*Jan Dzierżek*
*Jerzy Nitychoruk*

# 5.1. Orography and slope processes

The Wedel Jarlsberg Land is a mountainous region extending between Dunderdalen and Malbukta in SW Spitsbergen. The area borderd by the latitude 78°26' N to the south and by the waters of Bellsund to the north is heavily divided by glacier valleys. In some of the valleys the glaciers are better developed (Renardbreen, Scottbreen) and occasionally a glacier fills the valley completely (Recherchebreen); other valleys are nearly completely free of ice (Dunderdalen, Chamberlindalen). The character and course of relict and modern glaciations and geology of the region (lithology, tectonics) determine the geomorphology of this part of Spitsbergen (Orthophotomap, Appendix 1).

The relief of the NW part of Wedel Jarlsberg Land, is characterised by relatively flat areas such as: present-day glacier surfaces (ice tongue and firn fields), valleys left by retreating glaciers, coastal plains or raised beaches. Plain areas, comprised mostly of valley floors and marine terraces, that reach a maximum elevation of 130-150 m a.s.l. constitute the lowest structural and denudational level (Pękala & Repelewska-Pękalowa 1988a). The largest areas with a flat relief belonging to this level are located in Dunderdalen and Chamberlindalen (approximately 22 km$^2$ each), as well as Lognedalen and Lognedalenflya (20 km$^2$). Other similar areas, slightly smaller in size, can be found across the plains of Calypsostranda (12 km$^2$), Reinslatta (11 km$^2$), Dyrstadflya and Dyrstaddalen (7 km$^2$). The lowest orographic element of this area is the narrow strip of beaches and tidal flats (Zagórski 2004a, 2005). These are located at the mouth of the largest valleys (Dunderdalen, Chamberlindalen) and in the vicinity of Calypsostranda. The more elevated plains comprise a system of eight raised marine terraces, separated by a clearly distinguishable morphological step. Terraces occur in the following elevation ranges: 2-8 m a.s.l., 10-20 m a.s.l., 22-30 m a.s.l., 30-40 m a.s.l., 40-50 m a.s.l., 50-65 m a.s.l., 70-85 m a.s.l., 105-120 m a.s.l. (Zagórski 2002, 2004b).

Glaciers and firn fields, often connected with each other, form gently sloping areas in the modern landscape of Wedel Jarlsberg Land. These formations occur between the local sea-level (front of the Recherchebreen) and 700 m a.s.l. (Scottbreen, Renardbreen). In the central part of the research area they form the largest firn and ice plain (40 km$^2$). Equally large territories are covered by the Recherchebreen and Antoniabreen in the eastern part of the area. In the case of most glaciers the existence of firn fields indicates the location of two structural and denudation levels that are generally believed to be located at the level of 250-300 m a.s.l. and 450-500 m a.s.l. (Pękala & Repelewska-Pękalowa 1988a; Reder 2006) (Appendix 2, Geomorphological map).

In contrast to the relatively flat areas of land and ice are the sharp edged mountain peaks and slopes, showing features characteristic for alpine areas, despite the generally low absolute altitude. In the region between the Greenland Sea and Recherchefjorden, the mountain ridges spread radially away from the Stuptinden massif (755 m a.s.l.). Starting from this peak, one ridge leads away into the NE through Becketoppane (760 m a.s.l.) all the way to Activekammen (538 m a.s.l.) – it is approximately 8 km long and constitutes the border of the Renardbreen valley (Appendix 1, Ortophotomap, Zagórski 2005). Secondary ridges lead away from the upper parts of main rigde, surrounding smaller valleys or glacial cirques (e.g. Crammerbreane). To the N and NNE of Stuptindnen there are isolated mountain peaks with an altitude of over 700 m a.s.l. (Grytdalsnuten, Linuten, Dalkletten, Gløttnuten) – these constitute the southern border of Renardbreen. Starting at the Dalkletten, a clearly distinguishable ridge runs south all the way to Dunderfjellet (745 m a.s.l.), and further on in the direction of WNW to Dundrabeisen (435 m a.s.l.). Side crests lead away from it in the direction of NW, which constitute the borders of smaller valleys: Hamardalen, Vestre Lognedalsbreen, and Austre Lognedalsbreen. From the Storgubben (825 m a.s.l.), which is the highest summit in this part of Wedel Jarlsberg Land, originate regular ridges that separate the valleys of Dyrstaddalen (occupied in the west by Ringarbreane), Tjørndalen, Blomlidalen, Scottbreen and Renardbreen. The longest crest, the one that constitutes the western border of Dyrstaddalen, is about 7 km long and forms an arch from Storgubben through Ringaren (685 m a.s.l.) and Kolven (570 m a.s.l.) to Klokkefjellet (553 m a.s.l.). To the NNW of Kjerulfjellet, all the way to Bellsundhesten (490-388 m a.s.l.), there is a relatively straight ridge that constitutes the SW border of Tjørndalen. In the upper region a side ridge leads to Emil Nilssonfjellet (788 m a.s.l.). This peak is part of the ridge line that constitutes the amphitheatric borderline of the Scottbreen and Blomlibreen glacier system, running from Halvorsenfjellet (370 m a.s.l.) in the west to Bohlinryggen (555-515 m a.s.l.) in the east. Wijkanderberget (561m a.s.l.) – a vast massif with a visibly flattened summit – separates these two valleys in the north (Appendix 1).

The region of Wedel Jarlsberg Land that lies to the E of Chamberlindalen is definitely higher and mountain massifs and glacier valleys found here are significantly bigger. The ridge running from Palanderfjellet (725 m a.s.l.) through Solhøgda (663) to Observatoriefjellet (565 m a.s.l.) separates the non-glacierised Chamberlindalen from the Recherchebreen. It runs in the direction of NNW. Mountain slopes situated in the vicinity of glaciers are steeper and their relief is more diversed. A huge massif separates Recherchebreen from Antoniabreen. Its ridge runs generally in the direction of NNW and joins the following peaks: Gavltoppane (886 m a.s.l.), Bienaimétoppane (760 m a.s.l.), Durochertoppen (771 m a.s.l.), Magnethøgda (805 m a.s.l.), Okernuten (715 m a.s.l.), Jarnfjellet (696 m a.s.l.) and Maria Theresiatoppen (655 m a.s.l.). The highest mountain massifs in this part of Wedel Jarlsberg Land – Berzeliustinden (1205 m a.s.l.) and Hermelinberget (1065 m a.s.l.) close off Antoniabreen from the east, at the same time constituting the border of the studied area.

