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CLIMATE, SEA LEVEL AND GLACIATION CHANGES IN THE MARGINAL ZONE OF ANTARCTICA DURING THE LAST 50 000 YEARS

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The article integrates the results of half a century studies of Late Pleistocene–Holocene changes in climate, sea level, and glaciation in the marginal zone of Antarctica in order to identify the chronology, parameters, and mechanisms of these changes under the influence of global, regional, and local factors. During the interstadial (MIS 3), the natural conditions here resembled modern ones, and the sea level in some areas exceeded modern marks. The development of glaciation of the marginal zone from about 26 000 years BP went on when the temperature fell and the sea level dropped by 30–50 m. The growth of glaciation on the shelf outpaced the growth of ice on the outskirts of the continent leading to a moisture deficit in the interior regions. During the LGM, there was a thin (less than 300 m) glaciation of coastal and mountainous land areas, and a thick (more than 1000 m) glaciation on the shelf. Deglaciation of the marginal zone began about 17 000 years BP due to rising sea level and global warming. Holocene climate changes in most areas had a general trend: warming in the early Holocene to about 8000 years BP and 4000–2000 years BP, cooling 2000–1500 years BP, but also had local differences. The relative sea level rose in the regions from the Early Holocene to 8000–6000 years BP; then it fell with a decrease in speed or even with a possible rise about 2500–1300 years BP. Local differences in the amplitudes and direction of sea level changes were determined by local tectonics and dynamics of deglaciation. Deglaciation rates were high from the Early Holocene to about 7500 years BP due to warming and marine transgression and then decreased. The advance of outlet and shelf glaciers 6500 and 4500 years BP was associated with a decrease in sea level and cooling. In the period 4000–1000 years BP, outlet and shelf glaciers could also respond to changes in sea level, and ice domes expanded according to the "warming – increasing humidity – increasing snow and ice accumulation" pattern. During the Little Ice Age, moraines were created in some areas indicating a slight increase in glaciers due to cooling.

Keywords: *marginal zone of Antarctica, climate, sea level, glaciation, interstadial, last glacial maximum, Holocene, relief, Quaternary deposits, paleogeographic reconstruction*

INTRODUCTION

Marginal zone of Antarctica extends from the coast to 200–300 km inland and includes shelf and outlet glaciers and nearby islands. It differs from the glacial region within the continent by a significant influence of cyclonic activity and adjacent marine areas and, in general, by a warmer, wetter, and windy climate with increased accumulation of snow. In this zone, large irregularities of the bedrock and sea level largely determine the diversity of glaciers, borders of the continent, and presence of ice-free land areas. The mass exchange of Antarctic glaciation is the most dynamic here: ice losses account for almost 98 % of the total annual ice loss due to melt runoff, wind drift, melting at the bottom of shelf glaciers, and iceberg chipping [*Kotlyakov et al., 2003*]. Under the mutual influence of the components of the atmosphere–ocean–glacier system, the marginal zone has experienced the most rapid and large-scale restructuring of natural environments over the last 50 000 years following global climate changes, glaciations, and sea level changes. In turn, these reconstructions were important for the evolution of the Antarctic glaciation as a whole, and therefore for the global climate and the balance of water in the World Ocean.

Transformations of the marginal zone are baked in the relief and sediments of the ice-free territories (Fig. 1) – Antarctic oases [*Sokratova, 2007*]. The slope of the ice sheet is broken through by mountain oases with mean annual air temperatures (MAAT) below –20 °C. Mountain-valley oases stretch from the glacial slope to the coast. These oases are dry and windy; the MAAT is still about –20 °C, but summer temperatures in the valleys reach –1 °C. The land and sea bays of low-lying onshore oases are separated from the ocean by shelf glaciers and are bordered from the other side continental ice sheet and outlet glaciers. The MAAT in them is about –11 °C, and summer soil temperatures may be above 15 °C, air humidity is 50–55 %, and annual precipitation is less than 200 mm. The low-lying coastal oases are only partially bordered by glaciers, and the MAAT in them is about –10 °C; in the summer, mean air temperatures are up to 1 °C (often. above 10 °C); air humidity is about 60 %, and annual precipitation is 200–

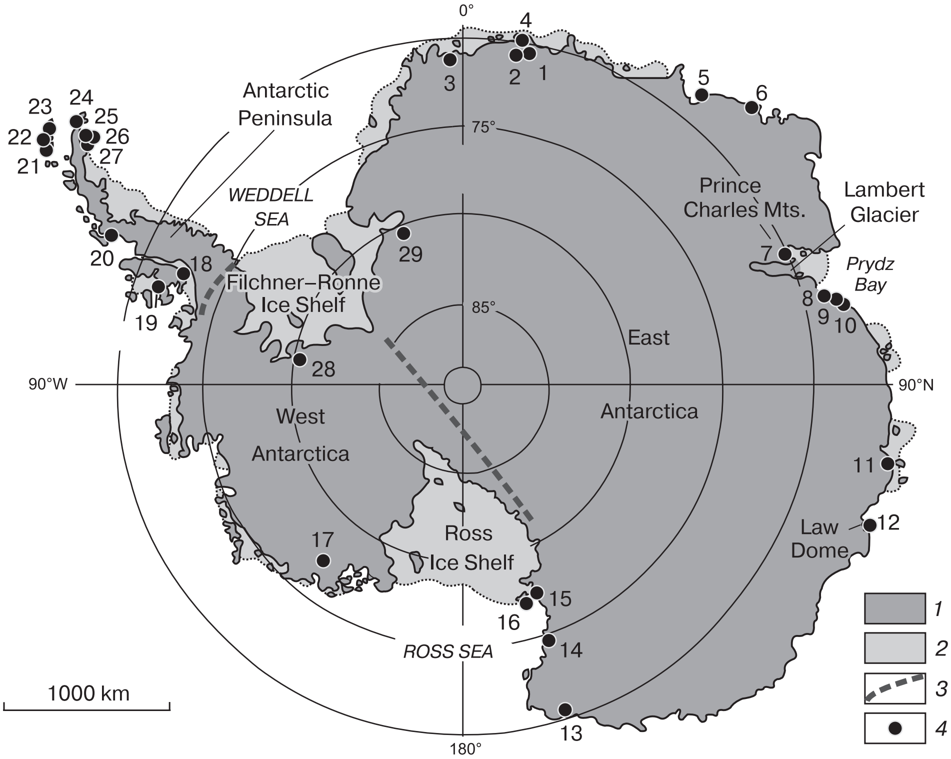


Fig. 1. Location of considered Antarctica objects.

(*1*) Ice sheets and domes, (*2*) ice shelves, (*3*) conventional boundaries between the main parts of glaciation, and (*4*) areas of the study of relief and sediments (1 – Untersee Oasis, 2 – Inzel Mountains, 3 – Robertskollen nunataks, 4 – Schirmacher Oasis, 5 – Soya Coast, 6 – Thala Hills, 7 – Amery Oasis, 8 – Larsemann Hills, 9 – Reuer Islands, 10 – Vestfold Oasis, 11 – Bunger Oasis, 12 – Windmill Islands, 13 – Little Rocks Area, 14 – Coast of Terra Nova Bay, 15 – Dry Valleys of Victoria Land, 16 – McMurdo Bay, 17 – Ridge Ford, 18 – George IV Ice Shelf, 19 – Alexander Island, 20 – Muller Ice Shelf, 21 – Livingston Island, 22 – Greenwich Island, 23 – King George Island, 24 – North of the Antarctic Peninsula, 25 – Beak Island, 26 – Vega Island, 27 – James Ross Island, 28 – Ellsworth Mountains; and 29 – Shackleton Ridge.

250 mm. In the Antarctic Peninsula, there are territories with maritime climate. In the west of the peninsula and nearby islands, the MAAT is about –3 °C; mean summer air temperatures are about 2 °C; air humidity is above 80 %, and annual precipitation reaches 600 mm on ice-free areas and is up to 1500 mm on glaciers. To the east of the ridges of the peninsula, which prevent western atmospheric transfer, the MAAT is –9 °C; summer temperatures are hardly above 0 °C, and annual precipitation is only about 150 mm.

Antarctic oases have been studied by paleogeographers for more than half a century. The accumulated data have been generalized for some parts or for the marginal zone as a whole. The issues of the Late Pleistocene–Holocene development of glaciation [*Stuiver et al., 1981; Ingólfsson et al., 1998; Anderson et al., 2002; Verkulich, 2009, 2010; Hodgson et al., 2014*], the relationships between climate changes and glaciers [*Hjort et al., 2003; Verleyen et al., 2011*], changes in sea level, and the state of the offshore zone and glaciers [*Berkman et al., 1998*] have been considered. As a rule, one or two of the main components of changes (climate, sea, glaciation) have been analyzed in these works for the period limited to the last glacial maximum (LGM) or Holocene. Russian research data are often missing in foreign publications.

