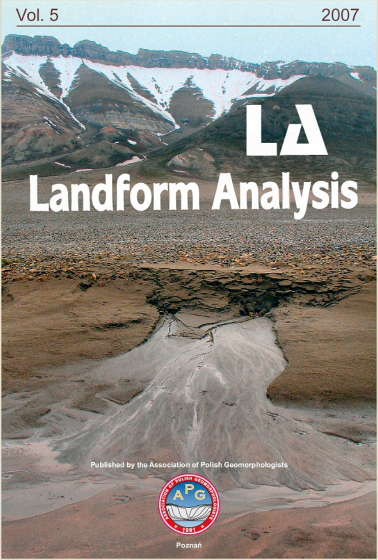
Vol. 5 2007

LA Landform Analysis



Published by the Association of Polish Geomorphologists

Poznań

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Poznań, Dzięgielowa 27, 61-680 Poznań, Poland www.sgp.org.pl, sgp@sgp.org.pl

Information from the Executive Committee of the Association of Polish Geomorphologists

Dear Readers, Fellow Geomorphologists,

Landform Analysis, a periodical of the Association of Polish Geomorphologists (APG) established in 1997, is changing its formula. Despite hard work and efforts from the editors, especially Professor Jacek Jania and the Editorial Board, which are gratefully acknowledged, it proved hardly possible to maintain *Landform Analysis* as a regular journal and only a few issues have appeared. Following the discussions among the members of the Executive Committee of the APG, we decided to adopt a model of thematic volumes. Each volume will consist of carefully selected papers on a given theme, and will be edited by a guest editor approved by the Executive Committee and the Editor-in-Chief of *Landform Analysis*. Occasionally, extended abstracts and papers introducing a particular geomorphological area may be considered. Each volume will undergo rigorous process of peer-reviewing, so that the high quality of the journal is maintained. The numbering of the issues will be retained, but there is flexibility concerning the language, which may be either English or Polish.

The present *Landform Analysis* issue no. 5, the first one under the new formula, appears in association with the Regional Conference of the International Association of Geomorphologists, organized by the Association of Polish Geomorphologists in Longyearbyen, Svalbard. Three further issues are planned for the next two years and we envisage publication of at least two issues per year. Therefore, further initiatives from fellow geomorphologists are welcome and will be considered by the Executive Committee of the APG. Potential guest editors will be provided editorial requirements upon request.

We hope that the new model will prove successful and *Landform Analysis* re-establishes itself as a widely known and highly regarded addition to the range of existing geomorphological publications.

President
of the Polish Association of Geomorphologists

Dr hab. Piotr Migoń

Cover photo: South side of Ebbadalen, Spistbergen (photo Zb. Zwoliński 2006)

Image from the World

Rocky (limestone) stream channel with cylindrical microforms in Mawsmai village, near Cheerapunji, Meghalaya, east India (photo Zb. Zwoliński, 2006)

Bogucki Wydawnictwo Naukowe, ul. Górna Wilda 90, 61-576 Poznań, Poland
tel. +48 61 8336580, fax +48 61 8331468, e-mail: bogucki@bogucki.com.pl, www.bogucki.com.pl

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Foreword

The operation of present-day and ancient geo-ecosystems of the Arctic polar zone is a subject of research under a great number of international projects. The regularities found in the contemporary development of relief in the polar zone provide a good basis for paleogeographic reconstructions and forecasting studies concerning the nature of landscape change in the Earth's surface at a variety of spatial and temporal scales.

The region around the North Pole embraces mostly frozen sea basins and archipelagos of glaciated islands featuring oases and coastal plains with marine terraces as well as rock and ice cliff shores. The Arctic occupies a special place in the international scientific discussion of changes in the individual elements of the Earth's surface induced by global climate change.

The Svalbard Archipelago, especially its largest island of Spitsbergen, is a unique area with all kinds of glaciers, a morpholithological diversity of marginal zones, inland and coastal oases, different types of slopes, an abundance of proglacial streams of various courses and level of formation, extensive periglacial plains, as well as diversified coasts with fjords and rock and ice shores.

Spitsbergen is a site of international interdisciplinary studies to which Polish polar research has made a substantial contribution. Established 50 years ago, the Polish Polar Station at Hornsund offers an excellent opportunity for research. The multi-year studies carried out in different areas of Spitsbergen by teams from a variety of academic centres in Poland have contributed a lot to our understanding of the geographical environment of this highly interesting place. The Geomorphological Workshop (2003) and the Glaciological Workshop (2004) organised in Spitsbergen by the Association of Polish Geomorphologists were good occasions for a field discussion and a survey of the research done on the past and present morphogenetic environment of Spitsbergen.

The 4th International Polar Year 2007/2008 provides an excellent opportunity for an overview of polar research and an appraisal of its cognitive and application merit. Special attention should be paid to the formulation of new research tasks to be undertaken in international co-operation.

The IAG/AIG Regional Conference on Geomorphology "Geodiversity of polar landforms" (Longyearbyen, 1–5 August, 2007) being now organised by the Association of Polish Geomorphologists is part of the International Polar Year celebrations. The present volume of *Landforms Analysis* contains its reviewed papers and posters as well as original descriptions of field trip sites within selected Spitsbergen areas.

Polar regions, including Spitsbergen, have been undergoing ever greater transformations; the range of glaciated areas keeps shrinking, not only as a result of global climate change, but also of man's multi-directional activity. We are convinced that the Spitsbergen meeting is going to be a good opportunity to experience the natural uniqueness of the polar landscapes and to discuss the character and current state of glaciation of the various parts of the world.

Andrzej Kostrzewski, Zbigniew Zwoliński

Guest Editors

Landform Analysis, Vol. 5: 5–8 (2007)

The functioning of Scott Glacier in conditions of climate global changes

Stefan Bartoszewski

Department of Hydrology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland

Andrzej Gluza, Krzysztof Siwek

Department of Meteorology and Climatology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland

Piotr Zagórski*

Department of Geomorphology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland

Introduction

North-western part of Wedel Larlsberg Land has become the site of comprehensive natural studies since 1986. They were connected with polar expeditions of scientists from Maria Curie-Skłodowska University in Lublin. At the study area, there have been systematic meteorological and topoclimatical measurements made and a review of hydrographical cartography during which the identification of way of alimentations of rivers were made, the extend of current glaciers and waterlogged areas and some patrol checks of water flow. Cartographical results were the basis of the choice of the drainage basin for further stationary studies. Among the group of glacial drainage basins as the basic object for studies the Scott Glaciers was chosen (Fig. 1).