The summit surface located in the NW part of Wedel Jarlsberg Land at the level of 650-850 m a.s.l. takes up 3% of the studied area, and the flat regions ranging in height between 400 and 600 m a.s.l. account for approximately 25% of the area. The main characteristic of relief of north-western Wedel Jarlsberg Land described above (e.g. direction of ridges and valleyes) is determined by local geology (rock resistance, tectonic lines) (Birkenmajer 2004, 2010). Recently the landscape is modified by periglacial and glacial processes (Repelewska-Pękalowa & Pękala 2006; Reder 2006). Due to intensive frost weathering of mountain walls and gravitational processes, talus fans and rock glaciers are formed (Fig. 5.1.1).

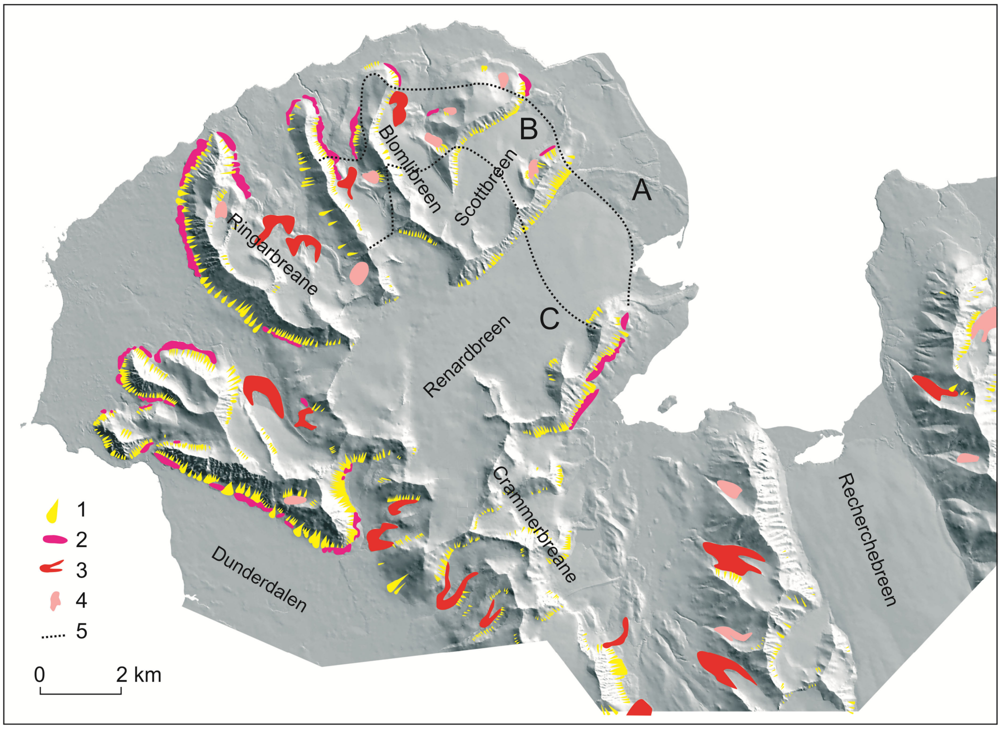


Fig. 5.1.1. Talus fans and rock glaciers in NW part of Wedel Jarlsberg Land (Dzierżek & Nitychoruk 1986; Nitychoruk & Dzierżek 1988a; background: Zagórski 2002): 1- talus fans: 2- subslope rock glaciers or nival moraines; 3- moraine rock glaciers; 4- cirque rock glaciers; 5- boundaries of morphoclimtic zones (A,B,C).

Studies of slope sediments and rock glaciers in the NW part of Wedel Jarlsberg Land were carried out in the years 1986-1987 (Dzierżek & Nitychoruk 1987ab, 1990; Nitychoruk & Dzierżek 1988abc) and revisited in 2007. Among the slope forms occurring in the area the most important ones are the talus fans. The data on their basic parameters such as inclination, exposure, longitudinal profile, petrographic composition of debris, roundness and microrelief, were studied in relation to the local geology. NW part of Wedel Jarlsberg Land is made up of Hecla Hoek rocks, i.e. metamorphic tillites with interbedding of quartzite sandstone and phyllites (Birkenmajer 2004). These rocks form a syncline, gently sloping in the north, that runs through the western part of the studied region, which has a big influence on the emergence of talus fans. On slopes conformable with the dip of the beds, talus fans occur sparsely and are poorly developed – e.g. on the eastern and south-eastern slopes near Scottbreen and Blomlibreen. On the other hand, on slopes positioned perpendicularly to the rock layers, talus fans are numerous and large, which can be observed on the western and north-western slopes near the glaciers Blomlibreen, Scottbreen (Photo 5.1.1) and Renardbreen (Szczęsny *et al.* 1989). Apart from the composition of rock layers, another factor that heavily influences the development of slope processes is the slope exposure. The microclimate conditions over slopes with a south-eastern and eastern exposure are usually more severe than in the case of slopes with a north-western and western exposure. This is mainly due to the influence of warm air masses flowing from the sea. Slopes with a south-eastern and eastern exposure, lying in the shadow of warm air masses, are characterised by higher temperature differences, particularly more frequent changes between sub-zero and above-freezing temperatures (Nitychoruk & Dzierżek 1988b). This intensifies the effects of frost weathering, which leads to faster breaking down of rocks that constitute these slopes, which in turn provides materials for talus fans. Warm air from above the sea influences the development of slopes with a north-western and northern exposure, ensuring higher thermal stability of soil, which limits the effects of frost weathering and the development of gravitational processes (Nitychoruk & Dzierżek 1988b; Bartoszewski *et al.* 2006). The specific topoclimate of glacier valleys in the region also results in uneven development of glaciers, especially in their front area, which is clearly observable in the case of, e.g. Scottbreen, Blomlibreen, or Tjørndalsbreen. What we see here is a feedback reaction, i.e. additional cooling influence of glaciers and intensification of gravitational processes on the slopes adjacent to them.

A detailed analysis of slope processes was carried out with regard to slopes surrounding the Renardbreen, Scottbreen and Blomlibreen (Nitychoruk & Dzierżek 1988abc). Most of the talus fans in this area are formed at the openings of gullies and rock chimneys, which most probably originate from impact fracturing and tectonic discontinuities (Photo 5.1.1A). This is another factor that determines the intensity of debris supply and, as a result, influences the formation of talus fans and debris covers. The fact that large amounts of snow accumulate in gullies and chimneys, combined with the fact that snow melts in the summer, leads to activation of rapid debris-flows and mud-flows. Corrasion channels created during debris-flows are often 1.5 m deep and up to 3 m wide, and end in fingerlike debris ramparts (Photo 5.1.1B). Similar accumulation of debris occurs after melt-out of snow delivered by avalanches. Some of the talus fans are mature and are usually covered in vegetation. The surfaces of these landforms are shaped also by solifluction processes.

