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Evolution of marginal zones during continued glacial retreat in northwestern Wedel Jarlsberg Land, Spitsbergen

ABSTRACT: Complex analysis of geomorphological glacial processes in forefields of retreating glaciers in the southern Bellsund region was undertaken. Field studies and archival cartographic materials indicate continued glacial retreat, at least since the turn of the 19th and 20th centuries. Its average present rate of about 20 m/y is rather high and no moraine ridges are formed, except for the fluted moraines. Forefields of large glaciers are characterized by typical relief and sediments of frontal deglaciation only. Intensity and extent of glacial forefield remodelling depends mainly on activity ablation waters. At present glaciofluvial erosion predominates.

Key words: Arctic, Spitsbergen, glacial geomorphological processes.

Introduction

Geological and geomorphological studies in forefields of several glaciers in the Bellsund region were carried out in 1986–1994. Collected material allowed for an analysis of contemporary processes, responsible for relief development in glacial marginal zones. Seasonal character of studies made estimation of annual changes of glaciers dynamics difficult and presentation of glacier mass balance impossible. Field studies of glacial marginal zones in the southern Bellsund have been carried out several times. The geological expedition headed by Różycki (1959) and the expedition of the Silesian University (Wach 1981) collected some observations in this region. Single data, particularly on the surging phenomena, were carried out by the Norwegian glaciologists (Hagen 1988a, Hagen and Liestøl 1990, Hagen, Lefauconnier and Liestøl 1991). Intensification of studies took place after 1986, owing to a geological mapping by the Norwegians (Dallmann *et al.* 1990) and the Geographical Expeditions organized by the Maria Curie-Skłodowska University of Lublin (Poland).

During the latter, a preliminary study was done in forefields of glaciers close to the station at Calypsobyen (Pękala 1987), with special attention to marginal zones of Renard, Scott and Blomli Glaciers (Szczęsny *et al.* 1989). Examination focused on ground moraines, often the fluted ones (Merta 1988b, 1989) and on sandurs (Łanczont 1988a, 1988b, Warowna 1994). Glacier retreat was roughly evaluated (Reder 1991, 1993). Studies of the Quaternary stratigraphy were of special importance (Pękala and Repelewska-Pękalowa 1990). Recognition and dating of organic sediments and archaeological finds allowed a preliminary palaeogeographical reconstruction of glacier forefields (Dzierżek, Nitychoruk and Rzętkowska 1989, 1990, Jasiński 1994).

Aim, range and methods

The studies focused on marginal zones of four large and few small glaciers in the southern Bellsund region (Fig. 1). Main aim was to evaluate deglaciation and its rate on the basis of field materials, TL datings and archaeological finds. Results were compared with other regions of Spitsbergen.

Geomorphological mapping with use of topographic maps in various scales was the main field method. In case of forefields of Renard, Scott and Blomli glaciers, a photogeological map was prepared in scale of 1 : 10,000, with a use of air photos of 1960 (Szczęsny *et al.* 1989). In other areas, a topographic Norwegian map in scale of 1 : 100,000 and enlarged to 1 : 50,000 was used. Additional information was obtained from analysis of the Norwegian air photos of 1960 and 1990, rendered accessible by the Norwegian Polar Institute in Oslo.

Glaciation of Spitsbergen and its morphological effects

During the Middle Vistulian, an ice sheet of the Barents Sea occupied the Barents Sea basin and the surrounding islands, one of which was Spitsbergen. Maximum extent of the ice sheet occurred at 46–42 ka BP (Baranowski 1977, Mangerud *et al.* 1992). The western coast of Spitsbergen was then a boundary zone of a large glaciated area. Glacial cover of this part of archipelago consisted of outlet glaciers, whereas thick ice filled up valleys and fiords. Maximum isostatic loading of the western coast, the amplitude of 75–80 m, should be also associated with this period (Landvik, Mangerud and Salvigsen 1987). The Barents Sea shelf ice sheet glacier decayed completely at about 16–20 ka BP (Baranowski 1977, Boulton 1979). Since that time, an independent trace ice bowl

(Nordaustlandet) as well as system of outlet and valley glaciers connected through the alimentary area, have been formed. The following climatic warming before the Holocene, caused gradual decrease of glacier range and thickness. At 11–12 ka BP, the glaciation in Bellsund occupied the whole Van Mijen Fiord up to the island Akseløya as well as the Van Keulen Fiord, including the Recherche Fiord (Mangerud *et al.*1987, 1992). Quick retreat of glaciers and complete melting of ice in the Van Mijen Fiord took place at the begining of the Holocene, between 11.2 and 9.6 ka BP (Mangerud *et al.* 1987, 1992). The same process in the Van Keulen Fiord and the Recherche Fiord took place probably at the same time. Isostatic loading of the coast, compared with the contemporary state, was equal to 50–55 m (Landvik, Mangerud and Salvigsen 1987, Salvigsen, Elgersma and Landvik 1991). Quick retreat of glaciers caused intensive land uplift — marine terraces at 25 m a.s.l. are dated at 9 ka BP (Salvigsen 1976, Salvigsen, Elgersma and Landvik 1991). Most glaciers became isolated due to increased ablation and decreased ice thickness. The most important change was reflected by change of a thermal regime of glaciers from cold to polythermal ones. During such change, some glaciers underwent sudden surge-type advances indicating that increased ice-covered areas have not resulted directly from climatic oscillations. There is no evidence for such glacier advances in the southern part of Bellsund but there is some for the Is Fiord region.

The Holocene climatic optimum with high temperature, humidity and precipitation has not resulted in significant fluctuations of the glacier extent — increased ablation was compensated by intensified precipitation and accumulation. These forms were created at large snow accumulation on slopes in a wetter climate than the present one. Baranowski (1977) found the Holocene optimum to be a main period of rock glacier formation. During colder but still wet climate just after the climatic optimum, glaciers in Spitsbergen advanced at 3.5–2 ka BP (Baranowski 1977). This advance resulted in development of large ice-cored moraines which constitute at present a predominant element in forefields of most glaciers that terminate on land.

Vast glaciological and geomorphological literature reports a glacier retreat from 2000 to 750 years BP (Baranowski 1977, Boulton 1979, Miller *et al.* 1989), due to which glaciation of Spitsbergen was minimal and extent of most glaciers was much smaller than at present. Next cooling from 750 to 100 years ago, known as the Little Ice Age in all glaciated areas due to global change of climate, caused another — the last advance of glaciers in Spitsbergen. Advancing glaciers formed intramorainal systems of ice-cored moraines.

Datings of organic sediments and driftwood from the raised marine terraces as well as numerous geomorphological observations indicate that the coast of Spitsbergen raised slowly during the Younger and the Middle Holocene. The rate of land uplift during the last thousand of years is assumed at 0.2–0.3 m/100 years (Baranowski 1977, Salvigsen, Elgersma and Landvik 1991).

Glacier retreat in Spitsbergen has been recorded since the first geodetic measurements on Svalbard and probably started during the Little Ice Age. Many researchers took interest in changeability of Spitsbergen glacier extents which resulted in many general and detailed papers (Ahlmann 1953, Koryakin 1967a, b, Jania 1986, 1987, 1988 and others). First comparison of changing extents of some glaciers were done by the Norwegians from the Norwegian Polar Research Institute in the fifties (Ahlmann 1953). A large number of glaciers was compared by Koryakin (1967a, b), based on his own investigations in 1965–1966, the Russian (1900–1901) and the Norwegian (1939) archival cartographic materials; he analysed also extents and thickness of 165 Spitsbergen glaciers. The results proved a significant retreat of most glaciers (96), though they did not allow for explicit conclusions as 34 glaciers advanced and 33 stagnated. This research proved changes in retreat rate and in most cases a drop of withdrawal rate in 1936–1966, if compared with 1912–1936. Small glaciers of up to 2 km^2 in area, underwent the largest reduction. According to Koryakin, withdrawal of glaciers in the Van Keulen Fiord region (just in the whole central part of the western coast) in 1936–1958 was much smaller than farther to the north. Retreat of glaciers in Hornsund was much more distinct in the same time. Greater changes in glacier extents (generally retreat but also some advances) were observed at the eastern than at the western coast (slow retreat, few advances). In most cases, these were the surge-like advances.

Surging phenomena in Spitsbergen are known from over 80 glaciers (Liestøl 1993). Many more glaciers are known to have surged in the past. The greatest surge was observed on the Brasvell Glacier (a part of the Nordaustlandet ice cap) in 1936–1938 when the glacier front advanced by 20 km. The Negri Glacier, inside the Stor Fiord in eastern Spitsbergen, surged 12 km with a velocity of 35 m a day in 1935–1936 (Liestøl 1969). In Bellsund, the surge was recorded at the glaciers: Recherche, Scott, Sieger, Martha, Martin, Bakanin, Hylling, Skut, Luncke, Are, Charpentier, Penck, Fridtjov, Finsterwalder and Hess (Liestøl 1993). The active surge phase for the Spitsbergen glaciers has usually lasted a few years (3–10) but there are some cases of over 10 years advances, equal to 85 m a year — the glaciers Hyllinge and Kjelstromdalen (Dowdeswell, Hamilton and Hagen 1991).