A special attention in hydroclimatical research was paid to conditions of outflow, including the size and spatial diversity of ablation of the Scott Glacier (Bartoszewski et al. 2003). Possibilities of making those studies were seriously limited by their expeditionary character. It was impossible to observe some phenomena and processes systematically in full year periods. However the extensive material of measurement from ten-year period was collected. The length of the series was related to time-limits of beginning and finishing the expeditions. The range of time when the studies were conducted before the first decade of June and October. The basic material was consisted of hydrological data (water level in watercourse, flows in water-gauges and patrol measurements of flow) and meteorological (like temperature and air humidity, cloudiness, speed and direction of wind, fall and ground temperature at different depths).

Characteristics of study area

The drainage basin of Scott Glacier comprises of the area of 10.1 km^2 with the part of maritime plain (Calypsostranda) and partly glaciated mountain valley. The glacier covered 4.5 km^2 in 2006 which was 44.4% of the whole area. That type of valley glaciers is alpine and it fills almost the whole mountain valley surrounded by Bohlinryggen i Wijkanderberget (Fig. 1). The highest parts of the glacier reach 600 m a.s.l., and the front meets the height of 90 m a.s.l. The length of the glacier is 4 km, width of 1.1–1.8 km, and the mean decrease of the longer axe is 8°.

The Scott Glacier has well developed drainage system which brings water flowing down the surface, inside and under it. The system of drainage of the Scott Glacier is made by subglacial, inglacial and supraglacial streams of different size. In some exam-

* e-mail: pzagorsk@biotop.umcs.lublin.pl

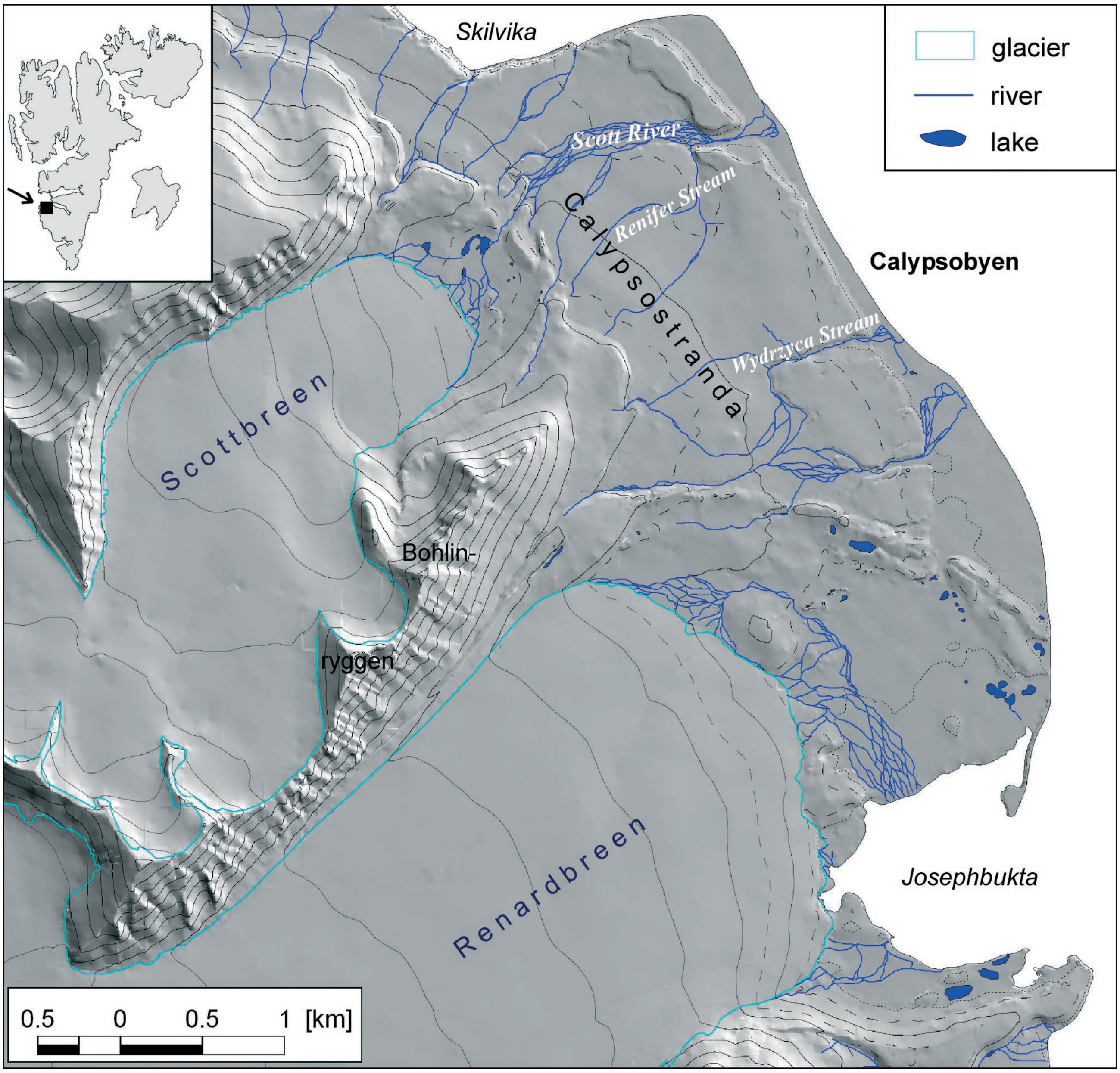

Fig. 1. Localisation of the study area (Zagórski 2002). The shadow relief map made on the basis of digital terrain model from the aerial photos of 1990 (Zagórski 2002)

ples the streams that cut the surface of the glacier vanish in glacial wells. Most of them group at the height of about 300 m a.s.l. in the zone of strong deformation of the bottom of the glacier. There is the rocky step linking two neighbouring mountain massives: Bohlinryggen and Wijkanderberget. A similar situation was also found in neighbouring valleys covered by Blomli i Tjørn Glaciers (Bartoszewski 1998). The system of drainage of the Scott Glacier can be considered as stabile and easy according to temporary studies.

The Scott Glacier, like most of Spitsbergen glaciers is at the stage of recession because of the removal of the fronts and changes of their longitudinal sections.

Being based on archival materials and GPS measurements, the recession of the Scott Glacier was characterised. It was stated that the intensivity of those processes is diverse in time. The information gathered during the analysis of changes of the range and geometry of that glacier confirm the negative balance of mass. In 1987–2002 the glacier was removed with the speed of 30 m a year, and its surface was being gradually lowered (Zagórski, Bartoszewski 2004). In the summer season of 2006 (3rd July – 31st August) the lowering of the glacier surface at the front was almost 2.5 m, and 1.0 m in the firn field. It was closely related to meteorological conditions.

Lowland part of the basin is built by lifted sea terraces of diverse height (Zagórski 2002). At the foreground, they were cut by erosion and aggradated with sandurs. The Scott River starts from a marginal lake that was created in an end depression between the present front of the glacier and the front moraine, the waters from the side cracks, supraglacial outflow and numerous subglacial streams flow right into that depression. The Scott Glacier has one subglacial channel located in the central part of the glacier. A structure of the outflow, understood as a proportion of components in total outflow is a changeable phenomenon depending on time. In the first half of the polar summer the outflow of the origin of in-glacier component was much lower than the mean what was the result of intensive ablation of superficial snow cover in the lower and middle part of the glacier.