In the north-western part of Wedel Jarlsberg Land the talus fans occur between the level of 200-550 m a.s.l. (fan summits) and 130-500 m a.s.l. (fan bases). The longest of these formations reach the length of 250 m. The width of fan bases ranges between 25 and 130 m, and is usually around 55-60 m. The inclination angle of fan surfaces is between 25° and 42°, with the average being 32°, and longitudinal profiles are significantly varied (Dzierżek & Nitychoruk 1987ac). The most common ones are convex profiles – over 50%, and straight profiles; with only a few concave profiles (Photo 5.1.1C). Older formations are accompanied by nival ridges and subslope rock glaciers – these, however, occur rarely (Fig. 5.1.2, Photo 5.1.2).

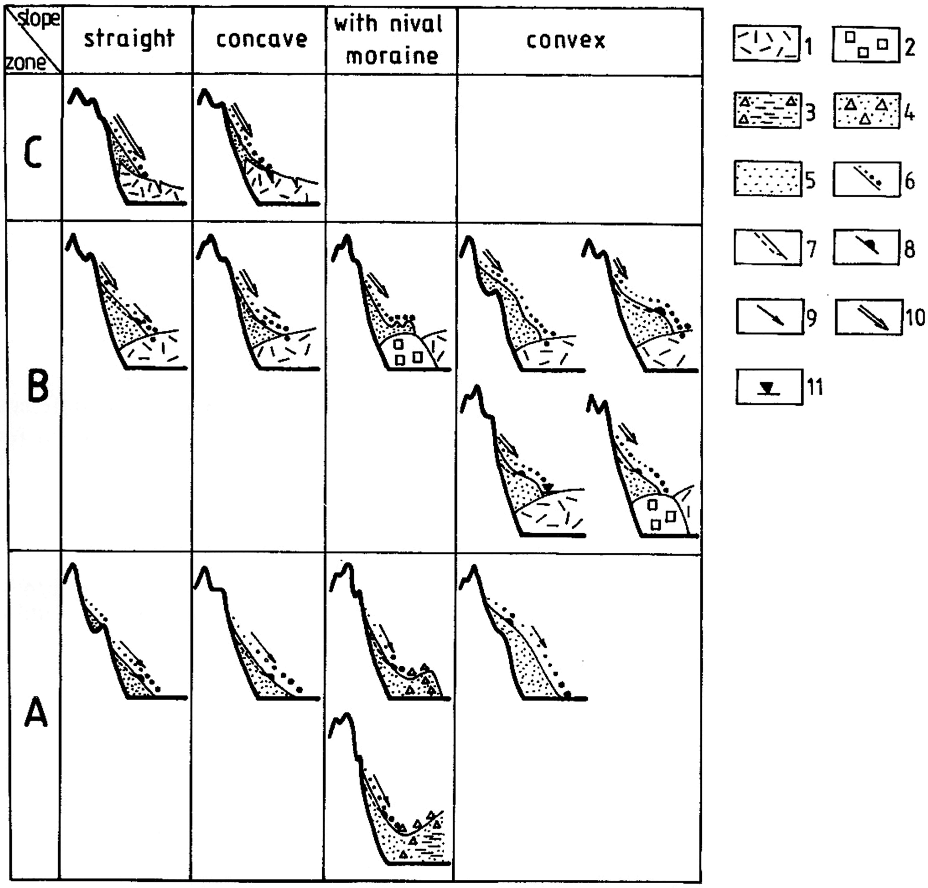


Fig. 5.1.2. Profiles and slope processes in northwestern Wedel Jarlsebrg Land dependent on morphoclimatic zones (Nitychoruk & Dzierżek 1988a). A- zone outside a direct glacier influence, B- zone influenced by glacier fronts, C- zone influenced by firn fields: 1- compact glacier ice, 2- lateral moraines, 3- rock glaciers, 4- nival moraines, 5- talus cones, 6- fractional segregation debris, 7- furrows of debris-mud flows, 8- bulges of mud-debris flows, 9- solifluction, 10- avalanches, 11- glacial streams.