It is very difficult to determine the glacier mass balance which is a resultant of accumulation and ablation during a hydrological year. Accumulation and ablation changes in successive years: winters have different snowfalls and the average temperature of ablation periods is inconstant. The glacier balance changes from year to year (Hagen 1988b, Hagen and Liestøl 1990, Hagen, Lefauconnier and Liestøl 1991, Lefauconnier and Hagen 1990, Jania 1994). Studies of glacier mass balance started in Spitsbergen in the fifties. A negative balance dominated during the last forty years (Fig. 2). Such studies were carried out in the Bellsund region at the Finsterwalder Glacier in 1950–1969 only

(Hagen and Liestøl 1990). A positive mass balance of this glacier occurred in 1952/54 and 1956/58 only. The Soviet studies carried out for 5 glaciers (Is Fiord region as well as northern and eastern coast of Spitsbergen) in 1966–1984 indicated absolute predominance of the negative net balance. The positive values were obtained for the hydrological year 1981/82 only (Hagen and Liestøl 1990).

So far, a complex model of relations between climatic conditions, mass balance and changes in glacier geometry have not been worked out. It is mostly due to delay of glacier response to changing conditions, accumulation and thermal regime of individual glaciers, fitting a longitudinal profile every time. Changing extents of glaciers can be good indices of climate and mass balance for the very long intervals only.

Contemporary morphodynamic processes in subpolar forefields of glaciers in western Spitsbergen are very complicated. They are to be subdivided into two groups: the ones, connected with direct ice movement and the others, resulting from deposition or erosion by glacial waters. Another problem is the overlapping of glacial and glaciofluvial forms, the earlier elements which are not a direct effect of glacier activity in the environment but are explicitly connected with a glaciation. Glacioisostatic movements and eustatic changes of a sea level formed raised marine terraces. However, coasts were not raised uniformly and altitude of terraces does not determine their age.

Glacier marginal zones

Antoniabreen

The Antonia Glacier (Figs 1, 3) is a large outlet glacier. It joins the Amundsen Plateau in the upper part. Many a time it has been described in literature and vastly mapped, examined by Różycki in 1959 and by Wach in 1977 (Wach 1981) what enables a comparative analysis at present.

The Antonia Glacier has been actively retreating, at least since the beginning of this century. The earliest data about the glacier front come from the Norwegian maps and refer to the period 1918–1920 (Różycki 1959). Comparison of these materials with maps based on air photos of 1939 indicates significant topographic changes. The glacier front retreated afterwards by about 150 m *i.e.* 9.3 m a year. Measurements on the Antonia Glacier by A. Marcinkiewicz in 1958 indicated a retreat of 150–180 m if compared with 1936 *i.e.* on the average 8.1 m a year (Wach 1981). According to the measurements by Wach, the glacier retreat was equal to about 270 m in 1958–1977 *i.e.* its average rate increased to 13.5 m a year. The present glacier front moved more to the south as compared to the measurements by Wach. The glacier retreat in 1977–1992 could be esti-

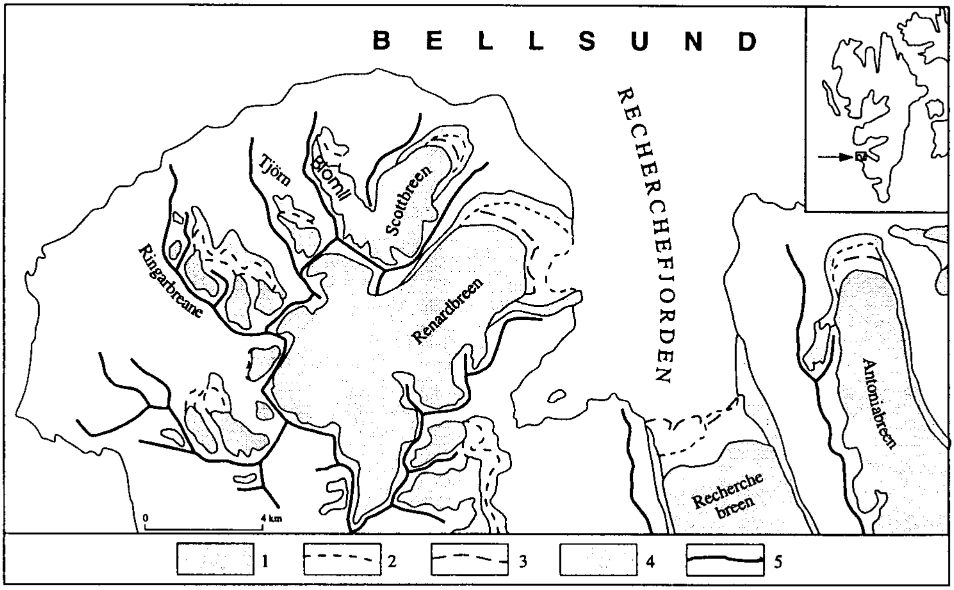

Fig. 1. Location of the studied area and recent changes of glacier front extent.
1 — present glaciers; glacier snouts in: 2 — 1936, and 3 — 1960; 4 — ice-cored moraines, 5 — main mountain ridges.

mated at 200–250 m and therefore, its rate increased to 15–19 m a year. These data indicate a growing retreat rate of the glacier.

At present, the Antonia Glacier is 7.5 km long, its average width is 2 km but the front of regular shape is 1.5 km wide. The glacier front proves a mixed deglaciation type. In the western part it is flat, without distinct separation from a marginal zone with a predominant areal deglaciation. In the east, the front becomes steeper and assumes a form of a convex bowl which is distinctly separated from the forefield with predominant frontal deglaciation. Glacier surface is covered by large patches of surface and ablation moraines close to the front. Inglacial and subglacial outflow is active but surface outflow is of secondary importance. Inglacial rivers carry away much mineral material, deposited just after leaving the glacier in its close forefield where some small inner sandurs are formed. They are long, narrow taluses at outlets of inglacial tunnels. Melting of ice occurs commonly at proximal parts of the taluses.

Ancient subglacial drainage disclosed in the forefield of the Antonia Glacier creates joints and particularly N-type channels in bedrock (Nye 1973, Jania 1993). Course of main channels depends on tectonics: their direction agree with cracks, dislocations and tectonic loosening predominant in this region. The subglacial drainage has a distributive-braided character with numerous embranchments, probably due to overcharging of streams. The channels are slightly winding, being cut into the bed to 1.5–4 m. Few flatnesses — remains of former

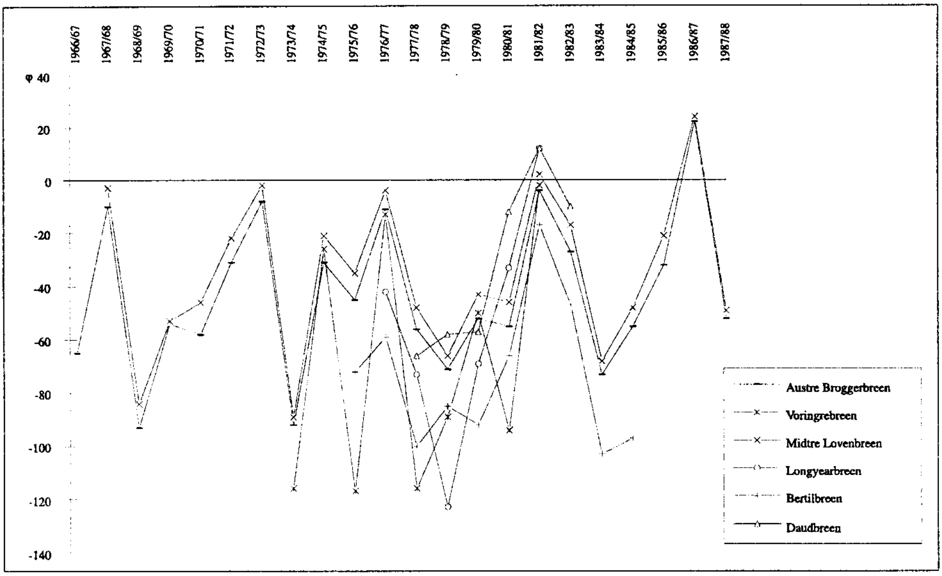

Fig. 2. Mass balance of several glaciers in 1966–1988 (after Hagen and Liestøl 1990).

subglacial stream bottoms, distributed in an irregular way at different altitudes occur at steep channel sides. Channel bottoms are uneven with numerous evorsion cavings and edge undercuts. Contemporary relief of the forefield presents a drainage system, covered by glaciofluvial material similar in petrography and grain size to the material of sandurs and moraines. It has been probably deposited later when the area was a glaciofluvial depositional zone.