Through the gorge of the moraine ridge, glacier water flows onto sandur where the river makes wide braided system. That system means considerable stability at least during some years. Easter part of the drainage basin is drained by little streams of snow-permafrost origin of supply. The biggest of them is The Renifer Stream (Fig. 1). A water-gauge in the Scott River is located 200 m above the mouth to a fiord, in the middle part of the gorge. The length of the river to that place is 2.56 km and the mean decline is 35‰.

Results and discussion

Hydrographical studies in the Bellsund regions allow us to state that so called active hydrological period, with the phenomenon of river outflow, last for about four months. It begins in June with spring freshet caused by melting out of snow cover on maritime plains, and finishes at the beginning of October with the end of melt-out of glaciers and permafrost. Processes of polar drainage basins show clear seasonal differentiation thanks to which it is possible to distinguish some periods of special hydrological features. The detailed characteristic of that phenomenon has been presented in publications of Paulina's team (Pulina et al. 1984, Pulina 1986).

It is characteristic for glacial rivers during polar summer, that they have huge dynamic of outflow. It can be one supporting example – presence of violent freshet of ablation-rain origin. The single outflows during the freshet cumulation of the Scott River exceed 1200 dm^3 s^{-1} km^{-1}. Specific feature of glacier rivers during rainless periods is twenty-four hours rhythm of flow as the effect of thermal and ablation cycle (Bartoszewski et al. 2006).

Hydrographical and climatological studies were conducted in 2006, between 1st July and 31st August. During that period of studies, the mean twenty-four hours air temperature was 4.9°C. The highest mean temperature was +7.8°C on 23rd July, the lowest was 2.8°C on 31st August. Absolute maximum air temperature at the height of 200 cm was 10.1°C on 23rd July, while the absolute minimum was 0.0°C on 31st August. Total of meteorological fall for the whole measurement period was 41.7 mm. The number of days with the fall was 29 (it is 47% of all days) with 11 days with the fall ≥1.0 mm. The highest fall was 8.6 mm on 25th August (Fig. 2).

The volume of outflow in summer season (6.4 mln m^3) was similar to one recorded in previous years (Bartoszewski et al. 2006). The mean flow was 1201 dm^3 s^{-1}, what is correspondent to specific runoff index of 118,6 dm^3 s^{-1} km^2. That time the runoff was 636 mm. The studied period was characteristic of no major freshets. Flows range was between 574 and 2719 dm^3 s^{-1} (Fig. 2).

The biggest freshet was on 26th August – the day after the highest fall. All rainfalls of twenty-four hours totals over 1.0 mm influenced the rise of flows.

The lowest flow of the Scott River on 31st August was connected with the fall of air temperature. The previous day and the day of the minimum flow, the temperature below zero (-1,5°C) was measured near the surface of the ground.

Twenty-four hours rhythm of discharge shows the influence of twenty-four hours run of air temperature at the height of 200 cm a.g.s. (Fig. 3). The maximum flow of the Scott River is about four hours after the thermal maximum.

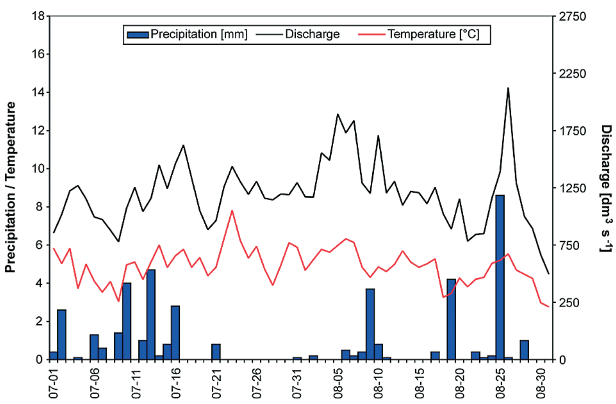

Fig. 2. The discharge of the Scott River, the precipitation and the temperature in Calypsobyen in 2006 summer season

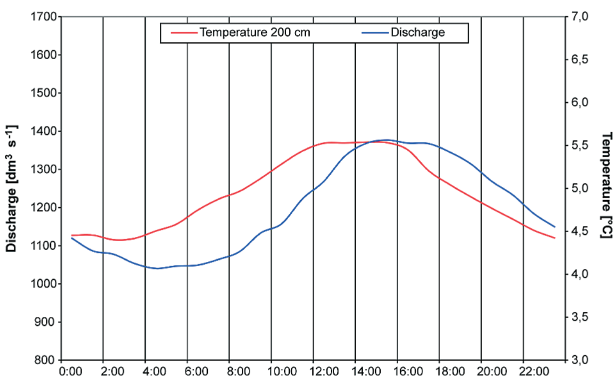

Fig. 3. Twenty-four hours rhythm of the discharge of the Scott River and air temperatures (200 cm above ground level)

Summary

Expeditionary character of hydrographical studies makes difficult the evaluation of the magnitude of the mean outflow from the analysed drainage basin. The registration of the outflow contained only part of the active hydrological period and the length of measurement series at certain years was diverse. On the basis of a series of ten summer seasons the factor of the outflow can be evaluated at about 900 mm.

The regime of the outflow of the Scott River is conditioned by the set of factors of which the most important is meteorological conditions, especially of an course of air temperature and precipitation in accumulation and ablation seasons. The analysis of meteorological conditions results that all freshets of the Scott River in the season of 2006 were of rain-ablation character.

The character of collected data makes difficult to do the equation of the water balance for the Scott drainage basin. Analysing the changes of geometry of the glacier it can be supposed that the size of the outflow exceeds the fall and the total is negative for years.

Acknowledgment

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Landform Analysis, Vol. 5: 9–10 (2007)

Quantitative studies on sediment fluxes and sediment budgets in changing cold environments – potential and expected benefit of coordinated data exchange and the unification of methods

Achim A. Beylich*

Geological Survey of Norway, Landscape & Climate group, Trondheim, Norway
Norwegian University of Science and Technology, Department of Geography, Trondheim, Norway

Polar and mountainous regions are among the most sensitive regions to climate change. The European Science Foundation (ESF) SEDIFLUX Network has been analysing the impact of climate change on landforms in high-latitude and high-altitude cold environments, via the mobilisation, movement and deposition of sediments by slope processes, rivers, glaciers, coastal processes and wind.