The presented review includes the results of foreign and Russian paleogeographic studies in most of the ice-free Antarctic oases. In addition to the most complete presentation of data, refinement and updating of local reconstructions, this review aims to identify the course of changes in the natural environment of the marginal zone over the last 50 000 years as a whole and to determine the mechanisms and parameters of the interaction of climate, sea level, and glaciation under the influence of global, regional, and local factors.

DATA AND THEIR PALEOGEOGRAPHIC SIGNIFICANCE

Paleogeographic information on the marginal land area of Antarctica is mainly derived from the study of the relief and sediments of glacial, water-glacial, marine, lake, and ornithogenic geneses.

Evidences of glacial activity (U-shaped valleys, glacial hatching, moraines, etc.) can be found everywhere along the Antarctic periphery. Their location, morphology, and composition provide information about the extent, parameters, and stages of evolution of the glaciers. The correctness of data interpretation depends on the correctness of temporal estimates of glacial formations. Estimates of the age of moraines based on their morphology, the presence of an ice core, weathering stage, and vegetation characteristics are relative [*Adamson and Pickard, 1986; Borman and Fritzsche, 1995*]. In recent years, estimates based on measuring ^{3}He, ^{10}Be, ^{21}Ne, and ^{26}Al in stone surfaces have appeared [*Stone et al., 2003; Nicholls et al., 2019*]. Sometimes the moraines contain organic matter, the age of which indicates the time of the expansion of glaciers [*Adamson and Colhoun, 1992; Hjort et al., 2001*].

Water-glacial landforms and sediments (glaciofluvial terraces, deltas, standing water levels, and others) usually characterize the degradation of glaciers. However, in the Dry Valleys of Victoria Land, the levels of glacial lakes indicates the extent of glaciation during the LGM [*Clayton-Green et al., 1988; Hall et al., 2001*]. Organic remains in sediments allow us to date the course of deglaciation. Radiation dosimetry methods are also used for this purpose [*Gore et al., 2001; Mahesh et al., 2017*].

The occurrence, morphometry, and structure of marine formations; species composition and age of the flora and fauna in marine sediments correlate with fluctuations in relative sea level (RSL), climatic and ice conditions of sedimentation, and with changes in the boundaries of glaciers. Determination of the chronology of marine events is complicated by the need to correct radiocarbon data on marine Antarctic organics [*Gordon and Harkness, 1992*]. In addition, the wide range of depths and the diversity of habitats of marine organisms reduce the accuracy of the estimation of sea level at the time of their burial in sediments [*Ahn, 1994*].

Organic deposits accumulated in birds' nests for thousands of years are a specific data source in Antarctica [*Verkulich, 2008*]. Correlating the height and location of the nests of snow petrels (*Pagodroma nivea*) with the time of their settlement, spatial and temporal marks of the past glacial surface and the course of deglaciation processes can be reconstructed [*Ryan et al., 1992; Verkulich and Hiller, 1994*]. Adelie penguins (*Pygoscelis adeliae*) build nests above the storm-affected zone and near ice-exposed water areas, so the position and age of the nests and the intensity of nesting are indicative of sea level changes, periods of coast deglaciation, and climatic fluctuations [*Baroni and Orombelli, 1994a,b; Emslie and Woehler, 2005*]. However, paleogeographic reconstructions based on ornithogenic material are not quite reliable, because time intervals between the appearance of nesting conditions and the real nesting of the birds are unknown, and correction of radiocarbon dating is also required.

Bottom sediments of water basins represent an archive of the most detailed information about the development of the natural environment in the marginal zone of Antarctica. Lithology, granulometry, mineralogy, geochemistry of sediments inform about the regime and composition of waters, the flow of material from the catchment area, i.e., about the conditions and the course of deglaciation. The results of biogeochemical and isotopic analyses of organic matter reflect changes in the bioproductivity and regime of the reservoir, vegetation, and bird populations in the catchment area [*Hodgson et al., 2004; Wasiłowska et al., 2017*]. Diatom analysis informs about the past water exchange, chemistry, temperature, depth and ice regime of the reservoir, about the climatic conditions of sedimentation, about changes in RSL [*Whitehead and McMinn, 1997; McMinn, 2000; Verleyen et al., 2003; Roberts et al., 2004*]. Studies of mosses, shrimps, and rotifers in the sediments and pigment analysis provide data on the past dynamics and depth of the reservoir and on phototrophic communities [*Björck et al., 1991; Swadling et al., 2001; Verleyen et al., 2004*]. Dating of organic matter in sediments often provides a good chronology of paleoevents. However, paleolimnologists often meet difficulties in Antarctica because of poor knowledge of aquatic ecosystems, rarity of representative reservoirs, and the need to introduce various corrections to the results of dating.

The chronology of changes in the natural environment in the marginal zone of Antarctica is most often established using the radiocarbon (^{14}C) dating method. The correctness of dating depends on the use of various corrections of up to thousands of years [*Gordon and Harkness, 1992; Berkman and Forman, 1996*]. Most of the reconstructions considered by the author were carried out using such corrected dates. If uncorrected radiocarbon definitions were presented in original publications, the author introduced the necessary corrections in accordance with information about regional errors of radiocarbon dating and the conditions of formation and nature of the dated material.

RESULTS

The results obtained by the author and other researchers are analyzed with the aim to clarify and refine ideas about the conditions and the course of development of certain areas in the marginal zone of Antarctica.

In the area of **the Untersee oasis** (Fig. 1), the conditions of the interstadial (MIS 3) and LGM were revealed when analyzing the nesting patterns of snow petrels. In nests at 900–1300 m asl, the age of organic matter was 32 600–26 800 BP and about 17 000 BP;

in three nests in **the Inzel Mountains**, the dates of 27 100–23 000, 31 180, and 36 200 BP were determined. This suggests that birds lived on the rocks before and during the LGM [*Hiller et al., 1995*]. A comparison of the time of settlement and the height of the studied nests shows that during the LGM, the ice surface in this area rose by 250–300 m. Deglaciation began around 17 000 years ago and accelerated in 13 000–9000 BP. By about 8000 BP, birds mastered heights of 700–1150 m asl, close to the modern limits of nesting. The retreat of the glacier in the Untersee oasis 12 000–9000 BP led to the appearance of Lake Untersee [*Schwab, 1998*]. The maximum number of nests in the Untersee oasis on the slopes of **the Robertskollen** and Inzel mountains was used by birds 4000–3000 BP, probably during the warming period [*Verkulich, 2008*]. Around 2000 BP, many nests were abandoned (cooling); during the last 1500 years, birds have been actively occupying new nests (warming?). A study of bottom sediments from Lake Untersee indicates that 9000–7000 years deglaciation of this area slowed down because of climate cooling. Later, lateral moraines were deposited on slopes near the lake at the heights below 700 m asl [*Bormann and Fritzsche, 1995; Schwab, 1998*]. Their settlement by birds began around 3200 BP, which means that the glaciers grew between 7000 and 4000 BP The second group of moraine deposits near the lake reflects the fluctuations of local glaciers by tens of meters in the last hundreds of years.

In **the Schirmacher oasis** ^{14}C dating and diatom analysis of cores of frozen rocks and bottom sediments of lakes showed the presence of shallow freshwater reservoirs during MIS 3 under conditions resembling modern ones [*Verkulich et al., 2012a*]. During the LGM period, the oasis was covered by a glacier of 100–150 m in thickness [*Bormann and Fritzsche, 1995*], and even thinner over lake basins [*Mahesh et al., 2017*]. Glacial masses of the oasis on the edge of its northern steep slope met with glaciation of the shelf with a thickness of about 600 m, i.e., glacial bodies of different thicknesses developed in the oasis and on the shelf [*Verkulich et al., 2011*]. Deglaciation of the oasis began around 9000 BP [*Mahesh et al., 2017*]. Holocene conditions are reconstructed from data on the relief and lake sediments [*Bormann and Fritzsche, 1995; Schwab, 1998; Verkulich et al., 2011, 2012a*]. In addition to glacial forms, the levels of standing water–markers of the stages of development of the local lake system–can be traced in the relief. Since the beginning of the Holocene, ice melting has led to the formation of huge glacial lakes. The rapid decline of the glacial surface on the shelf led to a rapid runoff of waters from the oasis about 7000 BP, after which disjointed glaciers and large lakes remained on a larger part of the oasis. Between 7000 and 4000 BP, there was a slow melting of residual glaciers against the background of relative cooling. The warming of 4000–2000 BP led to a reduction in the area of local glaciers, the volume of lakes, and general drying of the territory. The interruption of this trend is indicated by the ridges of moraines on the edge of the glacier sheet. Their formation probably coincides with a decrease in the content of diatoms in lakes 2000–1000 and 400–150 BP.

