Jan Rodzik
Grzegorz Gajek
Jan Reder
Piotr Zagórski

5.2. Glacial geomorphology

The NW part of Wedel Jarlsberg Land is less glacierised than other parts of Spitsbergen. The ice masses cover over 50% of the island, whereas the glaciers in the study area cover ca. 20%. The low degree of glacierisation is associated with specific climatic and orographic conditions. The largest valleys are deeply incised inland and their bottoms are located above the equilibrium line altitude, which in western Spitsbergen has recently risen over 400 m a.s.l. (Jania & Hagen 1996; Hagen *et al.* 2003; Sobota 2005). The western and north-western exposure of major valleys in the study area facilitates the migration of snow cover towards the fjords with strong eastern winds, which dominate during the winter season. Valley are also affect by strong foehn winds from the eastern direction as well as summer flux of air masses from the south and south-west (Piasecki & Rodzik 1988; Rodzik 1989, 2006). Southern air masses bring precipitation dominated by rain (Łupikasza 2007). For those reasons local glaciers are small-sized and form between 200-500 m a.s.l.

The largest glacier in the study area is over 20 km long and 3-4 km wide Recherchebreen – together with Bjørnbreen (Orthophotomap, Appendix 1). This is the only glacier in the study area representing the Spitsbergen-type glacier (transitional stage between valley glacier and ice cap). The wide ice field located between 500-700 m a.s.l. links Recherchebreen with large Vestre Torellbreen to the south and with Antoniabreen to the east. Recherchebreen terminates in a small lagoon, linked by an inlet between two inactive outwash plains, with Fagerbukta – the largest bay of Recherchefjorden. One of the Recherchebreen's alimentation basins – Varderyggfonna feeds also Dunderdalsbreen which flows towards Dunderdalen in the west. Other glaciers are represented by smaller land-terminated valley glaciers, cirque glaciers and glacierets (Fig. 5.2.1). Glaciers configuration, shape and the dynamics are to a large extent controlled by bedrock's relief, conditioned by its lithological and tectonic structure.

Structural conditioning of bedrock beneath glaciers

Recherchebreen. The functioning of Recherchebreen is strongly controlled by the structural conditioning of bedrock, referring to the course of main tectonic structures of the Wedel Jarlsberg Land (Fig. 5.2.2). The study area is located within the Tertiary West Spitsbergen fold- and thrust belt, exposing Precaledonian basement of the Hecla Hoek Succesion (Harland 1971).

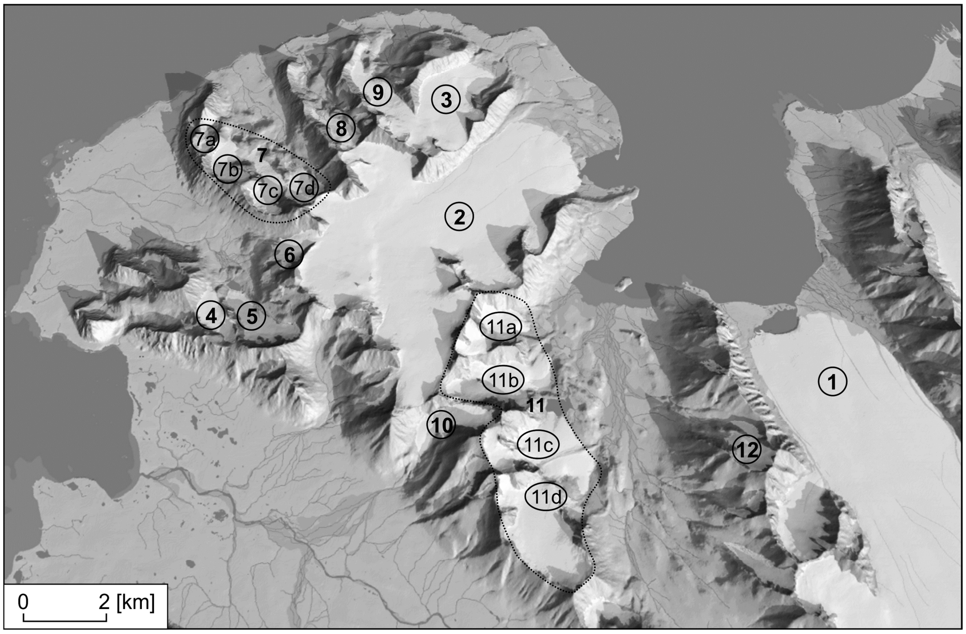

Fig. 5.2.1. Location of glacier of NW part of Wedel Jarlsberg Land (numbering according to Table 5.2.1): 1- Recherchebreen, 2- Renardbreen, 3- Scottbreen, 4- vestre Lognedalsbreen, 5- austre Lognedalsbreen, 6- Gløttfonna, 7- Ringarbreane, 7a- Ringarbreen I, 7b- Ringarbreen II, 7c- Ringarbreen III, 7d- Ringarbreen IV, 8- Tjørndalsbreen, 9- Blomlibreen, 10- Dölterbreen, 11- Crammerbreane, 11a- Crammerbreen I, 11b- Crammerbreen II, 11c- Crammerbreen III, 11d- Crammerbreen IV, 12- Bøckmanbreen. Source of background: Topo Svalbard, Norwegian Polar Institute).

The submeridional orientation of long-lived Recherche strike-slip fault – possible active also in Tertiary (Birkenmajer 2004) determine the development/the elongation of a major landscape elements: Recherchefjorden and Recherchedalen. However the position of the Recherchebreen front is strongly influenced by crooping out to the East three Caledonian thrust-sheets (low-angle reverse faults, Birkenmajer 2004). Russian ground-penetrating radar (GPR) surveys carried out in 1980's (Macheret & Zhuravlev 1982), suggest that the major part of the bottom of Recherchebreen tunnel valley is located below present sea-level (unpublished data after: G. Gajek, D. Puczko, M. Grabiec, J. Jania[10] from 2008 and 2009, P. Zagórski[11] from 2011, 2012) partly confirmed their results during fieldwork 2009. Bathymetric measurements and pilot GPR surveys along the Recherchebreen cross-section showed an asymmetry and glacial

[10] Special grant of the Ministry of Science and Higher Education, Republic of Poland, No. IPY/ 296/2006 –'GLACIODYN' – 'The dynamic response of Arctic glaciers to global warming'

[11] N N 306 703840 'Morphogenetic and morphodynamics conditions of development of the coast of the NW part of Wedel Jarlsberg Land (Spitsbergen) in the late Vistulian and Holocene' (MORCOAST), project leader Ph.D. Piotr Zagórski, duration of the 2011-2013.

remodelling of eastern part of the valley. One of the detected features was an over 100 m deep overdeepening formed by intensive glacial egzaration in tectonically cracked rocks along the Kongsfjorden-Recherchebreen-Hansbreen dislocation. In the last 20 years the overdeepening predisposed the eastern part of Recherchebreen ice-cliff to intensified calving and retreat.

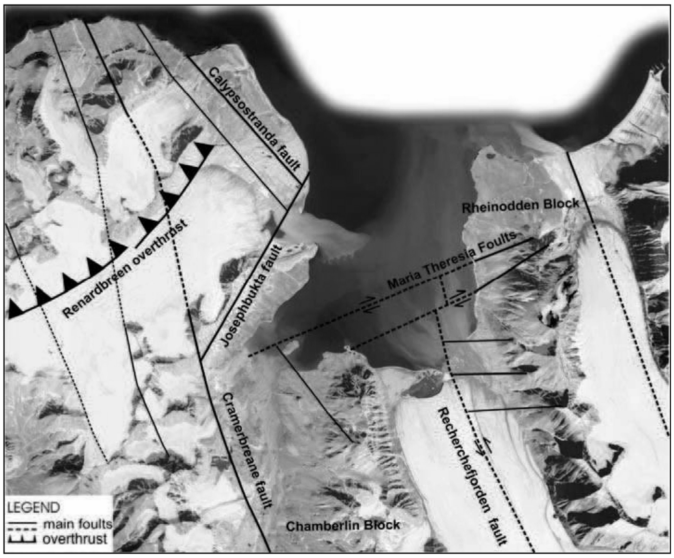

Fig. 5.2.2. Simplified tectonic map of NW part Wedel Jarlsberg Land (after: Birkenmajer 2004, 2006; Gajek *et al.* 2010). Base: Ortophotomap (Zagórski 2005).

Renardbreen. They is the largest glacier laying entirely within the study area (Fig. 5.2.1). The glacier's alimentation basin is opened to the NE and surrounded from other directions by mountain ridges: from the SE by Becketoppane, from the S, W and NW by series of peaks (over 700 m a.s.l.), including Storgubben massif (831 m a.s.l). Alimentation basin is divided into four asymmetrical lateral sub-basins and two small sub-basins located at the base of Activekammen and feeding the glacier tongue (Appendix 1, Photo 5.2.1.).

The GPR surveys showed that the bottom of sub-basin (A) is inclined at an angle of 20 degrees towards the main tunnel valley (Photo 5.2.1). Survey did not detect any clearly visible bedrock knickpoint. The inclination of basin mirrors the dip of rock layers on the lowest structural level located at 120-150 m a.s.l). The bedrock in this area bears no traces of glacial overdeepening. Second sub-basin (B) is characterised by typical cirque features such as overdeepened bottom and moutonised sill. Sub-basin is formed on the structural level of 250-300 m a.s.l. overdeepened ca. 80 m. The sub-basin is up to 1100 m wide and the direction of its axis is related to the Crammerbreane dislocation zone (Birkenmajer 2004).

The largest sub-basin (C) is characterised by longitudinal extent and covers almost entire southern part of the main basin. The length of the sub-basin is ca. 4 km and the widest part reaches ca. 2 km. Two cirques divided by sill with steep distal slope were eroded in the bedrock. The higher cirque (southern) is related to the denudational-structural level of 400-500 m a.s.l., whereas the lower and larger cirque is located between 130 and 190 m a.s.l and corresponds with the structural level of 200-250 m a.s.l. According to Birkenmajer (2004) the bottom of the cirque is eroded in outcrops of upper diamictites from Kapp Lyell Formation. This part of the sub-basin is limited to the east and west by strike-slip faults. Due to the bedrock relief the glacier reaches the largest thickness in this area (340 m).

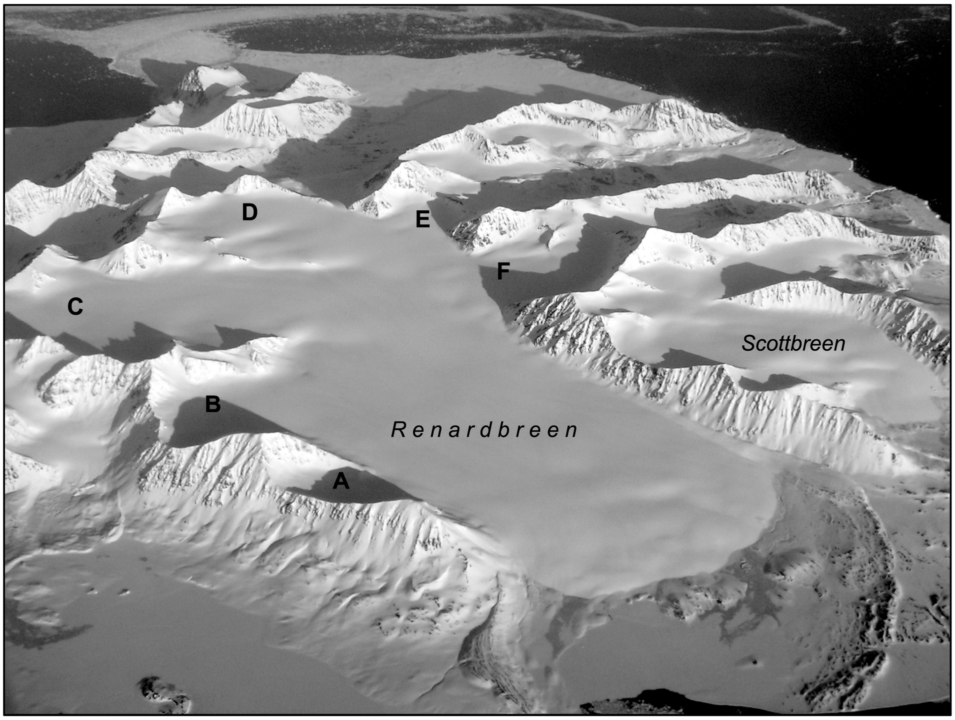

Photo 5.2.1. Arrangement of alimentation basin of Renardbreen (Photo M. Grabiec 2005).