SEDIFLUX has over the last years evolved into a coordinated multidisciplinary and multinational effort to monitor the changing structure of landforms in cold environments and has led to a series of coordinated research initiatives. The efforts conducted within the SEDIFLUX Network were urgently needed, given the critical importance of studying the impact of projected climate change on the land structure of such sensitive environments. There is especially strong synergy with the International Tundra Experiment (ITEX), whose focus is to determine the relationship between changing climate and circumpolar plant species. The impact of climate on plant species would be mediated partly through changes in land structure brought about by sediment transfers.

SEDIFLUX has established a sustainable framework for long-term research to coordinate multinational, interdisciplinary monitoring networks, a first in the field of geomorphology. SEDIFLUX is now providing the basis for further research. One of the major outcomes produced by the SEDIFLUX group will be the SEDIFLUX Manual, which will provide guidelines and protocols for monitoring and sediment budget studies in selected globally distributed cold environment key test catchments. These long-term monitoring campaigns will apply unified approaches and standardized methods to generate comparable datasets from different cold environments for the development of a metadata database and for modelling the impact of climate change on sediment transfers and sediment budgets.

The new I.A.G./A.I.G. Working Group SEDIBUD (Sediment Budgets in Cold Environments) (http://www.geomorph.org/wg/wgsb.html) builds on, continues and extends activities, which have been started within SEDIFLUX. There is a wide range of high-latitude and high-altitude cold environments that need to be studied, from high Arctic / Antarctic to sub-Arctic / sub-Antarctic, alpine and upland sites. This provides a great opportunity to investigate relationships between climate, vegetation cover and sedimentary transfer processes across a diverse range of cold environments, with the ability to model the effects of climate change and related vegetation cover adjustments through space-for-time substitution.

Climate change affects Earth surface systems all over the world but with arguable the greatest impact in high-latitude and high-altitude cold environments. In these areas climate change shapes earth surface processes not just by altering vegetation and human activities but also through its impact on frost penetration and duration within the ground surface layers. Climate change also exerts a strong control on cryospheric systems, influencing the nature and ex-

* e-mail: Achim.Beylich@ngu.no

tent of glaciers and ice sheets, and the extent and severity of glacial and paraglacial processes. Changes within the cryosphere have major knock-on effects on glacifluvial, aeolian and marine sediment transfer systems. All of these factors influence patterns of erosion, transport and deposition of sediments. However it is a major challenge to develop a better understanding of how these factors combine to affect sedimentary transfer processes and sediment budgets in cold environments. As a starting point our baseline knowledge of the sedimentary transfer processes operating within our current climate and under given vegetation cover, as a basis for predicting the consequences of future climate changes and related vegetation cover changes needs to be extended. Only when we have these reliable models will we have fuller understanding. It is therefore necessary to collect and compare data from different cold environments, and use this to assess a range of models and approaches for researching the relationships between climate change, vegetation cover and sediment fluxes.

Results from ongoing geomorphologic studies on sediment fluxes and sediment budgets in selected SEDIBUD key test sites are presented. Quantitative longer-term studies on sediment transfers and sediment budgets are carried out in five selected small cold environment catchments (<30 km^2) in Iceland, Swedish Lapland, Finnish Lapland and Norway. Investigations in East Iceland (Austdalur and Hrafndalur), Swedish Lapland (Latnjavagge) and Finnish Lapland (Kidisjoki) have been conducted for over six years whereas studies in Western Norway (Erdalen) have just been started three years ago. The five catchments are seen as clearly defined landscape units where detailed studies on sediment transfers and sediment budgets using unified techniques and approaches (including monitoring of present-day denudative processes as well as quantitative analysis of storage elements) – providing comparable data sets from the different cold environments – are possible. The five catchments are considered to be representative for the selected target areas in East Iceland, Swedish Lapland, Finnish Lapland and Western Norway.

Main focus of the research programme is on analysing the role of the factors morphoclimate, vegetation cover, ground frost, human impact, relief and lithology for present-day sediment fluxes, denudation rates, sediment budgets and relief development in the five different study sites. Direct comparison of the data collected in the different cold environment target areas provides information on variations in the absolute and relative importance of different denudative processes and helps to get more insight into the spatial differentiation of cold environments.

The two selected catchments in subarctic-oceanic East Iceland are characterized by very steep alpine relief and a partly destroyed vegetation cover (as caused by direct human impact). Mechanical denudation dominates over chemical denudation. Austdalur (basalt) is showing lower mechanical denudation rates than Hrafndalur (less resistant Rhyolithes). The slightly less steep Latnjavagge in arctic-oceanic Swedish Lapland (mica schist) is characterized by clearly lower mechanical denudation rates, which is mainly due to a very stable and closed vegetation cover and stable step-pool systems developed in the creeks. In this valley chemical denudation appears to be slightly higher than mechanical denudation. Kidisjoki in subarctic Finnish Lapland (gneisses) is situated in the area of the Baltic Shield and shows very low chemical and mechanical denudation rates. Chemical denudation dominates over mechanical denudation. All four catchments are characterized by altogether low denudation rates. Chemical denudation ranges from 2.6 t km^{-2}yr^{-1} in Kidisjoki to ca 8 t km^{-2}yr^{-1} in East Iceland. All four valleys are characterized by restricted sediment availability. More than 90% of the annual fluvial sediment transport occurs within a few days during snowmelt and/or rainfall generated peak-runoff. Only in the very steep catchments with partly destroyed vegetation cover in East Iceland mechanical denudation dominates over chemical denudation.

Erdalen is a characteristic and very steep U-shaped valley in the fjord landscape of western Norway (Nordfjord). The sub-Arctic Erdalen catchment is connected to the Jostedalsbreen ice cap and is in its uppermost areas glaciated. Current investigations in this key test site include the quantitative analysis of storage elements like talus cones, valley fillings and lake sediments by using different geophysical techniques, the year-round monitoring of meteorological parameters, ground temperature, permafrost, runoff, fluvial solute and sediment transport as well as the analysis of slope processes like rockfalls, avalanches and debris flows by combining different monitoring and dating techniques.

The possible potential and expected benefit generated by coordinated data exchange and the unification of methods and techniques applied to long-term process monitoring/analysis, the quantitative investigation of storage elements and for sediment budget studies in cold environments is presented. Comparable data sets generated in other cold environment key test sites in polar and alpine regions that follow the guidelines and protocols provided in the SEDIFLUX Manual will be added to a metadata database developed within the global I.A.G./A.I.G. SEDIBUD programme. The SEDIBUD metadata database will be used to model effects of projected climate change on solute fluxes, sediment fluxes and sediment budgets in sensitive cold environments worldwide.