In the intramorainal zone there are patches of ground moraine, large areas covered with icings throughout a summer as well as concentrated non-braided stream beds. Ice-cored hillocks are the most common landforms. They are of different size, from little features to the ones 20 m long and 5–6 m high. Most of them have longitudinal esker-like shapes.

The most distinct forms in a marginal zone are frontal ice-cored moraines (Pl. 1). They are located at 51 m a.s.l., being 32 m high at the side of the outer-morainal sandur. Analysis of maps indicates that the present ice-cored moraines were formed in 1920–1936. The frontal moraine at the geological map of Różycki (1959) is composed of small hills (without ice-core), 2 m high. They adjoin the main moraine rigdes. This frontal moraine does not semicircle closely a front of the Antonia Glacier. Glacial rivers use 4 ravines to flow out into the outer-morainal sandur. At distal part of the frontal moraine there are three sandur levels, cut into the raised marine terraces (Pl. 1). The oldest sandur is found at the southern side and is the highest. Close to the active glacial gate, the height difference in relation to the present sandur is equal to 3–4 m. The surface of active extramarginal sandur is at its lowest level similar to the two neighbouring

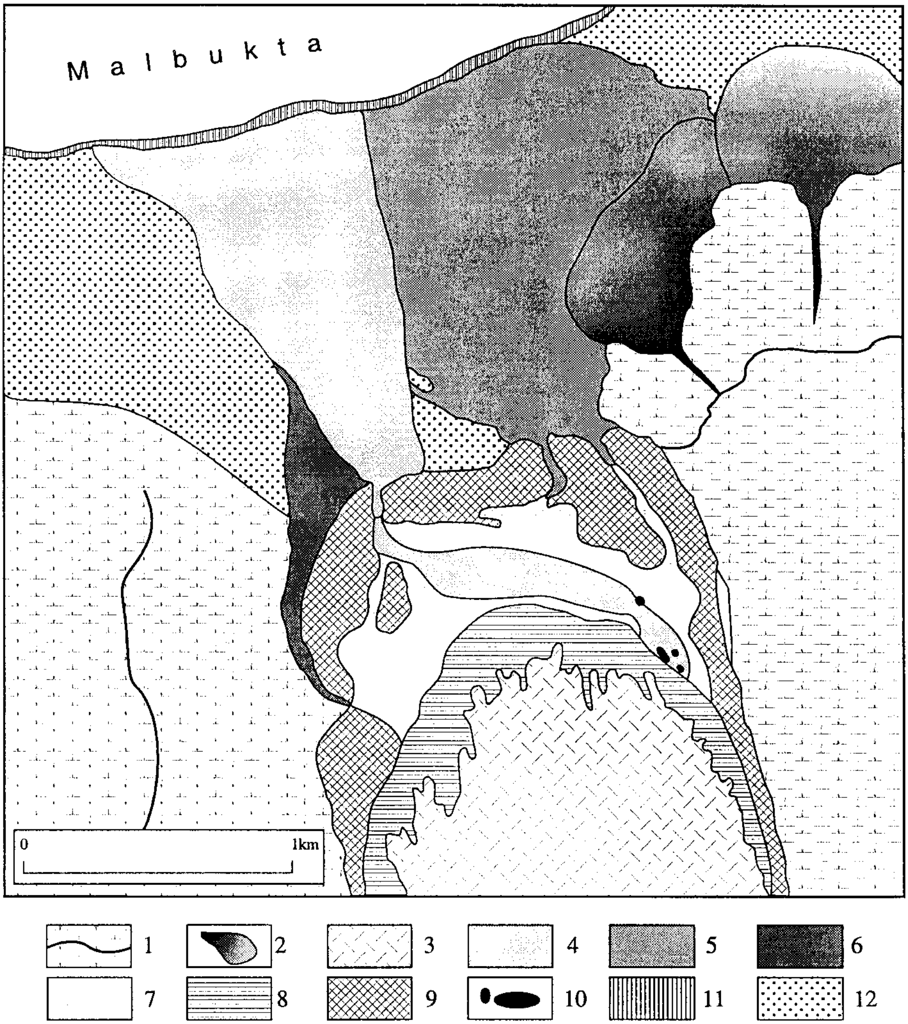

Fig. 3. Marginal zone of the Antonia Glacier.
1 — slopes and main mountain ridges, 2 — talus cone, 3 — glacier, 4 — youngest sandur, 5 — younger sandur, 6 — older sandur, 7 — ground moraine, 8 — ablation till, 9 — frontal and lateral moraines, 10 — eskers, 11 — present beach, 12 — marine terraces.

ones. There are two relic hills of the older level, indicating intensive erosion there. On the opposite side, a dead sandur plain is built of thick stony material. The beds in its upper part occur to a depth of several centimetres. Some of them run water occasionally only. This sandur was active when the glacier reached the frontal moraine, then it was abandoned as shown by lack of fine sediments from the phase of flow decay (Warowna 1994). The border part of the sandur along the estuary to Malbukta has a 4 m wide storm ridge, which checks outflow and formation of small lagoons in front of it.

Recherchebreen

The Recherche Glacier (Fig. 4) is the largest in the Bellsund region and constitutes a southern termination of the Recherche Fiord. It is about 15 km long and 3.5–4 km wide. Its upper part attains 450–500 m a.s.l. and joins the firn field of the Torell Glacier. The glacial front is flat what is typical for retreating glaciers. The Recherche Glacier presents a mixed deglaciation type. In the central part of the glacier, there is a several metre ice-cliff (Pl. 2) where ablation by calving is predominant. A frontal deglaciation prevails in the east and an areal one in the west. During the last several dozen of years, a glacier front underwent numerous fluctuations. In 1936, the glacier occupied an area of about 3 km^2 larger if compared to the present state, the glacier front moved northwards by about 550 m in the west and almost 1000 m in the east. Air photos of 1960 present quite another situation, with a more extensive glacier front. The Recherche Glacier experienced significant advance, caused by changing thermal regime between 1936 and 1960.

The first surge of the Recherche Glacier is dated at 1838 (Liestøl 1969). The next glacier surge took place between 1936 and 1948 (Koryakin 1985) or in 1945 (Liestøl 1993). Very little traces remained after this advance. The material was deposited at bottom of Fagerbukta. Constant and relatively quick retreat of the Recherche Glacier front has been observed since 1960. During the last thirty years over 3.5 km^2 of the forefield was disclosed and the glacier front moved southwards by about 1300 m. The average rate of retreat was equal to 43 m/y in that time. Such high average retreat rate (1960–1990) was not confirmed by the observations carried out at the turn of the eighties and nineties. Very quick changes of the glacier front took place probably just after surging.

In the area of glacial wasting, there are the surfaces which have been mostly shaped by action of ablation waters (Pl. 2). Glacial meltwaters get out mainly from inglacial tunnels (Pl. 3). The largest outflow occurs at contact of the glacier and lateral moraines. In glacier forefield, the streams are braiding, spilling sandur fans on a ground moraine. During intensive ablation, meltwater cut a sandur surface with numerous beds.

In the west, a gentle slope of the front made favourable conditions for sandur development not only in the forefield, but also partly on a snout surface. This material forms a mantle that locally reduces ablation and most sandur sediments are deposited on dead ice. The waters outflowing from glacial gates at contact with lateral moraines, assume a direction parallel to the front. At the same time they remove glaciofluvial material, reaching soon a dead ice below. This ice undergoes very quick ablation and created depressions are a few metres deep. At the same time, inselbergs of dead ice covered with sediments of the former sandur level, remain between the river beds. The material slips down into the depressions, where deformation of layers and complete isolation of ice-cores