The second largest sub-basin (D) is located in the western part of the main basin and constitutes the extension of Renardbreen tongue (Photo 5.2.1). This is the only sub-basin separated from the glacier by clearly visible bedrock threshold, marked by the series of nunataks. The threshold was formed in resistant tillits from Kapp Lyell Formation. In the northern part of the sub-basin the threshold dips under the glacier and gently transforms into the central zone of the overdeepening eroded in Deilegga Group shales.

Main tunnel valley of Renardbreen is bordered by Bohlinryggen and Activekammen ridges and is up to 6 km long and 2.6 km wide. The structural conditioning determines its SW-NE direction. The direction of valley coincide with the strike of main rock layers of Renardbreen Block and the limit of Lognedalen-Renardbreen Overthrust (Fig. 5.2.2).

Scottbreen. They fills the valley between three massifs: Wijkanderberget, Bohlinryggen i Emil Nilssonfjellet (Appendix 1, Photo 5.2.2). The valley has developed in lithologically uniform zone of Kapp Lyell diamictites. Scottbreen shares the firn field with small Blomlibreen. The accumulation part of the glacier is directed to the N and the ablation part is exposed to the NE. Glacier valley has a subsequent character. The tunnel valley eroded by Scottbreen is cut by Crammerbreane strikes-slip fault (Birkenmajer 2004). The field of moutonised structural monadnocks is associated with the fault. The maximum thickness of Scottbreen (160 m) has been documented over the central overdeepening located in the northern part of tunnel valley, approximately 800 m to the west from the Crammerbreane strikes-slip fault. The development of overdeepening is related to the asymmetry of glacial valley, caused by the northern dip of Kapp Lyell diamictites. Structural conditionings determine the configuration of glacier alimentation basins. The bottom of three cirques where alimentation basis are located, evolved in a structural level of 250-300 m a.s.l.

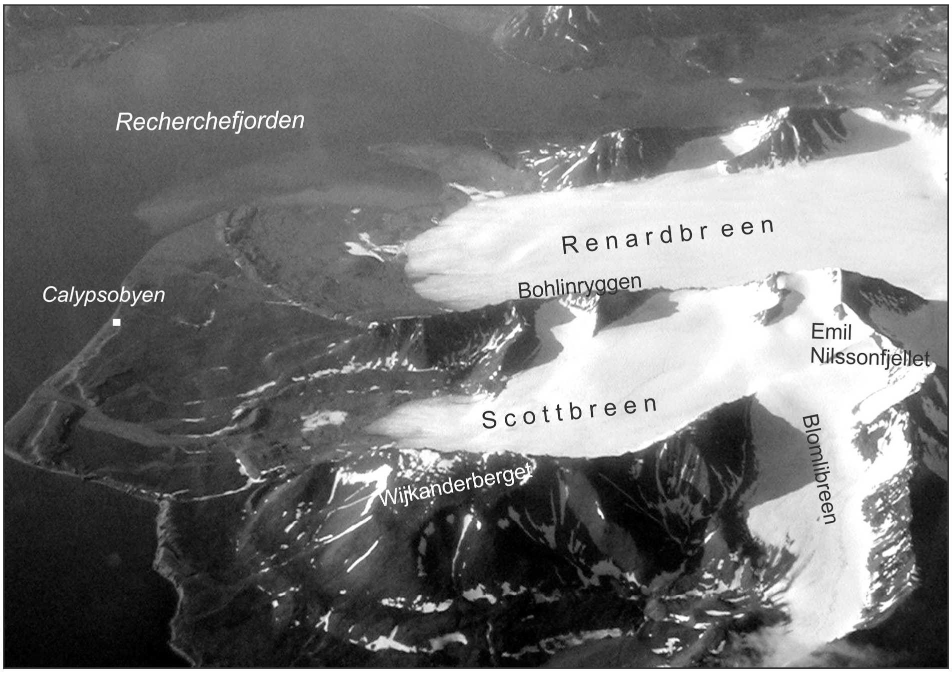

Photo 5.2.2. View of the forefield of Scottbreen and Renardbreen (Photo M. Górska 2003).

Small glaciers (Blomlibreen, Tjørndalsbreen, Ringarbreane, vestre Lognedalsbreen, austre Lognedalsbreen) are located in upper sections of partly glacierised valleys of the NW part of Wedel Jarlsberg Land: Blomlidalen, Tjørndalen and Dyrstaddalen (Appendix 1, Fig. 5.2.1). Glaciers are characterised by northern exposure and their position can be linked with the structural level located at 400-500 m a.s.l. The high level of division of Ringarbreane is associated with a system of longitudinal tectonic faults formed during the Tertiary.

The western part of Chamberlindalen is covered by system of Crammerbreane (Fig. 5.2.1). Glaciers are characterised by eastern exposure and form a group of cirque glaciers with accumulation fields developed in structural levels of 400-500 m a.s.l. and 250-300 m a.s.l. The surface of glaciers perfectly mirrors the structural conditioning of glacier basins. Relatively flat accumulation fields refer to denudational-structural plains formed in Kapp Lyell diamictics, whereas series of crevasses and seracs are linked with the structural threshold associated with the Crammerbreane dislocation zone (Birkenmajer 2004).

Glacier dynamics

The evolution of glaciers in the Late Pleistocene

During the Pleistocene glaciations the Barents Sea and Svalbard Archipelago was covered by an ice-sheet – Svalbard-Barents Sea Ice Sheet (SBSIS). The western coast of Spitsbergen constituted the border of a large ice sheet (Lindner & Marks 1993a; Ingólfsson & Landvik 2013) (see: Fig. 2.2.11). During the period of maximum extent of SBSIS (46-42 ka BP) the margin of an ice-sheet was located several kilometres off the western coast of Spitsbergen (Baranowski 1977a; Mangerud *et al.* 1992, 1998; Landvik *et al.* 1992). During the Late Weichselian the major fjord systems of Spitsbergen (Hornsund, Bellsund and Isfjorden) were still fully glaciated but the margin of an ice-sheet did not extend beyond the western coast (Landvik *et al.* 1987; Kłysz & Lindner 1981; Niewiarowski *et al.* 1993; Musiał *et al.* 1993).

The Weichselian glaciation is well-documented in the NW part of Wedel Jarlsberg Land. Weichselian deposits were dated in Skilvika profile (55±8 ka and 26 ka BP) and the bottom of Dyrstaddalen (59±9 ka BP) (Pękala & Repelewska-Pękalowa 1990; Reder 1990) (see: Figs. 2.2.2 and 2.2.10). The information on the existence and extent of an Late Weichselian glaciation can be extracted from several features including glacial trimlines along valley slopes and numerous allochtonic erratics scattered over uplifted marine terraces outside the belt of moraines formed during the Holocene glacial advances. Undoubtedly, the heavily-weathered morainic deposits covered by slope sediments (talus fans, nival moraines and rock glaciers) are of Late Weichselian age (Dzierżek & Nitychoruk 1990).

In the Recherchedalen the Late Weichselian glaciation is documented by trimline eroded in the eastern slopes of Observatoriefjellet and western slopes of Martinfjella at 100-150 m a.s.l. The older trimlines, poorly preserved and covered by heavily weathered morainic deposits, are visible at 300 m a.s.l. and may indicate the vertical fluctuations of an ice stream that drained the large ice cap located at the border of Torell and Wedel Jarlsberg Lands (Zagórski 2007d).

The petrology of erratic boulders found on uplifted marine terraces in Calypsostranda region differs from local bedrock (Lindner & Marks 1993a). Their petrographic characteristics suggest that they might be transported by an ice stream from the central part of Wedel Jarlsberg Land. Late Weichselian erratics differ also from those found in the Holocene moraines. They are smaller, more rounded (suggesting longer transport) and have heavily weathered surfaces (suggesting an old age). The spatial distribution of erratics, in relation to the location of bedrock outcrops they have been eroded from, allows to reconstruct the main directions of ice-streams flow and ice sheet advances.

The Weichselian glaciers were much larger than present-day glaciers and in the southern part of Bellsund flowed from the local ice cap located in central Wedel Jarlsberg towards the NW (Lindner & Marks 1993a; Ingólfsson & Landvik 2013). The extensive Recherchebreen was pushing Renardbreen towards NW along the Bohlinryggen and further, continued to push it towards later on towards the Skilvika (Zagórski 2007d). The formation of a depression in northern Calypsostranda is associated with the concentration of the erosive action of Renardbreen on a narrow zone that linked glacier basin with Skilvika (Zagórski 2007c) (see: Fig. 5.5.2A).

At the same time the Recherchebreen was pushed by a large ice-stream flowing from the Van Keulenfjorden towards the west. The medial moraine that was formed between Recherchebreen and Renardbreen ran parallel to the contemporary dead-cliff of Calypsostranda. The zone of convergence of two ice streams is marked by accumulations of erratics found across Calypsostranda (Zagórski 2007c).

The Late Weichselian glaciers that filled Recherchefjorden, Van Keulenfjorden and Van Mijenfjorden merged together in the widening of Bellsund and formed a common ice-cliff across the entrance to the bay (Mangerud *et al.* 1987).

In relation to modern sea-level the Late Weichselian glacio-isostatic loading of land reached ca. 60 m. Coastal plains in southern part of Bellsund, Lyellstranda, Dyrstadflya and Lognedalsflya as well as bottoms of Lognedalen and Dunderdalen were submerged. Glaciers filling Blomlidalen, Tjørndalen, Dyrstaddalen were marine-terminated (e.g. Pękala &Reder 1989; Szczęsny 1987). Their ice-cliffs were located at the entrances to present-day valleys and did not reach zone of raised marine terraces (Appendix 2, Geomorphological map).

<u>The stages of Late Weichselian/Holocene deglaciation</u>

The warming of climate associated with the end of the Weichselian and beginning of the Holocene resulted in rapid reduction of glacier thickness and extent, particularly in fjords. Around 12-11 ka BP the glaciation covered most of the fjord system: Van Mijenfjorden (glaciated up to the Akseløya), Van Keulenfjorden (glaciated up to Eholmen) and Recherchefjorden (Mangerud *et al.* 1987; Nitychoruk & Dzierżek 1994). According to Mangerud *et al.* (1998) the Van Mijenfjorden was deglaciated between 11.2 and 9.6 ka BP. It is highly probable that deglaciation of Recherchefjorden and Van Keulenfjorden occurred at the same period.

Glacioisostatic loading of coastal zone, in relation to present sea-level, reached 50-60 m. The Late Weichselian marine limit in southern Bellsund oscillated around

60 m and in Lognedalen is located at 55 m, in Dyrstaddalen at 58 m Tjørndalen 59 m and between Blomlidalen and Scottbreen valley 60 m (Salvigsen *et al.* 1991; Zagórski 2002).

Rapid recession of glaciers triggered the intensive glacioisostatic uplift. The marine terraces that are currently at 25 m a.s.l were formed around 10-9 ka BP (Troitsky *et al.* 1979; Salvigsen 1979). Radiocarbon ^{14}C dating of shell fragments (8,150±70 BP and 9,930±70 BP) found in raised terrace at 20-35 m a.s.l. around Renardodden suggests the Early Holocene age of terrace development (Troitsky *et al.* 1979). Similar dates were obtained by Salvigsen (1977), who dated the driftwood and shell fragments from marine terraces at 29 m a.s.l. (10,310±200 BP) and at 10 m a.s.l (9,400±120 BP). Those dates correspond with dates obtained by Landvik *et al.* (1987) in the northern part of Bellsund.

Extensive valleys of southern Bellsund, currently devoid of larger glaciers, are characterised by one common feature – a flat bottom located at a low elevation. The part of Dunderdalen located almost 15 km from the coast lays below 50 m a.s.l. and its mean gradient slightly exceeds 3‰. The bottoms of Chamberlindalen and Lognedalen are only slightly steeper, but also located below 50 m a.s.l. At the end of the Weichselian all valleys were submerged and formed bays with marine-terminated glaciers.

The warming of climate at the turn of the Pleistocene and the Holocene caused the reduction of sea-ice cover and influx of warm Atlantic waters to the western coast of Spitsbergen (Svendsen & Mangerud 1997; Bauch *et al.* 2001; Lubinski *et al.* 2001). The influx of warm Atlantic waters intensified the melting and retreat of marine-terminated glaciers. Flat glaciers located at low altitude decayed around 10-9 ka BP. During the Early Holocene glacier valleys were transformed into shallow bays filled with marine sediments. For instance, in Dunderdalen the marine sediments of Early Holocene age are found up to 6 km from the present-day shoreline.