Landform Analysis, Vol. 5: 11–12 (2007)

Coordinated quantitative studies on sediment fluxes and sediment budgets in changing cold environments – examples from three SEDIBUD key test areas in Canada, Iceland and Norway

Achim A. Beylich*

Geological Survey of Norway, Landscape & Climate Team, Trondheim, Norway
Norwegian University of Science and Technology, Department of Geography, Trondheim, Norway

Scott F. Lamoureux

Queen's University, Department of Geography, Kingston Canada

Armelle Decaulne

Laboratory of Physical and Environmental Geography GEOLAB, CNRS, Clermont-Ferrand France
Natural Research Centre of North-western Iceland, Saudarkrokur, Iceland

The new I.A.G./A.I.G. Working Group SEDIBUD (Sediment Budgets in Cold Environments) (http://www.geomorph.org/wg/wgsb.html) builds up on activities which were started within the European Science Foundation (ESF) Network SEDIFLUX (Sedimentary Source-to-Sink Fluxes in Cold Environments, 2004–2006)
(see: http://www.ngu.no/ sediflux,
http://www.esf.org/ sediflux).

Changes in climate have a major impact on Earth surface systems, especially in high-latitude and high-altitude cold environments. Such changes have a major impact on sediment transfer processes. The major aim of I.A.G./A.I.G. SEDIBUD is to provide an integrated quantitative analysis of sediment transfers, nutrient fluxes and sediment budgets across a range of key cold environments. Such an analysis has so far been lacking. The primary focus is on the impact on sediment transfer processes in response to a variety of climate change scenarios at a scale, which incorporates sediment flux processes from source to sink. In order to perform a fully integrated study of source to sink sediment fluxes and sediment budgets in cold environments, SEDIBUD analyses the key components of weathering, chemical denudation, erosion, aeolian processes, mass movements, fluvial transfers/transport, glacial sediment transfers, and sedimentation in lakes, fjords and coastal areas. SEDIBUD is also considering the impact of human activity on the environmental sites being studied and how this might relate to climate change.

Results from ongoing quantitative geomorphologic studies on sediment fluxes and sediment budgets in selected SEDIBUD key test sites in Arctic Canada, sub-Arctic Iceland and sub-Arctic Norway are presented and discussed in the context of possible effects of projected climate change on present-day process frequencies, intensities, process rates and sediment budgets in sensitive cold environments.

Cape Bounty is located in the Canadian High Arctic Archipelago and is representative of the low-relief, unglacierized landscape found in much of this region. Research is underway in paired watersheds with emphasis on suspended sediment delivery processes and fluxes, particulate and dissolved car-

* e-mail: Achim.Beylich@ngu.no

bon and nutrient fluxes, and linkages between fluxes and periglacial slope processes, active layer disturbances, and hydrological routing. Additionally, each watershed drains into similar lakes that contain annually-laminated sediments that will provide long term measures of sediment and particulate organic material delivery.

Fnjóskadalur is a representative U-shaped valley in sub-Arctic Northern Iceland and is characterized by a wide range of different denudative surface processes. Current research in this key test area is focused on:

- (i) the analysis and quantification of sediment fluxes from slope processes, especially snow avalanches and debris flows, and
- (ii) the investigation of the magnitude-frequency relationship of snow avalanches and debris flows. Currently applied methods cover topographical and geomorphologic (underlining erosion and accumulation areas, extreme reach of slope dynamics as well as their lateral spreading) purposes. The used dating techniques (phytogeographical techniques: vegetal cover, lichenometry, dendrochronology; weathering; tephrochronology) reveal the rhythms of present-day slope activity as well as during the Upper Holocene period.

Erdalen is a very steep U-shaped valley in the fjord landscape of western Norway (Nordfjord). The sub-Arctic Erdalen catchment is connected to the Jostedalsbreen ice cap and is in its uppermost areas glaciated. Current investigations include the analysis of storage elements by using different geophysical techniques, the year-round monitoring of meteorological parameters, ground temperature, permafrost, runoff, fluvial solute and sediment transport as well as the analysis of slope processes like rock falls, avalanches and debris flows by combining different monitoring and dating techniques.

The potential and expected benefit generated by coordinated data exchange and the unification of methods and techniques applied to long-term process monitoring/analysis, the quantitative investigation of storage elements and for sediment budget studies in cold environments is presented.

Landform Analysis, Vol. 5: 13–15 (2007)

Physicochemical characteristics of land waters in the Bellsund region (Spitsbergen)

Stanisław Chmiel, Stefan Bartoszewski

Department of Hydrology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland

Andrzej Gluza, Krzysztof Siwek

Department of Meteorology and Climatology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland

Piotr Zagórski*

Department of Geomorphology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland

Studies of physicochemical properties of polar environment waters were carried out in individual stages of water hydrological circulation in the north-west part of Wedel Jarlsberg Land on Spitsbergen. The aim of the research was estimation of changeability of physicochemical characteristics of waters during the polar summer in the rivers supplied from the glacier and from the permafrost. In the period of 15 July – 21 August 2005 there were taken up 290 water samples including 15 of precipitation, 9 from patches of melting snow, 39 from the glaciers Scott and Renard, 132 river samples 31 from streams on tundra, 56 from springs and 8 from small lakes. Time changeability was recorded from everyday observations of water levels and physicochemical analyses of the samples taken up from the river supplied from the Scottbreen and Wydrzyca Rivers, its tributaries and the springs draining the unglaciated area. In the water samples there ware signified: suspension, pH, electrolytic conductivity, BOD, COD, TOC, SiO_2, PO_4, F, Cl, NO_2, SO_4, Ca, Mg, Na, K, NH_4, Li, Sr, Fe, Mn, Zn, Cd, Pb, Cu. The analyses were made based on potentiometric, spectrophotomertic, available in the Hydrography Department, Maria Curie-Skłodowska University in Lublin.

Table 1. Contents of metals in the studied waters

Radical	Value mg dm^{-3}	Limit of detection
Fe	< 0.5	0.01
Mn	< 0.05	0.05
Zn	< 0.01	0.001
Cd	< 0.001	0.001
Pb	< 0.005	0.005
Cu	< 0.001	0.001

Table 2. Typical contents of ions in the precipitation waters

Parameter	Rainfall, wind E	Rainfall, wind WS
pH	5.88	5.27
Conductivity [μS cm^{-3}]	45.60	10.90
Alkalinity [mval dm^{-3}]	0.05	<0.05
Cl [mg dm^{-3}]	10.50	0.67
SO_4 [mg dm^{-3}]	1.80	0.38
NO_3 [mg dm^{-3}]	0.05	0.27
Ca [mg dm^{-3}]	0.96	1.10
Mg [mg dm^{-3}]	0.28	0.14
Na [mg dm^{-3}]	5.40	0.37
K [mg dm^{-3}]	0.19	0.05

* e-mail: pzagorsk@biotop.umcs.lublin.pl

The precipitation waters studied in Calypsobyen were characterized by slightly acidic reaction and low contents of mineral substances of the order from a few to several dozen mg dm^{-3}. Atmospheric precipitation chemistry was largely formed by marine aerosols which resulted in predomination of Na and Cl ions. Their concentration was connected with the direction of air mass influx.