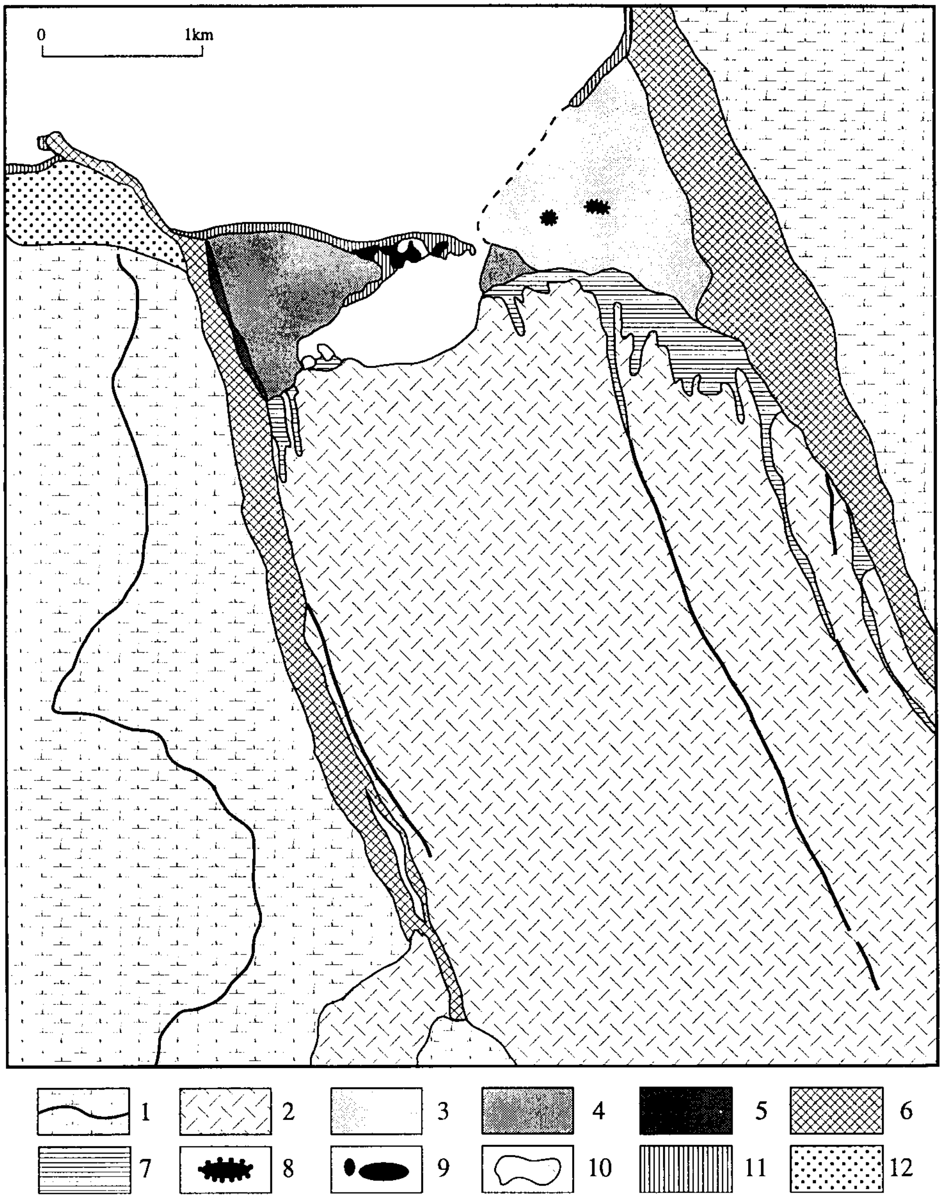

Fig. 4. Marginal zone of the Recherche Glacier.
1 — slopes and main mountain ridge, 2 — glaciers, 3 — youngest sandur, 4 — younger sandur, 5 — older sandur, 6 — lateral moraines, 7 — ablation till, 8 — push moraine, 9 — eskers, 10 — kame terraces, 11 — present beach, 12 — marine terraces.

take place, resulting in their conservation and slower ablation. Recurrent processes formed a system of kame hills with an ice-core inside (Pl. 2) and ablation water outflowing from a glacier front. Their location is unstable: they can translocate during each highwater period. Their beds are covered with chipping material of varying thickness, including also a fine-grained one. At present, the glacial river flows almost parallel and not far from a glacier front, entering a bay at the ice cliff.

A vast abandoned sandur is generally a single level. Only at contact with lateral moraine, remains of an older level are observed. The level is dual (Warowna 1994). The upper part which is 1.4 m high, forms a ledge close to the moraine. The lower one rises 0.5 m above the youngest sandur, occupies a larger area and has a triangle-like shape. A system of 1.2 m high and 12 m wide coastal ridges occurs at distal side of the sandur fan in a coastal zone.

An active fan-wise sandur which enters the bay, constitutes an eastern part of the forefield. Relics (Pl. 2) of loam and sand from the bay are found in its area due to a quick glacier advance. Intensive ablation and erosion caused almost complete destruction of the former morainal zone.

Renardbreen

The central part of the northwestern Wedel Jarlsberg Land is occupied by the vast Renard Glacier (Fig. 5). It is fed from a few firn fields at 450–500 m a.s.l. The glacier is 8 km long and 3.5 km wide but slightly more at the front. The latter occurs at several metres a.s.l. whereas its central part reaches a sea level at Josephbukta. The front is flat-bulging and only over the bay waters, there is a several metre high ice cliff. Recently, the glacier retreated intensively (Reder 1991) and typical frontal deglaciation occurs. It ocuppies a larger area in 1936 as far as the ice-cored moraine ridges. Josephbukta was a small bay surrounded by ice cliffs in the south, west and north. In 1936–1960, owing to glacier calving, the bay extended mainly to the west and north. Then, the glacier retreat was relatively small, estimated from about 250–300 m in the south to about 500–600 m in the central part of the tongue. In a northern part of the glacier, the adjoining ice-cored moraines without a contact with the bay, remained stabile. Different situation was observed in 1960–1994, when the bay expanded slowly and the ice cliff retreated only about 100–150 m to the west. However, significant increase in glacier retreat took place in a previously stable part of the glacier. During the last thirty years, the retreat of the glacier front was equal to 1200–1500 m *i.e.* 40–50 m a year. Observations during the last few years proved that the contemporary values were smaller, about 20 m a year.

The Renard Glacier is drained by supraglacial and inglacial streams. Inglacial tunnels provide abundant meltwaters, particularly in southern and central parts of the glacier. At present, several tens of small and large supraglacial meandering gullies and large water discharge occur at the glacier surface. These waters are responsible for development of depressions in the glacier front area, characteristic relief of ground moraine in the forefield and also on the ice tongue surface (Merta 1989). Ridges of a fluted moraine constitute a cast of a glacier

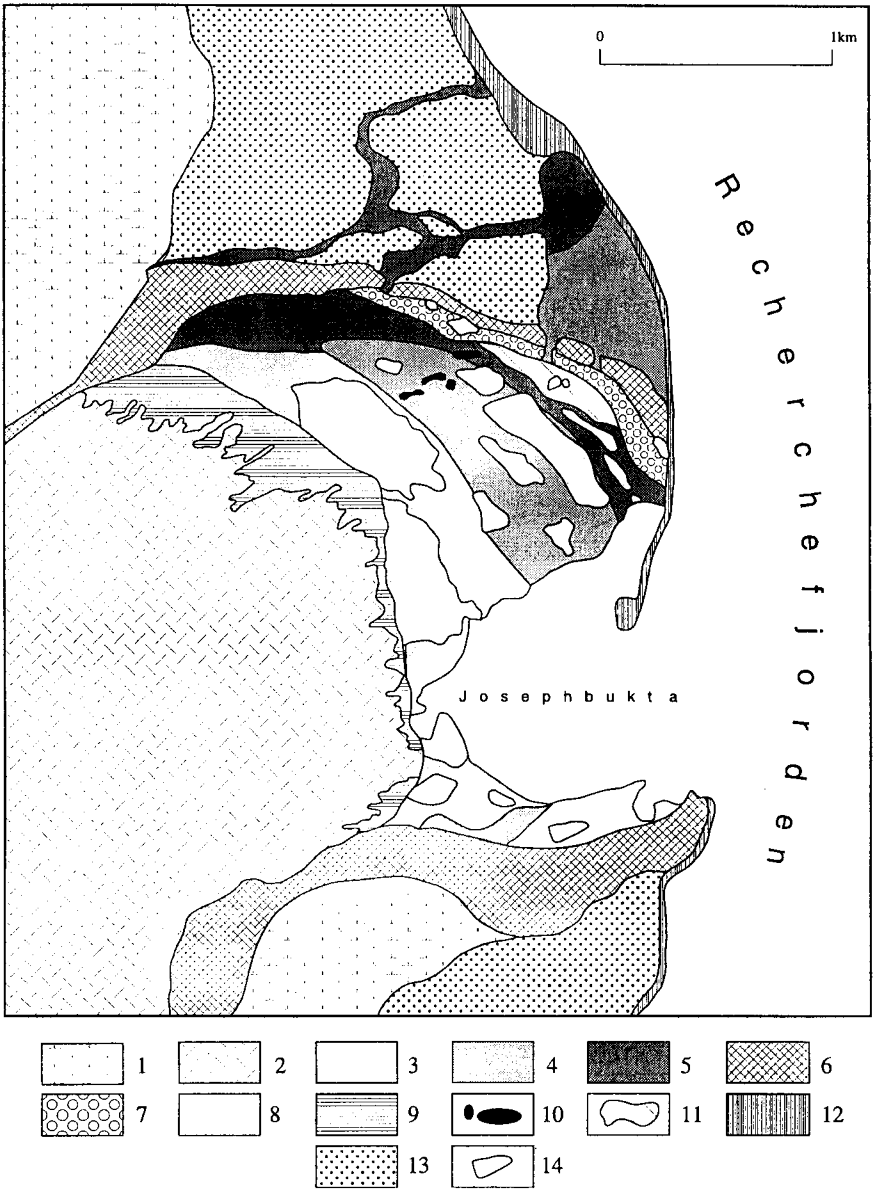

Fig. 5. Marginal zone of the Renard Glacier.
1 — mountain slopes, 2 — glacier, 3 — youngest sandur, 4 — younger sandur, 5 — older sandur, 6 — frontal and lateral moraines, 7 — ice-cored moraine, 8 — fluted moraine, 9 — ablation till, 10 — eskers, 11 — kame terrace, 12 — present beach, 13 — marine terraces, 14 — lakes.

front surface relief and their orientation determines direction of glacier withdrawal. The grooves behind large stones were considered for the effect of ablation gully development (Merta 1989).