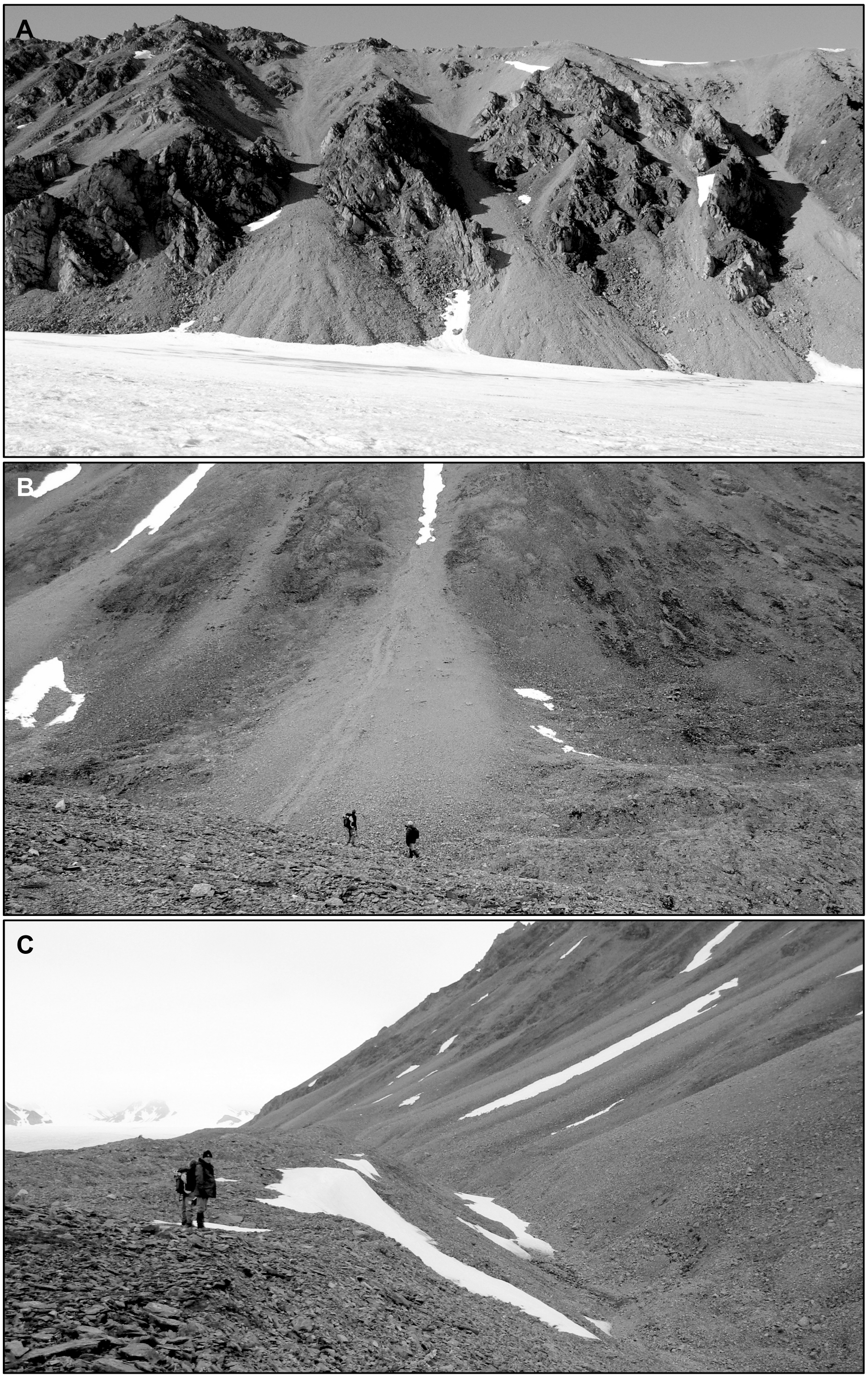


Photo 5.1.1. A- talus fans in Scottbreen area (Photo P. Zagórski 2006), B- corrasion channels in the Renard-breen area (Photo J. Nitychoruk 2007), C- talus fans with convex profiles in the Renardbreen area (Photo J. Nitychoruk 2007).

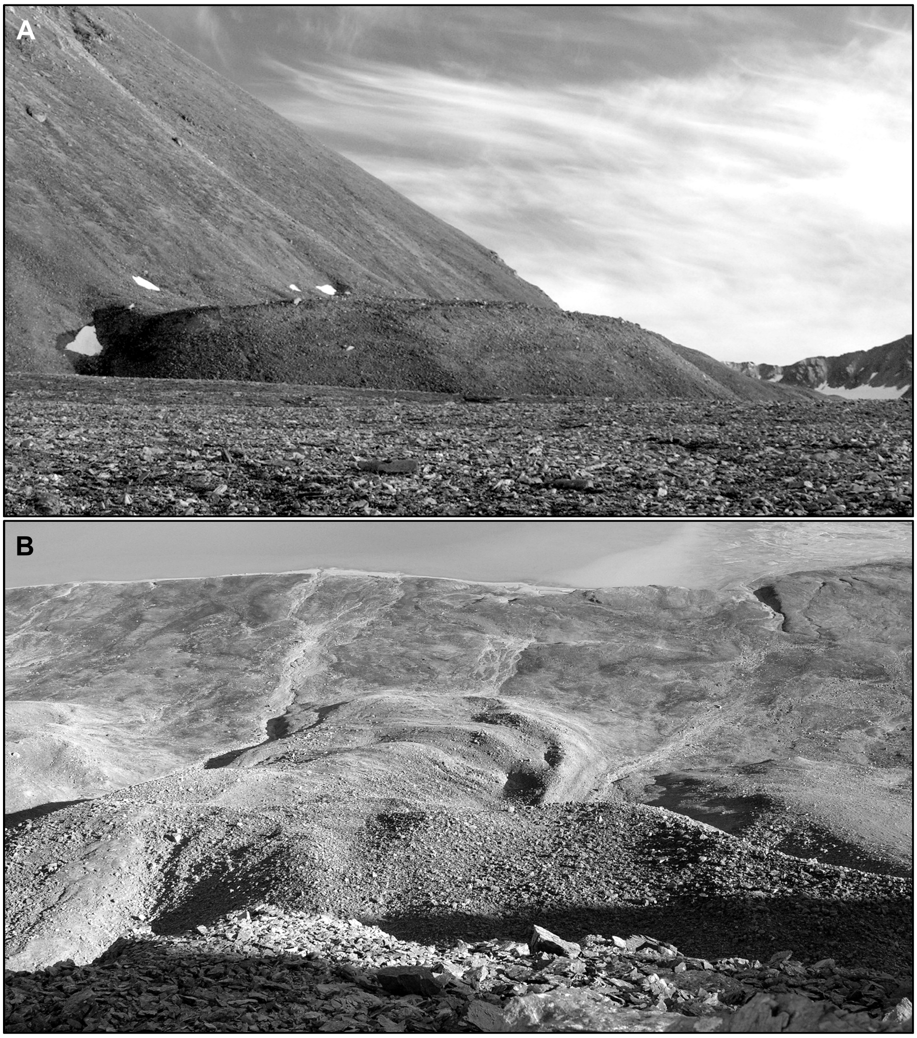


Photo 5.1.2. Subslope rock glaciers: A- in Dyrstaddalen (Photo P. Zagórski 2007), B- in Chamberlindalen (Photo P. Zagórski 2006).

Many factors affect the shaping of a talus fan's profile. The most important include: geological structure (lithological factors, layer composition, tectonic factors) and the shape of the bedrock, on which the fan is formed (so-called step slope, see: Jahn 1970); whether or not the fan is supported by a glacier, melting snow patches or a glacial moraine; possible undercutting of fan by glacial waters; and local microclimate conditions dependent on the exposure, altitude above mean sea level, or contact with the glacier's tongue (Photo 5.1.1C). Talus fans with a straight profile are characterised by an inclination angle equal to the angle of the natural state of rest of debris, which indicates that gravitational supply of debris is constantly taking place. Fans with a concave profile are formed as a result of remodeling of their surfaces by downflows. However, the profile of convex cones may result from undercut of its base by the flowing water, disappearance of the glacier support or the shape of the landform under the cone (Photo 5.1.1C).

In the north-western part of Wedel Jarlsberg Land there is a distinct interrelation between the process of forming talus fans and their contact with glacial tongues, firn fields or non-glacierised areas, i.e. morphological and microclimate conditions. Based on the aforementioned features, it is possible to distinguish three zones of talus fans occurrence in the region of the Renardbreen, Scottbreen and Blomlibreen. Zone A lies beyond the reach of glaciers, below 150 m a.s.l. Zone B is strongly related to glacier tongue and lies between 150 and 350 m a.s.l. Zone C, which covers the area above 350 m a.s.l., is affected by the influence of firn fields (Fig. 5.1.2).