The waters flowing within the Scott and Renard glaciers showed low mineralization and slightly acidic reaction similar to the precipitation waters. The lowest concentration of mineral substances and reaction was observed in the forefield, particularly in the waters loaded with the suspension. Hydrochemical type of waters in the upper part of glacier was a result of Na and Cl ions predomination

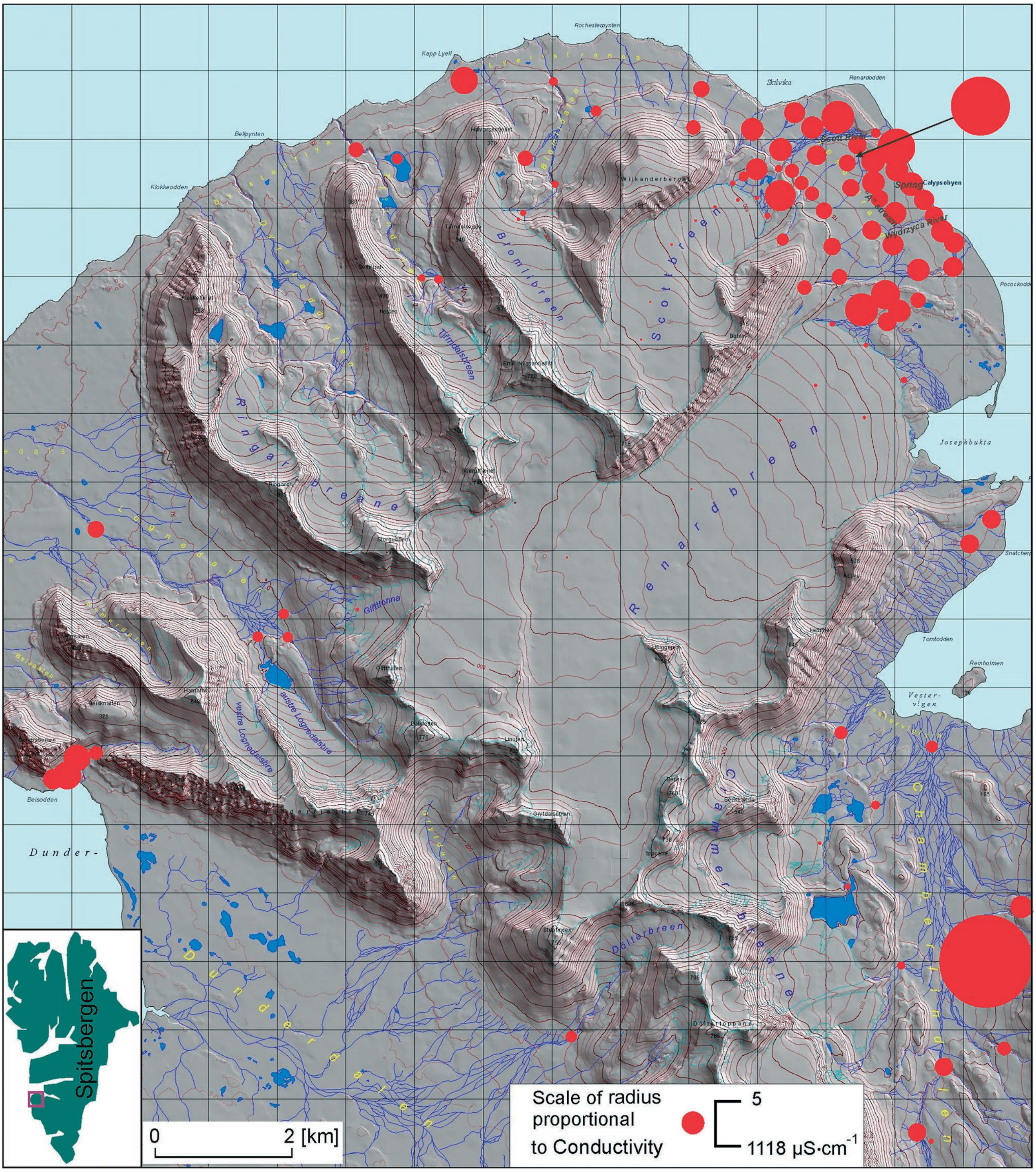

Fig. 1. Electric conductivity of waters in the Bellsund region. The shadow relief map made on the basis of digital terrain model from the aerial photos of 1990 (Zagórski 2002)

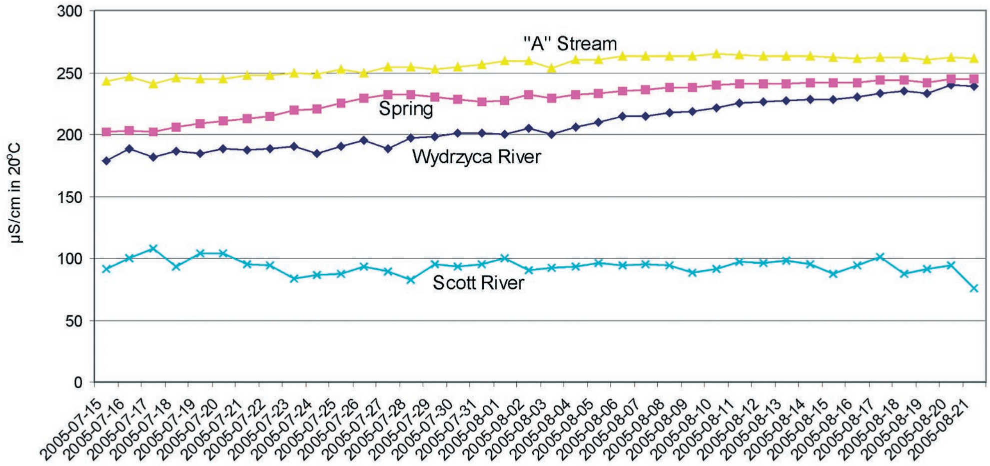

Fig. 2. Changeability of conductivity in the period of research

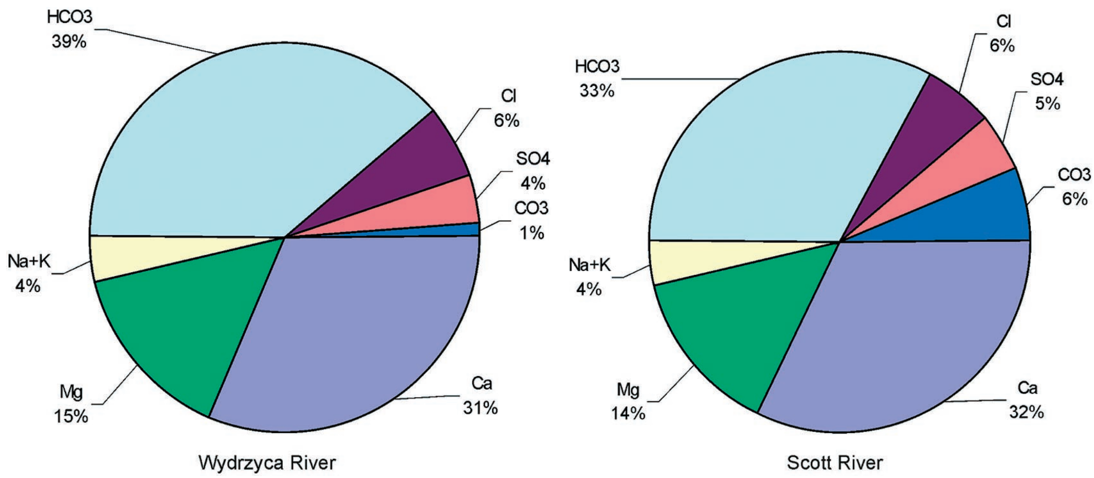

Fig. 3. Contribution of percentage share of ions in the waters of the Wydrzyca River end Scott River

