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ROLE OF CRYOGENESIS IN THE FORMATION OF LATE QUATERNARY DEPOSITS COMPOSITION IN ANTARCTIC OASES AND NORTH-EAST YAKUTIA

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The data pioneered by Cryology Laboratory of Institute of Physicochemical and Biological Issues in Soil Science from Antarctic oasis and North-East Yakutia have been analyzed. Two different types of cryogenic weathering are defined on the basis of analyses of current temperature regime of the active layer and mineral of composition of deposits. The detailed investigations of Late Quaternary deposits in northeast Yakutia and Larsemann Hills Oasis (Antarctica) have ascertained the cyclic structure of cryolithogenic strata induced by the changes in the environmental conditions during the process of sedimentation. It has been revealed that the most favorable conditions for cryogenic weathering occurred at the turn of the Late Pleistocene and Holocene for the studied regions.

Cryogenesis, Pleistocene, Holocene, Antarctic oases, North-East Yakutia

INTRODUCTION

The sediments transformation into deposits in the cryolithozone is accompanied by specific processes of cryogenesis, where cryogenic weathering is a key factor, consisting in selective destruction of minerals caused by the recurring freeze-thaw cycles over long periods of geological time. As a result, the mineral composition of rocks eventually acquire peculiar features, which defines the specificity of the cryolithogenic strata, as was discussed in [*Konishchev, 1981*].

The selective destruction of minerals occurring at temperatures crossing 0 °C threshold and during alternating freeze-thaw processes have been discussed in detail in many papers [*Konishchev, 1981; Rogov, 2000; Konishchev et al., 2006*]. During cryogenesis, mineral grains are destroyed due to changing thicknesses of water films, ice pressures and rupture of gas-liquid inclusions. The growth of resistance of most widespread minerals to cryogenic weathering can be illustrated by the following series: quartz → amphibole → pyroxene → feldspars → mica [*Konishchev, 1981*].

Multiple repeated freeze-thaw processes in the active layer produce specific mineral distributions in the grain-size fractions. The intensity of cryogenic transformation is assessed by the cryogenic weathering index (CWI) [*Konishchev, 1981*]:

$$\mathrm{CWI} = Q_1/F_1 : Q_2/F_2, \qquad (1)$$

where Q_1, F_1 is quartz and feldspar contents in 50–10 μm fraction respectively; Q_2, F_2 – quartz and feldspar contents in 100–50 μm size fraction respectively.

In rough approximation CWI values >1 bear the evidence of active cryogenesis during sedimentation, and the greater the value, the more severe conditions were responsible for the sedimentation. The CWI values <1 show that the sediments deposited outside the permafrost zone. This approach has been successfully applied to the reconstructions of paleo-permafrost conditions for various regions of Eurasia [*Konishchev, 1981, 2013; Konishchev et al., 2006; Schwamborn et al., 2007; Shmelev et al., 2013*].

The paper aims to determine the role of cryogenesis in the formation of loose sediments composition in the Pleistocene–Holocene time by the example of NE Yakutia and East Antarctica. With this in view, the following tasks were formulated:

– to identify characteristic features of cryogenic processes in the Antarctic oases and northeastern part of Yakutia in the Late Pleistocene and Holocene;

– to assess the influence of area-specific cryogenic processes on the formation of Late Quaternary deposits composition in the Antarctic oases and NE Yakutia.

CHARACTERISTIC FEATURES OF CRYOGENIC WEATHERING IN THE ANTARCTIC OASES AND IN NE YAKUTIA

Materials and methods of investigations

The data on temperature regime of the active layer (AL) for the study areas were collected with *Hobo Onset* thermal logger, which measures soil temperature every 2–6 hours at different depths in the temperature monitoring sites. The upper sensor is installed directly on the ground surface under the soil-vegetationt cover, and at the depth of 1–2 cm below surface, in the absence of soil-vegetation cover (Antarctica).

Samples were collected in the 57th and the 58th Russian Antarctic Expeditions and “Beringia” expedition in the summer 1991 (Cape Chukochi) and

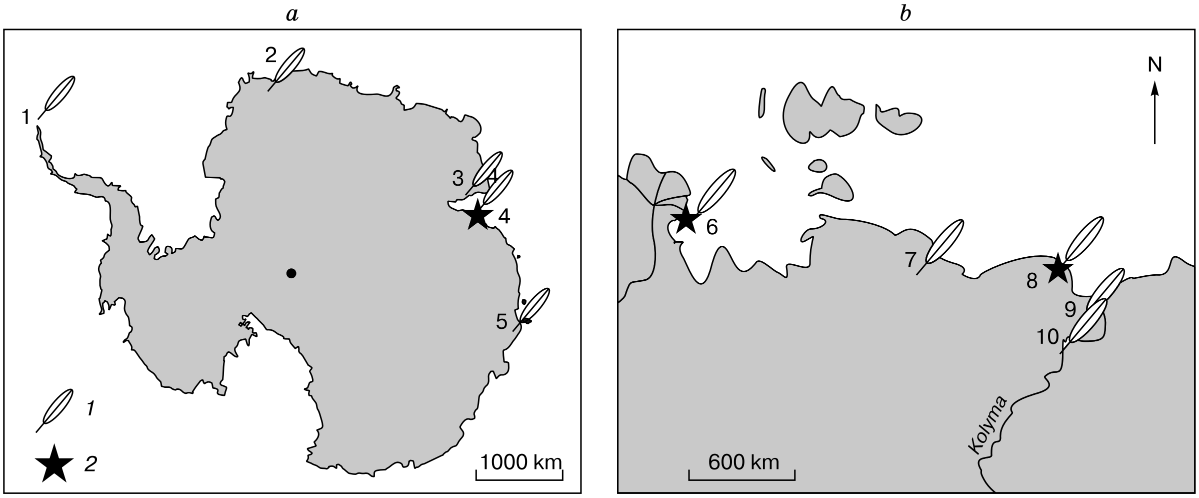


Fig. 1. Schematic map of sites for temperature regime monitoring (*1*) and paleogeographic studies (*2*) in Antarctica (*a*) and in North-East Yakutia (*b*):

1 – King George island; 2 – Schirmacher Oasis; 3 – Druzhnaya-4 Field Station; 4 – Larsemann Hills Oasis; 5 – Bunger Oasis; 6 – Bykovsky Peninsula; 7 – Allaikha; 8 – Chukochi Cape; 9 – Lake Akhmelo; 10 – Omolon River mouth.

2000 (Bykovsky Peninsula) (Fig. 1). The results of these investigation have been published for the first time. The core drilling was done using drilling rig UKB-12/25 (Vorovskiy Machine-Building Plant, Ekaterinburg) and MGBU-800 "Termite" (NPO "Gemmash" Ltd, Istra). The drilling was done "in dry", without the use of drilling fluids, the diameter of core samples was 70–120 mm. The air-dried samples were delivered to the laboratory for analyses. Particle size determinations were performed with Laser Particle Sizer "Analysette 22".

The mineral compositions in 100–50 and 50–10 μm grain-size fractions were determined by X-ray analysis with Dron-3M [*Shlykov, 1991*]. The morphology of 250–100 μm particles was studied with LEO 1410 and Tesla Scanning Electron Microscopes (SEM). The samplings of 30–40 grains were used for the analysis.

Roundness of quartz grains was assessed in five-gradations according to the Khabakov scale, where 0 is for sharp-angular shape, 1 – angular, 2 – angular-rounded, 3 – rounded, 4 – well-rounded. For estimations, we used average and the most frequently encountered (mode) values. The selection of this grain-size fraction was determined by the dominating cryogenic destruction of minerals, especially quartz and feldspar [*Konishchev et al., 2006*]. The separation of heavy and light fractions of minerals was done in the centrifuge using a heavy liquid (bromoform with density 2.89 g/cm^3). Based on the data obtained, the coefficient of heavy residue (CHR) was calculated as the function of heavy residue of minerals in 50–10 μm to 100–50 μm grain-size fractions. Key indices for assessment of sedimentation and freezing conditions are CWI and CHR [*Konishchev, 1981*].

This complex of research has been applied to the deposits in East Antarctica and NE Yakutia, both classified as areas with exceedingly severe geocryological conditions in the Late Quaternary.