<u>Glaciers changes during the Holocene</u>

It is still difficult to assess to what extent the Holocene climate changes influcenced the dynamics of glaciers in this part of the Arctic. According to Baranowski (1977a) the Holocene development of glaciers occurred around 3,500-2,000 years BP, during the Subboreal phase, corresponding with the beginning of Neoglacial (Denton & Karlén 1973). The investigations of subglacial relict vegetation and soil microbes found underneath of Longyearbreen led Humlum *et al.* (2005) to the conclusion that between 5,000-1,100 years BP the areas of central Spitsbergen located below 400-450 m a.s.l. were free of glaciers. The study of lake sediments accumulated in Linnévatnet, at the entrance to the Isfjorden, carried out by Svendsen & Mangerud (1997) suggests that the glaciers in western part of Spitsbergen reached their Holocene maximum extent during the Little Ice Age (LIA). Mangerud & Landvik (2007) claimed that the LIA frontal moraines of Scottbreen mark the glacier maximum extent since the late Allerød.

The last major advance of glaciers in the NW part of Wedel Jarlsberg Land occurred at the end of LIA and is recorded in the landscape in the form of belt of ice-cored moraines. The advance of Recherchebreen, Renardbreen and Scottbreen at the end of the 19th century is a result of glacier surge (Hagen *et al.* 1993; Liestøl 1993). One of the evidences for the occurrence of glacier surge are pushed and glaciotectonised frontal moraines of Renardbreen and Scottbreen. Archival materials collected during that period also suggest the occurrence of a surge (e.g. Bertrand 1852; Hamberg 1932).

The change of thermal and humidity conditions progressing since at least 130 years led to the significant transformation of local glaciers. The negative mass balance of local glaciers results in their intensified retreat that has been a subject of study of the UMCS scientific expeditions. The Tables 5.3.1. and 5.3.2 present data on area changes of and the rate of recession of glaciers for the periods for which they were available archival materials. The method of determining the recession value was used by Zagórski *et al.* (2008c).

The main focus was paid on description of post-LIA changes of Scottbreen and Renardbreen (e.g. Piasecki 1988; Merta et al. 1990; Zagórski & Bartoszewski 2004; Gajek & Reder 2008; Gajek et al. 2008, 2010; Zagórski et al. 2008bc, 2012). Since the end of LIA Renardbreen experiences frontal recession with intensified mechanical ablation of marine-terminated part of the front and the areal deglaciation of land-terminated part. The lower part of the Renardbreen valley that is currently not glacier covered is made up of a vast internal zone delimited by a bow of pushed moraine (ice-cored moraine ridges). During the period from the late 19th century (LIA) until as late as the 1940s, the surface of the glacier was higher and the catchment also covered an area located outside the moraine ridges. From the 1936 to 2011, the Renardbreen has been observed to continually retreat at a rate of 11.5 $m \cdot a^{-1}$, in the years 1936-1960 at 12 $m \cdot a^{-1}$ (it was the result of deglaciation of Josephbukta) and in the years 1990-2009 at 10.8 $m \cdot a^{-1}$ (Fig. 5.2.3, Photo 5.3.3B). The contact with waters of the fjord in the region of Josephbukta played an important role in the retreat of the glacier, as its front retreated mainly, as a consequence of thermo-abrasion. In the 1990s the front of the Renardbreen lost contact with the waters of the fjord and, at the same time, the ability to influence the coastal zone (Gajek & Reder 2008; Zagórski *et al.* 2008b, 2012). In the last years (2007-2011) have seen the high variability of recession of Renardbreen (Fig. 5.2.3). Second glacier the Scottbreen today is at a stage of recession to (Zagórski *et al.* 2008c, 2012). From the end of the LIA to 2011 the average retreat rate of the glacier's front reached 7.2 $m \cdot a^{-1}$ (and in the years 1990-2009 as much as 17 $m \cdot a^{-1}$) (Fig. 5.2.3, Photo 5.2.3A). In recent years there is also similar variability Scottbreen recession as in the case Renardbreen. In both cases, is made clear exemption recession in the years 2007-2008 (due to high spring snowfall was delay summer glacier ablation) and the acceleration in the years 2010-2011 (Fig. 5.2.3).

Table 5.2.1. The glacier areas of in selected years on the basis historical data and GPS measurements.

No.	Glacier	Area [km^2]				Decrease of area [%]**
		Max LIA	1936	1990	2011	
1	Recherchebreen***	-	-	-	-	-
2	Renardbreen	38.0	36.4	31.1	30.4*	16.5
3	Scottbreen	6.16	5.98	5.32	4.61*	22.9
4	vestre Lognedalsbreen	-	0.95	0.82	0.79	16.8
5	austre Lognedalsbreen	-	1.58	1.27	1.17	25.9
6	Gløttfonna	-	0.96	0.73	0.43	55.2
7	Ringarbreane	-	4.26	3.33	2.53	40.6
7a	Ringarbreen I	-	0.30	0.22	0.19	40.6
7b	Ringarbreen II	-	1.23	0.94	0.81	36.7
7c	Ringarbreen III	-	1.60	1.28	0.95	34.1
7d	Ringarbreen IV	-	1.13	0.89	0.58	40.6
8	Tjørndalsbreen	-	1.26	0.75	0.60	48.7
9	Blomlibreen	-	2.57	2.28	1.96	52.4
10	Dölterbreen	-	1.30	1.02	0.84	23.7
11	Crammerbreane	-	11.55	8.97	8.16	35.4
11a	Crammerbreen I	-	1.83	1.33	1.23	29.4
11b	Crammerbreen II	-	2.77	2.13	1.85	32.8
11c	Crammerbreen III	-	2.25	1.79	1.66	33.2
11d	Crammerbreen IV	-	4.70	3.72	3.42	26.2
12	Bøckmanbreen	-	1.19	0.71	0.49	27.2

*GPS measurement; ** the area decrease in realtion to 1936, *** the value of area is difficult to determine.
Sourse of data:
- maximum of LIA– geomorphological mapping;
- 1936 – B11 Van Keulenfjorden (1952);
- 1990 – Orthophotomap (Zagórski 2005);
- 2011 – aerial photographs taken from website: Topo Svalbard, Norwegian Polar Institute

Table 5.2.2. The values of the glacier recession in period 1936-2011.

Glacier	Mean recession of glacier front [m]	Maximum recesion of glacier front [m]	Annual recesion [m·a^{-1}]
Renardbreen	865	1,920	11.5
Scottbreen	540	1,340	7.2
vestre Lognedalsbreen	290	470	3.9
austre Lognedalsbreen	405	580	5.4
Gløttfonna	335	665	4.4
Ringarbreen I	215	290	2.8
Ringarbreen II	295	620	3.9
Ringarbreen III	385	665	5.1
Ringarbreen IV	315	620	4.2
Tjørndalsbreen	320	735	4.3
Blomlibreen	290	660	3.9
Dölterbreen	255	740	3.4
Crammerbreen I	355	730	4.7
Crammerbreen II	340	900	4.5
Crammerbreen III	290	725	3.9
Crammerbreen IV	515	605	6.8
Bøckmanbreen	415	1,240	5.5

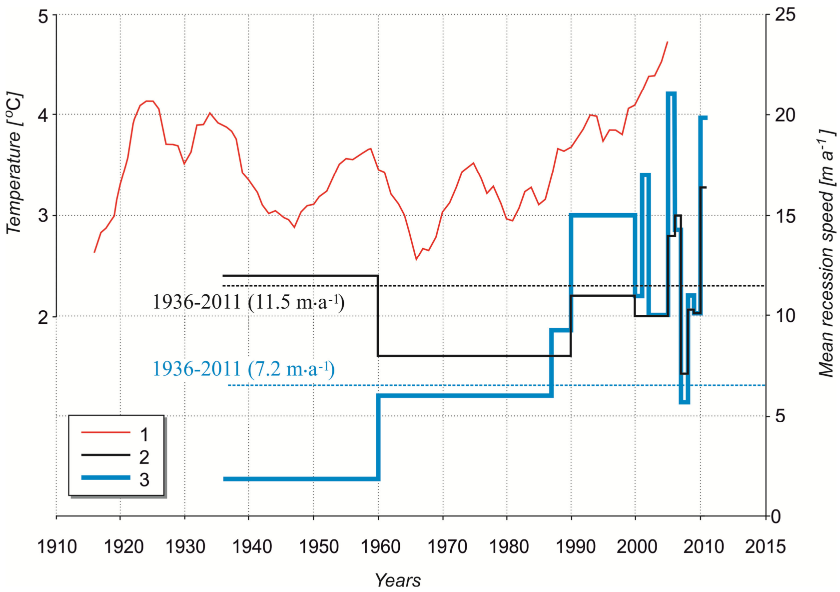

Fig. 5.2.3. Average recession rates of Renardbreen and Scottbreen for following periods on the background of avarage air temperature in periods June-September on Svalbard-Lufthavn station: 1- course of temperature equalize to 5-point moving average, 2- average recession rates of Renardbreen front for following periods, 3- average recession rates of Scottbreen front for following periods (after: Zagórski *et al.* 2008ab, modified and supplemented).

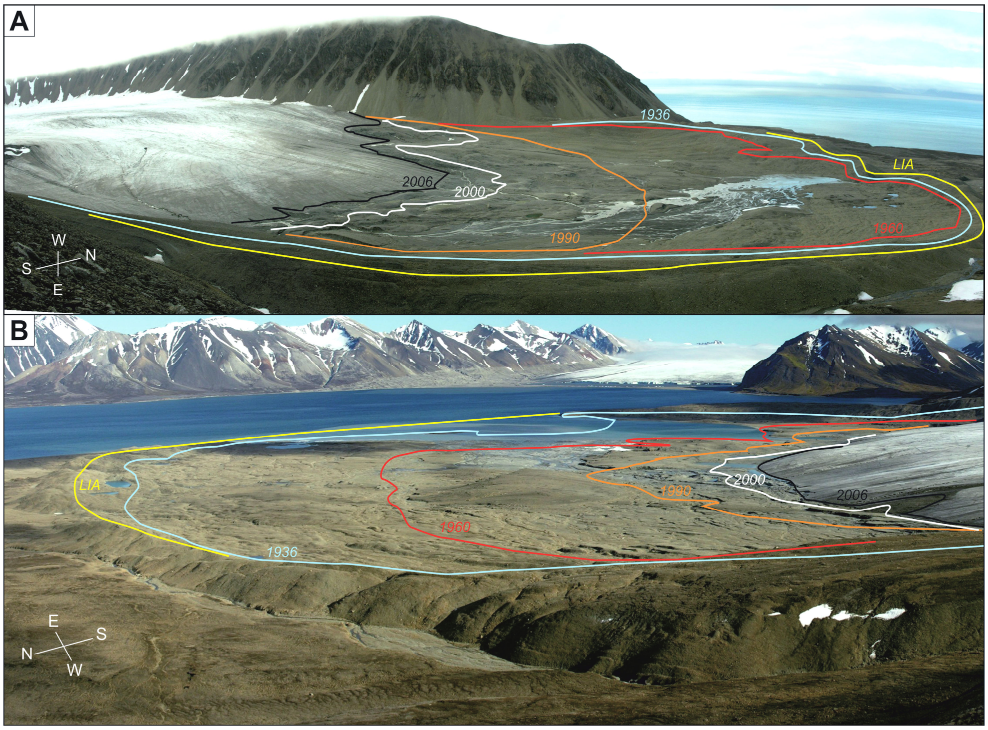

Photo 5.2.3. Changes of the front extent of: A- Scottbreen, B- Renardbreen (Photos P. Zagórski 2006).