Supraglacial drainage is responsible for large geomorphological changes at the glacier edge and its direct forefield. In the extraglacial area, rivers of this system join into an enormous marginal river. It remodels intensively the glacier forefield, manifested by complete supression of slot relief of fluted moraine and dissecting the sandur into single, isolated patches. The glacier forefield loses its smooth character and becomes a badland, with a very thick system of old beds as well as thermokarst depressions and crevasses.

In the intramarginal zone, intensive glaciofluvial deposition occurs. There are at least three sandur levels, kame terrace and eskers (Pękala 1987, Łanczont 1988a,b, Szczęsny *et al.* 1989). High level of sandur sediments creates terrace-type forms, adjoining the proximal sides of ice-cored moraines or deposited in depressions of a fluted moraine. Slope of glacier forefield towards the contemporary front (from the northwest to the southeast) resulted during glacier retreat in development of the lower and lower levels of intramarginal sandur.

Several 10 m high roches moutonneés composed of tillite bedrock occur between a sandur and a glacier snout. Proximal slopes of the roches moutonneés are gentle. They are covered with fluted moraine (Merta 1989). Distal slopes are steep and usually without sedimentary material.

A terminal moraine, passing into lateral moraines, is a predominant relief element of the Renard Glacier forefield (Pl. 3). At present this zone is not a continuous one, it breaks off at the contact with a fiord. Abrasive processes destructed a central part of the frontal moraine and cut a cliff in it. Duality of forms is distinct in the frontal moraine zone (Szczęsny *et al.* 1989, Dzierżek, Nitychoruk and Rzętkowska 1990). An outer ridge consists of mixed marine sediments and two layers of till. The older till was dated at 13.2±2 ka but the younger glacial sediments — at 8.6±1.3 ka (Pękala and Repelewska-Pękalowa 1990). Mixing and accumulating of marine and glacial sediments was caused by a quick glacier advance. Inner ice-cored moraines were formed during the Little Ice Age. Sharp, pyramidal hillocks and numerous thermokarst fissures and depressions (often filled with water) indicate a relic ice inside. Deformed organic sediments are found in sediments of ancient and ice-cored moraines. Their upper part was dated by ^{14}C method at 660±80 and the lower part at 1130±80 years (Dzierżek, Nitychoruk and Rzętkowska 1989, 1990). These organic sediments are concentrated within a cultural layer of the archaeological stand. The collected artifacts are referred to the West European whalers and Norwegian hunters, probably from the turn of the 16th and 17th centuries to the fifties of the 18th century (Jasiński 1994).

Dating and field observations enabled partial reconstruction of landscape development in the glacier forefield. The outer moraines were formed during the glacier advance of the Early Holocene. The glacier is covered by a younger moraine material due to a long ablation. Another glacier advance took place during the Little Ice Age and resulted in formation of inner ice-cored moraines.

Dating of translocated artifacts determines precisely the glacier maximum advance to the second half of the 18th and first years of 19th century. In that time, a glacial drainage system consisted of two gates cut in a frontal moraine (Reder 1993). In the first stage of glacier retreat, intensive glaciofluvial deposition occurred and a kame terrace was formed at contact of ice and moraine (Szczęsny *et al.* 1989). It is composed of sands, with inserts of gravel and flow till. Checking of outflow in the northwest direction formed a marginal river, parallel to a glacier front and flowing towards Josephbukta. The waters cut a deep valley in the moraine; its dry bed is covered at present with sand-and-gravel sediments. Further retreat caused development of a fluted moraine, sandur sediments and fissures, in which eskers were formed (Pl. 4). The older sandur series is covered by a ground moraine, several dozen centimetre thick (Merta 1989). There are eskers on a distal side of the roche moutonneé — on surface of the sandur series. They were formed during glacier retreat in middle of the 20th century. Their course resembles glacier crevasses and direction of ground moraine grooves in the forefield.

A southern part of the frontal moraine, relics of which are preserved only, has longitudinal bulges and depressions ("microflexures") which are parallel to one another. Frontal moraines of surging glaciers have similar morphological features. During the active surge, compression resulted in development of frontal moraine flexures (Jania 1993). As the older part of the Renard Glacier frontal moraine displays some features of the accumulated moraine, it can be regarded as effect of a former glacial surge.

Scottbreen

The Scott Glacier (Fig. 6) is only 3.5 km long. It is to 1.8 km wide but to 0.9 km at the front. It has two firn fields at 400–450 m a.s.l. The Scott Glacier has a detailed photogrammetric map (Merta, Ozimkowski and Osuch 1990), done during the 2nd Spitsbergen Expedition organized by M. Curie-Skłodowska University in 1987. This examination of the actual glacier front enabled a comparison with a photogeological map of the same area, prepared with use of the Norwegian air photos of 1960. During the past 27 years, the Scott Glacier underwent significant changes. The glacier front receded by 530 m *i.e.* 20 m every year and this trend is still observed. The glacier front dropped at about 75 m *i.e.* 2.7 m/y on the average. The glacier is an example of a typical frontal decay.

According to Liestøl, the Scott Glacier surged at about 1880 (Liestøl 1993). It did not overpass however the maximum extent line of the Middle Holocene, indicated by a frontal moraine series. Recent glacial sediments cover the ancient moraine.

A typical sandur does not occur in the intramorainal zone. It is rather a zone of cut and washed ground moraine, with pouring small taluses. Inner part of the

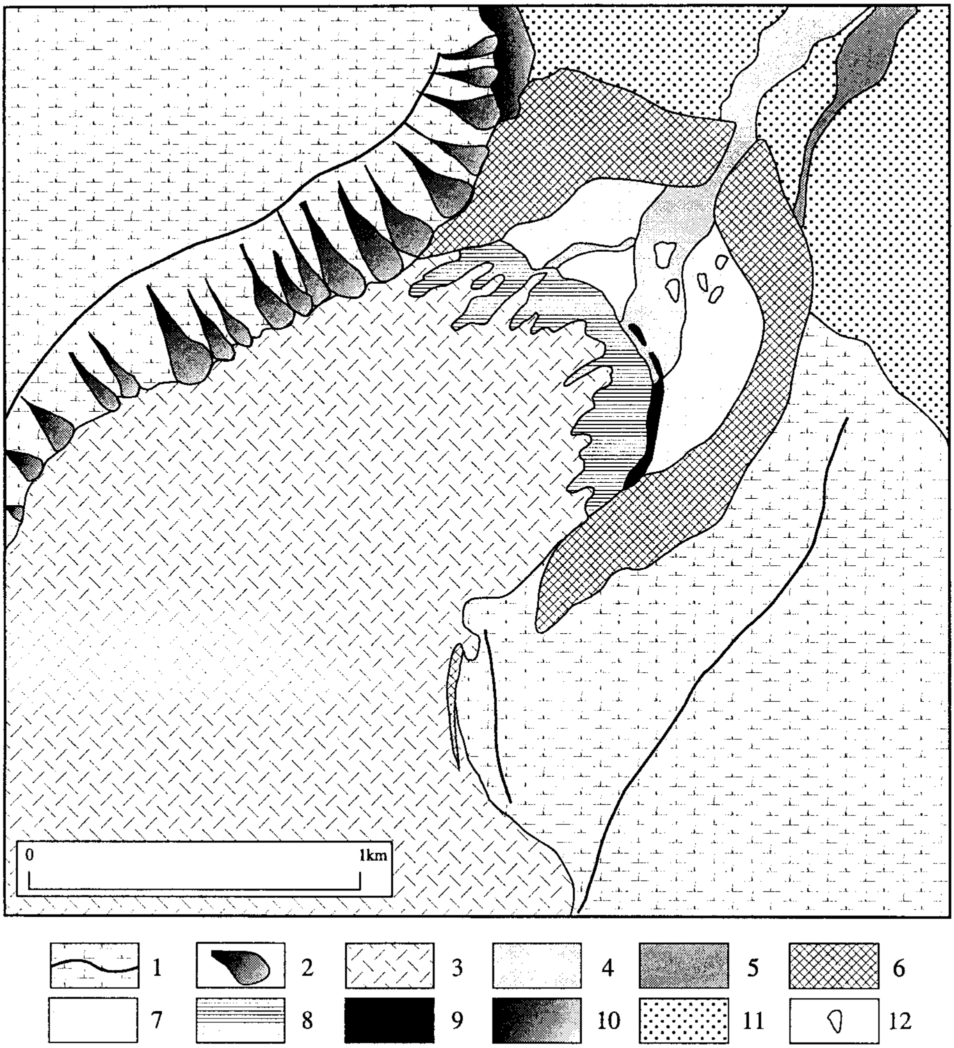

Fig. 6. Marginal zone of the Scott Glacier.
1 — slopes and main mountain ridges, 2 — talus cone, 3 — glacier, 4 — youngest sandur, 5 — younger sandur, 6 — frontal and lateral moraine, 7 — fluted moraine, 8 — ablation till, 9 — kame terraces, 10 — rock glaciers, 11 — marine terraces, 12 — lake.