but on its edge it was determined by HCO_3, Ca, Mg ions.

Further increase of mineralization of waters flowing down from the glaciers was observed in their forefield. Mineralization of waters reaching the level ~50 mg/l and the reaction became weakly alkaline. The waters of the Scott River in the estuary profile had mineralization reaching 100 mg dm^{-3} and concentration of the suspended material of the order several hundred mg dm^{-3}.

In the unglaciated areas the surface and underground waters were characterized by low concentration of suspension (mostly below <10 mg dm^{-3}). However, they exhibition much higher mineralization, ~200 mg dm^{-3} and their reaction was weakly alkaline. The surface waters under investigation were related to existence of patches of melting snow which caused decrease of dissolved substances level. In the river waters of glaciated and unglaciated basins HCO_3, Ca and Mg ions were predominant.

Twenty-four hour cycle studies of physicochemical features of waters of the unglaciated Wydrzyca River showed their significant stability. Relatively large dynamics of changes was recorded for the Scott River taking away the waters from the glacier.

As follows from the studies of waters in individual stages of hydrological circulation, they exhibited low contents of substances of biogenous character. The low levels were also found for organic carbon, silica and synthetic indices: biochemical and chemical demands of waters from oxygen. Concentrations of heavy metals were found to be low from a few to several dozen µg dm^{-3} and their concentration scheme is as follows: Fe>Mn>Zn>Pb>Cu>Cd.

The collected material indicates significant hydrological differentiation of waters in individual stages of hydrological circulation as well as time and spatial changeability resulting from the extent of basin glaciation and geochemical conditions.

Acknowledgment

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Geodiversity of polar landforms

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Landform Analysis, Vol. 5: 16–17 (2007)

Age and weathering status of granite tors in arctic Finland

R.G. Darmody*

Department of Natural Resources and Environmental Sciences, University of Illinois, Urbana, USA

M. Seppälä

Department of Geography, University of Helsinki, Finland

C.E. Thorn

Department of Geography, University of Illinois, Urbana, USA

Y.K. Li

Department of Geography, University of Missouri, USA

S.W. Campbell

Independent Scholar

J. Harbor

Department of Geography and Environmental Sciences, University of Colorado at Denver and Health Sciences Center, Denver, USA

Cold-based ice along the Fennoscandian glacial ice divide in northern Finland allowed preservation of older landscape features. Because there was little or no movement at the ice-land surface contact, erosion was at a minimum, and relict landscape features such as tors can be found. We investigated two such granitic tors located at Pyhä-Nattanen (27° 22.207' E, 68° 07.335' N, 485 m a.s.l.) and Riestovaara (27° 09.003' E, 68° 02.613' N, 365 m a.s.l.) in subarctic Finland. At Pyhä-Nattanen, the sampling sequence included bedrock material and grus taken from within horizontal cracks, which are so prevalent at the site as to make the tor resemble a stack of pancakes. At Riestovaara, where the outcrop is more subdued, in addition to bedrock and grus samples, soil samples were also extracted from a pit dug in an embryonic soil forming on the bedrock surface.

Based on cosmogenic dating, both tors greatly predate recent glaciation. The tor at Pyhä-Nattanen, which is a more prominent landscape feature, had a longer minimum estimated total exposure age, 994 kyr, than did the tor at Riestovaara, 857 kyr. Analyses of the 10Be and 26Al cosmogenic data in accordance with marine oxygen isotope records indicates that the tors have survived at least 14–16 episodes of glaciation. Weathering, as measured by porosity determined with a microprobe, was somewhat more advanced in the Pyhä-Nattanen granite samples than in the Riestovaara granite. However, with both granites, rock porosity did not change to a depth of 4 cm below the rock surface, or vary by lichen cover/non-covered surfaces, indicating that weathering had progressed to a stage where recent lichen growth is overwhelmed by the long weathering history. Other measures of weathering, including total elemental analyses, did not detect significant differences among at-a-site samples, perhaps because all samples are highly weathered and we did not have

* e-mail: rdarmody@uiuc.edu

a sample of unweathered granite for comparison. All samples, rock, grus and soil, were within the grus weathering range as indicated by chemical weathering indices. The soil forming adjacent to the tor at Riestovaara exhibited only slight development despite the great apparent age of the landscape. The regolith in which the soil is forming must therefore be much younger than the exposed bedrock and most likely represents a post-deglaciation accumulation of grus spanning only the last 9k years.

Landform Analysis, Vol. 5: 18–19 (2007)

The white streaks of Kärkevagge

Robert G. Darmody*

Department of Natural Resources and Environmental Sciences, University of Illinois, Urbana, USA

Colin E. Thorn

Department of Geography, University of Illinois, Urbana, USA

John C. Dixon

Department of Geosciences, University of Arkansas, Fayetteville, USA

Conventional wisdom once held that weathering in cold climates was overwhelmingly due to physical processes. Rapp (1960) challenged that convention with his publication on landscape denudation in Kärkevagge, Swedish Lapland where he made the startling discovery that chemical weathering exceeded any single measured physical denudation process. His interpretation was based on limited analyses of water chemistry where he found total flux of dissolved solids accounted for most mass loss from the watershed. The dominant anion he observed was sulfate. He also observed other features of chemical weathering in the valley including "white streaks of lime" in stream channels on the valley flanks (Fig. 1). While insightful, Rapp offered no mechanism to explain these findings. Our subsequent work in Kärkevagge has revealed the driving mechanism of chemical weathering to be acid production from pyrite oxidation.