The Soya Coast area (Lützow-Holm Bay) includes Ongul Island and peninsulas bordering a glacial slope or outlet glaciers lying in valleys extending into the bay and having a depth of about 500 m. On the Ongul Islands and the north of the Langhovde Peninsula, traces of exaration are rare; rocks are strongly weathered. There are marine deposits aged 46 000–23 000 BP. To the south, Glacial landforms are numerous, and rock weathering is weaker. Such differences may be indicative of the absence of LGM glaciation in the north of the area [*Hirakawa and Sawagaki, 1998; Miura et al., 1998a,b*], but this conclusion casts doubt. The top of ancient marine sediments is crumpled and transformed by meltwater – traces of glacier degradation. It there were no glaciation, how should we explain the compensatory elevation of the land by 20–25 m in the Holocene? It is more logical to assume that during the LGM there were both low-thick (100–200 m) glacial domes on the northern parts of the land with preserved ancient sediments under them and active glaciers of 400–500 m in thickness on the shelf. Glaciomarine sediments on the shelf have a thickness of up to 1.5 m, and the age of the oldest organic matter in them is about 14 000 BP [*Moriwaki and Yoshida, 1983*]. This confirms the erosion of the shelf by the glaciers of the LGM and indicates the beginning of deglaciation of the area around 14 000 BP. Hundreds of age determinations of fossil organic matter in sediments on the shores [*Hayashi and Yoshida, 1994; Hirakawa and Sawagaki, 1998*] allow us to construct an RSL curve: rise up to 6000–4500 BP and subsequent decline (fast 4500–3500, 2500–1500 BP, and slow 3500–2500 and about 1500 BP). A peak in dating of shells at 4000–3000 BP may mean an improvement in ice conditions, i.e., warming [*Verkulich et al., 2007*]. The warming around 4500–3500 BP is also indicated by the appearance of deltaic sediments, a signal of increased meltwater runoff [*Miura et al., 1998b*]. There are limited data on deglaciation of this area. Probably, the glaciers melted faster during the warming phase 4000–3000 BP. On the shore of the Rocky Peninsula, a push moraine of up to 10 m in height records the advance of the outlet glacier by about 200–300 m after 2000 BP with its following retreat [*Hayashi and Yoshida, 1994*].

In **the Tala Hills oasis (Enderby Land)**, studies of lake sediments [*Dolgikh et al., 2017*] indicated deglaciation no later than 11 000 BP (moraine at the base of sediments). The middle and upper layers of sediments have accumulated over the last 6500 years under harsh conditions for the development of diatom flora. Favorable (warmer?) conditions for the mass development of several types of diatoms existed in the reservoir about 4000–2000 BP, when the rate of sedimentation increased.

In the region of the **Prince Charles Mountains–Lambert Glacier–Prudes Bay**, the overlap of the edges of the LGM glaciers on the bottom of Prudes Bay is recorded by linear ridges [*Leitchenkov et al., 1994*]. Massive glacial deposits were found between the ridges and the shore; the continental ice did not move into the deep part of the shelf, and sedimentation took place under the glacier shelf. Judging by the age of the sediments, the closing of part of the bay by glaciers occurred after 20 000 BP, and deglaciation began earlier than 13 800 BP [*Domack et al., 1998*]. On the slopes of the Prince Charles Mountains in the middle reaches of the Lambert outlet Glacier, moraines reflect the rise of the glacier surface by 100–120 m [*Mabin, 1991*]. In the interior of the continent, LGM moraines are developed on the slopes of nunataks up to a height of 20 m above the ice surface. The maximum increase in ice on the Lambert outlet glacier is associated with a shift to the north of the line of its overlap with the bottom of the enclosing valley [*Adamson et al., 1997*].

The Amery Oasis lies at heights of about 200 m asl at the contact of the Lambert Glacier and the shelf glacier. The absence of glacial exaration features attests to the absence of the LGM glaciation in the oasis [*Adamson et al., 1997*]. However, moraine lies at the base of Terrasovoe Lake, and the height of the moraine ridge near the lake indicates an increase in the thickness of the nearby glacier by 150 m during the LGM. The rest of the territory probably had a snow-ice cover [*Wagner et al., 2004*]. Based on the results of studying lake sediments, glaciers had already left the lake's catchment by 12 400 BP, i.e., deglaciation started even earlier. The period of 12 400–10 200 BP was marked by the cold climate, harsh ice regime of the lake, and slow deglaciation. An increase in the content of biogenic material and diatoms in sediments aged 10 200–8600 BP indicates the beginning of warming. Further, up to 6700 BP, a diatom complex developed under warm conditions, and local glaciers decreased. Cooling and, probably, the expansion of glaciers are recorded in sediments aged 6700–3600 BP. During the last 3600 years, relatively warm (3200–2400 and 1600–900 BP) and cold (2000 and 600 BP) periods are distinguished [*Wagner et al., 2004*].

The Larsemann Hills Oasis is a group of low islands and peninsulas bordered by the glacial sheet slope from the south and by an outlet glacier from the southeast. The age of organic matter at the base of bottom sediments sampled in several lakes of the oasis corresponds to MIS 3 and LGM stages. Marine sediments of MIS 3 accumulated in Lake Kirisjes when sea level rose to 8 m asl (modern topography) [*Hodgson et al., 2001, 2009*] and in Mochou Lake, up to 10 m [*Gao et al., 2020*]. The study of frozen rock cores to a depth of 11 m showed that sea water during MIS 3 reached even modern levels of about 30 m asl [*Demidov et al., 2013*]. Studies of lake sediments and glacial hatching indicated that the LGM glaciation began later than 22 000 BP and covered a larger part of the oasis; the thickness of the glaciers reached the first hundreds of meters, and their extension to the shelf was small. On the Broknes Peninsula, glaciers moved through valleys; outlet glacier affected its eastern coast. Two lake depressions in the center were covered with snow and ice fields. The variety of glacier types was determined by the local relief conditions and the diverting role of the Dålk outlet glacier [*Hodgson et al., 2001, 2005; Verleyen et al., 2005*]. The beginning of deglaciation of the oasis marks the age of lake sediments in Kirisjes Pond – 13 600 BP [*Verleyen et al., 2005*]. Changes in the accumulation of sediments under marine, freshwater, and transitional conditions allowed us to recreate the curve of RSL changes: a rise in the early Holocene until about 7000 BP; then, a drop interrupted by a small rise in 2000–1500 BP [*Verleyen et al., 2004, 2005*]. Lake sediments also provided paleoclimatic information [*Verleyen et al., 2004*]. About 13 500–11 500 BP, Kirisjes Pond was covered with ice under cold conditions. In 11 500–9500 BP, due to warming, the flow of meltwater increased, and Lake Stepped appeared. The accumulation of sediments in this lake took place under relatively cold conditions in 9500–7500 BP and under relatively warm conditions in 7500–5230 BP. In 5230–3000 BP, the climate was similar to the modern climate, except for some warming around 3800 BP. The warm conditions of 3000–2000 BP are indicated by sediments in Kirisjes Pond Pap Lagoons. Around 2000 BP, the warming was replaced by a short cooling. Diatoms in the sediments of Kirisjes Pond reflect the cooling of 760–690 and 280–140 BP. The rapid deglaciation of the oasis in the early Holocene was caused by climate warming and sea level rise [*Hodgson et al., 2005*]. Intensified ice melting under the impact of warming could take place around 6000, 4000, and 1500 BP There are no traces of glacial growth in the oasis in the Holocene, but a slowdown in deglaciation around 7500 and 3000 BC is not excluded [*Verleyen et al., 2005*].

In the area of **the Reuer Islands**, reconstructions are based on studies of bottom sediment cores from local bays [*Berg et al., 2010*]. The accumulation of material with organic residues took place about 45 000 BP at one point of core sampling and 10 000 BP at another point. The characteristics of the sediments attest to the marine non-freezing conditions in the area in MIS 3 (MIS 5?) confirming that the ice sheet during the LGM did not have a continuous distribution and depended on the topography of the shelf and land. Deglaciation of the area began about 11 200 BP and proceeded under conditions of

Early Holocene optimum until 8200 BP, when a colder stage began. Warm conditions were established in 5700–3500 BP. Then, sedimentation proceeded under colder conditions again [*Berg et al., 2010*].

The Vestfold Hills Oasis is bounded by the slope of the ice sheet on the east and by the outlet Sørsdal Glacier. Glacial hatching in two directions, the thick moraine cover in the center and west of the oasis, and moraine ridges in the center, west, and near the Sørsdal Glacier are indicative to the LGM glaciation, but the mechanism of its development is open to argument. Most of the hatches are in the east–west direction characterizing the past expansion of the glacial sheet, and the rest are associated with the Holocene onshore advance of the Sørsdal Glacier [*Adamson and Pickard, 1986*]. Other researchers considered the coincidence of the orientation of moraine ridges in the west and center of the oasis with the ridges near the Sørsdal Glacier as evidence of the extension of this glacier to the north and northwest of the oasis during the LGM [*Hirvas et al., 1993*]. The analysis of rock fragments and weathering features and data on fossil fauna in the moraines gave a new reconstruction [*Gore, 1997*]: (1) before the LGM, land areas existed in the center and in the west of the oasis, and sea bays occupied local depressions; (2) glacial sheet expanded during the LGM; (3) moraine ridges in the center of the oasis were created upon stops in the retreat of the glacial sheet; (4) rock hatching and moraine ridges near the Sørsdal Glacier mark its advance in the Late Holocene. The scenario suggested by the author of this article assumes the onshore expansion of the outlet Sørsdal Glacier that carried material from the bottom of the sea and from the weathered land surface at the beginning of the LGM. The glacial sheet carrying relatively little material and slightly eroding the surface in the eastern part of the oasis also expanded. These glaciers met in the center of the oasis and the material of future moraine ridges was concentrated at the meeting point. Then, the two glacial bodies merged: then the hatching of the western direction appeared at high levels of the oasis. The thickness of the ice could reach only 300 m [*Gore, 1997*].