Cryogenesis conditions in the Antarctic oases

The inception of permafrost in Antarctica occurred synchronously with the formation of the ice sheet in the Eocene–Oligocene time [*Hambrey et al., 1989; Ehrmann and Mackensen, 1992*]. The Antarctic ice sheet continued to grow in the Late Pliocene (3.2–2.4 million years ago) [*Naish et al., 2007*]. The permafrost development in Antarctica was closely linked to the ice sheet–ocean system dynamics, and this relationship was most pronouncedly expressed in the Pleistocene–Holocene. The changes in the Antarctic ice sheet terminus zone and the sea level triggered the alternation episodes when the coastal oases were covered by the ice glacier, or flooded by the sea, interminently with their being in the ice-sea-free state. By the end of the Late Pleistocene, the Antarctic ice sheet had reached its maximum extent over the entire Quaternary period [*Verleyen et al., 2011*]. Only a few oases in East Antarctica remained ice-free (the Larsemann Hills and Bunger Oases) and the Dry Valleys [*Verkulich, 2011; Verleyen et al., 2011*]. The processes of weathering, transportation and accumulation of loose sediments were controlled by the specific low temperatures conditions. The formation of deposits was largely dictated by the processes of cryogenic weathering and transformations. The Antarc-

Table 1. **Temperature conditions of cryogenic weathering in the active layer of the Antarctic oases**

Study area	Years of observations	Characteristics of the daylight surface of soils				μ_{ef}	μ		
		$T_{m.an}$	τ_{col}	A_{max}	$T_{av.m}$		0 cm	20 cm	ALB
King George Island	2009–2010; 2011–2012	−1.6	264	24.6	+2.8 (Jan.)/−6.3 (Aug.)	7	32	n/a	n/a
Schirmacher Oasis	2009–2012	−9.5	268	66.1	+6.0 (Dec.)/−21.9 (Aug.)	32	111	n/a	n/a
Druzhnaya-4	2011–2013	−12.3	308	46	+1.8 (Dec.)/−21.1 (May)	48	114	n/a	n/a
Larsemann Hills Oasis (moraine line)	2012–2013	−11.0	296	43.2	+2.6 (Dec.)/−19.8 (July)	25	52	n/a	n/a
Larsemann Hills Oasis (valley floor)	2011–2013	−9.8	303	46.3	+2.0 (Dec.)/−17.2 (May)	63	81	10	1
Bunger Oasis (valley slope)	2011–2014	−10.5	342	39.3	−0.6 (Jan.)/−16.9 (June)	6	30	n/a	n/a

Notes to Tables 1, 2. $T_{m.an}$ – mean annual temperature of the day surface of soils, °C; τ_{col} – duration of the cold period with mean diurnal temperature of the day surface of soils less than 0 °C, day; A_{max} – maximum range of temperature on the day surface of soils, recorded in the observation period, °C; $T_{av.m}$ – annual (within the observation period) average monthly temperatures of the most warmest and coldest months, °C; μ_{ef} – number of effective freeze/thaw cycles on the day surface of soils; μ – transitions across 0 °C threshold; ALB – bottom of the active layer; n/a – no data available.

tica ice sheet had been decreasing since the Holocene and many coastal oases became free of ice, with the sea level fluctuation observable during this time. The periods of cooling were accompanied by the ice sheet surges [*Verkulich, 2011; Verleyen et al., 2011*].

Modern mean annual surface temperatures in the Antarctic oases are: −0.6...−1.0 °C for the north of the Antarctic Peninsula; −7.0...−8.0 °C for the Larsemann Hills and Bunger oases, Druzhnaya-4 field station; below −8.0 °C for Schirmacher Oasis; the lowest temperatures account for the inland oases (the Dry Valleys) and nunataks [*Abramov et al., 2011*]. Given that mean annual thawing depth in all the monitoring sites does not exceed 1 m, it largely depends on the date of loss of snow cover and influences of meltwaters. The active layer starts freezing mid-February, accompanied by its upward propagation from the permafrost table. The AL closure and freezeup is normally completed in late February–early March. Moisture content of AL proves low and is counted by a few first per cent. In the surface horizons (at a depth of 5–10 cm) daily freezing-thawing is caused by the change in diurnal insolation [*Abramov et al., 2011*].

Mean annual surface temperatures range between −7...−10 °C in the Antarctic oases (except King George Island), and their amplitudes can reach 60 °C (Table 1). A large number of temperature transitions across 0 °C threshold, up to 100 and more was observed on the surface within one year period, due to the persisting daylight during the polar day (Druzhnaya-4 Field Station area, and Schirmacher and Larsemann Hills Oases). The snow cover appears to be yet another factor contributing to the cryogenic weathering, so the temperature sensors are located in the Bunger oasis stream valley, which is covered by a snow field most of the time. Accordingly, the above 0 °C surface temperatures were recorded during 75 days only, over the three years of observations. Schirmacher Oasis is characterized by harsh environment conditions and the greatest ranges between summer and winter months, whereas good insolation of the surface contributes to long-lasting warming up during the summer (the temperatures can be as high as +20 °C, and more). The number of temperature crossing 0 °C threshold has reduced to a minimum (30) for King George Island and Bunger Hills Oasis.

In the first case, it was influenced mostly by a combination of the marine climate and low-amplitude temperatures, while in the second, by the snow cover. A great number of freeze-thaw effective cycles (from +2 to −2 °C) [*Matsuoka, 1990*] ranging from 30 (Schirmacher Oasis) to 50–60 (Druzhnaya-4, Larsemann Hills Oasis) were observed on the day surface, whereas in less severe conditions their number equaled 6–7 (King George Island, Bunger Hills Oasis) (Table 1). This parameter tends to decrease with depth and equals zero at a depth of 20 cm.

Pleistocene–Holocene sediments in the Larsemann Hills Oasis

The Larsemann Hills Oasis is located in East Antarctica on the Prydz Bay, its area is approximately 120 km^2. The age dating values derived from the lacustrine sediments suggest that at least partly the oasis avoided being affected by the glaciation in the past 40 thousand years. During MIS 2–MIS 3 the sea level was 30–40 m higher than at present, and a part of the oasis was flooded by sea [*Hodgson et al., 2009; Verleyen et al., 2011; Demidov et al., 2013*]. During 2011–2013 field woks, the drilling on the lakes Reid–Nell strait exposed the thickest section of loose sediments in Antarctica (Fig. 2).

The cross-section of the sediments consists of two units, represented by a terminal moraine overlying the lacustrine-lagoonal sediments dated to

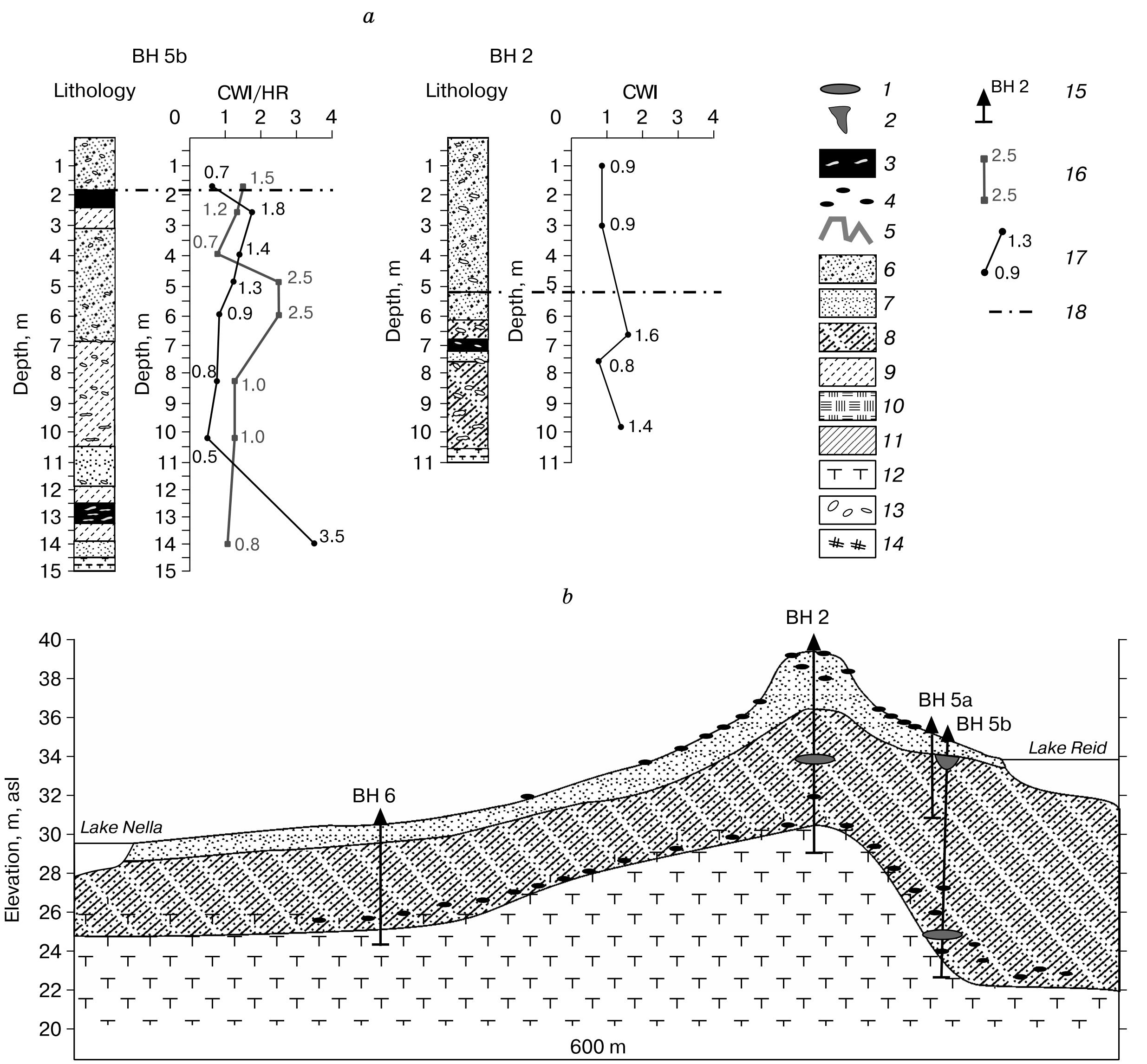


Fig. 2. Results of the Larsemann Hills Oasis deposits studies:

a – cryolithological structure of BH 2, 5b; *b* – the geological cross-section; *1* – ground ice bodies, *2* – ice wedges, *3* – ice-rich ground, *4* – boulders, *5* – baidzherakhs (relic mounds), *6* – coarse-grained sand, *7* – fine-grained sand, *8* – interbedding of sand and sand loam, *9* – sand loam, *10* – peaty sand loam, *11* – light clay loam, *12* – bed-rock, *13* – pebble inclusions, *14* – plant remains and peat inclusions, *15* – borehole (BH), *16* – heavy residue (HR), *17* – coefficient of cryogenic weathering (CWI), *18* – stratigraphical boundaries.