The recession of Recherchebreen differs from other glaciers in the study area. The post-LIA glacier retreat has been interrupted by episodes of advance resulting from glacier surge (Liestøl 1969; Koryakin 1975; Jania 1988; Hagen *et al.* 1993; Jiskoot *et al.* 2000). In the scale of the enitre Spitsbergen the Recherchebreen is an exceptional glacier with regard to the first documented active phase of glacier surge and associated processes within the glacier front (Bertrand 1852). According the quantitative model of glacier evolution on Spitsbergen (Jania 1993), every new phase of surge is smaller than previous one as a result of decreasing of ice mass in the glacier system. The evolution of Recherchebreen can be treated as a model example of such a process. For instance, the 1838 surge of Recherchebreen was over 2 km longer than one observed in 1945. The retreat of Recherchebreen is strongly controlled by the topography of glacier forefields. The slower rate of retreat (ca. 8 m a^{-1}) observed between 1898 and 1918 was associated with passing through resistant bedrock outcrops. In years 1895-1898, before reaching the zone of structurally uplifted bedrock, the glacier was retreating with a very rapid rate 127 m a^{-1} (Hoel & Werenskiold 1962; Gajek *et al.* 2010). Further melting resulted in retreating of the western part of glacier front to deeper water in the contemporary lagoon (position in 1936) in an average rate of ca. 36 $m \cdot a^{-1}$. The next surge event in the forties resulted probably in overpassing of the threshold. An advance of the front by an average 10 $m \cdot a^{-1}$ was noted between 1936 and 1960. In 1960 glacier front was situated on land. Subsequent retreat into relatively deep waters of the Recherchebreen lagoon continuing up to nowadays with increasing intensity (13 $m \cdot a^{-1}$ in 1960-1990 and 50 $m \cdot a^{-1}$ in 1990-2008). Glacier flow velocity in the quiescent phase was less than 15 $m \cdot a^{-1}$ (Błaszczyk *et al.* 2009). For this reason, despite the contact with the sea water in the lagoon, calving is very rare and mass loss due to calving is very low in comparison to superficial ablation (Zagórski *et al.* 2012) (see: Fig. 5.5.9).

The post-LIA retreat rates of small glaciers was slower and ranged between 5.1 and 5.9 $m \cdot a^{-1}$. The progressive recession and lowering of glacier surface , particularly in their ablation zones, led in case of some glaciers to the inversion of ablation. For instance the thick debris layer that covers the ablation zone of Blomlibreen effectively protects the glacier surface from intensive melting and slows down the rate of retreat.

The changes in glacier extent have entailed significant decrease of glacier area. Since the end of the LIA Renardbreen has lost 8 km^2 of area (1/5 of the total area). Neighbouring Scottbreen lost 1.5 km^2 (1/4 of the area at the end of the LIA). The largest area loss was observed in case of small cirque glaciers e.g. between 1936-2011 Bøckmanbreen, Gløttfonna and Tjørndalsbreen lost over 50% of the area.

Apart from the loss of area local glaciers have experienced significant changes in thickness. Between 1990-2011 the surface of Renardbreen has been lowering ca. 1.6 $m \cdot a^{-1}$ in the frontal part and ca. 0.04 m a^{-1} in the upper part of the glacier located above 450 m a.s.l. The thinning of glaciers accelerates the upward migration of equilibrium line altitude (ELA). At the turn of the 1980's and 1990's the ELA was located at 300 m a.s.l, whereas currently it is estimated to ca. 460 m a.s.l. The systematic increase of ELA is a characteristic for all glaciers located along the western coast of Spitsbergen (Jania & Hagen 1996; Hagen *et al.* 2003; Sobota 2005). Due to the small contribution of accumulation in the form of firn, at the concurrent domination of superimposed ice, the precise estimation of ELA for glaciers of the NW part of Wedel Jarlsberg Land is difficult (Piasecki 1988). Nevertheless, besides Recherchebreen, the ablation zones of local glacier are larger than their accumulation zones.

The geomorphology of glacier forefields

The present-day relief of glacier forefields in superimposed on the Pleistocene egzaration landforms and raised marine terraces. In the zones of glacial and glaciofluvial relief of land-terminated glaciers the characteristic zonal configuration of landforms is observed, often modified by local factors including: irregular overdeepenings, blocking of glacier front by rocky sills or monadnocks (e.g. Pękala 1987; Szczęsny 1987; Pękala & Reder 1989; Reder 1991, 1993, 1996; Zagórski 2002; Reder & Zagórski 2007ab; Zagórski 2007acd; Zagórski *et al.* 2012). The hypsometrically dominant landforms are arches of latero-frontal moraines. Two generations of frontal moraines have been determined. The younger (inner) zone in the form of ice-cored moraines and the older (outer) zone in the form of pushed moraines (Pękala & Repelewska-Pękalowa 1990). The relief of pushed moraines is undulating with microfolds observed on the surface. Distal slopes are steep and stabilised, whereas proximal slopes are covered by thick layer of ice-cored morainic deposits. The zone of ice-cored moraines the relief is characterised by big denivelation, with depressions filled by thermokarstic lakes and slopes remodelled by solifluction. The hypsometric dominance of ice-cored moraines is often smoothened by adherent kame terraces. The frontal moraines are incised by proglacial rivers forming deep gorges (Reder 1996; Reder & Zagórski 2007ab).

Icings and blocks of dead-ice covered by outwash plain deposits are a common feature found in the proglacial zone (see: Fig. 5.5.5C). Between retreating glaciers and LIA moraines the zones of ground moraines, ablation moraines and inner outwash plains were formed. Fragments of ground moraines – often as fluted moraine – are particularly well-preserved on the elements of older relief such as roches moutonnées. The hollows between roches moutonnées are shaped by proglacial rivers which form and incised extensive outwash plains (Merta 1988c, 1989; Reder 1996; Reder & Zagórski 2007a).

Outside the belt of frontal moraines lays the zone of outer outwash plains (sandurs). Due to the differences in glacier sizes and amount of released meltwaters the outwash plains vary in size and type of formation. Large, extramarginal outwash plains are located in front of Renardbreen. Outwash plains formed in the forefields of smaller land-terminated glaciers are small and their configuration is controlled by the topography of valley bottoms (Reder 1996; Zagórski 2011).

Recherchebreen

The front of Recherchebreen is flat and typical for glaciers in recession. On the surface of the glacier tongue, four middle moraines are seen. They are a few kilometres long and a few tens of metres wide straps of mineral material melted out on the surface of the ice. It comes from nunataks that diverse ice streams from boosting glaciers (Photo 5.2.4A, Appendix 1). The longest and best seen middle moraine is being made beneth the Kvartsittkammen Nunatak, in the place of junction of ice masses of Bjørn-breen and Tverrbreen. The middle moraines of Recherchebreen show wavy course that proves the surge in 1945 (Hagen *et al.* 1993; Liestøl 1993). The glacier front then reached one of the structural edges along the line Observatoriefjellet-Okernuten.

The marginal zone of Recherchebreen is formed in the typical way for glaciers that ends in the sea (Jania 1993). Nowadays the glacier is showing a mixed type of deglaciation. In western and central part of the head there is a glacial cliff about 20 m high, where ablation by calving is dominant. From Recherchebreen ablation water gets out through subglacial tunnels. The highest outlets, which were still in 1990s within the contact zone of glacier and side moraines, accumulated outwash plains (see: Photo 5.5.10ABC). During the time of intensive ablation and higher flows, these waters washed and also slit the outwash plain surface system of beds. Currently, only the out-side outwash plain is under the influence of sea processes, because the whole outflow of the Recherchebreen waters is hapenning towards the inside lagoon. In 1990 the sur-face of the lagoon was only 0.56 km^2 and within next 19 years it grew to 3.31 km^2.

In eastern part the mild slope of the front created conditions for deposition of outwash plain material not only on the forefield, but also partly on the surface of the glacier tongue. This material is the cover the slows down ablation process of some are-as of the glacier. As a result a lot of fluvioglacial sediments is now depositioned on dead ice. The waters that flow out from the glacier cut into mineral mineral sediments and dead ice. As a result in front of the head of the glacier it was created the whole system of ice-cored kame hills, that are divided by system of melt-water channels (Photo 5.2.4B).

Vast, inactive outwash plain in the western part of the forefield forms one level, only in contact with latteral moraine there are remnants of older and higher by 1.5 m level. The lower part is elevated 0.5 m above the contemporary outwash plain and has larger area and triangular shape (Photo 5.2.4B). On the surface of that cone, in its southern part, there are a lot of crack forms. Distal part of outwash plain in the shore zone is prograding by the system of beach ridges (Reder 1996; Zagórski *et al.* 2012).

The Recherchebreen tongue is accompanied by laterral moraines on almost the whole of its length. The right one stretches along eastern side of Recherchefjorden, up to Lægerneset – 4.5 km to the north comparing to today localisation of the glacier front (Photo 5.2.4C). Whereas the left moraine stretches along western shore to Rubypynten (3 km to the north). It is built from glacio-marine sediments, its inner structures show some traces of glaciotectonics, that allows to treat that moraine as a pushed one.

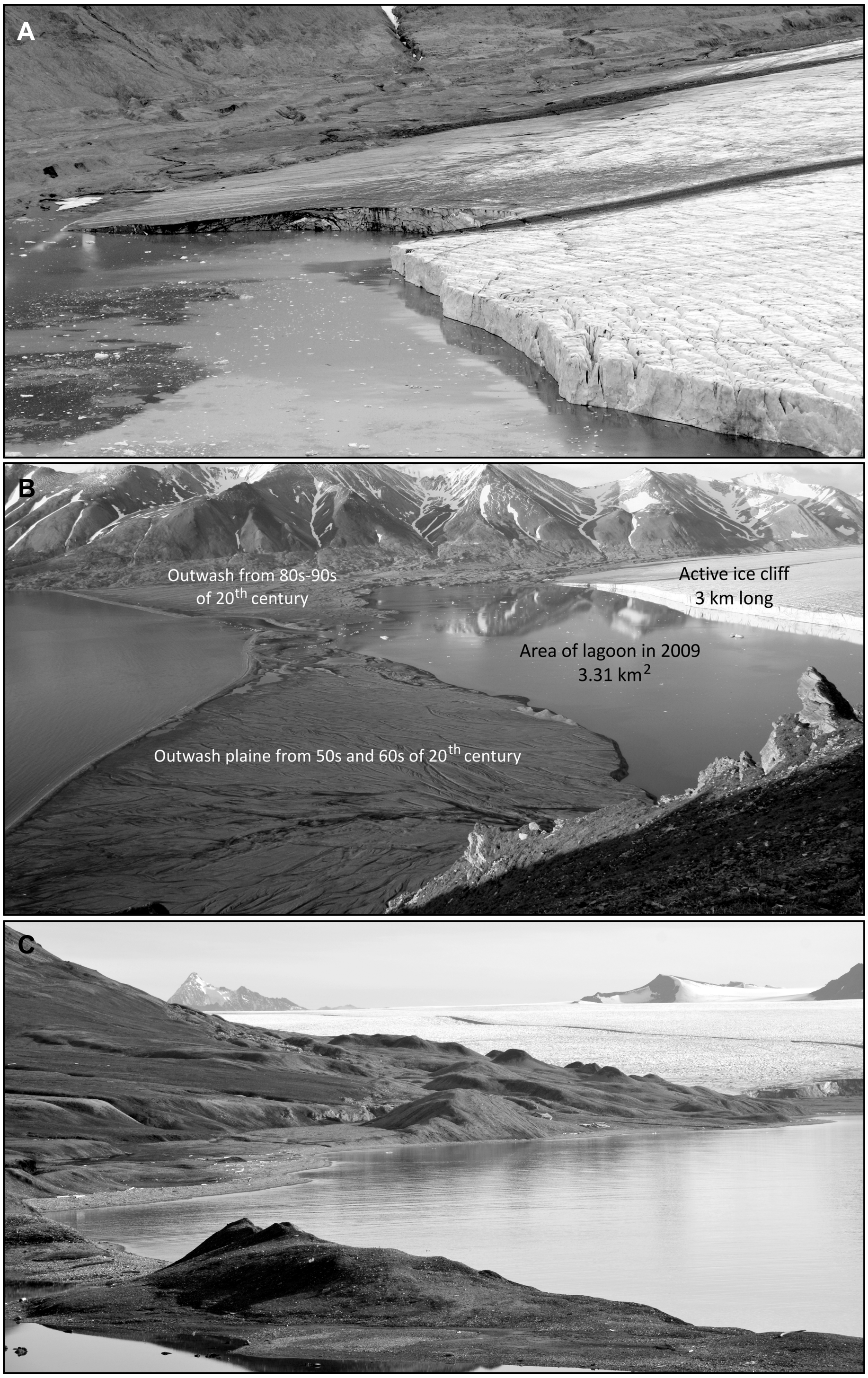

Photo 5.2.4. A- middle moraines of Recherchebreen (Photo P. Demczuk 2009), B- view of Recherchebreen forefield with system of outwash plains (Photo J. Jania 2008), C- lateral moraines of Recherchebreen – eastern coast of Recherchefjorden (Photo P. Zagórski 2011).

On both sides, left and right, two generations of moraines can be distinguished: older, built from mineral material and younger, containing relict ice-cores. The older moraine is further from the glacier and has higher hipsometrical position. It is built from mineral material of various fractions (domination of fine-grained materials). The top of moraine, on the western side of the glacier is supplied in sediments fromtalus fans. High hipsometrical position, domination of fine material, lack of ice-cores, and above all, sediment cover from of talus fans allow to associate that landform with the Pleistocene period (Reder 1996).

