A – 20000–18000 years ago, B – present time. 1 – sea level as related to modern;
 2 – soil average annual temperature; 3 – bedrock; 4 – the Quaternary deposits genesis index;
 5 – permafrost boundary; 6 – ice complex; 7 – sheet ice; 8 – cryopeg lenses; 9 – deposits containing free gas;
 10 – deposits with gas hydrates; 11 – thermokarst lake; 12 – fast ice; 13 – young ice; 14 – open lead;
 15 – cold salt brine; 16 – greenhouse gases streams.

For example, sodium sulfate precipitates from solution at a temperature of $-7,6^{\circ}\text{C}$ rather than at $-3,5^{\circ}\text{C}$, which takes place in the pure solution of this salt. The only components precipitating from solution soon after solution freezing – carbonates and hydrocarbonates (HCO_3) – are absent in cryopegs.

THE EASTERN SECTOR OF THE ARCTIC SHELF

In contrast to the shelf's western sector, the Arctic shelf eastern sector cryolithozone consists entirely of relict permafrost (Fig. 1

and Fig. 9). Here, there is evidence of an ancient icy wedge complex of the Zyryan and Sartan Epochs flooded by a postglacial transgression [Romanovskii et al, 1997; Romanovskii and Tumskoi, 2011; Schirmeister et al, 2008]. Those are ice wedges, which are wedge-shaped or column-shaped in section and form a polygonal system in plan view. They form only in continental conditions on periodically flooded surfaces. The basic conditions for ice wedge formation are the polygonal fracture system formation as a result of repeated soil surface thermal-contraction cracking and subsequent water penetration into the cracks.

On the Eastern Siberia plains, in contrast to the western sector, ground ice is almost entirely represented by polygonal wedge ice that forms the so-called icy complex. Polygonal wedge ices are most widespread in the following areas – the Yana-Indigirka and Kolyma lowlands, the Central Yakut lowland, and the New Siberian islands. Here, unlike in the western Arctic sector, the ice enclosing sediments are continental and belong to the alluvial, alas, slope, coastal-marine, and lagoon types. Starting with the Pliocene, the Prymorskaya Lowland deposits are represented by the lacustrine-alluvial, alluvial, and lacustrine-bog deposits. The Olerskaya Suite deposits (Early Pleistocene) contain ice-wedge pseudomorphs [Arkhangelov et al, 1989; Nikolsky and Basilyan, 2002]. On the Dmitri Laptev Strait coast and on Great Lyakhovsky Island, the strata with polygonal wedge ice age has the Middle Pleistocene age – 200000–180000 years ago, by $^{230}\text{Th}/\text{U}$ [Schirmeister et al, 2002; Tumskoi, 2012]. The Late Pleistocene, starting from the Kazantsev time, is also represented exclusively by continental deposits [Alekceev et al, 1992]. Only in the Middle Pleistocene, there was a small sea transgression that flooded a narrow coastal strip approximately from the Lena River mouth to the Chaun Bay [Alekceev et al, 1992] and formed a marine terrace, 138000 years old. [Bolshiyarov et al, 2009].

This points to the region's continental (as opposed to the western sector) development during the whole Pleistocene in the conditions of consistently severe climate and the absence of the glacial cover.

CONCLUSIONS

The main conclusion follows from the performed analysis: the cryolithozone with massive ice beds can form under shelf submarine conditions and at a rather large depth.

Frozen dislocated sequences of the apparent submarine origin are observed within marine plains formed in northern Russia (on

the Arctic western sector) during almost all stages of the Pleistocene. This means that these sequences were formed on ancient shelves during all epochs, both glacial and interglacial. Therefore, we can assume that the submarine cryolithozone also forms on the Arctic shelf.

At present, the main problem is to find differences between the cryolithozone, newly forming under submarine conditions, and the relic cryolithozone, which formed under subaerial conditions and was subsequently flooded. Certain criteria of differences between the newly forming shelf cryolithozone and the relic zone can be proposed even at this stage of research. a) Newly forming (recent) permafrost can be encountered only at depths larger than 40 m. Permafrost at smaller depths can be only relic. b) Depending on the salinity of bottom sediments and the temperature gradient in these sediments, the recent permafrost thickness can vary from 50 to 80–100 m, whereas the thickness of the relic permafrost can reach several hundred meters. c) The recent permafrost temperature should be no lower than $-1,5 \div -1,6^\circ\text{C}$, whereas the temperature of the relic permafrost can be even lower. d) The cryogenic structure should reflect the syngenetic type of freezing: uniform alternation of ice and soil laminae and high conformity of layers even if their configuration is complex. Since the relic permafrost is a subaerial formation, it should reflect the epigenetic type of structure: it should be observed rarefaction downward ice layers e) The chemistry of sediments containing emerging ice should reflect the marine type of salinization. Ice should be fresh.

The ice spatial patterns in relation to their genesis indicate the absence of ice sheets in the Pleistocene and Holocene on the Russian North plains, most likely, to the East of the Kanin Peninsula. Mountain-valley glaciation, sometimes changing into reticulate glaciation, occurred only in mountain areas.

The Arctic and Subarctic eastern and western sectors evolved differently in the Pleistocene, and transgressive–regressive regime was

manifested in different ways. During almost the entire Pleistocene (excluding the Sartan Epoch), the western sector plains formed under predominantly marine sedimentation conditions, while the eastern sector plains formed under continental, predominantly lacustrine-alluvial and lagoon sedimentation conditions. Only the regression of the last part of the late Pleistocene (the Sartan Epoch) and the Holocene transgression proceeded simultaneously.

The currently accepted synchronism “glacial period – sea regression”, “interglacial period

– sea transgression” is not detected. All the facts cast doubt on glacioeustatic leading role in sea level variations, bringing to the forefront the impact of tectonic processes. The complex heterogeneous tectonic structure of the Arctic Basin supports this statement. It is obvious that the mid-oceanic ridge (the Gakkel Ridge) within the Arctic basin and the junction of the Eurasian and Amerasian tectonic plates influence the irregular fluctuations in the Arctic basin, and their influence appears to be considerably higher than the effect of the glacioeustatic processes. ■

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