The deglaciation of the oasis began 12 000–13 000 BP [*Fabel et al., 1997*], and its course was determined by changes in RSL and climate. About 7500–6500 BP, RSL reached a maximum (10–13 m asl) and then began to fall (with a slowdown in speed and even some rise in 3000–1500 BP) [*Zwartz et al., 1998*]. The oasis had relatively warm conditions in the Early Holocene, warming and humidity growth in 3500–2500 BP, cold and dry conditions in 2000–1800 BP, and then relatively cool conditions [*Fulford-Smith and Sikes, 1996; McMinn, 2000*]. By about 8000 BP, deglaciation occurred over 50 % of the oasis; in 8000–5000 BP, the retreat of the glaciers slowed down. It intensified again in the second half of the Holocene with a decrease in RSL and under relatively warm conditions. Moraines near the Sørsdal Glacier record its progress due to changes in sea level 3000–1500 BP and cooling around 2000 BP. Along the border of the glacier, ridges formed after 700 BP [*Adamson and Pickard, 1986*].

The Bunger Hills Oasis carries traces of glacial activity, but the activity of glaciers in the LGM is a matter of discussion. One of the opinions is that the oasis was covered by the sheet ice moving to the west and northwest and having a thickness of more than 500 m [*Adamson and Colhoun, 1992*]. The study of rock weathering and OSL dating of the sediments served as the basis for the conclusion about the presence of ice-free areas during the LGM period [*Gore et al., 2001*]. The dates, together with the errors of the method, fall into time intervals of about 40 000–19 000 BP and near the beginning of the Holocene. The values of ^{14}C dating of organic matter in the moraine base of bottom sediments of local lakes fall into the first interval [*Melles et al., 1994, 1997*]. There are no dates for the period 19 000–14 000 BP. Such facts indicate the presence of glacier-free territories in the area from MIS 3 to 19 000 BP and the short-term glaciation of the LGM. The variety of directions of glacial hatching contradicts the idea of the development of glaciation of the LGM only by the expansion of the glacier sheet, and the preservation of hatches is difficult to explain against the background of long-term weathering.

The thickest moraines in the west and northwest of the oasis, in the author's opinion, were formed when outlet glaciers moved to land and merged there with the glacial sheet with mass accumulation of moraine sediments in this area. Further, the movement of the ice cover depended on changes in the RSL and the subglacial relief, which created hatching in different directions. Glaciers on land were no thicker than 100 m in high areas and 300 m in depressions [*Verkulich, 2010*]; glaciers in the sea basins of the oasis had a thickness of more than 500 m [*Melles et al., 1997*]. Dating of organic sediments indicates the development biota in the reservoirs and the settlement of birds 13 550–9470 BP, the minimum time of the beginning of deglaciation [*Melles et al., 1994; Verkulich and Hiller, 1994*]. The climatic conditions of deglaciation were clarified by studying the bottom sediments in local reservoirs [*Kulbe et al., 2001; Verkulich, 2007; Berg at al., 2020*]. The RSL changes left traces on the coast up to the height of 10–11 m, as well as in the sediments of reservoirs [*Adamson and Colhoun, 1992; Verkulich et al., 2002*]; on this basis, the curve of RSL changes was constructed [*Poleshchuk and Verkulich, 2014*]. Already by 10 000–8000 BP, large areas of land and many lakes of the oasis were freed from glaciation, and oceanic waters penetrated into local bays [*Melles et al., 1997*]. With the cooling of 8000–5000 BP, the melting of glaciers on land slowed down, but due to the rise of sea level, seawater penetrated even to the southeastern glacial margins of the oasis. From about 6000 BP, the RSL began to fall, which, against the background of cooling, led to the activation of outlet glaciers on the shores of the oasis and moraine accumulation along its western edge (6000–5000 BP). Warm conditions of 4000–2000 BP contributed to deglaciation, and by 2000 BP the boundaries of the oasis began to resemble modern ones. Then, against the background of deglaciation, the extension of the outlet Edisto glacier and the appearance of push moraine took place [*Adamson and Colhoun, 1992*] due to cooling around 2000 BP and fluctuations of the RSL [*Poleshchuk and Verkulich, 2014*].

The Windmill Islands area is bounded by the Law Dome from the north and east, and by outlet glaciers from the south. During the LGM, the northern islands were under the glacial dome, and the southern ones were eroded by outlet glaciers; the edges of the glaciers could extend into the sea for 7–15 km, the estimated ice thickness on the islands did not exceed 200 m; on the shelf, 300–400 m [*Goodwin, 1993*]. Information about the development of the area is contained in the sediments of local bays: at their base, there are organic-mineral sediments aged 46 000–26 000 BP, overlain by a moraine; above, there are glacial-marine sediments and Holocene mainly organic sediments [*Kirkup et al., 2002; Cremer et al., 2003*]. The accumulation of marine sediments in 10 500–4000 BP proceeded under relatively cold conditions; in 4000–1000 BP, conditions were generally warmer, although 2000–1700 BP the temperature began to decrease; conditions remained cold in the recent millennium [*Cremer et al., 2003*]. The RSL curve was reconstructed during the study of beaches and lake sediments: the rise in sea level up to 32 m by 6000 BP, then the decrease in sea level with a slowdown in 2500–1800 BP (in 1900–1800 BP, RSL could even rise a little) [*Goodwin, 1993; Roberts et al., 2004*]. The beginning of deglaciation in the area dates back to about 12 00 BP according to dating of organic matter in lakes and penguin rookeries [*Goodwin and Zwek, 2000*]. In the Early Holocene, the destruction of ice was facilitated by the rise of RSL, and relatively cold conditions did not prevent the development of lakes and bird nesting [*Emslie and Woehler, 2005*] until 7000 BP. With a maximum rise of RSL and cold climate of 7000–6000 BP, deglaciation slowed down. In 6000–4000 BP, the number of rookeries increased due to the beginning of the fall of the RSL, that is, the expansion of the land. The continuation of the fall of the RSL and the warming of the climate 4000–2000 BP led to a reduction in glaciation. However, at the same time, the Law Dome created moraines, because of the increased precipitation [*Goodwin, 1993*]. It was relatively cold in the area during the last 2000 years, but penguins have been actively inhabiting the islands [*Emslie and Woehler, 2005*].

In the Ross Sea–McMurdo Bay–Dry Valleys region of Victoria Land, data on LGM conditions are collected in marine basins, on the coast, in valleys, and in mountainous areas. In the Ross Sea, most of the shelf was covered by the ice sheet [*Karl, 1989; Anderson et al., 2014*]. About 20 000 BP, its boundary was found 100 km south of the edge of the continental shelf at depths of 300–500 m [*Licht et al., 1996*]. The ice sheet also closed the McMurdo Bay and entered the lower parts of the Dry Valleys in Victoria Land blocking them with ice of 200–300 m in height. From 26 000–23 000 BP, most of the valleys were free from cover glaciation; dammed lakes developed there [*Stuiver et al., 1981; Clayton-Green et al., 1988; Hall et al., 2001*]. In **the area of Terra Nova Bay**, the surface of glaciers in the mountains rose by tens of meters; in the middle parts of the valleys, by the first hundreds of meters; on the coast, by about 400 m (coastal ice strata were part of the Ross Ice Shelf). According to the dating of the shells in the moraines (37 500–25 300 BP), the glaciers that captured and moved them developed later than 25 000 BP [*Orombelli et al., 1990*]. On Ross Island near the eastern shores of the Ross Sea, the height of the ice surface reached 720 m, on the western continental margins of the Ross Sea, it was 800 to 950 m asl [*Anderson et al., 2014*]. Glaciers were somewhat thicker on **the Ford Ranges** at heights up to 950 m asl [*Stone et al., 2003*]. Data from the region confirm the presence of ice sheet on the shelf, coast, and in the mouths of mountain valleys from 25 000 BP with its maximum expansion 21 000–18 000 BP [*Stuiver et al., 1981; Clayton-Green et al., 1988; Hall et al., 2001; Anderson et al., 2002; Oberholzer et al., 2003; Stone et al., 2003*]. At the same time, the thickness of the ice sheet sharply decreased in the direction from the shelf to the continent, probably due to the interception of a significant part of the moisture of atmospheric masses by the ice shelf [*Orombelli et al., 1990*].