The younger latteral moraine is placed nearer the glacier tongue. It is shaped as ice-cored moraines in which ice core fills almost the whole structure of the landform and is covered only by thin mineral layer of thickness less than 0.2-0.3 m (Photo 5.2.4B). These landforms are under fast degradation, which in present climate conditions, undergo fast melting out of mineral covers in spring and in early summer, and then their solifluction on a frozen surface. Ablation happens due to permanent revealing of inner ice-cores, their melting out and slow lowering of moraine surface.

Renardbreen

The Renardbreen catchment area is 41 km^2. The glacier area was 30.4 km^2 in 2011 (Table 5.2.1). Its length reached 8 km and width was from 2.5 km to 7.5 km. The highest parts of the glacier reach 720 m a.s.l., its flat-convex front almost comes down to the sea level. Until 1990 it reached Josephbukta, but now Renardbreen is undergoing front recession with some symptoms of areal deglaciation in its left part of the head (Bartoszewski 1998; Gajek & Reder 2008; Zagórski *et al.* 2008b) (Photo 5.2.3B).

Renardbreen is drained in three ways: supra-, in- and subglacial. Subglacial drainage takes place by the net of dozens of meandering channels and considerable discharge. Not all of them flow at the same time, because some are filled and blocked by snow patches even in summer season. High inclination of the glacier front let transport not only fine material, but also even large rock blocks. Inglacial tunnels work in right and central part of the lake. Winding subglacial tunnels refer to cracks directions and have their outlets as 2-3 m high glacial gates (Bartoszewski 1998). Melt-waters carry plenty of suspended sediments as well as and gravel-dominated sediments, with clear dominance of middle gravel fraction (Łanczont 1988b).

Rivers that belong to supra-, in- and subglacial system, leaving the glacier, join at its edge into a proglacial river of concentrated flow, parallel to the glacier edge. Year by year its bed is moved towards the line of retreating glacier front. The surroundings of the glacier front are dominated by erosional processes. Across the glacier forefields, a proglacial river incises surface of outwash plain and divides them into single sediment patches and islands. The area remind s a badland landscape with plenty of abandoned channels and thermokrast hollows. In the lower section its waters carve contemporary outwash plain level with a system of braided channels. The proglacial river drains into the Josephbukta and forms a large delta.

In the intramarginal zone of the glacier, between ice-cored moraines and contemporary edge of the front, single patches of fluted moraine occur and numerous forms of glaciofluvial accumulation, for example three outwash plain levels. It is connected with an inclination of glacier forefields towards temporary front (Reder 1996; Merta 1988c, 1989) (Figs. 5.2.4 and 5.2.5; Photo 5.2.5). During the glacier recession lower levels of inner outwash plain are formed parallel to its edges. Between outwash plain zone and glacier front moutonised rock sill appears. Its proximal slope is mild and covered with fresh fluted moraine, but distal slope is steep and devoid of sediment cover (Pękala 1987; Łanczont 1988ab; Szczęsny *et al.* 1989).

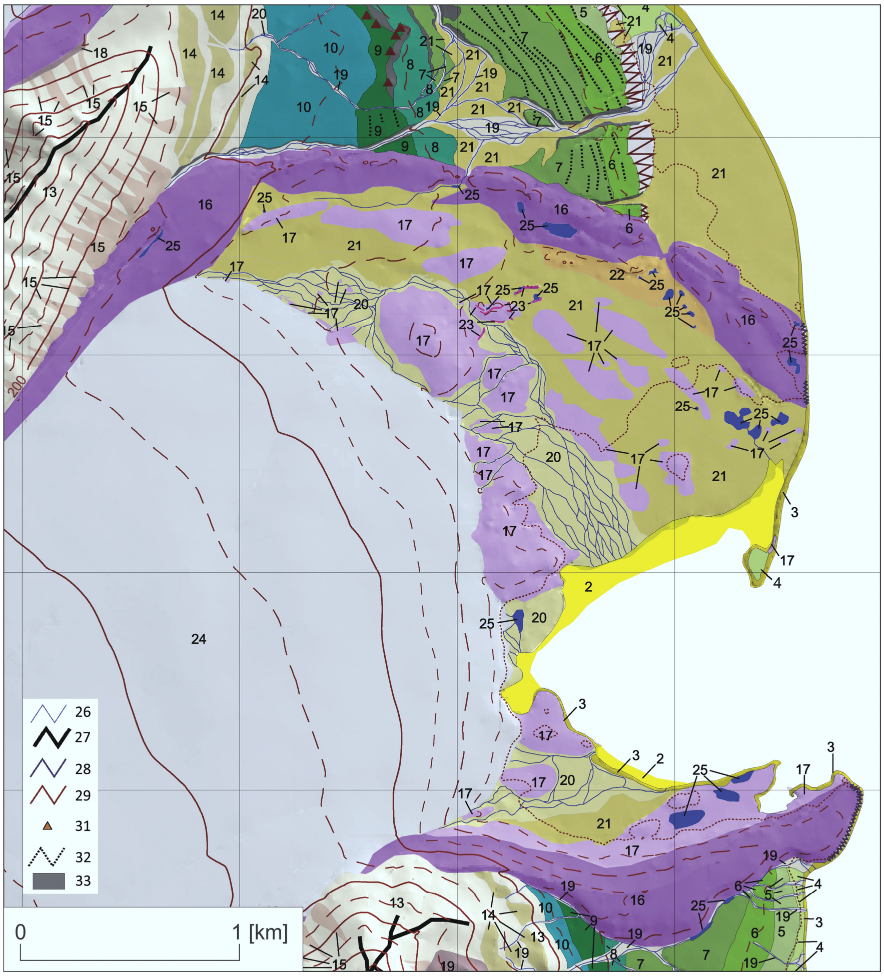

Fig. 5.2.4. Geomorphological map of the Renardbreen forefield (after: Zagórski 2002, legend according: Reder & Zagórski 2007a), numerical marking on the map: 2- tidal flat, delta, 3- modern storm ridge, 4- terrace I (2-8 m), 5- terrace II (10-20 m), 6- terrace III (25-30 m), 7- terrace IV (30-40 m), 8- terrace V (40-50 m), 9- terrace VI (50-65 m), 10- terrace VII (70-85 m), 13- slopes, 14- denudation-structure level, 15- talus fans, 16- ice-cored moraine, frontal and lateral moraines, 17- ground and ablation moraines, 18- rock glaciers (nival), 19- pronival river bottoms and alluvial fans, 20- modern outwash plains, 21- old outwash plains 22- kames, 23- eskers, 24- glacier, 25- lakes; other legende: 26- rivers, 27- ridges, 28- active cliffs, 29- fossil cliffs, 31- paleoskerries, 32- fossil storm ridges, 33- edges.

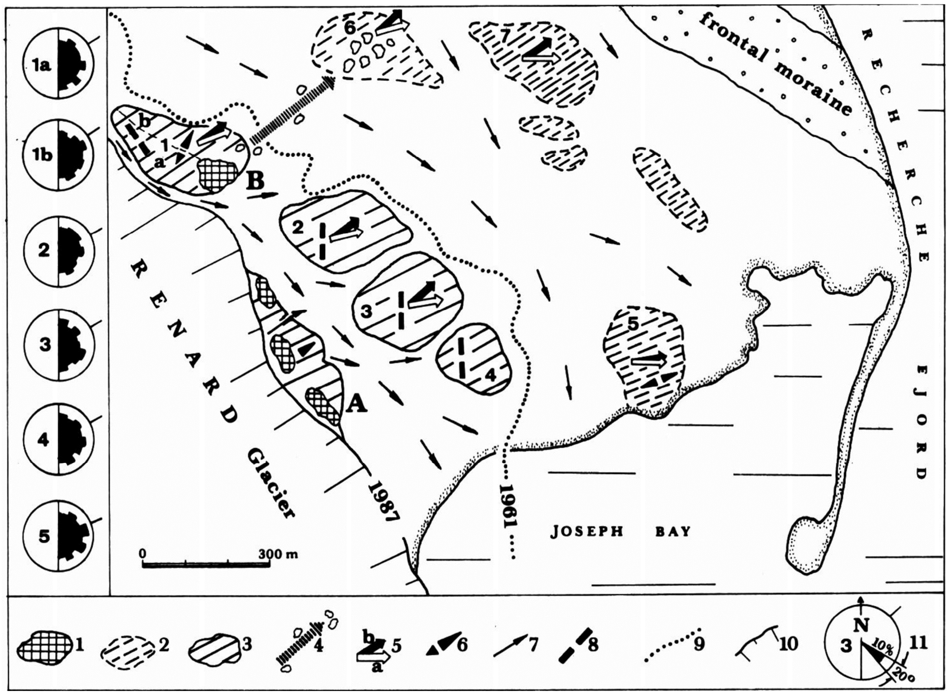

Fig. 5.2.5. Sketch map of the Renardbreen forefield zone (after: Merta 1989): 1- exposed Hecla Hoek Fm. rocks of the southeast (A) and northwest (B) thresholds, 2- patches of the ground moraine with the degradated top surface fluted relief; dashed lines indicate the crest directions, 3- ground moraine area with the distinct top surface fluted relief; the lines indicate the crest directions, 4- boulder transport direction from the glacier-eroded threshold B, 5- drumlin-like forms: a- boulder long axis orientation, b- tail orientation, 6- slope decline, 7- proglacial stream flow direction in the area of the outwash plain sedimentation, 8- transverse crest orientation, 9- glacier front position in 1961, 10- glacier front position in 1987, 11- measurements of the long axes orientations of boulders and stones in the half-circle projections; the number indicates the domain of the measurements and the external bar shows the crest elongations.

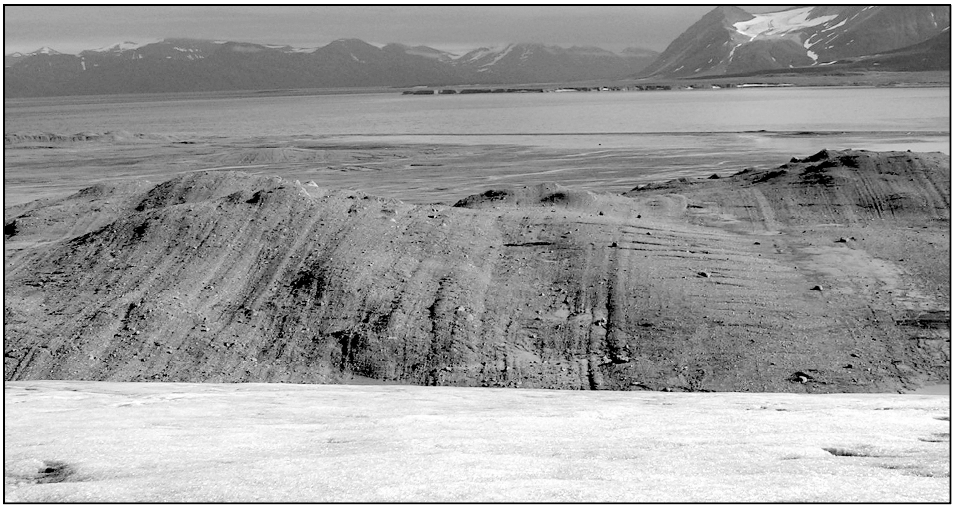

Photo 5.2.5. The fluted moraine covers the proximal slope of roche moutonnée – marginal zone of Renardbreen (Photo P. Zagórski 2006).

On distal side of roches moutonnées, there are two large eskers and a few mounds, which are the remnants of the third esker (Fig. 5.2.4, see: Photo 2.2.1B). Eskers are dozens of meters long and up to up to 5 m high. Their genesis might be linked with infilling of subglacial tunnel during recession in the second half of the 20th century. The set of glaciofluvial forms is completed by a kame terrace which is 100 m wide and adjoin ice-cored moraines on the distance of about 1 km. It was made during the early stage of glacial recession right after the termination of the Little Ice Age (Dzierżek *et al.* 1990a; Pękala & Repelewska-Pękalowa 1990).

The main landform in the Renardbreen proglacial zone is a frontal moraine that was accumulated on a marine terrace (6 m a.s.l.) that transforms, along glacier tongue, into a lateral moraines (Fig. 5.2.4, Photo 5.2.3B). The frontal moraine is eroded by waves and forms a cliff on both sides of the Josephbukta. Clear dichotomy of glacial forms is observed. The outer belt of moraines is formed as a pushed moraine with a of mixed glacio-moraine sediments at the bottom. Their age is estimated by TL method and it is 13.2±2 ka BP, in upper part is 8.6±1.3 ka BP (Szczęsny *et al.* 1989; Pękala & Repelewska-Pękalowa 1990) (see: Figs. 2.2.5 and 8.5).