Zone A includes mostly mature fans, with a straight or, less commonly, concave profile, often accompanied by subslope rock glaciers or nival moraines. These are made up of weathered debris and their surface is partially covered in vegetation that makes other formations stand out – e.g. tongue, solifluction terraces, or old erosion clefts. The highly weathered debris and the presence of solifluction forms indicates that current landforms are not being built-up very rapidly. This is also confirmed by lichenometric dating of boulders taken from the surface of the studied landforms as well as from subslope rock glaciers – according to the results the formation of landforms originated 3.5-2.0 ka BP (Dzierżek & Nitychoruk 1990) (Fig. 5.1.3).

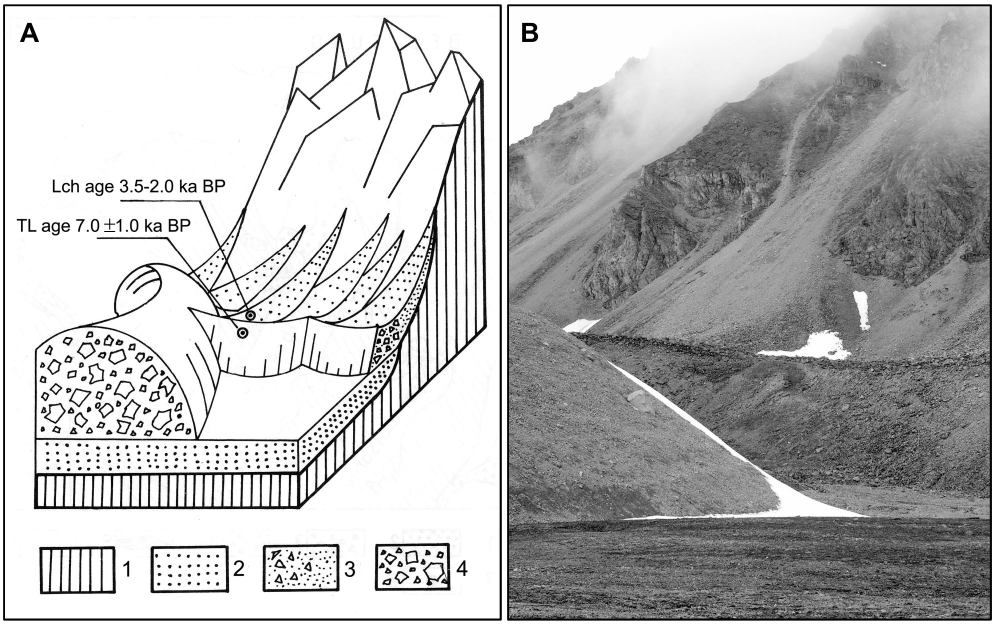


Fig. 5.1.3. A- location of the subslope rock glacier – Wijkanderberget (Lch 3.5-2.0 ka BP – lichenometric date; 7.0±1.0 ka BP –TL date) and the glacial moraine of Scottbreen from Little Ice Age (Dzierżek & Nitychoruk 1987a): 1- rock substrate, 2- deposits of the marine terrace od 70-80 (100) m altitude, 3- deposits of talus cones and rock glacier, 4- moraine deposits of Scottbreen; B- view of the sampling place for dating (Photo P. Zagórski 2006).

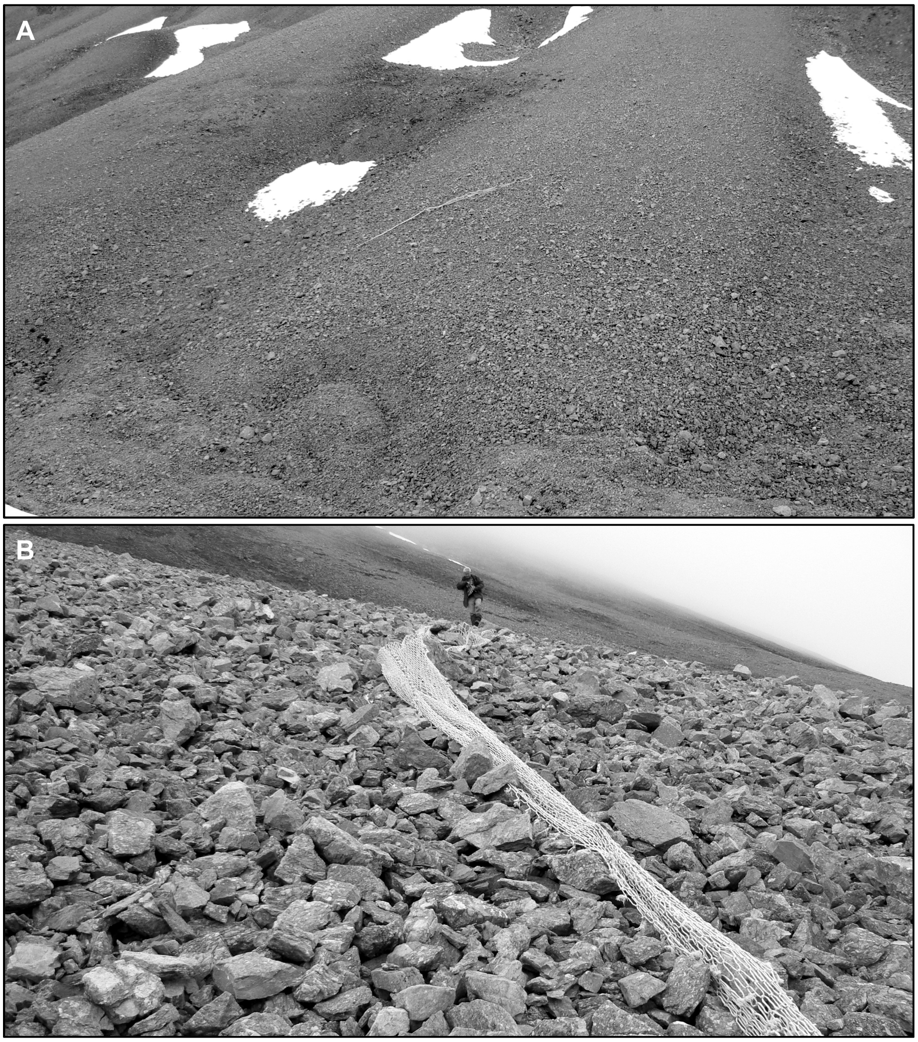


Photo 5.1.3. A, B- very dynamic debris movement have been observed in the years 1986-2007 by mounting fishing nets on the fans in the Renardbreen area (Photo J. Nitychoruk 2007).