Deglaciation of the region began in 17 000–10 000 BP: in the Ross Sea, the retreat of the northern edge of the ice sheet began about 17 000 BP; degradation of glaciers blocking Dry Valleys dates back to 16 000–13 000 BP; penguins appeared on the shores 13 000–11 000 BP. Deglaciation in **the Little Rocks** Ford ranges started 11 000–10 000 BP [*Jordan and Wateren, 1993; Baroni and Orombelli, 1994a; Doran et al., 1994; Stone et al., 2003*]. The curve of Holocene changes in the RSL in the region is based on the study of coastal forms and penguin rookeries: sea water level reached 30–40 m around 7500 BP; then, the RSL rapidly decreased; in the last 3000 years. this process has slowed down [*Baroni and Orombelli, 1991, 1994a*]. The paleoclimate on the coast, as evidenced by the dynamics of penguin rookeries, had an optimum 4300–2900 BP; after 2900 BP there was an increase in the severity of ice conditions (cooling); in the recent millennium, the number of has increased

[*Baroni and Orombelli, 1994b*]. The study of lakes in Dry Valleys revealed the presence of warm conditions 3000–2000 BP, cold and dry conditions in the period from 2000 to 1200–1000 BP, and relatively warm and wet conditions in the recent millennium [*Smith and Friedman, 1993; Lyons et al., 1998*]. Dating of organic matter in sediments indicates rapid deglaciation of coastal areas up to 8000–7000 BP. By 7500 BP, outlet glaciers on the shores of the Terra Nova Bay retreated beyond their modern borders [*Baroni and Orombelli, 1994c; Hall et al., 2001*]. Between 5000 and 1500 BP, the edges of the glaciers moved back blocking part of the beaches; in 1200–500 BP, glaciers were smaller than modern ones, and then reached their current boundaries [*Baroni and Orombelli, 1994c*].

West Antarctica, including the **Weddell Sea, the Antarctic Peninsula,** and islands is the region with a lot of ice-free land areas. The natural conditions during MIS 3 were studied in detail on the Fildes Peninsula **of King George Island**, where marine sediments with shells, algae, bone remains, and diatoms date back to 50 000–19 000 BP. During MIS 3, the peninsula was under seawater to the level of 40 m asl and turned into an archipelago. Seawater was no colder than at present. The LGM glaciation was relatively thin, so that loose deposits of MIS 3 were preserved [*Verkulich et al., 2013, 2015*]. Shells of about 30 000 years old were found in the moraine on the coast of **Alexander Island.** They can be considered the remains of MIS 3 in the surroundings of the Antarctic Peninsula. Shells of about 34 000 years old were found in glacial deposits of **Vega and James Ross Islands** [*Clapperton and Sugden, 1982; Ingólfsson et al., 1992*].

The study of hatching and erratic boulders in the areas framing the **Filchner** and **Ronne** ice shelves indicated that the ice surface in coastal areas during the LGM was up to 400–650 m asl; on the slopes of the mountains facing the Weddell Sea, these features of glaciation can be found at heights of up 1000–1900 m asl; glaciers leaned on the bottom of the Weddell Sea at a distance of up to 800 km. The main growth of glacial masses was on the shelf [*Carrara, 1981; Waitt, 1983; Denton et al., 1992*]. Geological and geophysical studies have confirmed the development of a thick shelf ice on the western shelf of Weddell Sea, possibly, after 21 000 BP [*Elverhøi, 1981*]. The use of cosmogenic nuclides has clarified the growth of glaciers on the Weddell Sea coast; they reached the thickness of 310–650 m in the east (in the area of **Shackleton Ridge.** 800 m in the area of the **Ellsworth** mountains; at least 385 m on the western mountain margins of the Ronne Ice Shelf [*Nichols et al., 2019*]. During the LGM, glaciers of the Antarctic Peninsula and the adjacent shelf ice began to degrade approximately 18 000 BP. Deglaciation of the areas was asynchronous and depended on the subglacial topography. By the beginning of the Holocene the ice sheet approached the current configuration [*Cofaigh et al., 2014*].

The age of organic matter in sediments indicates the collapse of ice shelf in the west of the Weddell Sea before 11 270 BP [*Anderson et al., 2002*]. On the islands **of Vega, James Ross,** and **Beak**, deglaciation began 11 000–9500 BP [*Zale and Karlen, 1989; Ingólfsson et al., 1992; Sterken et al., 2012*]. On King George Island, it began at least 11 500 BP; by 9300 BP, glacier on the Fildes Peninsula had shrunk to its present size [*Verkulich et al., 2012b*]. The climate on the island was relatively warm from the beginning of the Holocene to 5300, 4000–2300 and 1400–600 BP, and cooling occurred 5300–4000, 2000–1400 BP and during the Little Ice Age (LIA) [*Verkulich et al., 2012a,b*]. The RSL was up to 19–20 m asl about 8000 BP, and then decreased (with a slowdown in 5000–4000 and 2500–1600 BP, and possibly with a slight rise 2000–1300 BP) [*Poleshchuk et al., 2016*]. On Beak Island , the study of lake sediments revealed relatively warm conditions from the Early Holocene to 6407 cal BP, then cooling to 3169 cal BP, a sharp warming of 3169–2120 cal BP, cooling 2120–543 cal BP, and then warming [*Sterken et al., 2012*]. The RSL on Beak Island after rise to a level of 15 m asl about 8000 cal BP began to decrease [*Roberts et al., 2011*].

On **Livingston Island** and the north of the Antarctic Peninsula, reconstructions of climate change and RSL are sketchy, so we will present them in the context of regional deglaciation. Until about 7000–6000 BP, the destruction of glaciers proceeded rapidly due to the rise of RSL and warming [*Zale and Karlen, 1989; Ingólfsson et al., 1992*]. The subsequent fall of the RSL and the beginning of the cold phase slowed down deglaciation, and even caused the extension of outlet and shelf glaciers that left traces on the shores of Alexander and James Ross Islands [*Ingólfsson et al., 1992; Hjort et al., 2001*]. Warming from about 4000 to 2500 BP against the background of a decrease in RSL and an increase in the area of the islands contributed to the deglaciation. However, on the James Ross Islands, the rise in temperature and humidity led to the growth of glaciers about 3000 BP [*Bjorck et al., 1996*]. Between 2000 and 1000 BP the climate of the area was, in general, colder than the modern one; this contributed to the growth of the glacier on the Fildes Peninsula [*Verkulich et al., 2012a,b*]. In the last millennium, climatic fluctuations have been frequent. The most notable was the LIA event, which caused the growth of glaciers and the extension of the edges of ice shelf and outlet glaciers [*Zale and Karlen, 1989; Domack et al., 1995; Hall, 2007*].

DISCUSSION

A comparison of reconstructions of the recent history of glaciation and deglaciation in coastal areas of Antarctica attests to some common and individual features in the changes of climate, sea level. and glaciation–deglaciation and permits the identification of correlation and dependence of these changes on global and local factors.

Conditions of MIS 3 interstadial in the marginal zone of Antarctica

The natural conditions of MIS 3 interstadial have been reconstructed only at a few points, but their location suggests the existence of habitable land and sea spaces along the Antarctica periphery at that time. The material with the age of MIS 3 *in situ*, as well as included in the LGM sediments, confirms the presence of seawater in MIS 3 in the areas of the Soya Coast, Larsemann Hills, Rauer Islands, Bunger Oasis, Windmill Islands, Ross Sea, George IV ice shelf, Antarctic Peninsula, and its surroundings. The location of the findings indicates that the sea/glacier boundaries during MIS 3 were close to modern ones. This is confirmed by the nesting of snow petrels in the Untersee area and Inzel mountains. Feeding of birds could only be provided if the line of open ocean waters did not shift significantly to the north in comparison with the current position of the edge of the shelf ice. Studies of lake sediments, permafrost, and cosmogenic nuclides in the Schirmacher, oases, Larsemann Hills, and Bunger oases indicate the absence of a cover glaciation during MIS 3 and the presence of organic life in the lakes. For example, in the Schirmacher oasis, local conditions of the sediment accumulation in lake basins during MIS 3 were no colder than those at present [*Verkulich et al., 2012a*].