The inner part of pushed moraine, which is an effect of a glacier advance during the Little Ice Age, consists of secondary ice-cored moraine elements. Their sharp peaks and high number of thermokarstic hollows (often filled with water) proves that relict ice is still preserved inside the landforms (Reder 1996). Between sediments of old pushed moraine and ice-cored moraine sediments the organic sediments was found. The organic material is deformed and moved from its original position. The upper part of organic sediment layer was dated on 660±80 ^{14}C BP, lower part was dated back to 1,130±80 lat BP (Dzierżek & Nitychoruk 1990; Dzierżek *et al.* 1990ab) (see: Figs. 2.2.5 and 8.5). Organic sediments were enriched by culture layer of archeological site linked with the 17th century West-European whaling activities (Jasinski 1994).

On the distal side of pushed moraine a vast and currently inactive extramarginal outwash plain is located (see: Photo 5.5.9). The plain is formed by to cones accumulated after incision of moraine belt. One cuts only older zone of pushed moraines. That is the trace of a river channel formed during the mid-Holocene glacier advance. The gorge in the moraine was closed by sediments deposited during the LIA glacier advance. The second gorse cuts the whole moraine zone – and it was created by river draining the glacier in its maximal LIA extent (Reder 1996).

Scottbreen

The Scottbreen fills a valley between Bohlinryggen and Wijkanderberget massifs. The Scottelva catchment covers an area of 10.1 km^2, whereas the area of the glacier was 4.75 km^2 in 2006. (Zagórski *et al.* 2008ac). The highest parts of Scottbreen almost reach 600 m a.s.l., while its face goes down to ca. 100 m a.s.l (Photo 5.2.6ABC). The glacier's length is 3.6 km, its width 1.0-1.5 km and its mean inclination is 8°. The Scottbreen's surface is highly undulating due to deformations in the glacier base.

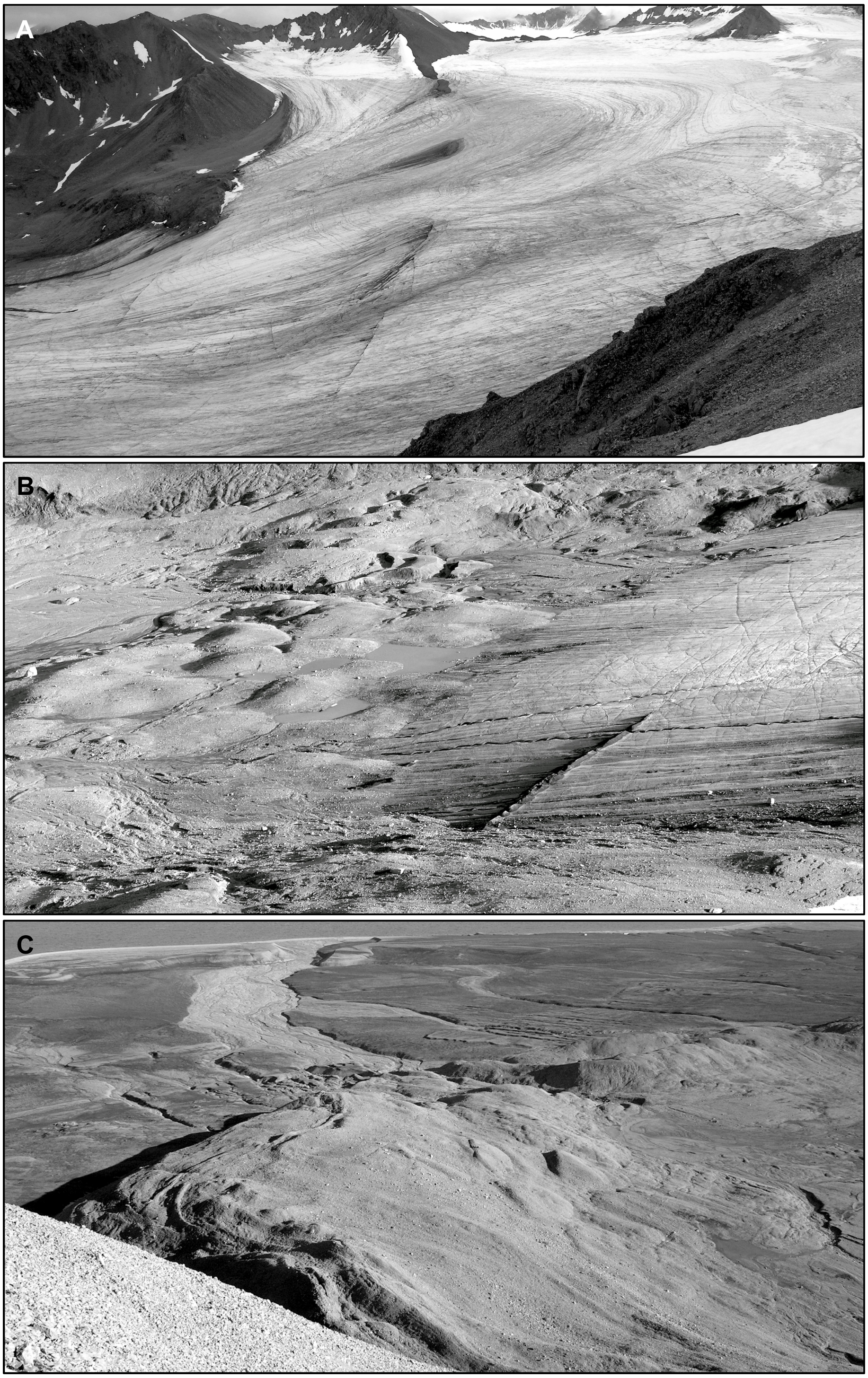

Photo 5.2.6. Scottbreen and Scottelva catchment: A- the upper and middle part of the glacier (Photo P. Zagórski 2005), B- the relief of marginal zone (Photo P. Demczuk 2009), C- the frontal moraine ridges (older and from LIA) and extramarginal outwash plain - Scottleva (Photo P. Zagórski 2007).

The ice-tongue surface features ridges reaching up to 2 m, which are transverse to the glacier axis and direction, and are made up of fine-grained material that was moved along the shearing panes in the ice (Photo 5.2.6B). These ridges are of asymmetrical nature: their proximal glacis makes a smooth extension of the glacier area and the distal glacis falls at an angle of 30-40°. A similar group of forms, with a similar direction in relation to the glacier face, can be observed to the left, at a distance of several dozen metres from the glacier face (Reder 1996; Bartoszewski 1998; Reder & Zagórski 2007b).

Scottbreen exhibits a typical frontal recession and is drained by the dispersed supraglacial run-offs. The only englacial channel, starting with a huge glacier basin with the diameter of 10 m, is found in the north-eastern part of the tongue, at the contact with the right lateral Bohlinryggen moraine, raising up to 60 m above the glacier surface. Between these two structures a kame terrace was formed (Pękala 1987; Merta 1988b; Reder 1996; Zagórski *et al.* 2008a) (Photo 5.2.6B). A marginal river, forming a small outwash fan, flows out of the channel. The second fan, which is oblique to the glacier face and shows only seasonal activity, is located in the centre of the glacier foreland. The area between the terminal moraine and the glacier does not make a typical outwash plain, but rather a zone where ground moraine is split and washed and small fans are accumulated. A more distinct landscape element is a series of small hills marking one of the stages of glacier retreat (Photos 5.2.6BC and 5.2.7). The south-eastern part of the intermarginal zone is a lower-lying and water-covered area, whereas the north-western part, from the front moraine line, lies a few metres higher. The area is mostly a ground moraine, which is locally fluted and intertwined by several active beds with the depth of 2-4 m, extending to the only active moraine face (Merta 1989).

Photo 5.6.7. Marginal zone of Scottbreen (Photo P. Zagórski 2006).

The ground moraine is smoothly transformed into the terminal moraine, with a ridge of up to 120 m a.s.l. as a dominant element of the Scottbreen foreland (Fig. 5.2.4, Photo 5.2.6C). The ridge lies on a roche moutonnée-type bedrock of old denudation levels, predominating it by 25 meters. In relation to the depression of the intermarginal zone, the height difference is only 15 meters (Reder 1996). Greater moraine ridge dimensions, when compared to the glacier's size, indicate cyclic glacial accumulation in the same zone, which may be determined by the bedrock shape. Both the terminal and lateral moraines are complex forms developed in two different phases of glacier transgression. Young, partly banked, glacial and moraine sediments were deposed at glacier advance in the 19th century, probably in 1890 (Liestøl 1993; Reder 1996). These sediments cover a bit older moraine series which might have been formed during Scottbreen transgression in the earlier LIA phase (Fig. 5.2.3A, Photo 5.2.6C).

At the outlet of the terminal moraine ridge breach the extramarginal outwah plain is found (Photo 5.2.6C). The outwash plain cuts several meters into elevated Calypsostranda marine terraces representing the area with the length of over 1.5 km and average width of 0.2-0.3 km. Close to where the Scottelva flows into the Bellsund, in a circa 50 m wide gorge that splits the dead cliff, the depth of the embayment is 10 m (Reder 1996; Zagórski *et al.* 2008a). The outwash plain is intersected by numerous channels of varying configuration, which cut into the stone and gravel sheets to the depth of ca. 0.5 m.

<u>Crammerbreane</u>

In 2011, the Chamberlindalen glaciers covered an area of 8.16 km^2, which accounts for 17% of the total area of its catchment (Table 5.2.1). There are several small glaciers at its eastern margin, the largest of which – Bøckmanbreen – covers only 0.5 km^2 (Photo 5.2.8A, see: Photo 5.1.4). In the western part of the valley four glaciers can be found, known by their collective name Crammerbreane. In the mid-20th century their total area was 6.75 km^2. Due to their location within the fault zone, the even glacial surface in the accumulation zone reveals a sudden fall in the ablation zone (Zagórski *et al.* 2012). Deep lateral fractures and seracs were formed at a break of the longitudinal profile of the glaciers. The above mentioned glaciers are now in their retreat. Between 1990 and 2009, they were retreating by 7-12 m·a^{-1}. Their forelands are characterised by numerous cave-in forms, such as eskers and kame terraces. The complex geological structure and its layout have both significantly contributed to the formation of the ground surface layout (Pękala & Repelewska-Pękalowa 1988b).

The terminus of the two southern glaciers (Crammer III and IV) has a wide internal outwash plain (Pękala & Repelewska-Pękalowa 1988b). In July 2005 ice-dammed lakes could be observed in front of the faces of both glaciers. The Crammer III lake covered an area of 0.3 km^2 (Photo 5.2.8B). The face in the southern part of the glacier ended with a 15 m high ice cliff whose ice blocks used to fall to the lake (Bartoszewski 1998).

Photo 5.2.8. A- view of the Crammerbreane forefields - Chamberlindalen (Photo P. Demczuk 2009). B- the lake on marginal zone of the Crammerbreen III (Photo P. Zagórski 2005).

The NW Valley Glaciers

The NW-trending valleys, which extend out to the Bellsund at its outlet to the Greenland Sea, are characterized by mild glaciation (Appendix 1). There are two small catchments in this zone, namely Blomlielva and Tjørnelva, with an area of 7.0 km^2 and 6.1 km^2 and glaciation of 25% and 15% respectively, and two larger catchments, Dyrstadelva and Logna, with the area of 14.8 and 20.4 km^2 and glaciation of 20% and 10% respectively (Table 4.1.1). Glaciers only occur in upper parts of these valleys. Most of them constitute intermixed types of valley and cirque glaciers. The ice thickness in the face zones that are usually covered by ablation moraine is strongly reduced (Bartoszewski 1998).