Zone B is currently the location of most intense gravitational processes: rock-falls, debris-flows, mud-flows, deflation, solifluction, and frost weathering – which is understandable when taking into account the contact with the glacier (Nitychoruk & Dzierżek 1988abc). The processes that shape the surface of fans in this zone are often catastrophic. Very dynamic debris movements, accompanied by debris-flows, have been observed in the years 1986-2007 by mounting fishing nets on the fans, using painted markings and a series of temporary measurement benchmarks (see: Dzierżek &, Nitychoruk 1987b; Nitychoruk & Dzierżek 1988ab). After 21 years all of the bench-marks that had been mounted on the fans in zone B in 1986 were either displaced, de-stroyed or covered with debris (Photo 5.1.3AB). This indicates that the gravitational processes are highly dynamic and that the slope covers are being built-up successively.

André (1986) pointed our attention to stress relaxation of mountain rocks in mountain massifs, associated with large debris supply, while Jahn (1976) believed that permafrost is the crucial factor in the shifting of rock debris. The talus fans in this zone are made of 'fresh' non-weathered, poorly selected debris.

Zone C has very few talus fans, because the intensity of frost weathering processes is severely inhibited above the snow line (Dzierżek & Nitychoruk 1987b). The only formations present here are initial forms, characterised by low activity, built-up by snow and stone avalanches. Such a low activity of gravitational processes also stem from the fact that in the surroundings of the firn fields there are no vast rocky surfaces with long slopes (Fig. 5.1.2).

At the beginning of August 1986, in order to determine the dynamics of gravitational processes the authors set up several measuring stations in zone A and B. However, after 3 weeks of observations no obvious traces of slope relief changes were noted. In our opinion the optimal time for the development of these processes is therefore the turn of Arctic spring and summer, when the ablation of the snow cover and frozen ground starts. According to the observation of Ph.D. Jan Rodzik (personal report) in the summer, most intensively are shaped cones in zone C. In zones A and B, the activity of the slope affected by heavy snow or rain during the summer season and early autumn (Figs. 5.1.1 and 5.1.2).

According to André (1986), the last significant stage of forming talus fans in Spitsbergen took place during the Little Ice Age. However, our observations from the past 21 years indicate that the activity of processes that form talus fans depends on the specified morphological/climate zones, as in each of these zones the pace of slope processes is different. The analysis of the ortophotomap (Appendix 1, Zagórski 2005) and aerial photographs of the north-western part of Wedel Jarlsberg Land shows that the pattern of slope processes activity is also applicable to other glacier-occupied valleys. Fans are formed slightly differently in large deglacierised valleys. The best developed fans are found on the slopes of Dunderdalen – they are the largest in size, usually mature and formed in large part due to surface water flows. Some of these are talus-alluvial fans. Their heads often protrude beyond the base of the slope, resting on higher marine terraces (Fig. 5.1.1).

During the expedition to Spitsbergen in 1986 a number of rock glacier were studied, what formed a basis of their new classification (Dzierżek & Nitychoruk 1986, 1987c) based on the ones developed by Wahrhaftig and Cox (1959), Nemčok & Mahr (1974), Johnson (1980), Serrat (1979), Humlum (1982) and Lindner & Marks (1985). The basic criterion of modified classification was the location of these landforms, which at the same time focused on their structure and origin (Fig. 5.1.1). In the north-western part of Wedel Jarlsberg Land on Spitsbergen the following types were distinguished: moraine rock glaciers, cirque rock glaciers and subslope rock glaciers (Fig. 5.1.4).

Moraine rock glaciers can be found in the surroundings of frontal moraines e.g. in the valleys of Lognedalen, Dyrstaddalen, Tjørndalen, Blomlidalen, Chamberlindalen, or Dunderdalen. Taking into account the manner of reshaping the end moraines, i.e. either partial or complete, two types of moraine rock glaciers can be distinguished. The first type includes glaciers with partially reshaped end moraines in the valleys of Dyrstaddalen, Tjørndalen and Blomlidalen, where the rock glacier lobe either bends or cut through the end moraine (Figs. 5.1.1 and 5.1.4). All the remaining landforms belong to the second type of moraine rock glacier, in which a series of end moraines were completely reshaped by the rock glacier formation and movement. Moraine rock glaciers of the first type form a massive, lobe-shaped rampart with a height of 30-40 m onto outwash plains located in front of modern glaciers fronts. For instance the length of moraine rock glacier lobe in Dyrstaddalen reaches 400 m, and its width – 200 m. The surface of these landforms includes numerous hillocks and hollows, sometimes filled with snow or water. In some places they form a system of similarly-shaped lobes, whose appearance reflects that of the first and largest lobe – this indicates that the forms are in motion. The ramparts are made of rock blocks with a size of 20-50 cm, sometimes reaching 5 m in diameter. Finer debris (diameter of 2-20 cm) can be found in the depressions between respective ramparts. The sandy and silty material is washed into deeper parts of the landform (Fig. 5.1.2, Photo 5.1.4).

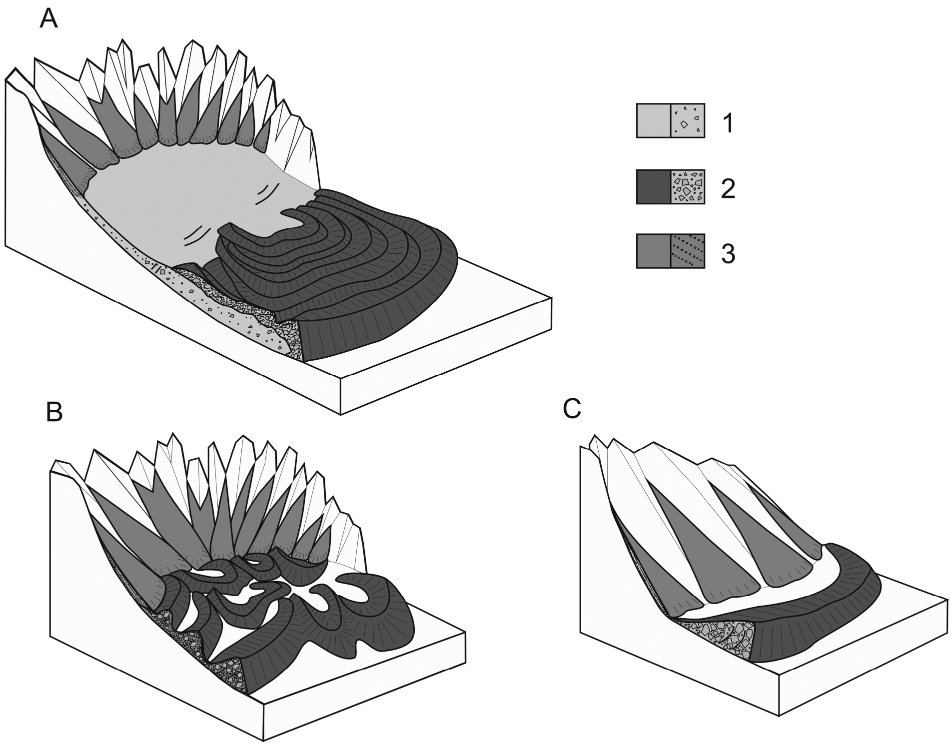


Fig. 5.1.4. A- structure of moraine rock glacier; B- structure of cirque rock glacier; C- structure of subslope rock glacier; 1- glacier ice; 2- moraine glaciers; 3- talus fans.