MIS 2–MIS 3, with inclusions of ground ice up to 1 m thick. The upper part of the cross-section is composed by coarse- and medium-grained sand with massive cryostructure and boulder inclusions.

On the surface, these sediments form an end moraine swell with a height of 5–7 m and a width of 30–50 meters, crossing the strait between the lakes [*Demidov et al., 2013*]. The lower part of the cross-section consists of rhythmically alternating sand and sand loam layers with inclusions of debris and gravel. A distinct cross-stratification with 30–45° dips, which is highlighted by up to 2–3 cm thick ice schlieren spaced at 5–10 cm in sand loam, and in sands – by discrete oblique ice schlieren (thickness: 1–2 cm) with distinctly sloping interlayers of garnet (up to 2–5 cm thick). The thicknesses of sand loam members range from 0.5 to 3.0 m, and those of sand members are 2–4 m. The contacts are formed by the horizontal cross-cutting unconformities containing large amount of gravel.

CWI value for the lower unit varies from 0.4 to 1.8, with the exception of the lowermost sample from borehole 5b, where its value is 3.5. The lower values, varying from 0.5 to 0.8, are characteristic of more dispersed (sand-loamy) sediments, whereas they range from 0.8 to 1.6 for coarser sediments. CWI values increase systematically from the bottom up, reaching 1.8 at a depth of 2.2–2.4 m in the sand loam member, underlying the buried ice strata.

CWI values in borehole 2 also tend to increase, reaching 1.6 at the top of the lower unit. The HR values are indicative of the elutriation throughout the deposition history, with HR values for sand growing slightly greater (2.5), than for sand-loam (1.0–1.2). At the same time, low (0.8) values of HR and high (3.5) CWI for sand occurring at 13.9–14.3 m depth interval are accounted for their nature being different. The abundance of heavy minerals proves appre-

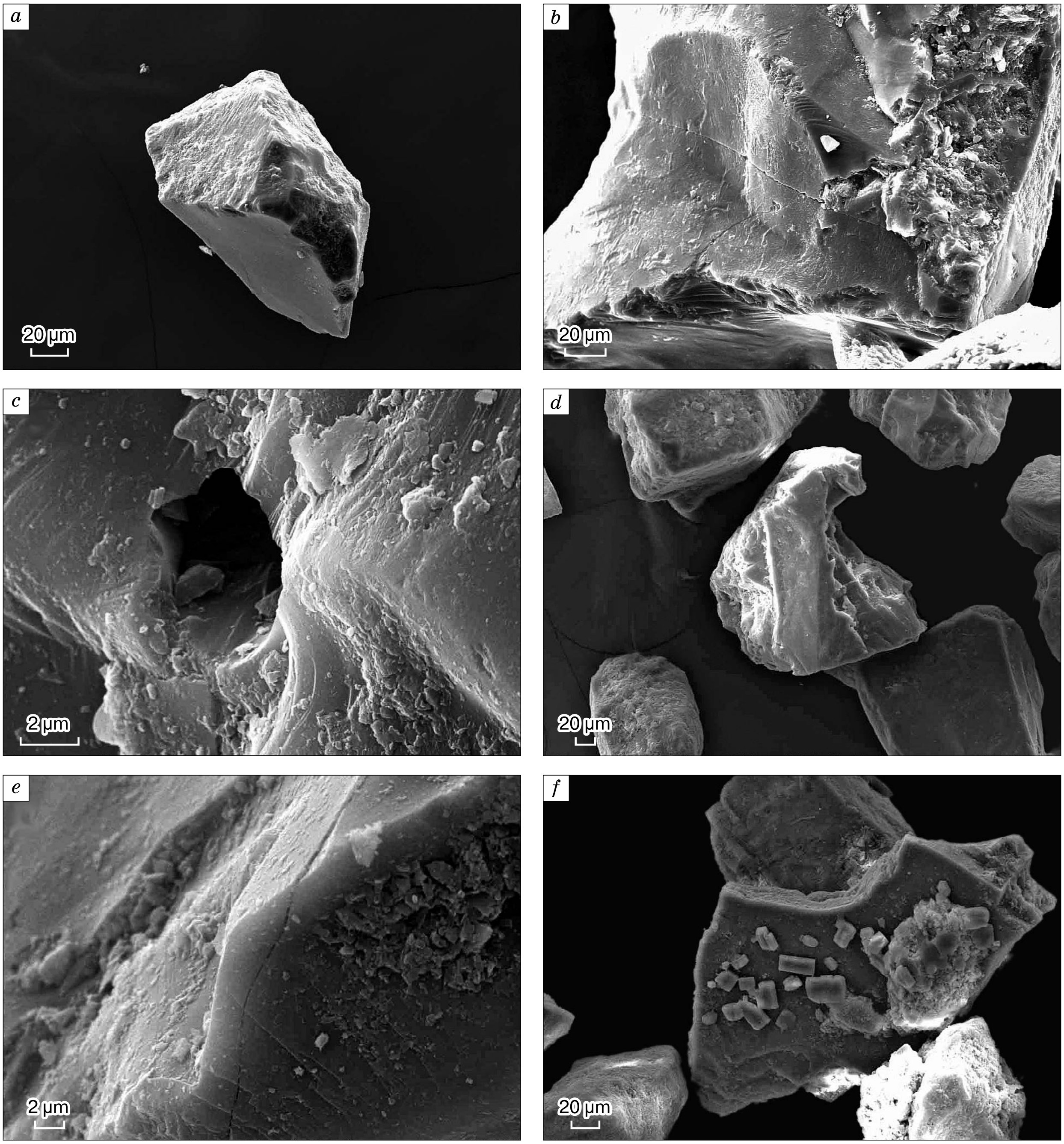


Fig. 3. Quartz grains morphology (Larsemann Hills Oasis):

a – main view of particle with fresh chip, *b* – cracks on the surface of quartz particle (*a*, *b* – sample from 2.2–2.4 m depth interval); *c* – cavern the surface of quartz particle, *d* – chip on the surface of particle (*c*, *d* – sample from 5.8 m depth); *e* – cracks on the surface of quartz particle, *f* – salt crystals on the particle surface (*e*, *f* – sample from 13.9–14.3 m depth interval).

ciable: in 100–50 μm fraction – from 12 to 28 %, in 50–10 μm fraction – from 13 to 55 % (garnet, marcasite, hornblende, grossular, siderite, pyrite). Their composition does not change throughout the entire cross-section, suggesting a single source of sediments. The shape of quartz particles varies from angular in the upper part of the section (average: 1.4–1.7, mode 1) to the angular-rounded, and even rounded in the middle (sandy) portion (average: 1.9–2.5, mode 2–3), and again becomes angular in the lower portion. More sharp-angled or subrounded particles are characteristic of samples with high CWI values, regardless of lithology. The particles bear the marks of multiple chips (Fig. 3, *a*), cracks (Fig. 3, *b*, *d*, *e*), small cavities (Fig. 3, *c*). Prism-shaped salt crystals (Fig. 3, *f*) were detected on the particles in the sample from 13.9–14.3 m depth interval.

Conditions of the cryogenic processes development in North-East Yakutia

The formation of permafrost in North-East Yakutia began in the late Neogene, which is evidenced by the presence of cryoturbation, ground ice, ground ice wedges, and pseudomorphs in different horizons [*Sher, 1971*]. Throughout the Pleistocene, the area was not subject to glaciation, given that the marine transgressions occurred only in the place of the modern rivers' mouths. The present-day topography had been formed by syncryogenic ice-rich sediments with thick polygonal ice-wedges, which constitute the Late Pleistocene Ice Complex (IC). IC aggradation was taking place in unique cryoarid tundra steppe (arid tundra-grassland) landscapes, featured by high biological productivity and diversity [*Sher, 1971; Arkhangelov et al., 1979; Schierrmeister et al., 2011*]. Within the modern estuary of the Kolyma Rv. accumulated silt deltaic sands, which compose the present-day accumulative delta plain. The sedimentation occurring at the Pleistocene–Holocene boundary produced the cover layer (CL). The thermokarst activity proved to be high at the beginning of the Holocene. The thawing of IC has lead to the formation of thaw lake basins and the accumulation of alas complex (AC) sediments [*Kaplina, 2009*].