frontal moraine passes gradually and gently into a ground moraine with the evident flow towards the only active snout. Small hills of a glacier retreat phase are the only outstanding element of the relief. The intramarginal zone has a depression character, a southeastern part of the area is lower and mostly flooded with water. A northeastern part is located a few metres higher somewhere from the moraine snout. This zone is formed by a ground moraine, in some places a fluted one, cut with a few 2–4 m deep active beds. Kettles in moraine surface are filled with meltwater. A thin layer of silts is deposited there. Remaining fragments of a moraine do not show any sign of washing away. Close to the narrow snout in a frontal moraine there are flood waters, in which deposition of fine-grained material occurs (Merta 1988a). Main outflow of subglacial water is observed in the south. A kame terrace occurs between the glacier and a lateral moraine.

There is only a supraglacial drainage from the Scott Glacier. The only inglacial tunnel, in the northeastern border area of the glacier tongue, forms a marginal river that actively creates a contemporary sandur. At present supraglacial gullies do not occur. The water flows down the inclined surface of the glacier tongue. The surface moraine which thaws out during the ablation season, is washed by these waters immediately. A coarse material (stones) is distributed disorderly on ice surface whereas fine material is carried towards the glacier lower parts or central depression of the forefield.

All these glaciers have thick icings in their forefields. The Recherche and Scott Glaciers occasionally surged (Liestøl 1969, 1993), this phenomenon has been probably also present for the Renard Glacier. So far, surges have not been noted for the cold glaciers. These are typical subpolar glaciers or, according to Jania (1993) — the polythermal ones. Another evidence is an extensive system of former subglacial drainage of well preserved Nye's channels in the forefield of the Antonia Glacier.

Large icings in glacier forefields indicate that they are formed by winter-active streams. The icings are 1.5–2 m thick at the beginning of a summer. Proglacial rivers on their surfaces cause melting of meander beds and deposit a poorly sorted material. Vast icings make usually a forefield transformation impossible. They check development of marginal sandur and less or more influence a proglacial outflow.

Blomlibreen

The Blomli Glacier is located to the west of the Scott Glacier and joins its firn field. It occupies the upper part of the Blomli Valley, its firn field occurs of 350–400 m a.s.l. and the residual, steep tongue moves down to 200–220 m a.s.l. A snout is mantled with a thick layer of rocky material from neighbouring mountain slopes and ablation till, which passes into a basal till in the forefield. The glacier marginal zone is small. It is enclosed by a steep roche moutonneé with an oblique course along the valley. A frontal moraine is extremely large if compared to the glacier itself. During the Holocene advances, the glacier has never overcrossed a rocky threshold. The ablation is insignificant and a retreat rate cannot be estimated by field observations during the last few years. Analysis of air photos has not given desirable effects either and indicates slow transformation of the tongue into a rock glacier. A frontal moraine is composed of ice-cored ridges. At present, there are no glacier outflows. The entire movement of ablation waters takes place in thermokarst canals inside the moraine material. Thus, the forefield area is non-typically shaped. The intramarginal zone is formed of narrow depressions between a steep snout and roche moutonneé, mantled with a till. Neither icings nor inner sandurs were observed here. Studies of Merta (1989) proved that the ground moraine was a fluted one. The material is regularly distributed, with longitudinal strips on the glacier surface what therefore confirms a supraglacial origin of the fluted moraine (Merta 1988b).

Tjørndalsbreen

The Tjørndals Glacier is similar to the Blomli Glacier and adjoins it in the west. However it is smaller, fills up only the upper part of the valley and is pushed westwards by a roche moutonneé. The residual tongue is closely mantled with thaw and slope material. Narrow ridges of median moraines are also disclosed. A lateral moraine is located at eastern side of the glacier. Lack of frontal moraine indicates poor dynamics during the Holocene. A huge rocky threshold in the central part of the valley axis is covered with a till. The latter does not create distinct ridges that would manifest extent and phases of glacier retreat. A rock valley bottom below the roche moutonneé barrier is deprived of glacial sediments. They were washed away by meltwater rivers and streams. Only in the bottom depression as well as permanent and temporary lakes, the sandur sediments could be preserved (Szczęsny 1987).

Ringarbreane

The Ringar Glaciers (Fig. 7) fill up potholes in the upper part of the Dyrstad Valley. The glacier is composed at present of four separate fragments in four mountain pot-holes above 250 m a.s.l. Lower parts of residual glacial tongues descend about 50–60 m. Firn fields occur at a structural flatness at 450 m a.s.l. The contemporary glacier keeps actively retreating. The frontal zone of glaciers is flat and covered with ablation moraine. A ground moraine with numerous tongues fed by supraglacial waters occurs between fronts and ice-moraine ridges in a glacial forefield. Frontal moraine zone is composed of three rows of ice-cored moraines. In some places, they are overlapping and passing into a lateral moraine in the east. Ice-cored moraines are composed of coarse, block material of an extreme diameter of 1–2 m. Ice-cored moraines are not cut by rivers. However, in some places, they are drained with water. Thus, intensive melting of relic ice takes place there.

Glacial sediments — till, rocky blocks and stones with glacial striae are found in the whole Dyrstad Valley. They are of different age and stratigraphic location (Reder 1990). Patches of the Pleistocene till occur at two levels and are noted in the erosive cuts of the marine terraces 80 and 65 m. The older till level on a

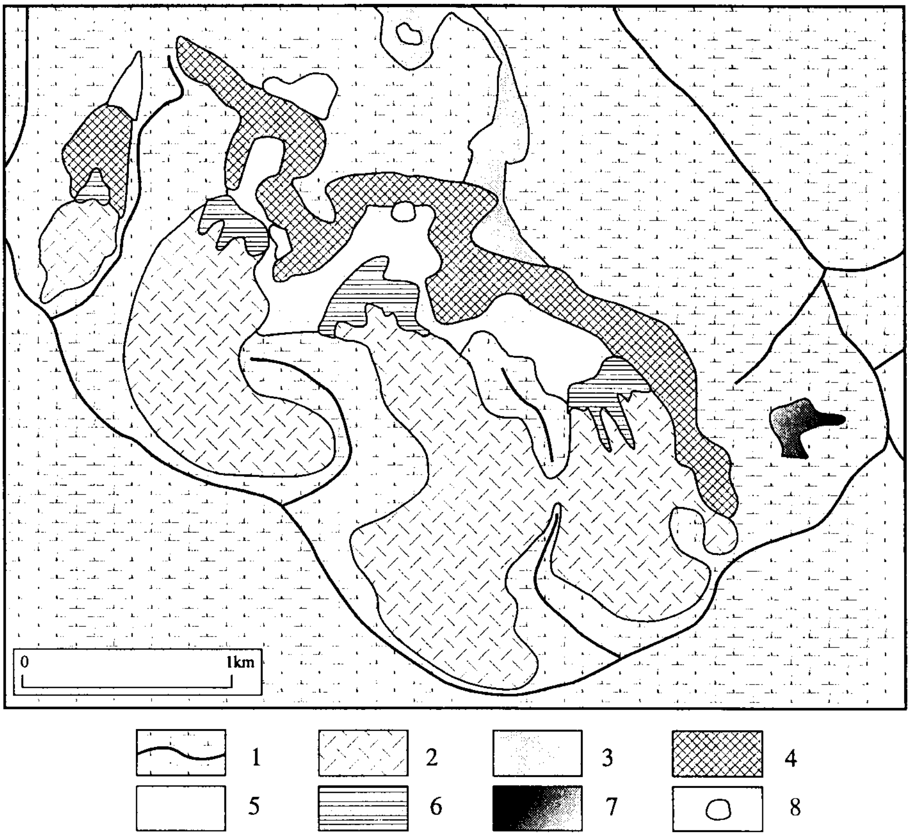

Fig. 7. Marginal zone of the Ringar Glaciers.
1 — slopes and main mountain ridges, 2 — glaciers, 3 — sandur, 4 — frontal and lateral moraines, 5 — fluted moraine, 6 — ablation till, 7 — rock glacier, 8 — lakes.

rocky bed of the terrace 80 m was TL-dated at 155±23 ka BP, the younger from the terrace 65 m was TL-dated at 59±9 ka BP (Reder 1990). The Dyrstad Valley is the only place in the southern Bellsund where the Pleistocene tills were recognized in a forefield of a small retreating glacier. They correspond in age to the glacial advances during the Riss and the Middle Würm. The Late Pleistocene marine sands and gravels, deposited in the fiord bay which was then the lower part of the Dyrstad Valley, are found up to 60 m a.s.l. at present. This value probably reflects the Pleistocene isostatic uplift of the area.