Photo 5.1.4. Moraine rock glacier – Bøcmanbreen, Chamberlindalen (Photo M. Świtoniak 2009).

Moraine rock glaciers of the second type are also connected with frontal moraines, but only those belonging to relatively small glaciers that fill narrow and short valleys surrounded by high rockwalls. Such a morphological situation is conducive to high supply of rock debris. High inclination of the surface, on which the moraine rock glacier travels, results in the creation of debris ramparts placed lengthwise in relation to the valley. The structural material of these ramparts includes similar fractions to those occurring in moraine rock glaciers of the first type. It is common for such rock glaciers to be split by a stream that allows for water runoff from the glacier located up the valley. In such situations, the stream can uncover old glacial ice filled with rock material from under a several-meter thick layer of debris.

The core of all moraine rock glaciers is usually made of ice (Fig. 5.1.4). The activity of these landfroms depends mainly on the supply of rock material, which in turn depends on the size of co-occurring glaciers and the pace of slope processes on the surrounding rockwalls.

Cirque rock glaciers occur in the higher parts of slopes of the following mountain valleys: Dyrstaddalen, Tjørndalen, Blomlidalen, Chamberlindalen and Dunderdalen and have no linkage to the system of currently glacierised valleys (Dzierżek & Nitychoruk 1987c) (Fig. 5.1.1). A characteristic landscape feature associated with glacial cirques are the steep talus fans, whose inclination often exceeds 35°, and debris ramparts located at the bottom of cirques or at their entrance. The direction of some ramparts is conformable to the cirque axis, others mimic the shape of the slopes, and yet others are lobe-shaped. The relative height of these forms reaches 30 m. They are made of debris, with the dominant fraction size being 50-100 cm, although there is also a large proportion of smaller-sized debris, particularly in the depressions between the forms. Respective ramparts or systems of ramparts lie at different heights and in different locations at the bottom of a cirque, and sometimes extend beyond the cirque threshold. The inclination of the outer edge of a rampart is approximately 37° (Fig. 5.1.4).

The origins of glacial cirques are connected with the accumulation of debris on the surface of decaying cirque glaciers and the shifting of debris along the surface or within the ice (Fig. 5.1.4). It is possible that the accumulation of snow in the debris-covered bottoms of cirques, and its later transformation into ice, facilitates further shaping of these forms. One of the conditions for the formation of cirque rock glaciers is that the intensity of debris supply needs to be higher than the intensity of accumulation of firn and ice in cirques. Two types of cirque rock glaciers were distinguished: active and fossilised. Active ones include rock ramparts in the threshold areas of vanishing firn fields. Fossil cirque rock glaciers fill the cirque floors devoid of pure ice. The degree of debris weathering, partial moss and lichens cover and the lack of movement all indicate that these features are fossil. The existence of ice below or within the debris cover is probable (Photo 5.1.5).

Third type of rock glaciers in the study area are the subslope rock glaciers formed at the base of mountain slopes covered by talus fans (Fig. 5.1.4). They can be several kilometers long, up to 50 m high and reach a maximum width of several hundred meters (Fig. 5.1.1), particularly in the valleys of Lognedalen, Dyrstaddalen, Tjørndalen and Dunderdalen (Dzierżek & Nitychoruk 1987c). Their surface is uneven, with several crests, depressions and flat areas. Height differences reach up to 10 m. They are separated from the slopes by undrained depressions, which results in a characteristic slope profile. While the average slope inclination is 32°, the outer edge inclination of the ridge is 40°. The point of contact with the base of the slope is clearly distinguishable. Ridges of subslope rock glaciers are composed of coarse and angular rock material. On the surface, almost exclusively, weathered blocks of 0.5-3 m in diameter are found. Finer fractions are accumulated at the bottom of undrained depressions. In most of these depressions snow patches survive throughout the entire summer season. They are often wet and overgrown with moss. Between the sharp edges of external subslope ridges it is possible to see debris with a fraction size of 20-30 cm and smaller. Most probably, the formation of subslope rock glaciers is related to the intensive accumulation of debris in the lower sections of slopes and subsequent movement of debris with the interstitial ice. Debris material coming from slopes and talus fans is unable to move on flat areas and builds up vertically. Snow or water accumulate in empty spaces within the debris, and in time turn into ice (Fig. 5.1.4). For some period the ice-cemented debris cannot move and, form wide shelves in the lower parts of slopes. Only once the critical thickness is exceeded debris start to be plastically remodeled. As a result, the ice-and-debris ridge slides down the slope, without losing its ability to move even on the flat surface at the foot of the slope. The different shapes of these features indicate that the process is not uniform and takes place in stages. Subslope rock glaciers are either arched or wave-shaped, and the height of their lobes is also different (Fig. 5.1.4).

Photo 5.1.5. Cirque rock glacier in the Scottbreen area (Photo J. Nitychoruk 2007).

In these types of landforms the outer ridges contain clefts, up to 1.5 m deep and 3 m wide, which suggest with the presence of ice cores. Most probably the presence of ice blocks the slope processes taking place on the steep outer slope of a ridge. As a general rule, on the outer slope of a debris ridge, approximately 2-3 m from the top, there is a slightly darker strip, about several dozen centimeters wide. This strip is saturated with water and is created due to thawing of permafrost within the ridge.

Another landform, similar to subslope rock glaciers, are nival moraines (Dzierżek & Nitychoruk 1987a). They are smaller in size, show no signs of movement and form close to talus fans (Fig. 5.1.1). These features can be found in many valleys, such as: Lognedalen, Tjørndalen, Blomlidalen, Chamberlindalen, Dunderdalen (Dzierżek & Nitychoruk 1987c).

The age of rock glaciers of the north-western part of Wedel Jarlsberg Land was determined using TL dating and lichenometric dating (Dzierżek & Nitychoruk 1986, 1990). It was estimated that the main stage of the formation of fossil cirque and subslope rock glaciers occurred around 8.0-7.0 ka B.P (Fig. 5.1.3). The renewal of rock glaciers formation happened in the period between 3.5 and 2.0 ka BP, during which a significant cooling of the climate took place, which resulted in the increased efficiency of frost weathering processes and led to the supply of large amounts of rock material to the glacier surface (Denton & Karlén 1973).