At the present time, the permafrost is characterized as continuous in the northeastern parts of Yakutia. Ground temperatures range between –4...–11 °C, depending on the landscape conditions and tend to decrease northward [*Romanovsky et al., 2010*]. The lowest temperature of permafrost (about –10 °C) was recorded in the tundra zone (Chukochi Cape, Bykovsky Peninsula, Allaikha). Within the bounds of the tundra and forest patches of the Kolyma lowland, ground temperatures vary from –3 °C (the Omolon river mouth) to –5 °C (Akhmelo Lake) [*Romanovsky et al., 2010*]. In the northern tundra areas (Chukochi Cape, Bykovsky Peninsula, Allaikha) the active layer depth is from 30 to 50 cm, whereas in the southern parts of the tundra and thin forest areas (in the vicinity of Akhmelo Lake) its depth varies from 60 to 100 cm, and 50–60 cm – in the northern taiga (the Omolon Rv. mouth) [*Fedorov-Davydov et al., 2004*].

The AL freezing begins in late September – early October and may last as long as 1.5 months, with temperatures fluctuating around 0 °C at that time. Moisture content of the active layer varies significantly in different years and from site to site. Moisture content peaks in the near-surface horizons and at the bottom of the active layer, where moisture weight accounts for a minimum of 30 %.

Mean annual surface temperature varies from –6 to –10 °C with the coldest and warmest month amplitude ranging from 30 to 35 °C (Table 2). The number of transitions across 0 °C threshold on the surface varies from 4 on alas sites (Bykovsky Peninsula) to as many as 50 times on the Yedoma sites in some years (Chukochi Cape). Their quantity range is dictated by the landscape (vegetation) and by year-specific conditions (Cape Chukochi site: from 16 to 52, in different years). Their number decreases with depth (subject to yearly variations). The value of surface-bound effective cycles range from 3 (Chukochi Cape, Bykovsky Peninsula (alas)) to 5 (Allaikha, Lake Akhmelo).

Table 2. **Temperature conditions of cryogenic weathering in the active layer in North-East Yakutia**

Study area	Years of observations	Characteristics of the daylight surface of soils				μ_{ef}	μ		
		$T_{m.an}$	τ_{col}	A_{max}	$T_{av.m}$		0 cm	20 cm	ALB
Bykovsky (Yedoma, tundra)	2008–2011	–9.5	250	58.1	+6.5 (July)/–23.8 (Feb.)	5	22	2	3
Bykovsky (alas, tundra)	2009–2010	–9.4	254	39.8	+8.3 (July)/–22.6 (Feb.)	3	4	1	1
Allaikha (Yedoma, tundra)	2008–2013	–9.3	246	66.3	+9.6 (July)/–25.3 (Feb.)	5	32	3	4
Omolon mouth (Yedoma, forest)	2008–2010; 2012–2013	–3.4	233	40	+7.0 (July)/–14.9 (Feb.)	5	15	8	14
Akhmelo (tundra)	2012–2013	–5.3	246	43.6	+8.8 (July)/–23.3 (Feb.)	5	14	2	1
Chukochi Cape (Yedoma, tundra)	2006–2010	–8.8	250	60.6	+5.8 (July)/–23.9 (March)	3	31	6	3

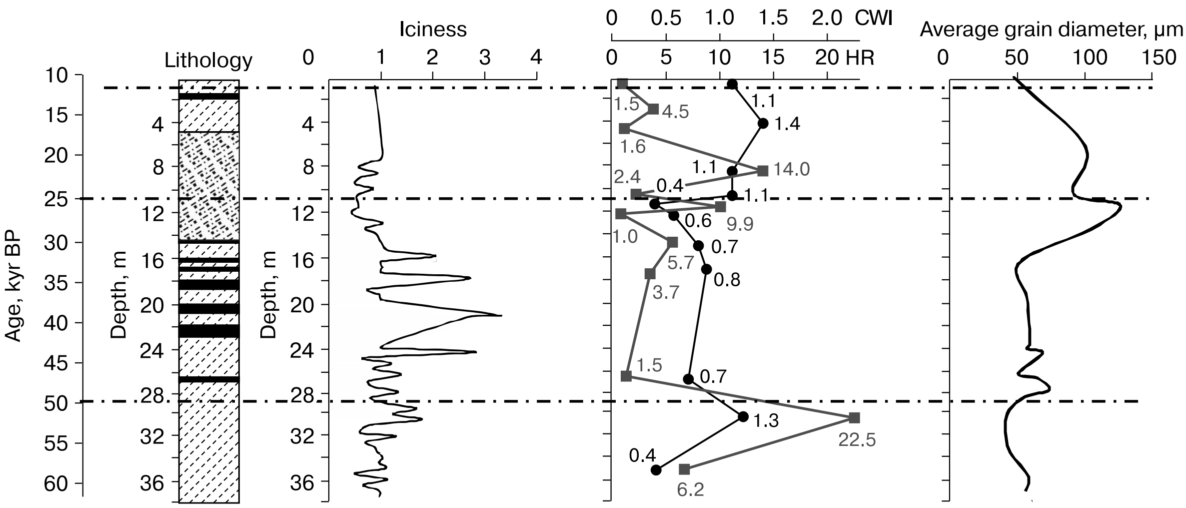
Ice Complex Sediments on the Bykovsky Peninsula

The Mamontov Khayt outcrops on the Bykovsky Peninsula are believed to represent the full-most IC cross-section for the Laptev Sea coast [*Schirrmeister et al., 2002, 2011; Sher et al., 2005*]. In 2001, BH 1/01 was drilled from the Yedoma residual outcrop, providing core samples as reference materials for the analyses (Fig. 4).

The samples age interpretation was done according to the data obtained by L. Shirrmeister [*Schirrmeister et al., 2002*], suggesting a relationship between the absolute height of sediment column and age of deposits. His research [*Schirrmeister et al., 2002*], A.V. Sher [*Sher et al., 2005*], supplemented by the data on sediments structure revealed by drilling, allowed to identify the following horizons within the IC sequence :

– the lower part of the cross-section formed approximately between 60 and 50 kyr BP (MIS 5–MIS 4) and is represented by clay-loams and sand-loams interbedded with horizontally stratified silty sands. Cryostructures are reticular, lenticular-reticular, ataxitic, breadcrust and massive, with ice content varying from 50 to 100 %;

– the middle part was formed between 50 and 25 kyr BP (MIS 3) and is composed of interbedded sand-loam with massive and thin-schlieren cryostructure, ice soil, autochthonous peat and buried soils. These layers are characterized by lenticular and non-



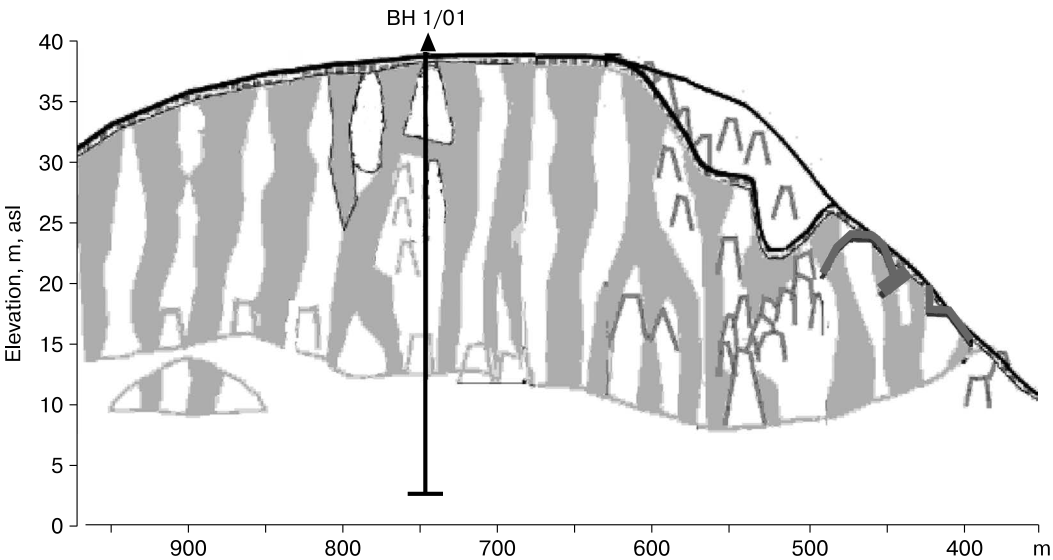


Fig. 4. Results of the deposits study on Bykovsky Peninsula.

a – cryolithological structure of BH 1/01; *b* – the geological cross-section of Mamontov Khayt [*Sher et al., 2005*] with marked location of BH 1/01. Legends cf. Fig. 2.

uniform bedding, rhythmic structure pattern, and are ice-rich;

– the upper part of IC accumulated between 25 and 12 kyr BP (MIS 2) and is represented by interbedded sand-loams, clay-loams and sands with sporadic grains of gravel; its cryostructure is layered and thin-schlieren, with ice content not exceeding 100 %.

During the episodes of IC accumulation the material was transported from the slopes of the Kharaurlakh Ridge [*Siegert et al., 2000; Schirrmeister et al., 2002*].

The role of cryogenic transformation of material has varied significantly throughout the period of Ice Complex depositional history. The upper unit is distinctly distinguished with CWI values ranging from 1.1 to 1.4 and is assigned by L. Schirrmeister [*Schirrmeister et al., 2002*] to MIS 2, as well as the middle member (MIS 3) with CWI < 1. The lower unit yielded two values of CWI: 1.3 and 0.4, and can be attributed to the lower part of Ice Complex dated to MIS 4–MIS 5. The content of heavy minerals within the sequence is less than 1.5 % (pyrite, hornblende, siderite). HR values indicate a good sorting of sediments throughout the deposition episodes, which is why the suggested stratigraphic boundaries between the IC members are marked by high HR values (14.0; 9.9; 22.5).