The Holocene glacial morphogenesis was confined to the upper part of the Dyrstad Valley. During the Middle Holocene, the glacier advance formed a frontal moraine on outcrops of the roches moutonneés. This moraine was partly destroyed and covered with younger ice-cored moraine sediments from the Little Ice Age.

The Ringar Glacier was fed from four firn fields during the Pleistocene. It transformed the Dyrstad Valley into an incomplete glacially eroded valley and deposited tills (only partially conserved) on its bottom during the retreat. During the Holocene, advances of this glacier built a frontal moraine along the line of the roches moutonneés. It was separated after the Little Ice Age. At present, there are four independent glaciers with residual tongues in the upper part of the Dyrstad Valley.

The Blomli, Tjørndals and Ringar Glaciers are located entirely in the upper, cool hypsometric level. They were subjected to intensive ablation and quick retreat, the evidence of which are huge canyons, cut into the raised marine terraces. They are 12–15 m deep and not very active, compared with amount of thaw waters. They were formed during quick retreat, intense ablation and large flows during the Early Holocene. Longitudinal profiles of the rivers flowing along the canyons, adapted themselves to the current erosion base.

Rock glaciers

In the southern Bellsund Region the rock glaciers are quite common (Dzierżek and Nitychoruk 1987). They occur mainly at non-glaciated bottoms, at present postglacial cirques as well as at the ridge feet, particularly in the western, wet part of southern Bellsund. Cirque rock glaciers are developed in the Bohlinryggen and Wijkanderberget. Single forms are also found in the upper parts of the Tjørn and the Dyrstad valleys. Their characteristic feature is a location on the westerly-exposed slopes only. Rock glaciers close to the slope are the ridges, 50 m high and outer slope inclination of 35–40°. They are the largest in the Wijkanderberget, Klokkefjellet as well as in the Dunder Valley. Their activity and dynamics in Bellsund is minimal and most forms became stabilized. It is difficult to establish their stratigraphy and chronological position. Only the rock glaciers at foot of the Wijkanderberget, the western part of this moraine passing into a lateral moraine, destroyed and partially covered sediments at foot of the slope. The frontal moraine of the Scott Glacier was formed during the Middle Holocene advance and then, an additional part was created during the Little Ice Age. The rock glacier ridge under the slope was formed earlier. The sediments of this ridge were TL-dated at 7±1 ka BP (Dzierżek and Nitychoruk 1990). Morphology and dating indicate glacier formation at the turn of the Boreal and the Atlantic periods.

A similar date was obtained from a rock glacier situated in the ice cirque of the Tjørn Valley. It is TL-dated at 8±1 ka BP (Dzierżek and Nitychoruk 1990). Cirque rock glaciers formed just after a fast retreat of glaciers in Bellsund at 9–10 ka BP. Rapid climatic warming and high humidity caused transformation of small cirque glaciers into rock glaciers as intensive periglacial processes at upper parts of slopes produced much waste which, covering the ice, controlled the ablation. Glaciers in large cirques and with large masses, carried away the waste and formed small lateral moraines (Ringarbreen IV).

Discussion and results

The paper presents a complex analysis of geomorphological glacial processes in forefields of glaciers in southern Bellsund during the Holocene.

Field investigations and analysis of ancient maps and of air photos indicate retrteat of glaciers at least since the turn of the 19th and 20th centuries. A retreat rate was varied, the greatest during the last 30 years. At present, the average rate of glacier retreat is about 20 m a year. Large glaciers, snouts of which are situated at a sea level or at small altitudes, undergo the most intensive retreat. The glaciers at higher altitudes with their snouts at 200–250 m a.s.l. undergo the smallest changes. Their ablation is small. No distinct changes in geometry of glacier front are observed. They are transformed into cirque glaciers, and residual tongues became rock glaciers. Marginal zones of large glaciers are characterized by a typical zone character of relief and sediments which are common for a frontal deglaciation (Szupryczyński 1963, Pękala 1987, Reder 1993). Frontal retreat in this area is predominant. Only the Recherche Glacier is characterized by mixed (areal and frontal) deglaciation.

Quick deglaciation of Bellsund took place at the turn of the Pleistocene and the Holocene as well as during the Early Holocene between 11 and 9 ka BP (Mangerud *et al.* 1987, 1992, Mangerud and Svendsen 1990). Then, fiords and lower parts of small valleys getting directly into the ocean, became ice-free what was accompanied by isostatic uplift of the coast with amplitude from over 60 m in the ocean coast region to 30 m inside the fiord. Then the glacier retreat was intensive but it might have had a smaller range than at present. Subslope rock glaciers formed in the open forefields and at raised marine terraces. Large delivery of waste from degraded slopes and high humidity caused transformation of the then smallest cirque glaciers into cirque rock ones which have remained almost unchanged.

Climatic changes during the Holocene resulted in common glacial advance at 3.5–2 ka BP (Baranowski 1977, Andrzejewski and Stankowski 1985, Andrzejewski and Błaszkiewicz 1991), creating large frontal moraine ridges. Transgressing glacier tongues destroyed rock glaciers on their way and covered the occupied area with a moraine material. Another common advance of glaciers of the same range took place during the Little Ice Age. It caused development of ice-cored moraines and complete remodelling of former marginal zones, indicated in translocation and deformation of the sediments from the warm "Viking period". The maximum advance took place in the 18th and 19th centuries.

Another warming up of the climate at the end of the 19th and in the 20th centuries caused change of the glacier thermal regime. Due to that, various surges of the Recherche Glacier (twice), the Scott Glacier and probably of the Renard Glacier took place. Modern glaciers adjusted their longitudinal profiles and thermal structure to the changed climatic conditions. Large glaciers at low altitudes and great dynamics of movement and ablation are polythermal. They are characterized by active subglacial and inglacial flows as well as icings in the forefields. Small high-altitude cirque glaciers can be cold glaciers, as indicated by minimum dynamics and lack of subglacial water.

At present, glacier forefields are formed at rapid frontal retreat when moraine ridges are not formed but only the till surfaces, often of a fluted type. They are well preserved on elevations (raised bedrock outcrops) only but in the depression they are quickly washed and cut by meltwaters. Two or three sandur levels in forefields were explained as the effect of isostatic uplift of a coastal zone; it does not seem however true if taking into account the large differences in altitudes of the sandur levels (even 2–3 m). Anyway such high values of isostatic uplift might not have taken place during the last retreat. It may result from quick melting of glaciers and a sandur level, abandoned by ablation waters. Altitude of in— and subglacial tunnel outlets changed, therefore also of proximal slopes of sandur fans.

Intensity and extent of glacier forefield remodelling depends on amount of meltwater outflows. Small, highly located glaciers are a source of poor melt-waters which flow away in covers or in thermokarst crevasses. Sandur fans are not formed there, forefield transformation is insignificant and changes of relief include slow degradation of ice-cored moraines, resulting in frequent changes of their morphology and appearance of cracks and ephemeral thermokarst lakes.

Intensive melting of large glaciers provides enormous amounts of meltwaters. Their forefields are shaped by marginal rivers, icings and morainal lakes of changing configurations. At present, predomination of glaciofluvial erosion is observed. Abandoned beds of marginal rivers at the fluted moraine and sandur fans prevail. Accumulation forms are observed rather on flattened glacier snouts or in dead ice. Intensive erosion by meltwaters destructed even young sandurs and ground moraines. Remaining fragments are usually intensively washed and enriched in coarse material.