# Streszczenie

## Orografia i procesy stokowe

Północno-zachodnia część Wedel Jarlsberg Land, położona w rejonie zachodniego wybrzeża Spitsbergenu, na południe od cieśniny Bellsund, ma cechy górskiej rzeźby glacjalnej, ze stromymi graniami, rozdzielającymi doliny wypełnione lodowcami. Najwyższy szczyt, położony w głównym grzbiecie oddzielającym Wedel Jarlsberg Land od Torell Land – Berzeliustinden – osiąga 1205 m n.p.m., jednak w obszarze badań, ograniczonym od wschodu przez Recherchebreen i Recherchefjord, zaś od południo-zachodu przez Dunderdalen, najwyższym szczytem jest Storgubben o wysokości 831 m n.p.m. Nisko położone, stosunkowo równinne tereny: dna dolin opuszczonych przez lodowce i wyniesione terasy morskie, występują na obrzeżu tego obszaru.

Ukształtowanie powierzchni terenu wynika zasadniczo z przebiegu struktur tektonicznych oraz odporności skał. Główne elementy rzeźby strukturalnej były modelowane przez procesy glacjalne, peryglacjalne, fluwialne, abrazyjne i stokowe, tworząc urozmaiconą rzeźbę powierzchni tego terenu. Silnie nachylone zbocza górskie są głównie polem działania wietrzenia mrozowego i procesów grawitacyjnych, zaś powierzchnie o małym nachyleniu noszą piętno soliflukcji. W efekcie tych procesów powstały między innymi takie charakterystyczne formy rzeźby jak: stożki usypiskowe i lodowce gruzowe.

W północno-zachodniej części Wedel Jarlsberg Land widać wyraźną zależność między wykształceniem stożków usypiskowych, a kontaktem ich – lub brakiem kontaktu – z jęzorami lodowcowymi i polami firnowymi, stwarzającymi szczególne warunki morfologiczne i topoklimatyczne. Na tej podstawie można wyróżnić trzy strefy występowania stożków usypiskowych. Strefa A jest poza zasięgiem lodowców, poniżej 150 m n.p.m.; strefa B jest ściśle związana z jęzorami lodowcowymi i znajduje się między 150 m a 350 m n.p.m., zaś strefa C, którą można wyróżnić powyżej 350 m n.p.m., sąsiaduje z polami firnowymi. Nieco inaczej wykształcone są stożki w dużych dolinach niezlodowaconych. Mają one znacznie większe rozmiary i najczęściej są formami dojrzałymi, ukształtowanymi przy dużym udziale spłukiwania.

Lodowce gruzowe, występujące w tym obszarze, dzielą się na trzy grupy: morenowe, cyrkowe i podstokowe. Główna faza ich powstania miała miejsce w okresie 8 000-7 000 ka BP. Odnowienie lodowców gruzowych nastąpiło w okresie między 3 500-2 000 ka BP., kiedy to doszło do znacznego ochłodzenia klimatu, skutkującego intensyfikacją wietrzenia mrozowego i dostawą dużej ilości, budującego je materiału skalnego. Morenowe i cyrkowe lodowce gruzowe podlegają nieznacznemu przekształcaniu również obecnie.

# Objaśnienia

### Ryciny

Ryc. 5.1.1. Stożki usypiskowe i lodowce gruzowe w NW części Wedel Jarlsberg Land (Dzierżek, Nitychoruk 1986; Nitychoruk, Dzierżek 1988a; podkład: Zagórski 2002): 1- stożki usypiskowe, 2- podstokowe lodowce gruzowe i moreny niwalne, 3- morenowe lodowce gruzowe, 4- cyrkowe lodowce gruzowe, 5- granice stref morfoklimatycznych (A, B, C).

Ryc. 5.1.2. Profile i procesy stokowe w NW części Wedel Jarlsberg Land w poszczególnych strefach morfoklimatycznych (Nitychoruk, Dzierżek 1988a): A- strefa poza zasięgiem bezpośredniego wpływu lodowca; B- strefa wpływu jęzora lodowcowego; C- strefa wpływu pola firnowego: 1- lód lodowcowy, 2- moreny boczne, 3- lodowce gruzowe, 4- moreny niwalne, 5- stożki usypiskowe, 6- frakcjonowanie gruzu, 7- rynny spływów gruzowo-błotnych, 8- jęzory spływów gruzowo-błotnych, 9- soliflukcja, 10- lawiny, 11- potoki proglacjalne.

Ryc. 5.1.3. A- położenie lodowca gruzowego – Wijkanderberget (Lch 3,5-2.0 ka BP – data lichenometryczna; 7,0 ± 1,0 ka BP – data TL) i moreny Scottbreen z Małej Epoki Lodowej (Dzierżek, Nitychoruk 1987a): 1- podłoże skalne, 2- osady terasy morskiej 70-80 (100) m, 3- osady stożka usypiskowego i lodowca gruzowego, 4- osady moreny Scottbreen; B- widok na miejsce poboru prób do datowań (fot. P. Zagórski 2006).

Ryc. 5.1.4. A- budowa morenowego lodowca gruzowego, B- budowa cyrkowego lodowca gruzowego, C- budowa podstokowego lodowca gruzowego: 1- lód lodowcowy, 2- moreny i lodowce gruzowe, 3- stożki usypiskowe.

**Fotografie**

Fot. 5.1.1. A- stożki usypiskowe w otoczeniu Scottbreen (fot. P. Zagórski 2006), B- rynny spływów w okolicy Renardbreen (fot. J. Nitychoruk 2007), C- stożki usypiskowe o wypukłym profilu w okolicach Renardbreen (fot. J. Nitychoruk 2007).

Fot. 5.1.2. Podstokowy lodowiec gruzowy: A- w Dyrstaddalen (fot. P. Zagórski 2007), B- w Chamberlindalen (fot. P. Zagórski 2006).

Fot. 5.1.3. A, B- na pułapkach z sieci rybackich zaobserwowano w latach 1986-2007 skutki dynamicznego przemieszczania gruzu po powierzchniach stożków usypiskowych w okolicy lodowca Renerdbreen (fot. J. Nitychoruk 2007).

Fot. 5.1.4. Morenowy lodowiec gruzowy – Bøcmanbreen, Chamberlindalen (fot. M. Świtoniak 2009).

Fot. 5.1.5. Cyrkowy lodowiec gruzowy w okolicy Scottbreen (fot. J. Nitychoruk 2007).