In terms of the grain-size distribution the upper member of Ice Complex appears more coarse (average diameter: about 100 μm) than the underlying sediments, with its stratigraphic boundaries marked by peaks of increased grain-size of sediments. The behavior of grain-size curves also exhibit difference between

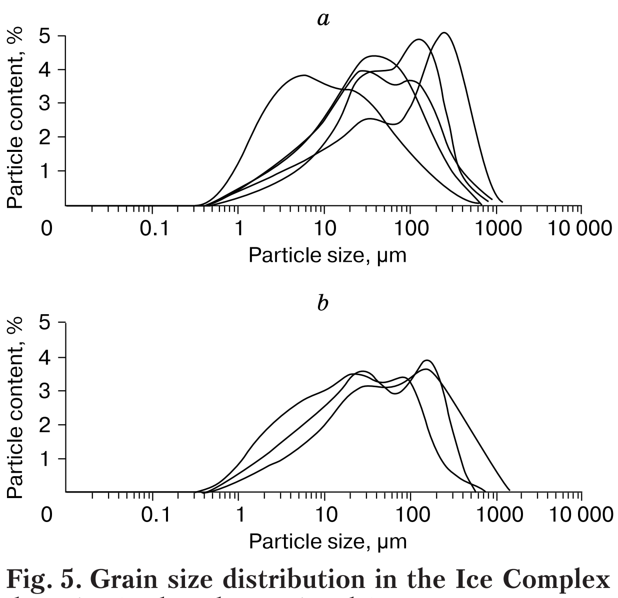


Fig. 5. Grain size distribution in the Ice Complex deposits (Bykovsky Peninsula).

a – the upper member: CWI > 1; *b* – the middle member: CWI < 1.

the upper and middle members (Fig. 5). The second peak, shifted to 100–10 μm fractions, is explicitly expressed in diagrams for samples with CWI > 1, despite the coarser grain structure, which is accentuated by the 500–100 μm peak, however, sometimes only one peak accounts for this fraction. At the same time, more flattened curves for the middle unit appear similar to each other, with the peak approximating values between 100–50 μm.

The shape of quartz particles varies from angular-rounded in the upper unit (average: 1.6–2.0, mode 2) to rounded in the middle unit (average: 2.1–2.8, mode 3). There's a great number of cryogenic aggregates that often have "chipped cavities" (Fig. 6, *c*). Cryogenic-weathering induced deformations in quartz grains manifesting themselves as fractures on their surface (Fig. 6, *b*) and particles torn "from the inside" (Fig. 6, *d*) were encountered in the deposits of the upper member of IC. Grains with fresh chippings (Fig. 6, *a*) are also observable therein. The middle level of deposits is dominated by cryogenic aggregates (Fig. 6, *e*), but they also show sporadic grains of quartz with traces of obvious cryogenic weathering (Fig. 6, *f*). The results agree fairly well with the earlier data on the morphology of grains in LC and support the inference made on the role of cryogenesis in the formation of deposits, particularly in the upper unit [*Siegert et al., 2000*].

The deposits of ice and alas complexes at Cape Chukochi

The outcrop at Chukochi Cape was described by *A.A. Arkhangelov et al.* [*1979*] and *Yu.V. Kuznetsov* [*1979*]. It represents the sedimentary cross-section of both Ice and Alas Complexes. The upper horizon of brown silts 10–13 m thick (Fig. 7, horizon **a**) with horizontal layering and rhythmic pattern is displayed in the outcrop. The succeeding layer is 8–9 m thick, its upper limit is marked by the sand interlayer (thickness: 0.20–0.45 m) at a height of 14–15 meters above the river level (Fig. 7, horizon **b**). The cross-section displays a gradual transition from brown silts to the interlayer of sands. Rhythmic alternation of units with various degrees of peat content increase and sanding become distinctly visible; buried thin ice-soil and ice veins are also encountered. Sands are brownish-yellow in color. They are underlain by dark gray silts, prone to sanding at times, with a lot of autochthonous roots. Thickness: 2–4 m. Cryostructure: massive- or micro-schlieren; syngenetic and epigenetic ice veins were encountered. The middle unit is underlain by ice-rich horizon of yellow-green silts (Fig. 7, horizon **c**). It contains lenses and interlayers of poorly decomposed peat. Lamination is not explicitly expressed. Cryostructures are lenticular and reticulate with thick parallel schlieren, with observable ice-soil interlayers up to 10 cm thick [*Arkhangelov et al., 1979*].

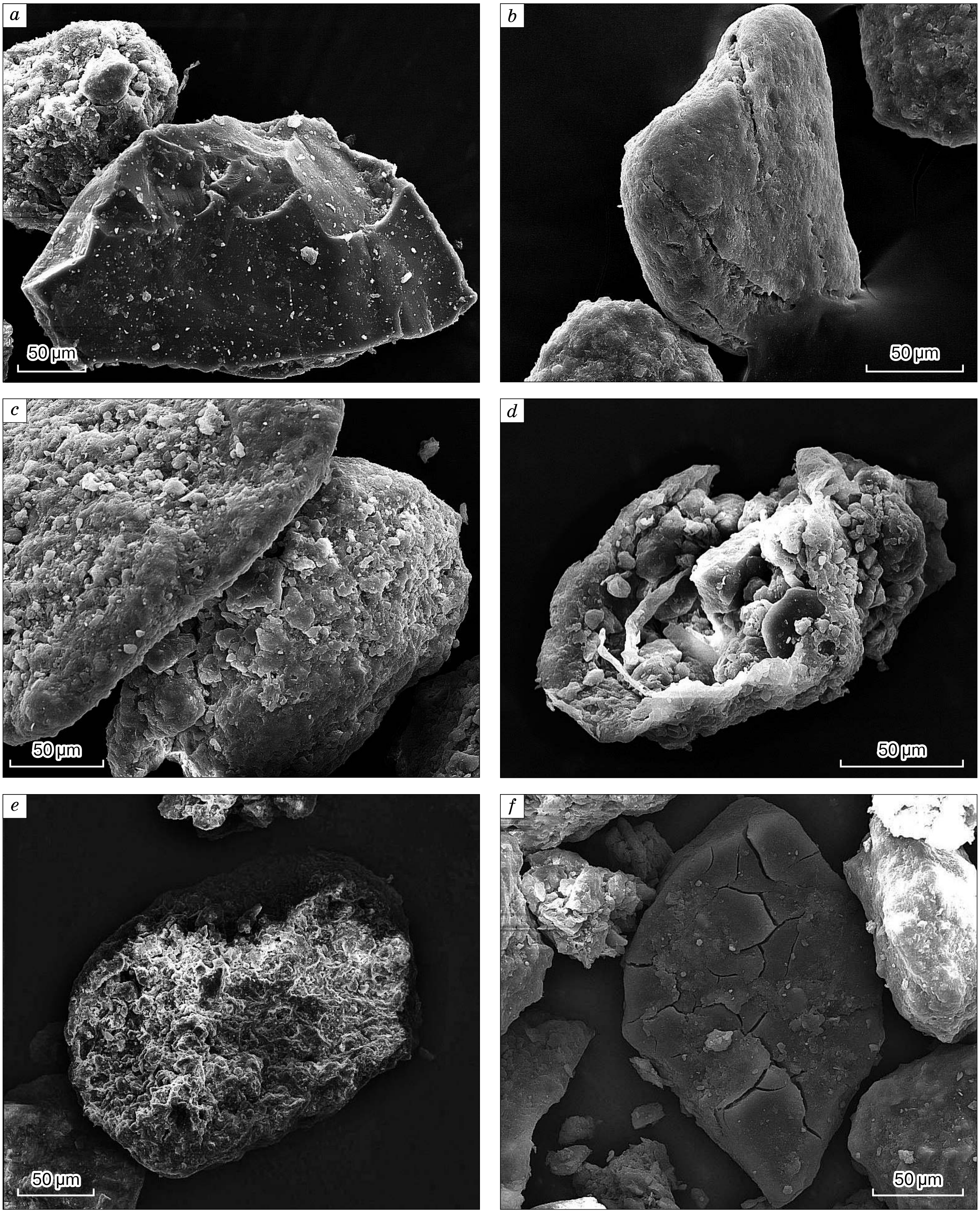


Fig. 6. Quartz grains morphology (Bykovsky Peninsula):

a – fresh chips on the surface of quartz particle (depth: 0.5–0.7 m); *b* – cracks on the surface of quartz grain (depth: 4.8 m); *c* – "chipped cavities" on the aggregate (depth: 10.5 m), *d* – " shattered" grain of quartz (depth: 10.5 m); *e* – cryogenic aggregate (depth: 11.5 m); *f* – cracks on the surface of particle (depth: 15 m).

The radiocarbon dating provided the following ages: for the middle unit – 39,000, for upper brown silts – 33,200 ± 200 years, for the lacustrine sediments in the alas basin – 8,400 ± 100 years [*Kuznetsov, 1979*].

When drilling the area in the summer 1991, the sediments of both IC (horizon **a**) and Alas Complexe (AC) were sampled. The alas deposits are represented by silts with interlayers of peat and organic matter, with traces of cryoturbation, their cryostructure is layered and reticulated (lacustrine – boggy deposits – horizon **e**, Fig. 7). The deposits are underlain by sandy silts and fine sands with fragments of wood and traces of cryoturbation and solifluxion flow, and cross-bedded massive cryostructure (taberal portion: horizon **d**, Fig. 7).