Detailed information was provided by studies of MIS 3 sediments s in situ. Paleolimnological data revealed the presence of freshwater reservoirs in the Larsemann Hills (in warm conditions) 53 000–47 000 BP, deterioration of biomass accumulation conditions by 38 000 BP, and ingression of seawater into lakes in 38 000–26 650 BP (shallow coastal zone conditions), when the RSL reached a height of about 10–12 m [*Hodgson et al., 2009; Gao et al., 2020*]. The study of cores of frozen rock from the Larsemann Hills showed that the rise of RSL in MIS 3 could reach 30 m [*Demidov et al., 2013*]. Approximately the same time interval of the MIS 3 transgression was established for sections of the Soya Coast with a rise of RSL up to 20 m [*Miura et al., 1998a*]. On King George Island, the rise of RSL in MIS 3 could reach 40 m, and most of the studied sections of marine sediments with rich flora and fauna were formed 30 000–20 000 BP [*Verkulich et al., 2013, 2015*]. The explanation for this height of the RSL at a time when the global eustatic level was about 60–70 m below modern sea level [*Lambeck et al., 2002*] is considered to be a significant increase in ice masses in the regions, which led to the deflection of the Earth's crust and, accordingly, the rise of the RSL [*Hodgson et al., 2009; Gao et al., 2020*]. However, it is quite difficult to imagine a picture of a simultaneous significant increase in glaciation of the margins of the continent and the presence of marine waters here in wider than modern borders. Such a situation, for example, is impossible for King George Island, which reduced its area and turned into an archipelago during MIS 3 [*Verkulich et al., 2013, 2015*]. The author believes that changes in the RSL depended to a large extent on regional tectonic processes that are not directly related to changes in glacial load. The considered regions are located in fault zones and other areas of tectonic activity and characterized by extreme fragmentation of the earth's surface, which indicates the probability of intense block tectonic movements.

The last glacial maximum

The period of LGM in the marginal zone of Antarctica can be estimated from minimum values of the most ancient dates of organic deposits (Fig. 2). They indicate that glaciers did not cover many mountain, coastal, island and sea areas 26 000–19 000 BP.

In low-lying areas of the coast and nearby islands, the glaciation of the LGM was relatively thin (ice thickness of 100–300 m); in the Schirmacher oasis, Larsemann Hills, Bunger Hills, and Dry Valleys of Victoria Land, there was no continuous glacial cover; part of the land in these oases was blocked only by snow and ice fields. An increase in the height of the ice cover in many areas of East Antarctica (the Untersee and Amery oases, slopes of the Prince Charles Mountains and the Terra-Nova Gulf) was less than 200–300 m. At the same time, the thickness of shelf ice during that period exceeded 500 m and even 1000 m in many areas. Deep depressions were occupied by outlet glaciers, often coming to land (the Vestfold Hills and Bunger Hills oases, Windmill Is-

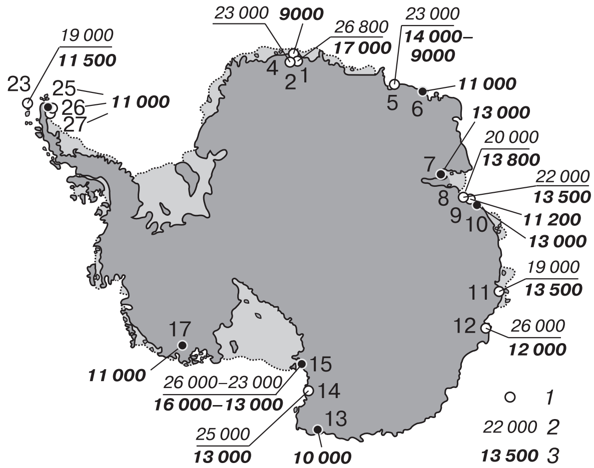


Fig. 2. Areas and time of recorded events of MIS 3 and LGM.

1 – Study areas of MIS 3 event, *2* – minimum dates of sediments before the LGM, years ago, *3* – maximum dates for postglacial events, years ago. Numbers of study sites are given in Fig. 1.

lands), or diverting continental masses of ice from land (Larsemann Hills, Ongul Island near the Soya Coast). Other areas of the shelf were blocked by thick glacial masses that made up the main part of the glaciation of the Ross and Weddell Seas, Prudes Bay, and other seas.

These features of glaciation during the LGM suggest a scenario for its development. The initiation of glaciation under cold conditions and sea level lowered by tens of meters [*Clark and Mix, 2002; Lambeck et al., 2002; Peltier, 2002*] created prerequisites for the rapid accumulation of ice masses on the drained shelf, where the line of glaciers' leaning on the bottom was shifted northward, and the water layer under the shelf glaciers was reduced. In these areas, climate cooling led to a rapid lowering of the snow line and an increase in snow accumulation in the presence of humidifying sea areas nearby. As a result, glaciers on islands and drained areas (glaciation cores) grew rapidly on the shelf, shelf glaciers thickened and began to lean on the drained bottom of shallow-water areas due to a decrease in ice melting at the bottom and an increase in accumulation on the surface. The edges of the ice sheet and outlet glaciers moved toward the shelf. Under such conditions, glaciers on the coast and in the interior of Antarctica could receive sufficient moisture supply only at the beginning of the LGM, when there were areas of open water (a source of moisture) on the shelf, and summer melting slowed down. At the same time, the drop in sea level led to an increase in ice discharge by outlet glaciers following the downward shifting of the line of glaciers to the bottom of the shelf. The further closure of the shelf by ice the spread of the shelf ice e cover to the north distanced the continental areas from the open ocean masses so strongly, that moisture shortage to feed the glaciers was established in most areas of the Antarctica. This scenario was most typical for East Antarctica, especially in the vast areas of the shallow shelf and areas of development of the system of outlet glaciers.

In West Antarctica, on the modern borders of the Filchner and Ronne ice shelves with the nearby slopes, on the western continental slopes on the border with the Ross Sea, the rise of the ice surface during the LGM was great. Assuming the above scenario of the development of the LGM, the active increase in the thickness of shelf ice and the low position of the ice sheet in these areas (at present, 1000–1500 m asl) created a situation, when shelf ice prevented ice flow from the continent, and accumulation was active at border of the shelf and continental glaciations.

Post-glacial development of the marginal zone

The beginning of deglaciation is marked by the oldest dates of the organic matter in postglacial sediments, the time of the release of rocky surfaces from under the ice and the formation of water-glacial sediments (Fig. 2). The first evidence is associated with the destruction of shelf ice in 17 000–14 000 BP, when marine sediments accumulated in bays near the Coast of Soya, in Prydz Bay, in gulfs of the Bunger Hills oasis, near the Windmill Islands, and in the McMurdo area. Snow petrels populated the Untersee oasis due to the shortening of the distance from the mountains to open ocean waters. Somewhat later (13 000–9000 BP), degradation of onshore glaciation in modern oases took place. The initial stage of deglaciation certainly depended on global changes in sea level and climate. By 14 000 BP, when the first sections of the shelf were freed from ice along Antarctica coast, sea level had already risen by 10–13 m (which is comparable with the contribution of the entire postglacial melting of Antarctic ice to sea level rise); in 12 000–10 000 BP, see level rose by 30 to 60 m [*Bentley, 1999*] causing widespread floating, disintegration, and rapid decline of glaciers. Evidences of deglaciation 16 000–14 000 BP are attributed to the period of air temperature rise recorded in ice cores. It was interrupted by a cooling of 14 000–12 500 BP (though temperatures remained higher than those during the LGM) followed by warming between 11 500 and 9000–8000 BP [*Masson et al., 2000; Jouzel et al., 2001*].

The curves ***of climate change*** in particular regions of Antarctica differ from one another in chronology and amplitude (Fig. 3), which is associated with problems of dating and time resolution of reconstructions. However, they reflect the dependence of these changes on global, regional, and local factors. The most reliable reconstructions indicate warming in the Early Holocene, up to about 8000 BP, which coincides with records in ice cores [*Masson et al., 2000*]. This is indicative of close relationship between local climate changes along Antarctic coasts and global processes: optimum of summer insolation in the Northern Hemisphere, warming of oceanic waters in the Southern Hemisphere, and changes in the atmospheric circulation [*Verleyen et al., 2003*].

The period of 8000–4000 BP was characterized by smoothing of climatic fluctuations and the proximity of climatic conditions to modern ones. According to data from ice cores, it has been warming since 6000 BP in the central regions of East Antarctica. However, on the margins of the continent, indications of a weak warming 7000–5000 BP were only found in the Larsemann Hills and Schirmacher oases. In many other areas, indications of cooling in that time were recorded. These differences seem to indicate the beginning of the regional and local conditions' influence on local climate.

For 4000–2000 BP, data on warming have been obtained almost everywhere. Ice cores produce warming peaks that are almost imperceptible in the center of East Antarctica and increase their amplitude to the edge of the ice sheet [*Masson et al., 2000*]. The increase in warming from the center to the margins of Antarctica and its discrepancy with the global climatic optimum of 6000–5000 BP [*Folland et al., 1990*] indicate its dependence on regional factors.