In its upper part, Blomlibreen is bound to Scottbreen, yet its highly inclined residual tongue ends already at an altitude of 270 m a.s.l. (Fig. 5.2.1, Photo 5.2.9A). The ablation moraine that covers changes into a ground foreland moraine, partly in the form of a fluted moraine (Merta 1989; Reder 1996) (Fig. 5.2.6). The marginal glacier zone is narrow and delimited by a steep roche moutonnée, cutting across the valley diagonally. It is overbuilt by a massive – in relation to the glacier – terminal moraine developed in the form of relict ice ridges (Photo 5.2.9A). The outflow of ablation waters from a small pool located in front of the glacier face occurs along thermokarst channels inside moraine ridges. The sedimentation of varved clay alone in this reservoir and absence of thicker material suggest very limited dynamics of morphogenetic processes at the present time (Merta 1988a). The stagnation of these processes is also indicated by the stabilisation of beds within the extramarginal outwash plain located 150 m below the present glacier face. The outwash plain begins in the outlet of a narrow and steep gorge that is cut in a ridge below the moraine, whereas the outflow to the sea occurs along a 30 m deep canyon, cut out in the elevated marine terraces.

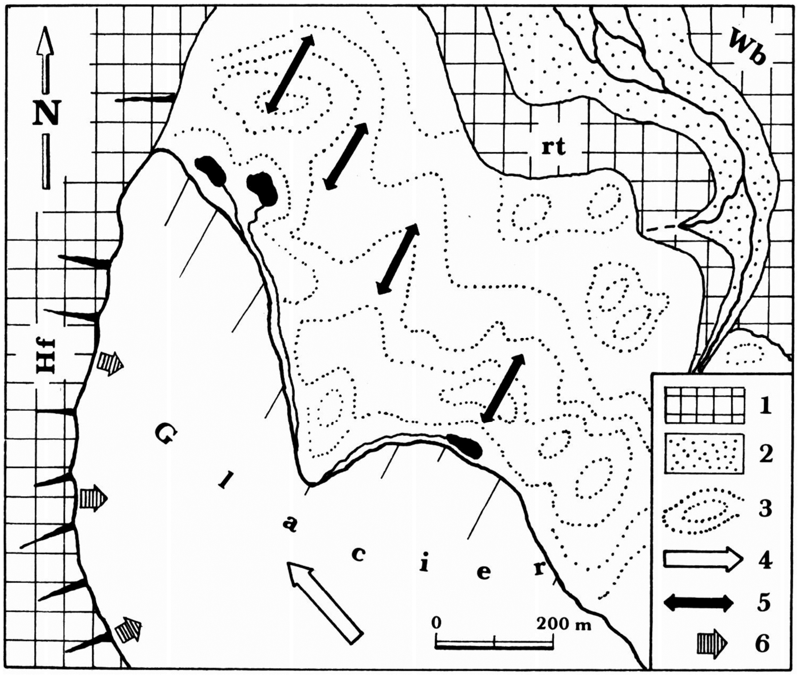

Fig. 5.2.6. Sketch map of the front zone and the forefield area of the Blomlibreen (after: Merta 1989): Hf-escarpment of the Halvorsenfjellet, Wb- Wijkanderberget, rt- rock threshold, 1- Hecla Hoek Formation, 2- outwash plain deposits, 3- morainic morphology outline in the forefield zone, 4- glacier movement direction, 5- crest elongation, 6- slope material yields.

Tjørndalsbreen fills out only the upper part of Tjørndalen and is moved to the west by a bulge in the bedrock. On the east, there is a lateral moraine that consists of two parallel and partly overthrust ridges. A residual tongue divides a narrow and 500 m long medial moraine (Fig. 5.2.1, Photo 5.2.9B). The glacier face, which goes down to only 300 m a.s.l., is thoroughly covered by mineral material: areal deglaciation has been observed here (Bartoszewski 1998). Despite relatively well-developed forms of glacial accumulation, the terminal moraine is not clearly structured. A sandur trail, laying just below, shows little activity. The 50-200 m wide trail ends in a shallow ice-dammed lake from which water is discharged to the sea along the canyon.

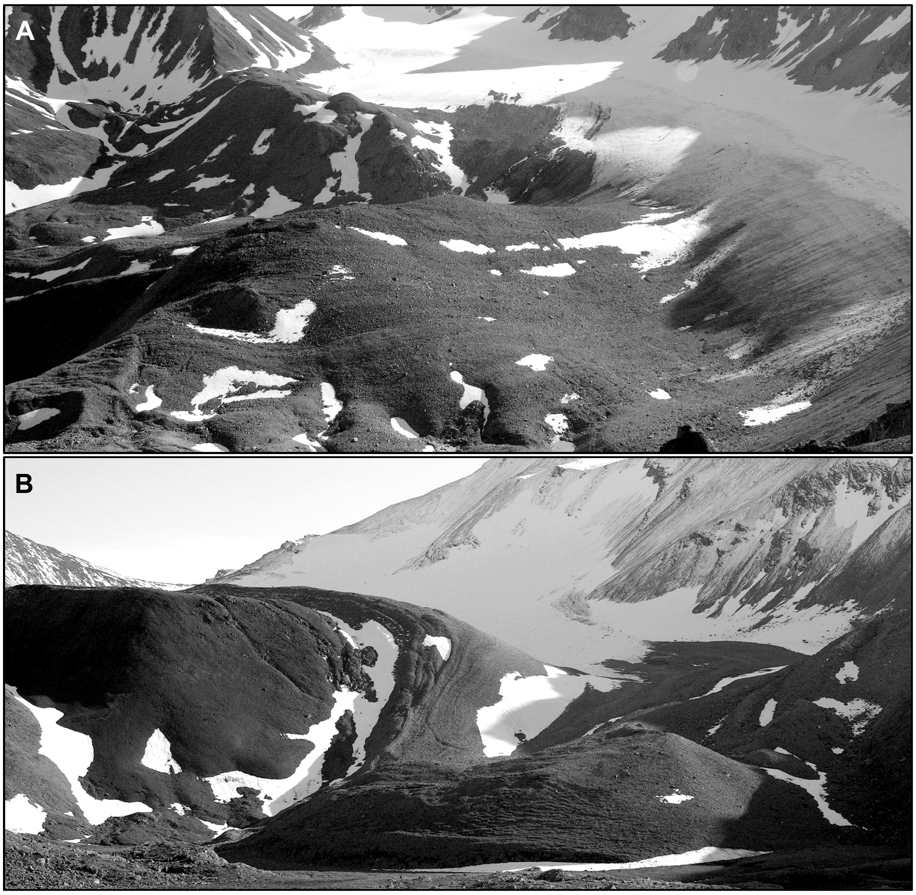

Photo 5.2.9. A- the marginal zone of Blomlibreen forefield (Photo P. Zagórski 2008), B- the marginal zone of Tjøndalsbreen forefield (Photo P. Zagórski 2007).

The upper part of the Dyrstaddalen occures a group of four glaciers – Ringarbreane- covering in 2011 a total area of ca. 2.5 km^2 (Figs. 5.2.1 and 5.2.7; Table 5.2.1). The glaciers are now in retreat The Ringarbreane faces, falling as low as to 220 m a.s.l., show little inclination or are flat and covered by ablation moraine (Pękala & Reder 1989; Reder 1990, 1996). Medial moraine ridges and single ice-cored moraine hillrocks are found locally. The glacier foreland incorporates the ground moraine with numerous ice-dammed lakes (Photo 5.2.10). The width of the marginal moraine zone is 200-350 m (Appendix 2). The terminal moraine features three series of partly overthrust ice-cored ridges that consist of dense rock material, with block diameter of up to 0.5 m. The ridges are characterized by significant elevation differences, are narrow and steep, and are not split by any river gorge. In some places water outflows have been observed that are indicative of thermokarst melting of relict ice inside the ridge. To the north of the terminal moraine ridge there is a typical valley outwash plain which ends in a canyon splitting the zone of elevated marine terraces.

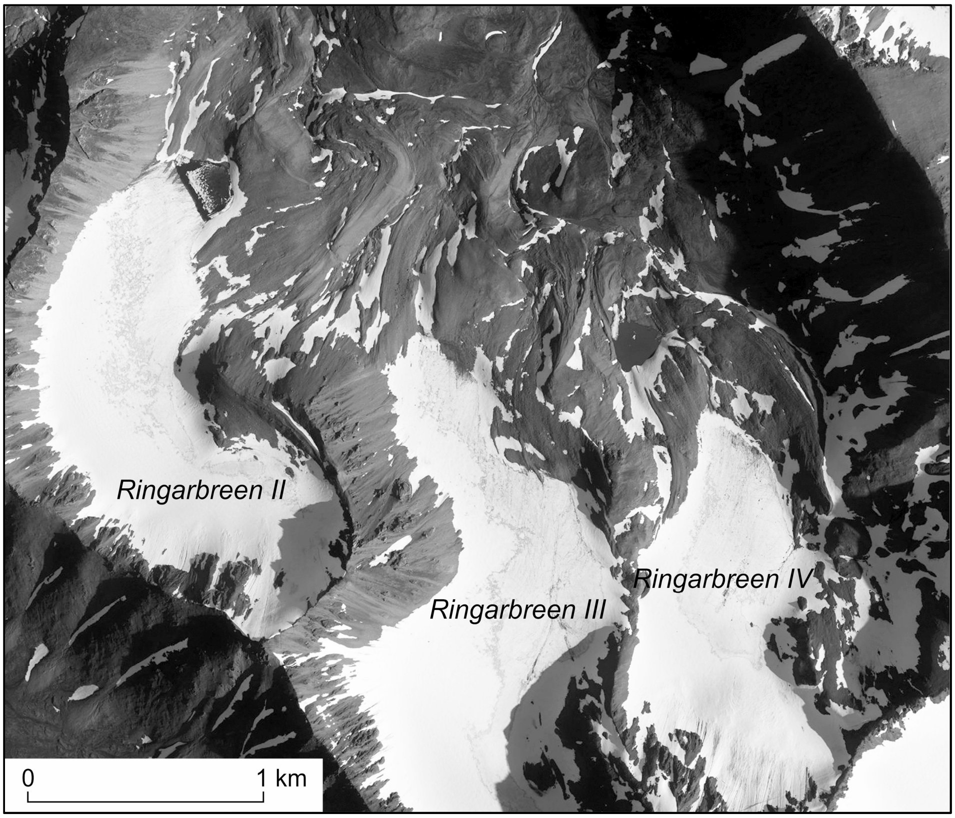

Fig. 5.2.7. View of the Ringarbreane forefields - aerial photo from 2011. Source: Topo Svalbard, Norwegian Polar Institute.

Photo 5.2.10. Lake on the marginal zone of Ringarbreen II (Photo P. Zagórski 2007).

The root part of the Lognedalen is covered by four small glaciers with a total area of 2 km^2: Gløtfonna, Vestre Lognebreen, Austre Lognebreen and a small glacier without a name (Figs. 5.2.1 and 5.2.8). Gløtfonna, that used to be connected to Renardbreen, lies in a basin at the Storgubben side and ends ca. 200 m above the Lognedalen bed, at the altitude of ca. 300 m a.s.l. The glacier surface is concave, and is covered with greater rock blocks and ablation moraine clay. Vestre Lognebreen has the characteristics of a field glacier. Its tongue, which falls down to 200 m a.s.l. and shows visible traces of retreat, hangs on a knickpoint. The lowest of all glaciers is Austre Lognebreen that falls down to 150 m a.s.l. Numerous eskers and kames in the foreland document old subglacial water outflows (Bartoszewski 1998).