All the samples are characterized by uniform mineral composition. About 90 % of the composition of light fraction accounts for quartz and feldspar, heavy minerals content does not exceed 5 %, averaging at 1–2 % (hornblende, siderite, and pyrite).

The following CWI values were obtained for the upper level of IC deposits (horizon **a**): 0.9 – gray-brown peaty silts underlying ice lens of; 2.3 – gray-brown peaty icy silts, overlying the ice-soil interlayer; 0.7 – sandy gray silts with massive cryostructure, underlying ice interlayer. The CWI values, thus, change cyclically in horizon **a**, and spread widely (from non-cryogenic to typically cryogenic values), which is governed by cyclically changing landscapes and conditions of the deposits transformation into permafrost state.

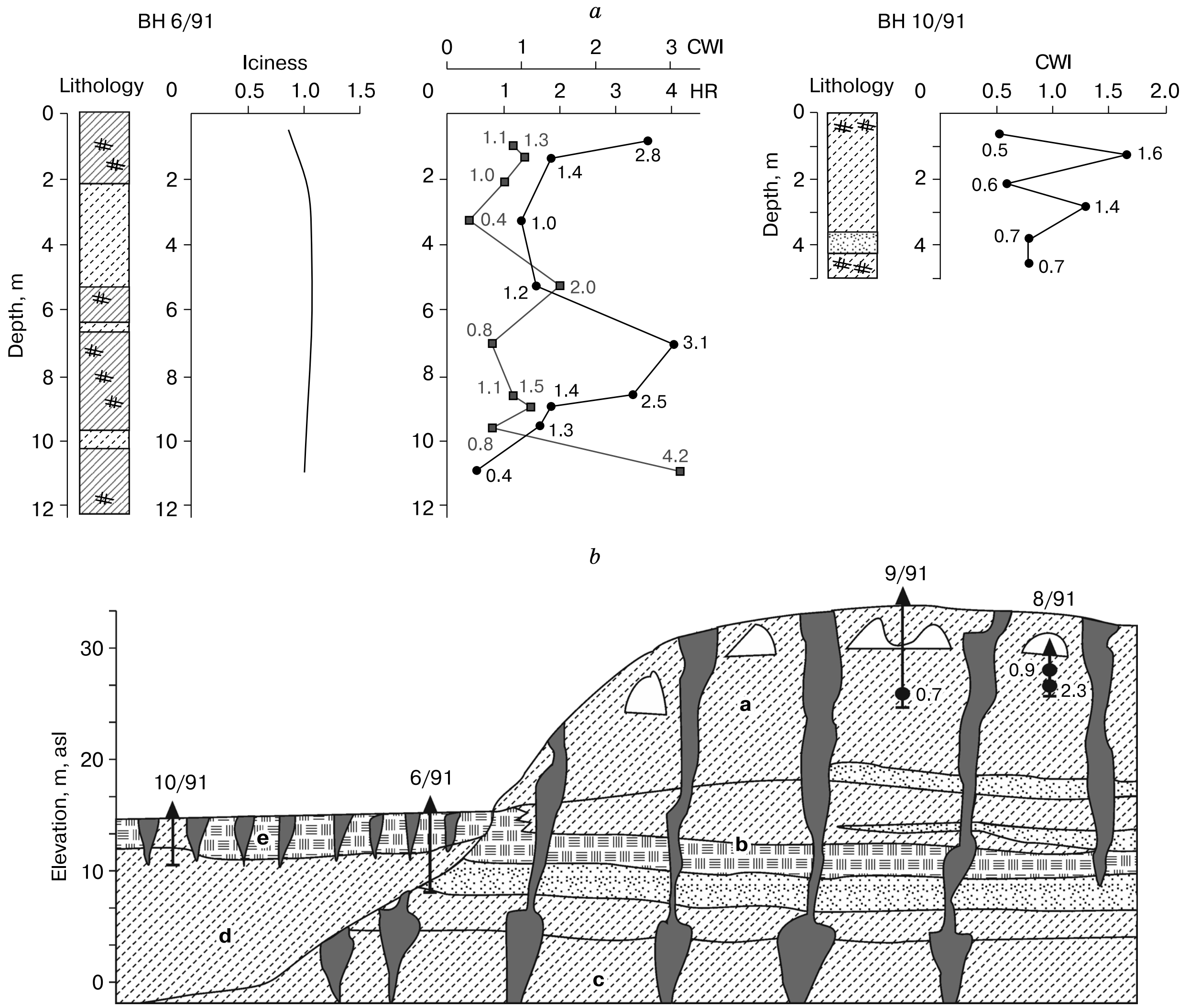


Fig. 7. Results of the Chukochi Cape deposits study.

a – cryolithological structure of BH. 6/91, 10/91; *b* – the geological cross-section of Chikochi Cape [*Arkhangelov et al.,1979*] on the basis of field records during the “Beringia” expedition. **a, b, c, d, e** – lithological horizons (as detailed in the paper). Legends cf. Fig. 2.

The alas complex was studied by drilling 2 boreholes. One of them (BH 10/91) is drilled directly in the valley between two thermokarst lakes, whereas the second hole (BH 6/91) – not far from the Yedoma slope. The first borehole profile is characterized by a diversity in lithological composition, from sand interlayers to silts and peaty horizons, with CWI values cyclically varying from 0.5 to 1.6 at 5 m depth.

Borehole 6/91 is characterized by a more uniform composition, featured by alternating light loam and sandy loam, with peat enrichment and cryoturbation traces and ice content ranging between 0.9–1.1. In terms of CWI values distribution, the two horizon can be distinguished: upper (up to 6 m) where CWI is from 1.0 to 2.8 (arithmetic mean: 1.6) and lower (6–10 m) where CWI ranges from 1.3 to 3.1 (arithmetic mean: 2.1), at the very bottom CWI value equals 0.4. HR values average at 1, varying from 0.8 to 2.0 for the alas sequence.

General view of the grain-size distribution curve for the deposits of IC and AC appear very similar, indicating that there exist a genetic relationship between them. At the same time, the curves are distinctly different for samples with low and high CWI values (Fig. 8). The deposits subjected to the active cryogenic processing, are more dispersed and better sorted, and the curve is more symmetrical, peaking in the 50–10 μm fraction. The curve is asymmetric for samples with CWI <1, and its line is flatter.

Angular and angular-rounded grain shapes (average: 1.6–2.2, mode 1) predominate in deposits with high CWI, however, among them rounded and well-rounded (23 % of total) grains are also encountered. Rounded grain shapes (average 2.0–2.3, mode 3) are characteristic for deposits with CWI <1 (including sample from 10.8–11.5 m depth interval), with the angular shaped grains adding up to 26 %. A large number of quartz grains with fresh chips were observed, with one surface often covered with aggregates, while the other tends to be completely clean (Fig. 9, *a*, *b*). Great many of these particles were found in the samples from BH 6/91 at a depth of 0.8–1.1 m. They showed a lot of chips and cracks (Fig. 9, *d*), which often cut a particle apart into several pieces (Fig. 9, *c*). The sampled lacustrine-boggy sediments exhibited a large number of cryogenic aggregates (Fig. 9, *e, f*) of rounded shape, but some of them have "chipped cavities" (Fig. 9, *e*). Samples with low values of CWI are characterized by a better preservation of quartz grains than samples with high CWI.

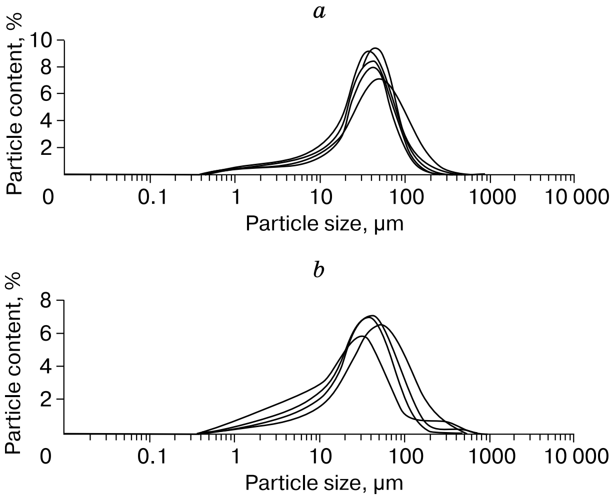


Fig. 8. Grain size distribution of Alas and Ice Complexes (Chukochi Cape).

a – deposits with CWI > 1; *b* – deposits with CWI < 1.

INFLUENCES OF CRYOGENIC PROCESSES ON SEDIMENTS COMPOSITION IN THE LATE PLEISTOCENE AND HOLOCENE

The modern natural conditions and their influence on cryogenesis development

The study areas are characterized by severe geocryological conditions that favor for the cryogenic weathering processes. Given that warm period can be up to 100–120 days long in NE Yakutia, the AL freeze-up takes longer than in the Antarctic oasis. These areas are characterized by a great number of transitions across 0 °C threshold, from a few tens to hundreds.