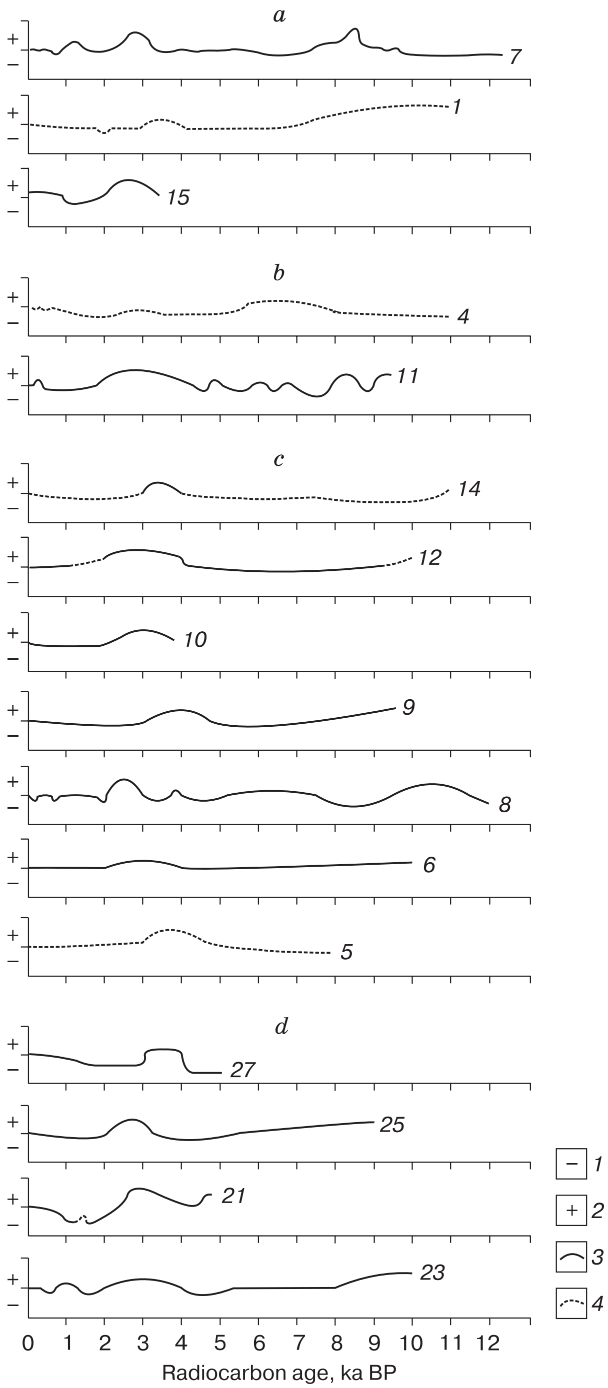


Fig. 3. Relative climate changes in the peripheral zone of Antarctica:

(*a*) mountains and mountain valleys, (*b*) onshore areas beyond ice shelf, (*c*) coastal areas, and (*d*) marine areas. *1* – Colder (drier?) climate, 2 – warmer (wetter?) climate, *3* – reliable reconstruction, and *4* – approximate reconstruction. Curve numbers denote key areas indicated in Fig. 1.

These could be changes in summer insolation in Antarctica [*Hjort et al., 1998*], oscillations of the high pressure region over the mainland [*Björck et al., 1996; Hodgson et al., 2004*], and destruction of shelf ice [*Masson et al., 2000; Verleyen et al., 2004*]. A slight warming for one of these reasons slightly changed the temperatures inside the mainland, and in coastal areas led to a reduction in sea ice cover and increased heat exchange and heat transfer, i.e., increased warming. This is confirmed by observations in the polar regions: in the depths of the continents, the amplitudes of warming are weaker than in the adjacent marine spaces [*Walsh, 2009*]. Local factors are latitude, distance from the sea, altitude and glaciated surroundings of the oases: warming is most pronounced in marine and coastal areas at relatively low latitudes.

Data on climate fluctuations over the past 2000 years allow us to identify only the traces of cooling common to most areas between about 2000 and 1500 BP. This cooling is also reflected in ice cores, i.e., it has a regional scale [*Masson et al., 2000*]. Other climatic events of the last millennium (including LIA) are difficult to find in the data from the marginal zone.

The curves of changes in the RSL (Fig. 4) demonstrate an increase in sea level from the early Holocene to maximum heights between 8000–6000 BP and the subsequent tendency for its fall. Many curves also indicate the presence of certain stages in the decrease in sea level: fast up to 3500–3000 BP, slowed

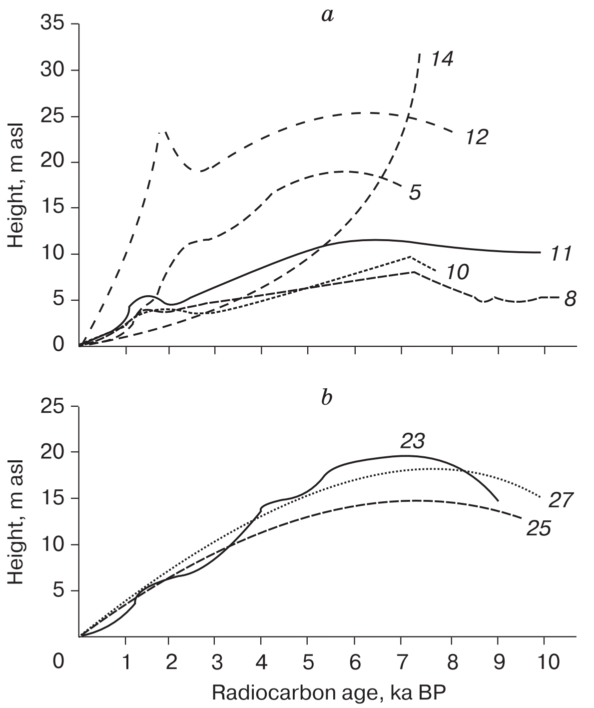
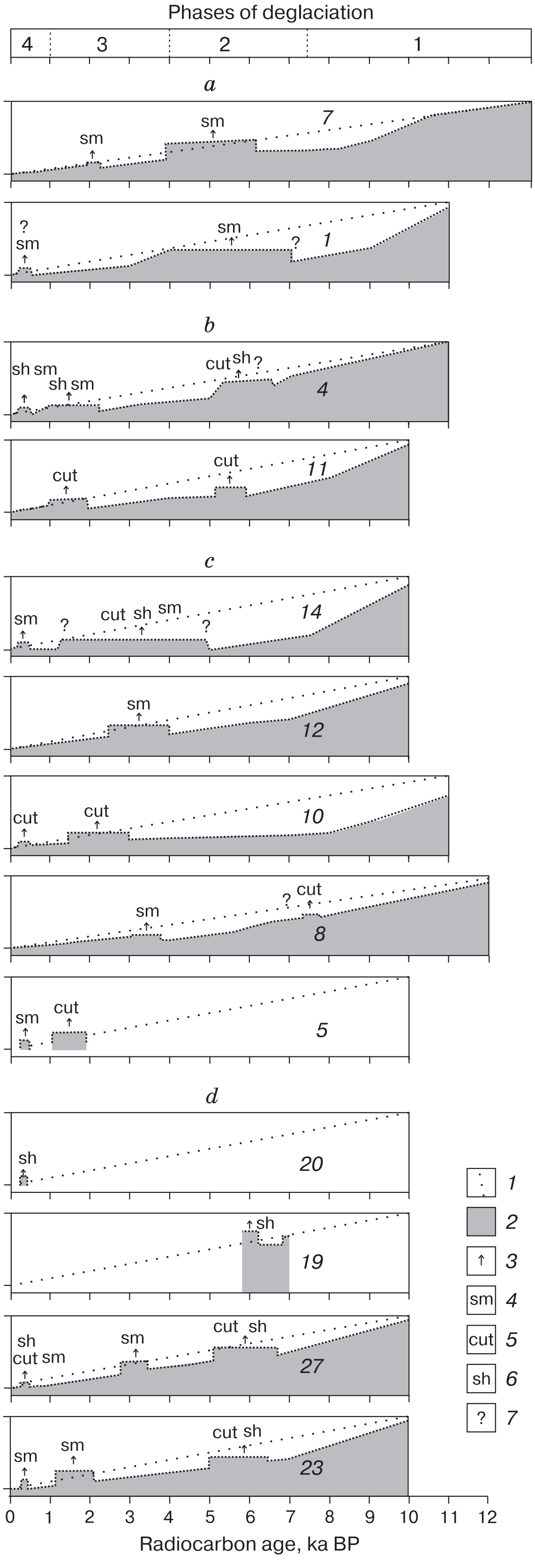


Fig. 4. Reconstructions of Holocene changes in the RSL (relative to modern absolute heights) in the peripheral zone of (*a*) East Antarctica and (*b*) West Antarctica.



down afterwards, stopped or even replaced by sea level rise in several areas in 2500–1300 BP, and then resumed again. The rise of the RSL 1500–1400 BP was confirmed by the study of coastal bars on **Greenwich Island** [*Santana and Dumont, 2003*]. A comparison of Antarctic sea-level curves with the ideas of eustatic oscillations and the contribution of glacier melting to them [*Bentley, 1999*] reveals similarities: a rapid rise in sea level by the middle of the Holocene (although in Antarctica, the peat of this rise was 1000–1500 years earlier). The main difference between the curves is in the heights of the maximum rise of the RSL in the areas of the marginal zone of Antarctica (from 7–8 to more than 30 m). Taking into account similarity of the estimates of the thickness of the LGM glaciation for these areas, such a difference is more likely due to regional tectonic features rather than to glacioisostatic compensatory uplift of the territory. Other differences are related to the correctness of paleoreconstruction; however, they also generally reflect the relationships between global eustatic oscillations and regional tectonic movements, including local glacioisostatic movement associated with the LGM and the course of deglaciation.