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Streszczenie

W opracowaniu podjęto próbę kompleksowej analizy geomorfologicznych procesów glacjalnych zachodzących na przedpolach lodowców południowej części Bellsundu w holocenie. Wyniki badań terenowych oraz analiza archiwalnych materiałów kartograficznych i zdjęć lotniczych wskazują na aktywną recesję lodowców przynajmniej od przełomu XIX i XX wieku (fig. 1, 2). Tempo recesji jest zmienne, największe nasilenie zjawiska obserwuje się w ciągu ostatnich 30 lat. Obecnie średnie tempo recesji lodowców wynosi około 20 m/rok. Najsilniejszej recesji ulegają duże lodowce półpokrywowe i dolinne, których czoła położone są na poziomie morza albo tylko nieco powyżej. Najmniejszą zmienność wykazują lodowce w całości położone w wyższym piętrze hipsometrycznym, z czołami na wysokości 200–250 m n.p.m. (fig. 7). Ablacja tych lodowców jest mała, nie obserwuje się wyraźnych zmian w geometrii czół lodowych. Zauważa się tendencję do przekształcania ich w lodowce cyrkowe, a szczątkowe jęzory przeobrażają się w lodowce kamieniste.

Strefy marginalne dużych lodowców charakteryzują się typową strefowością form rzeźby i osadów, występującą powszechnie przed lodowcami o frontalnym typie deglacjacji (fig. 3, 5, 6). Recesja frontalna w tym rejonie jest dominującą. Jedynie lodowiec Recherche charakteryzuje się deglacjacją arealną, a jego przedpole jest ukształtowane nietypowo, ponieważ dotychczas jęzor tego lodowca schodził bezpośrednio do morza (fig. 4).

Obecnie lodowce dopasowały już swe profile podłużne i termikę do zmienionych warunków klimatycznych — nisko położone duże lodowce, o znacznej dynamice ruchu i ablacji są lodowcami politermalnymi, z aktywnymi przepływami subglacjalnymi i inglacjalnymi oraz lodami typu naledzi funkcjonującymi na przedpolach. Małe, wysoko położone lodowce karowe mogą być przymarzniętymi do podłoża lodowcami zimnymi. Wskazuje na to minimalna dynamika ruchu i trudny do stwierdzenia odpływ subglacjalny.

Przedpola lodowców są kształtowane obecnie w warunkach dosyć szybkiej recesji frontalnej, w wyniku czego nie tworzą się wały moren recesyjnych, a jedynie powierzchnie moreny dennej często wykształconej w formie moreny żłobkowej (pl. 4). Moreny te zachowują się dobrze jedynie na pozytywnych elementach rzeźby (wyniesionych wychodniach skał podłoża), w obniżeniach są szybko przemywane i rozcinane przez wody ablacyjne.

Intensywność i stopień przemodelowania przedpoli lodowców zależy od ilości odpływających wód ablacyjnych. Małe, wysoko położone lodowce są źródłem niewielkiej ilości wód roztopowych, które odpływają w pokrywach, ewentualnie w szczelinach termokrasowych. Nie tworzą się tu stożki sandrowe, stopień przeobrażenia przedpola jest niewielki a zmiany rzeźby ograniczają się do powolnej degradacji wałów lodowo-morenowych, czego efektem są częste zmiany ich morfologii, pojawianie się spękań i efemerycznych jeziorek termokrasowych.

Aktywnie topiące się duże lodowce dostarczają ogromnych ilości wód ablacyjnych. Ich przedpola są kształtowane przez rzeki marginalne, pokrywy naledzi i zbiorniki zastoiskowe o zmiennej konfiguracji (pl. 1–3). Generalnie obserwuje się obecnie dominację procesów erozji fluwioglacjalnej. Dominującym elementem rzeźby przedpoli są porzucone koryta rzek marginalnych, rozcinających pokrywy moreny fluted i stożków sandrowych. Formy akumulacyjne obserwuje się współcześnie raczej na powierzchniach wypłaszczających się jęzorów lodowcowych bądź na martwym lodzie przedpola (pl. 3). Intensywna erozja wód ablacyjnych prowadzi obecnie do niszczenia nawet stosunkowo młodych pokryw sandrowych i osadów moreny dennej. Zachowane fragmenty są zazwyczaj silnie przemyte i wzbogacone w materiał grubszej frakcji.

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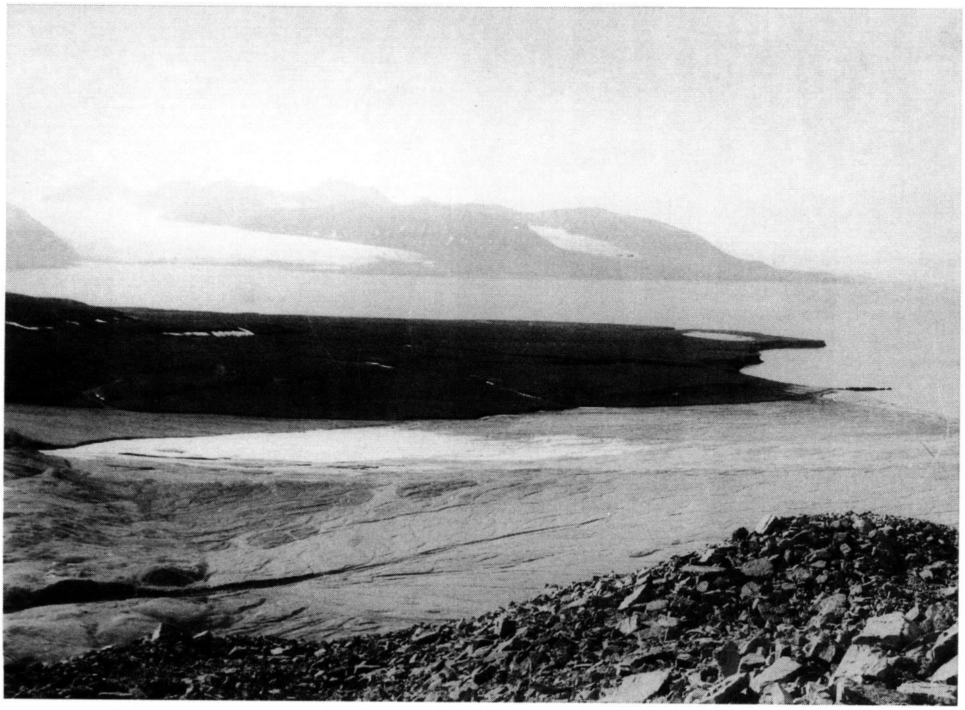
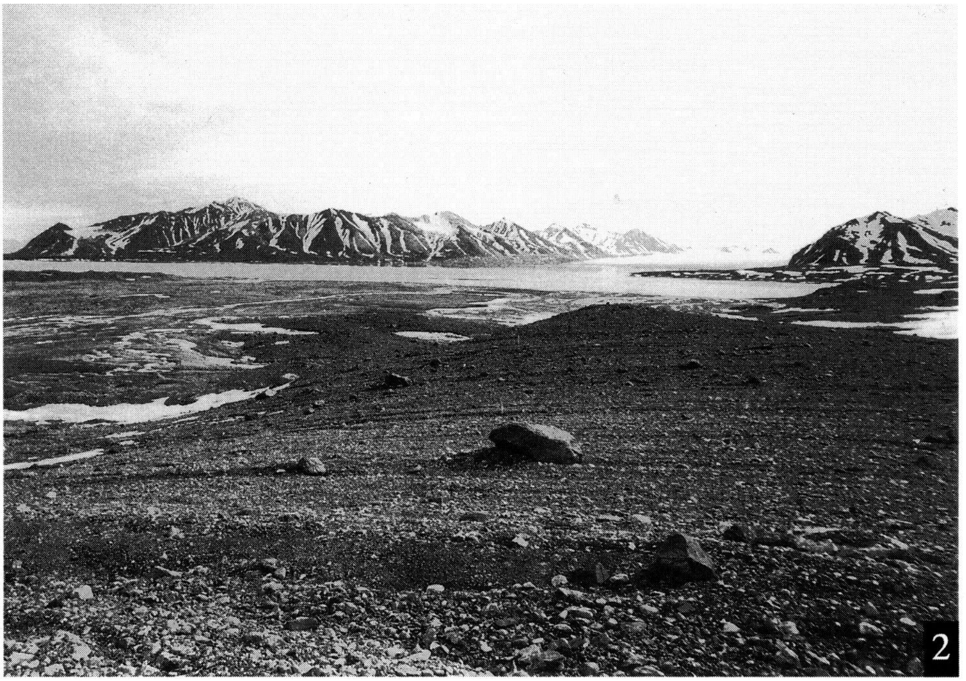


1. Morainal zone of the Antonia Glacier in 1992.
2. Extramorainal sandur in forefield of the Antonia Glacier in 1992.

1. Sandur with inselbergs of a push moraine of the Recherche Glacier in 1992.
2. Kame-type hills at snout of the Recherche Glacier in 1992.

1. Recherche Glacier: inglacial channel and lateral river in 1992.
2. Morainal zone of the Renard Glacier in 1990.



1. Eskers in front of the Renard Glacier in 1990.
2. Fluted moraine in front of the Renard Glacier in 1988.