Antarctic areas are normally subject to explicit rapid freezing from the bottom of AL, whereas in NE Yakutia the "zero veil" effect may last for a long time in the lower part of the AL profile, and as many as ten freeze-thaw cycles may take place at depths up to 20 cm during the freezing episodes. The amount of effective freeze-thaw cycles differs in the oases of Antarctica and in the northeastern parts of Yakutia, which is controlled by both climatic and landscape factors.

In Antarctica, a major transformation of the mineral part occurs in the subsurface horizons, driven by volumetric (hydrostatic)-gradient stress in the "particle – film of water – ice" system induced by temperature drops (effective freeze-thaw cycles). In NE Yakutia , transformation of the material part is largely attributed to the "zero veil" responsible for the prolonged freezing period. The diversity of mineral composition transformations is reflected in the morphology of mineral grains. In the samples from Larsemann Oasis a large number of sharp-angled particles are encountered (10 % of all the examined ones).

Mineral grains are subject to and bear the evidence of such deformations, as: chips, sharp edges, linear cracks. In North-East Yakutia, cryogenic disintegration is associated with the rupture of gas-liquid inclusions and destruction of cryogenic aggregates. Despite the abundance of deformations, like caverns, cavities, variously-shaped "chipped cavities" and ring-type cracks, the shape of grains tends to be smoother and rounded, though.

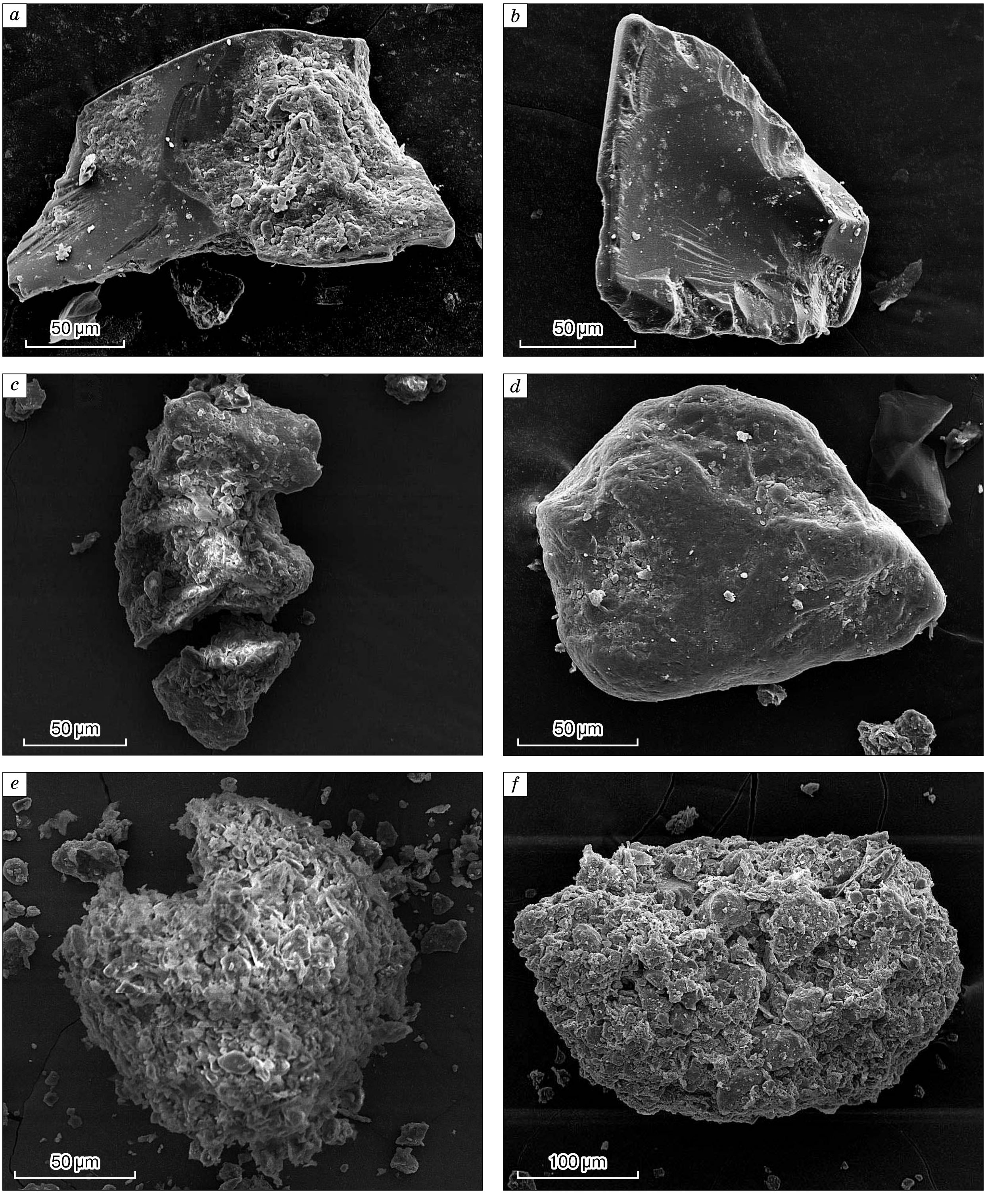


Fig. 9. Quartz grains morphology (Chukochi Cape):

a, *b* – fresh chips on the surface of quartz (BH 6/91, depth: 0.8–1.1 m); *c* – quartz grain “shattered” into two parts (BH 6/91, depth: 7.5 m); *d* – rounded quarts grain (BH 6/91, depth: 10.8–11.3 m); *e* – “ chipped cavity” in rounded aggregate (BH 10/91, depth: 1.8–2.5 m); *f* – cryogenic aggregate (BH 10/91, depth: 4.0–4.5 m).

Cyclic structure of Late Quaternary cross-sections, controlled by changes in the depositional environments and sediments freezing conditions

The cyclic structure of Ice Complex deposits in NE Yakutia has been pointed out and studied by many researchers. Rhythmic pattern of lithology and structure of polygonal ice wedges was marked by *E.M. Katasonov* [*2009*], referring to the description of the Mus-Khaya, at Cape Chukochi exposure by *A.A. Arkhangelov et al.* [*1979*]. *Yu.K. Vasil'chuk* [*2006*] discussed in detail the heterocyclic structure of polygonal ice wedges of IC and distinguished three levels of cycles: micro- (changes in the seasonal thaw depth), meso- (fluctuations of the water-level in water bodies) and macro- (climate change and related fundamental restructuring of sedimentation processes) levels. The cyclic structure of IC at the level of cryogenic processing of mineral substances, driven by climatic fluctuations was emphasized by *V.N. Konishchev* [*2013*].

The comprehensive studies of deposits in Larsemann Oasis, on Chukochi Cape, and Bykovsky Peninsula also allowed to define several types of the cycles.

1. *Cycles caused by the change of climatic epochs.* They are most distinctly expressed in the IC deposits on Bykovsky Peninsula. There can be distinguished the middle member attributed to MIS 3, and the upper member related to MIS 2 [*Schirrmeister et al., 2002*]. Sedimentation occurred in the warm interstadial environments during MIS 3.

During the summer season, soils warmed up to high temperature values, while winter temperatures remained cooler than the present-day's [*Schirrmeister et al., 2002; Sher et al., 2005*]. Peatlands are developed widely [*Schirrmeister et al., 2002*]. Ice content of the deposits attests to essential moisture content in the active layer. When transitions across 0 °C threshold reached their lower bound during the year, the soil thermal regime proved close to the alas sites in the study area (Table 2). Pedogenesis has been extensive within the area.

These conditions are reflected in the low values of CWI (0.4–0.7), suggesting that the delivered material had failed to undergo the cryogenic processing before its transition into the permafrost state. The time period of 25–12 kyr BP (MIS 2) is characterized as extremely cold and dry [*Schirrmeister et al., 2002*].

Nevertheless, the summer temperatures appear persistently higher, which provids for deeper thawing [*Kaplina, 2009*]. The thermal regime of the active layer is similar to the modern Yedoma sites in the area (Table 2), more favorable for cryogenesis, which is illustrated by the CWI values (1.1 to 1.4). Coarse grain size is indicative of more dynamic depositional environments. These conditions are preserved throughout the sedimentation in the upper member of IC.

2. *Cycles governed by the sea level change.* These appear most pronounced in the Larsemann Hills Oasis, which is lithologically corroborated by interbedding of sands and sand loams. The accumulation of lacustrine-lagoonal sediments is apparently related to the sea level change in the Late Pleistocene, except for the lower-most part of the sequence (Fig. 2), since it formed in the aerial or subaerial environments and represents by itself cryogenic eluvium, which is evidenced by the CWI, HR values, and particle shape.

Sand loams of the lower member (to a depth of 7.0 m) formed at a higher sea level in the MIS 3 era [*Hodgson et al., 2009; Verleyen et al., 2011*], with the seawater and glacier meltwaters intermingling [*Demidov et al., 2013*]. At this stage the freezing graduates into syncryogenic and subaqueous type, expanding inward from the coast and from beneath [*Romanovskii, 1993*]. The process developed fairly slow, with hardly any temperature fluctuations recorded. This therefore throws light on the lithological layering matching cryogenic textures. The full freeze-up of the member was accomplished already in the next stage.

Accumulation of the upper sandy part was associated with the sea level decreasing, gradually [*Verleyen et al., 2011*]. The freezing occured already by subaerial syncryogenic type, with the role of cryogenesis growing therein. At this stage, the connection with the sea has ceased, the material transport from the glacier by meltwater diminishes, while the erosion base level rises.