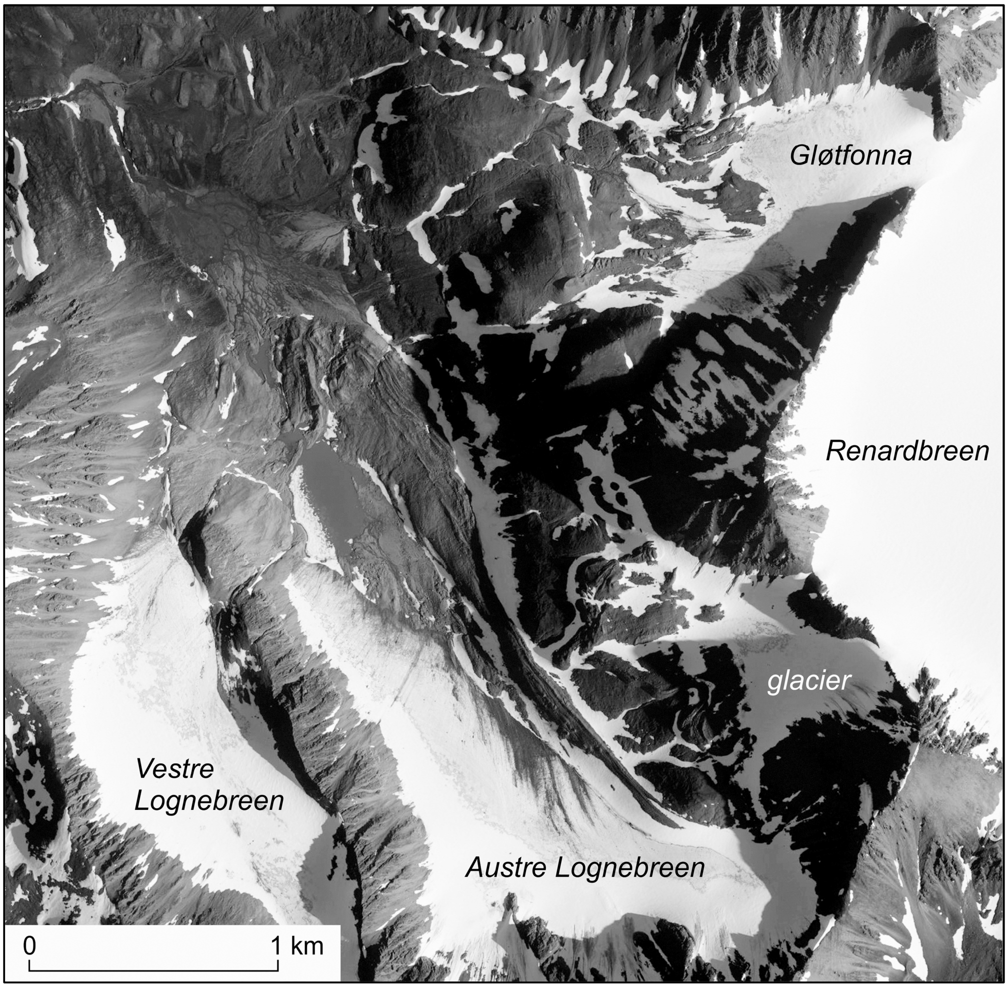

Fig. 5.2.8. Wiew of the forefields of Vestre Lognebreen Austre Lognebreen, Gløtfonna and small without name glacier. Source: Topo Svalbard, Norwegian Polar Institute.

Streszczenie

Rzeźba glacjalna

Lodowce zajmują obecnie tylko około 20% powierzchni północno-zachodniej części Ziemi Wedela Jarlsberga, jednak w okresach glacjalnych plejstocenu obszar ten, poza partiami grzbietowymi, był całkowicie zlodowacony, a zachodnie wybrzeże Spitsbergenu stanowiło partię brzeżną lądolodu Morza Barentsa. Zlodowacenie weichselian jest tu udokumentowane osadami w profilu Skilvika (55±8 ka i 26 ka BP) i w Dyrstaddalen (59±9 ka BP). Jeszcze około 11-12 ka BP lodowce wypełniały Recherchefjorden, zaś obciążenie izostatyczne wybrzeża Bellsundu w stosunku do stanu współczesnego wynosiło 50-60 m. Wszystkie doliny miały dna położone poniżej poziomu morza i były wypełnione lodowcami kończącymi się w morzu, które uległy rozpadowi ok. 10-9 ka BP. Największy ich holoceński zasięg wyznaczają moreny czołowe, datowane na Małą Epokę Lodową (MEL). U jej schyłku powszechna była transgresja lodowców, przeważnie typu „szarży". Świadczą o tym spiętrzone moreny czołowe, wały lodowo-morenowe oraz formy z wyciśnięcia.

W obrębie terenu badań znajduje się jęzor tylko jednego dużego lodowca sieciowego, Recherchebreen, uchodzącego do morza. Pozostałe lodowce północno-zachodniej części Ziemi Wedela Jarlsberga, dolinne lub karowe, kończą się na lądzie, przy czym Renardbreen utracił kontakt z morzem dopiero w końcu XX w. Położenie, kształt, a w znacznym stopniu także dynamika lodowców, zależą od rzeźby podłoża skalnego, uwarunkowanej jego tektoniczno-litologiczną strukturą. Pola alimentacyjne położone są zwykle w obrębie spłaszczeń strukturalno-denudacyjnych, na wysokości 250-300 oraz 400-500 m n.p.m. Obecnie linia równowagi podniosła się powyżej 450 m n.p.m.; w związku z tym powierzchnie ablacji są znacząco większe od pól alimentacji, która odbywa się głównie poprzez akumulację lodu nałożonego. Wynikiem ujemnego bilansu masy lodowców jest ich recesja, przebiegająca w tempie kilku metrów na rok, zaś w przypadku dużych lodowców przyspieszająca okresowo do kilkudziesięciu metrów na rok. Od schyłku MEL większe lodowce (Renardbreen, Scottbreen) zmniejszyły swoją powierzchnię o 10-25%, zaś małe o 30-50%.

Współczesna rzeźba przedpoli lodowców kończących się na lądzie jest nałożona na egzaracyjne formy plejstoceńskie i podniesione terasy morskie. Dominującym elementem są wały moren czołowych, przechodzące wzdłuż jęzorów w moreny boczne. Przecięte są one przełomami (bramami), u wylotu których znajdują się sandry zewnętrzne. Zewnętrzną, starszą strefę wałów moren czołowych tworzą moreny spiętrzone, zaś wewnętrzną, młodszą – wały lodowo-morenowe, do których nierzadko przylegają terasy kemowe. Pomiędzy cofającymi się jęzorami a morenami czołowymi wytworzyły się strefy moren dennych i ablacyjnych, sandrów wewnętrznych oraz towarzyszących im nieraz form szczelinowych. Fragmenty moren dennych, często wykształcone w postaci moreny żłobkowej, widoczne są przede wszystkim na wypukłych elementach starszej rzeźby, głównie zmutonowanych ostańcach skalnych. Po stronie dystalnej mutonów znajdują się niekiedy wały ozów. Obniżenia pomiędzy mutonami wykorzystują rzeki proglacjalne, na przemian tworzące i rozcinające powierzchnie sandrowe. Wszystkie te elementy rzeźby występują w różnych konfiguracjach i nie są reprezentowane na przedpolach wszystkich lodowców.

Objaśnienia

Ryciny

Ryc. 5.2.1. Położenie lodowców NW części Wedel Jarlsberg Land (numeracja zgodna z tab. 5.2.1): 1- Recherchebreen, 2- Renardbreen, 3- Scottbreen, 4- vestre Lognedalsbreen, 5- austre Lognedalsbreen, 6- Gløttfonna, 7- Ringarbreane, 7a- Ringarbreen I, 7b- Ringarbreen II, 7c- Ringarbreen III, 7d- Ringarbreen IV, 8- Tjørndalsbreen, 9- Blomlibreen, 10- Dölterbreen, 11- Crammerbreane, 11a- Crammerbreen I, 11b- Crammerbreen II, 11c- Crammerbreen III, 11d- Crammerbreen IV, 12- Bøckmanbreen. Źródło podkładu: Topo Svalbard, Norwegian Polar Institute).

Ryc. 5.2.2. Uproszczona mapa tektoniczna NW części Wedel Jarlsberg Land (Birkenmajer 2004, 2006; Gajek i in. 2010). Podkład: Orthophotomap (Zagórski 2005).

Ryc. 5.2.3. Średnia prędkość recesji Renardbreen i Scottbreen w poszczególnych okresach na tle średniej temperatury powietrza w okresie od czerwca do września na stacji Svalbard-Lufthavn: 1- przebieg temperatury wyrównany 5-punktową średnią kroczącą, 2- średnia prędkość recesji czoła Renardbreen w poszczególnych okresach, 3- średnia wartość recesji czoła Scottbreen w poszczególnych okresach (Zagórski i in. 2008ab, zmodyfikowane i uzupełnione).

Ryc. 5.2.4. Mapa geomorfologiczna przedpola Renardbreen (Zagórski 2002; legenda zgodna z Reder, Zagórski 2007a): 2- równia pływowa, delta, 3- współczesny wał sztormowy, 4- terasa I (2-8 m), 5- terasa II (10-20 m), 6- terasa III (25-30 m), 7- terasa IV (30-40 m), 8- terasa V (40-50 m), 9- terasa VI (50-65 m), 10- terasa VII (70-85 m), 13- stoki, 14- poziomy denudacyjno-strukturalne, 15- stożki usypiskowe, 16- moreny z jądrem lodowym, moreny: czołowa i boczna, 17- moreny denna i ablacyjna, 18- lodowce gruzowe, 19- dna rzek proniwalnych i stożki napływowe, 20- współczesne sandry, 22- kemy, 23- ozy, 24- lodowiec, 25- jeziora, 26- rzeki, 27- grzbiety górskie, 28- klify aktywne, 29- klify martwe, 31- paleo-szkiery, 32- stare wały sztormowe, 33- krawędzie.

Ryc. 5.2.5. Szkic stref przedpola Renardbreen (Merta 1989): 1- wychodnie skał formacji Hecla Hoek na progach południowo-wschodnim (A) i zachodnim (B) 2- płaty moreny dennej ze zdegradowaną górną częścią rzeźby żłobkowej; linie przerywane wskazują kierunki przebiegu grzbietów, 3- powierzchnia moreny dennej z wyraźną w górnej części rzeźbą żłobkową; linie wskazują kierunki grzbietów, 4- kierunek transportu głazów z progu B erodowanego lodowcowo, 5- formy typu drumlinowego: a- orientacja osi dłuższych głazów, b- orientacja ogonów, 6- nachylenie zbocza, 7- kierunek odpływu strumieni proglacjalnych w obszarze sedymentacji sandrowej, 8- orientacja poprzecznych grzbietów, 9- zasięg czoła lodowca w 1961, 10- zasięg czoła lodowca w 1987 roku, 11- pomiary orientacji dłuższych osi głazów i kamieni na projekcji półkola; liczba wskazuje domenę pomiarów a zewnętrzny pasek pokazuje wydłużenia grzbietów.

Ryc. 5.2.6. Szkic strefy marginalnej i przedpola Blomlibreen (Merta 1989): Hf- zbocza Halvorsenfjellet, Wb- Wijkanderberget, rt- progi skalne, 1- formacja Hecla Hoek, 2- osady sandrowe, 3- zarys morfologii morenowej w strefie przedpola, 4- kierunek ruchu lodowca, 5- ukierunkowanie grzbietów, 6- dostawa materiału zboczowego.

Ryc. 5.2.7. Widok przedpoli Ringarbreane – zdjęcie lotniczne z 2011 (źródło: Topo Svalbard, Norwegian Polar Institute).

Ryc. 5.2.8, Widok na przedpola: Vestre Lognebreen Austre Lognebreen, Gløtfonna i małego lodowca bez nazwy. Źródło: Topo Svalbard, Norwegian Polar Institute.

Fotografie

Fot. 5.2.1. Rozmieszczenie basenów alimentacyjnych na Renardbreen (fot. M.Grabiec 2005).

Fot. 5.2.2. Widok na przedpola Scottbreen i Renardbreen (fot. M.Górska 2003).

Fot. 5.2.3. Zmiany zasięgu czoła: A- Scottbreen, B- Renardbreen (fot. P. Zagórski 2006).

Fot. 5.2.4. A- morena środkowa Recherchebreen (fot. P. Demczuk 2009), B- widok przedpola Recherchebreen z systemem stożków sandrowych (fot. J. Jania 2008), C- morena boczna Recherchebreen – wschodnie wybrzeże Recherchefjorden (fot. P. Zagórski 2011).

Fot. 5.2.5. Morena żłobkowa przykrywająca stok proksymalny mutonu – strefa marginalna Renardbreen (fot. P. Zagórski 2006).
Fot. 5.2.6. Scottbreen i zlewnia Scottelvy: A- górna i środkowa część lodowca (fot. P. Zagórski 2005), B- rzeźba strefy marginalnej (fot. P. Demczuk 2009), C- wały moreny czołowej (starsze i z LIA) oraz sandr zewnętrzny – Scottbelva (fot. P. Zagórski 2007).
Fot. 5.2.7. Strefa marginalna Scottbreen (fot. P. Zagórski 2006).
Fot. 5.2.8. A- Widok stref marginalnych i przedpoli Crammerbreane – rejon Chamberlindalen (fot. P. Demczuk 2009), B- jezioro w strefie marginalnej Crammerbreen III.
Fot. 5.2.9. A- strefa marginalna Blomlibreen (fot. P. Zagórski 2008), B- strefa marginalna Tjøndalsbreen (fot. P. Zagórski 2007).
Fot. 5.2.10. Jezioro w strefie marginalna Ringarbreen II (fot. P. Zagórski 2007).

Tabele

Tabela 5.2.1. Powierzchnie lodowców w wybranych latach na podstawie danych archiwalnych i pomiary GPS.
Tabela 5.2.2. Wielkość recesji lodowców w okresie 1936-2011.