A comparison of reconstructions of ***deglaciation*** in different parts of Antarctic periphery (Fig. 5) with changes in climate and sea level outlines four main phases. In the Early Holocene (the first phase), deglaciation was rapid and was related to global factors: the rise of RSL and climate warming. The phase of rapid deglaciation ended earlier (9000–8500 BP) in the mountainous oases, where it depended only on the climate. On the coast, it lasted for another 500–1500 years. Here, the rise of the RSL continued due to the predominance of the eustatic component over the rate of uplift of territories freed from glaciers: land areas were submerged, the warming effect of expanding sea areas increased, as well as ice discharge from the continent through floating outlet glaciers. On the coast, despite the relatively cold conditions since 8000 BP, relatively thin glaciers and snow–ice fields have been degrading. Ice melting in the oases was facilitated by the warming effect of growing area of glacier-free surfaces. This means that already in the first phase, the dependence of deglaciation on regional and local factors (location, glacial surroundings, local climate) was clearly manifested. By 8000–7000 BP, from 50 to 70–80 % of the area of coastal oases was freed from glaciers.

Fig. 5. Relative assessment of the course of deglaciation in the marginal zone of Antarctica.

Areas: *a* – mountainous, mountain-valley, *b* – offshore, *c* – seaside, *d* – marine. *1* – The “ideal” line for reducing the area covered by glaciers; *2* – assessment of the area covered by glaciers; *3* – the advance, the expansion of glaciers; *4* – small glaciers, glacial domes; *5* – outlet glaciers; *6* – shelf glaciers; *7* – expected events or their time limits (area numbers in Fig. 1).

In the second phase, from about 7500 to 4000 BP, deglaciation slowed down. The climate became colder and drier. In the mountainous Untersee and Amery oases, climate cooling even led to a slight increase in the area of local glaciers and snow fields. On the coast, temperatures were higher, which, together with the warming effect of the territories freed from ice, ensured a slow degradation of glaciation. Push moraines in the periphery of some oases mark the advance of shelf ice and outlet glaciers over the oases. These moraines were deposited in different times. However, everywhere, the advance of glaciers began upon lowering of sea level. Apparently, since 7500–6000 BP, glacioisostatic land uplift began to outpace eustatic sea level rise, and glaciers on the shelf grew due to cooling. Their edges rested on the rising edges of the land and created moraines. Thus, responding to global changes in general, the course of deglaciation strongly depended on the location, glacial surroundings, and local climate of the oases, as well as on differences in parameters of their glacioisostasy. As a result of the second phase, no less than 80 % of the modern area of peripheral oases was freed from glaciers, and reconstruction of local glacial systems took place.

The third phase (from about 4000 to 1000 BP) was characterized by warming 4000–2000 BP followed by cooling 2000–1500 BP and by general lowering of the RSL interrupted by a slight rise in 2500–1500 BP. In mountainous oases (where there is a minimum dependence on the influence of the ocean), warming led to a reduction in glaciation, and cooling could cause the growth of glaciers and snowfields (Amery Oasis). On the coast, along with the continuation of deglaciation, local growth of glacial domes took place between 4000 and 2000 BP because on the increase in air humidity and precipitation; such areas are known of James Ross Island and Windmill Islands. Probably, the same was the mechanism of a small growth of glaciers in the areas of the Terra Nova Bay, and the Larsemann Hills and Vestfold oases, although the chronology of these events is not quite certain, and they cannot be unambiguously linked with climate warming. Glaciation of coastal oases beyond ice shelf did not increase during that period; the local climate remained dry due to the considerable distance of these territories from the humidifying influence of ice-free ocean (ice shelves had a width of up to 100 km).

On the shores of several oases, there are indications of the advance of glaciers (push moraines) between 2000 and 1000 BP. This could be caused by climate cooling 2000–1500 BP and by fluctuations in sea level, which led to the resting of the edges of the outlet and shelf glaciers on the rising shores. Other mechanisms are also possible. For example, the advance of the outlet glacier over the land surface in the Vestfold Oasis can be explained by the retreat of the ice sheet and the cessation of its damming action on the outlet glacier [*Adamson and Pickard, 1986*]. The cooling could also contribute to the growth of glacial domes on King George Island and in the Terra Nova Bay. In general, the influence of local conditions on deglaciation increased in the third phase. By the end of this phase, the environmental conditions in the considered regions began to resemble the modern conditions, which is confirmed, for example, by the mass settlement of penguins and snow petrels that began in 2000–1000 BP [*Verkulich, 2008*].

The fourth phase started about 1000 BP with a decrease in the RSL and climatic fluctuations. The environmental conditions were generally similar to the modern ones. However, there were at least two episodes of the expansion and reduction of glaciers in some areas. The reduction of glaciers is indicated by presence of sea and lake sediments in moraines along the edges of glaciers in the Vestfold Oasis, on the shores of Terra Nova Bay, and on King George Island. Their age varies from 1200 to 700–600 BP, which makes it possible to link the retreat of glaciers with a warming of about 1000 BP, which is recorded in ice cores [*Masson et al., 2000*]. Then, the glaciers in these areas advanced again to their modern borders. Indications of the advance of glaciers by 150–500 m exist in the form of moraines with an ice core that are known on James Ross Island, the Soya Coast, and the Schirmacher and Untersee oases. The advance of ***the Muller ice shelf*** took place later (after 400 BP) [*Domack et al., 1995*]. The exact time of glaciers' growth is unknown, but most scientists associate it with the LIA.

CONCLUSIONS

The analysis, interpretation, and synthesis of the results of paleogeographic studies make it possible to update the reconstructions of climate, sea level, and glaciation changes in particular peripheral regions of Antarctica and to identify the general pattern of the development Antarctic margins over the past 50 000 years under the influence of global, regional, and local factors.

During the interstadial period (MIS 3), the boundaries of onshore glaciers and ice shelf along the Antarctic periphery, as well as the conditions of sedimentation in seas, lakes, and on land were close to modern ones. During this period, the RSL could rise to a height of up to 30 m relative to modern marks, and its changes mainly depended on regional tectonic processes that are not directly related to changes in glacial load.

The active development of glaciation during the LGM (from about 26 000 BP) in the peripheral zone of Antarctica began under conditions of a drop in sea level by tens of meters and global cooling. The increment of glacial masses on the shelf was faster than the growth of onshore glaciers causing a shortage of atmospheric moisture supply in continental areas. During the LGM, there was a system of relatively thin (less than 300 m) glaciers on land and very thick (often, more than 1000 m) mobile ice shelves mobile in deep depressions and less mobile (anchored) on level shelf areas.

Deglaciation of the peripheral zone of Antarctica began around 17 000 BP and was not the root cause of the beginning of the planetary process of deglaciation; actually, it followed global changes in the climate and sea level.

Post-glacial climate changes in most areas had a general trend (warming in the Early Holocene until about 8000 BP and in 4000–2000 BP, cooling in 2000–1500 BP), but there were also local differences associated with factors such as latitude, altitude, features of the glacial surroundings of Antarctic oases, and distance from the sea.

Holocene curves of the RSL attest to a general rise in sea level in the Early Holocene up to maximum sea level heights in 8000–6000 BP with a subsequent tendency sea level staged drop: fast in 3500–3000 BP and slowed down afterwards (with a possible stabilization and even some rise in 2500–1300 BP in some areas). Regional differences in the amplitude of sea level changes were caused by local tectonics and deglaciation intensity.

In general, four main phases can be distinguished in deglaciation of the peripheral zone of Antarctica: from the end of the Late Pleistocene – the beginning of the Holocene to 7500 BP; from 7500 to 4000 BP; from 4000 to 1000 BP; and the last 1000 years. Deglaciation rates were high in the Early Holocene (up to about 7500 BP) due to warming and marine transgression. Then, deglaciation slowed down. The advance of the outlet and shelf glaciers in 6500–4500 BP was due to a decrease in sea level and climate cooling. In 4000–1000 BP outlet glaciers and ice shelves could react to changes in sea level, and onshore glacial domes grew according to the scheme “warming– increase in humidity–growth of snow and ice accumulation”. During the LIA, push moraines were deposited in some peripheral areas of Antarctic oases attesting to a slight expansion of glaciers due to cooling. Thus, the influence of global climate and sea level changes on the development of glaciation and the environment of the Antarctic periphery was the strongest in the Early Holocene; then, it weakened simultaneously with the increasing importance of regional and local factors.

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