This has prompted smoother sedimentation conditions, which is reflected in the decreasing values of HR and increased precipitates dispersion, while they are exceedingly subject to cryogenic reworking. Later on, the sequence becomes overlain by glacial moraine deposits, which formation is likely to have been controlled by the events of MIS 2 or of the Late Holocene [*Verleyen et al., 2011; Demidov et al., 2013*].

3. *Cycles ensuing from thermokarst lakes discharge and filling.* Fluctuations of water-level in the water bodies (discharging and filling) are most explicitly illustrated by the example of the AC deposits on Cape Chukochi. CWI values in BH 10/91, located at the bottom of the alas basin between the lakes, allowed to determine the stages of filling, discharging and vegetal invasion of lakes. The deposits formed in hydromorphic conditions with low CWI values, the latter tend to be growing, though, during the shallowing and lakes and their transition into subaerial conditions. A rhythmic pattern of sedimentation is clearly expressed in the accumulation of the lacustrine-boggy member in AC.

Borehole 6/91 is located on a high level of the alas formed at the beginning of the Holocene, during the first surge of thermokarst processes. Radiocarbon dating analysis allowed to determine the events timing for the area, at the turn of 10.5–10.0 kyr BP [*Kap-*

lina, 2009]. The lower part of the sequence characterized by the lowest CWI is a taberal unit, i.e. thawed, re-deposited and re-frozen Ice Complex, providing the best representation of the coldest peak of its formation. At this, the difference in CWI values were levelled as thermokarst – related processes developed.

The freeze-back of taberal layer is classified as epigenetic (parasyngenetic), and accomplished over very short period of time (several decades). The overlying unit with lower though cryogenic values of CWI (i.e. CWI > 1), is composed of lucustrine-boggy deposits. Given CWI values typify substantially warmer Holocene environments, some of them (1.0–1.2) can be assigned to a time when the thermokarst lake was exposed to minimal cryogenic impact.

A further increase in CWI values up-hole due to discharge of water from the lake with the surface became exposed to the subaerial and aerial conditions, more favorable to cryogenesis. The value of CWI equal to 2.8 at a depth of 1.1 m (cover layer, CL), are consistent with the conditions at the turn of the Holocene–Pleistocene.

4. *Cycles driven by variations in seasonal thaw depths.* Another factor that determines the degree of cryogenic processing of the deposits are the interannual fluctuations of seasonal thaw depths and related cryogenic rhythms featured by alternating layers with high and low ice content and differing in cryostructure.

When basal part of the active layer is exposed to subaerial syncryogenic freezing it is likely to result either in slow transition into frozen state to form a massive cryostructure, or in the permafrost table stepwise aggrading and rising due to ice segregation. This being the case, schlieren with thickness up to 10–12 cm may develop [*Romanovskii, 1993*]. This process is most pronounced within the IC deposits on Chukochi Cape, where horizons are characterized by higher ice content contributed by ice segregation, and greater values of CWI (up to 2.3). The underlying deposits show lower CWI (sometimes non-cryogenic, i.e. less than 1) values, though, due to their transition into frozen state by means of stepwise freezing.

Cryogenesis evolution in the Pleistocene–Holocene in East Antarctica and NE Yakutia

In the Northern Hemisphere, the traces of the permafrost dated to the Pliocene [*Sher, 1971*] have been encountered in the form of pseudomorphs, cryoturbations in the Kolyma lowland and Chukotka Peninsula areas. In the Southern Hemisphere, their inception is deemed to have occurred simultaneously with the formation of the Antarctic ice sheet, which commenced in the Eocene–Oligocene [*Hambrey et al., 1989; Ehrmann and Mackensen, 1992*].

Despite the permafrost in Antarctica has a longer history of development, the highest CWI values are attributed to NE Yakutia (up to 3.1). This accounts for the fact that the Antarctic oases were most time either covered by the ice sheet, or by the encroaching sea, therefore, the cryogenic weathering processes had often been suppressed. Cryogenesis is confined specifically to the time windows when the oasis is already free from the sea, but not yet occupied by the glacier.

The area of NE Yakutia during the entire period of the permafrost existence was free from ice cover, sea transgressions were limited to the river mouths. It noteworthy that CWI values for the Chukochi Cape tend to be far greater than for Bykovsky Peninsula owing to the slower sedimentation and more severe geocryological conditions in the Late Pleistocene.

The end of the Late Pleistocene (MIS 2) – Last Glacial Maximum (LGM) appears the coldest time period in the Quaternary [*Shakun and Carlson, 2010*]. According to the stable isotope content in wedge ice, winter temperatures in the Arctic prove to have been lower by 10–15 °C than modern temperatures [*Vasil'chuk, 2006*]. The formation of site-specificicy syncryogenic strata and the Yedoma superhorizon in the deposits of Ice Complex in NE Asia belongs to this period as well [*Sher, 1971; Schirrmeister et al., 2011*]. In Antarctica, the ice-sheet sizes proved the largest over the Cenozoic. Only a limited number of coastal oases in East Antarctica and the Dry Valleys remained ice-free [*Verkulich, 2011; Verleyen et al., 2011*].

Despite the paleogeographic settings in the Pleistocene–Holocene time were different in the oases of Antarctica and in NE Yakutia, in entire cross-sections an increase in CWI values is recorded along the section, with observable escalation to the end of the Late Pleistocene. At Chukochi Cape the highest values are attributable to Cover Layer (2.8), the taberal part of AC (3.1) and brown silts of IC (2.3).

An increase in average values of CWI is observable on Bykovsky Peninsula from the middle unit of the cross-section (MIS 3 – from 0.4 to 0.7) to the top (MIS 2 – from 1.1 to 1.4). The maximum values of CWI for Larsemann Oasis are also confined to the upper part of the lacustrine-lagoonal sand-loam member, reaching 1.8. Similar patterns are observable in the morphology of mineral grains, as the most angular, shattered particles are confined to the LGM deposits.

This specificity is observable in all types of sediments, regardless of their composition and origin. It is not dependent on lithology, either. So, on Bykovsky Peninsula the upper unit is much less dispersed than the middle part of IC, whereas the situation is the opposite in Larsemann Oasis.

The results obtained on the distribution of CWI values and on the changing role of cryogenesis in the formation of deposits agree with other similar research on the basis of application of the CWI calculation method conducted on the areas of: the Chukotka

Peninsula [*Schwamborn et al., 2007*], the Yana-Indigirka lowlands [*Konishchev, 1981, 2013*], the Kolyma lowlands [*Konishchev, 2013; Shmelev et al., 2013*], the tundra of Bolshaya Zemlya [*Konishchev, 1981*], Central Europe, and the Russian Plain [*Konishchev et al., 2006*]. All the areas of the Northern Hemisphere revealed the highest values of CWI in deposits dated to the end of the Late Pleistocene, with this trend being persistent in the areas beyond the present-day permafrost zone (Central Europe and the Russian Plain). From this it follows that throughout all the Quaternary period the role of cryogenic processes in the formation of deposits composition proved to consistently increase, reaching a maximum at the Pleistocene–Holocene boundary, which is associated with the global climate cooling in that time period [*Shakun and Carlson, 2010*].

CONCLUSIONS

The investigations of conditions and results of cryogenic destruction of mineral matter in the Pleistocene-Holocene in the Larsemann Hills Oasis (Antarctica) and in NE Yakutia allowed the inferences as described below.

1. The two mechanisms of cryogenic weathering of mineral matter have been established. With regard to Antarctic oases, the subsurface zone (the topmost 20 cm) is subject to the most intense weathering, in the context of the greatest number of transitions over 0 °C threshold (up to 100 times, and more) and effective freeze-thaw cycles (50–60), which is associated with volumetric-gradient stress, originated in the “particle – film of water–ice” system. This results in the formation of sharp-angular particles that show rough and straight chips. In NE Yakutia the weathering runs parallel to the long lasting freezing of the active layer, which is largely responsible for changes in film thickness of bound water and ruptures of gas-liquid inclusions. The implications are that disintegrated cryogenic aggregates and chipped grains predominate en masse.

2. The Late Quaternary deposits of the oases of Antarctica and NE Yakutia are characterized by cyclic cryogenic processing of the material, which is explicit in the cross-section. This cyclicity stemmed from systematic changes (cycles) in the conditions of sedimentation and freezing of rocks: climate and sea level changes in different geological epochs, thermokarst lakes discharge and filling, interannual variations of seasonal thaw depths.

3. Despite the longer time range of the permafrost existence in Antarctica, the deposits in NE Yakutia were subjected to more intense cryogenic transformation (CWI: less than 3.1), due to the absence of the ice cover and marine transgressions in the Quaternary. Given the fairly rapid sedimentation rates and softer geocryological conditions during accumulation of Ice Complex on the Bykovsky Peninsula, CWI values tend to be much lower (from 0.4 to 1.4) than for IC on Chukochi Cape (from 0.7 to 2.3).

4. The highest CWI values for the three study areas are confined to the deposits formed at the Pleistocene–Holocene boundary. Despite the different specific features of the natural environment in the Late Quaternary in the Northern and Southern Hemispheres, the conditions for cryogenic processes were the most favorable at the end of MIS 2 in NE Yakutia and in Antarctica.

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