In this work, we used scanning electron microscopy (SEM) and energy dispersive x-ray fluorescence (EDXF) to examine a variety of coatings found in the valley. Analyses revealed that the "lime coats" are primarily an amorphous aluminum oxyhydroxide sulfate such as basaluminite $[Al_4(SO_4)(OH)_{10}{\cdot}H_2O]$, which "paints" surfaces it contacts (Fig. 2). That this is an active process is demonstrated by efflorescence on seasonal vegetation in stream channels. We found no systematic spatial patterns along the valley axes or up the valley wall, either of the coating appearance under SEM or of chemistry by EDXF. Although the "white stripes" were not crystalline and did not contain appreciable amounts of Fe or Ca, in sheltered overhangs among boulders on the valley floor we found other well crystallized secondary sulfate minerals commonly associated with pyrite oxidation, including white crusts of gypsum $[CaSO_4{\cdot}2H_2O]$, yellowish coatings of jarosite $[KFe_3(SO_4)_2(OH)_6]$, and rust-colored amorphous Fe compounds $[Fe(OH)_3]$. This difference is due presumably to the pH of the associated waters, because Fe compounds tend to precipitate only at pH <5, and Al compounds at pH >5, the pH of the stream water. We believe that pyrite oxidation may be an important early weathering process in many environments. It largely goes unrecognized because it occurs rapidly and typically is only identified in recently disturbed landscapes associated with mining and other large-scale earth-moving activities. An additional implication of our findings is that Rapp may have unknowingly chosen an environment to do his work where the particular geochemistry evokes accelerated chemical weathering. In short, Kärkevagge demonstrates that sub-arctic conditions do not preclude intense chemical weathering where other conditions are favorable, but does not establish that strong chemical weathering is a widespread attribute of sub-arctic conditions.

* e-mail: rdarmody@uiuc.edu

Fig. 1. View of eastern cliff face of Kärkevagge showing "white streaks" of efflorescence along stream courses

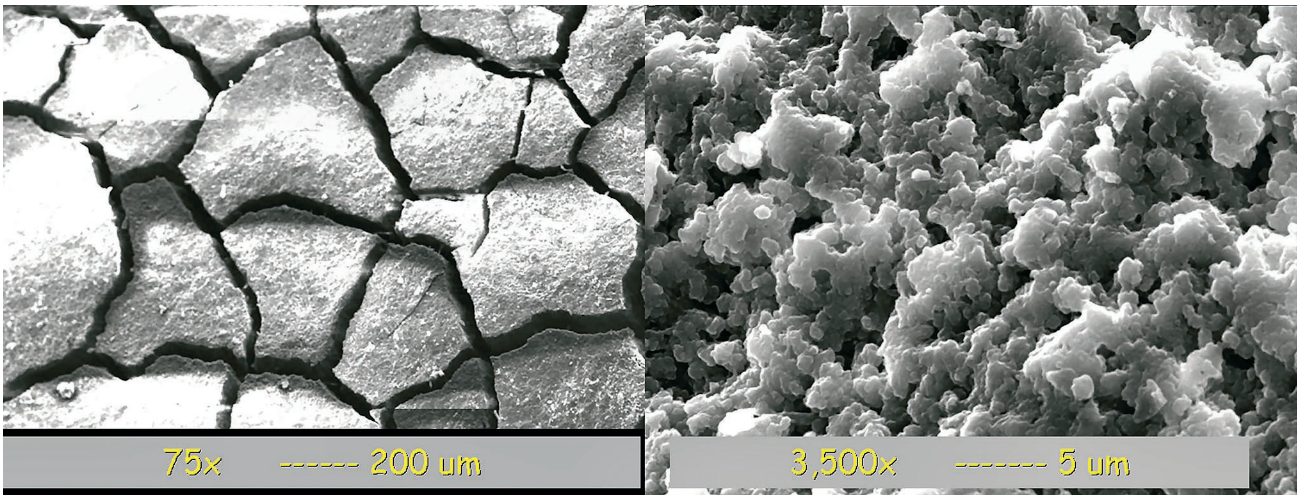

Fig. 2. SEM micrographs of "white streak" coating a rock outcrop in stream

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Landform Analysis, Vol. 5: 20 (2007)

Geomorphology and global environmental change

Christine Embleton-Hamann*

Institut fur Geographie und Regionalforschung, University of Vienna, Vienna, Austria

Olav Slaymaker

Department of Geography, University of British Columbia, Vancouver, Canada

A major statement on the state of our understanding of landscape scale geomorphology in relation to global environmental change is being prepared. A book will be published by Cambridge University Press in 2009 to inform the delegates to the 7th International Conference on Geomorphology in Melbourne, Australia and as a complement to the 4th IPCC Assessment (2007). The concept behind the book is that there is an unfilled niche in the climate change literature, namely the nature of landscape scale change in the face of anticipated climate change. The book will cover both zonal and azonal landscapes.

The first meeting of this IAG Working Group was held, through the good offices of Tom Spencer, at Magdalene College, Cambridge from September 19–22, 2006. A protected Web site for the use of Working Group members has been established in the Institut fur Geographie und Regionalforschung at the University of Vienna and a professional cartographer has been engaged in the Department of Geography at the University of British Columbia.

The chapter topics are as follows: Landscape scale change: the unfilled niche; Ice sheets and ice caps; Mountain environments; Lakes and their basins; Freshwater wetlands; Rivers; Sedimentary coasts; Sandy coasts and dunes; Coral reefs; Tropical rainforest; Savannas; Deserts; Mediterranean lands; Mid-latitude forest; Tundra and periglacial taiga; and Bridging the gap between science and policy.

The second meeting will be held as a joint meeting with the Austrian Commission on Geomorphology at the University Centre in Obergurgl, Austria, September 1–6, 2007 under the leadership of Christine Embleton-Hamann. Lead authors will be presenting their preliminary findings at that meeting.

The present paper reviews the progress of each of the chapters to date and their distinctive emphases.

* e-mail: christine.embleton-hamann@univie.ac.at

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Glacier distribution and direction in the Arctic: the unusual nature of Svalbard

Ian S. Evans*

Department of Geography, Durham University, Durham, United Kingdom

Local glaciers in the Arctic, as elsewhere, are valuable climatic indicators. On a regional scale (Table 1), their varying ELAs (Equilibrium Line Altitudes) or mid-altitudes reflect the dominant direction of snow-bearing winds. In the Eurasian Arctic, this is mainly from the Atlantic and ELAs rise eastward (Grosval'd and Kotlyakov 1969; Dowdeswell and Hambrey 2002; Dowdeswell and Hagen 2004). Here I will focus on the effects of accumulation area slope aspect on the numbers and altitudes of local glaciers. Local slopes affect several components of glacier mass balance, and we expect that there will be both lower glaciers, and more glaciers, facing directions (azimuths, aspects) with more positive mass balances. North: south contrasts due to solar radiation receipts are great in middle latitudes, especially in dry, sunny climates, but diminish toward the Poles (Evans and Cox 2005).

Analyses of World Glacier Inventory data confirm the expectation that poleward tendencies in both greater numbers and lower altitudes would be weaker for Arctic glaciers, if defined as those regions above 70°N (Evans 2006; Table 2). However, several Arctic regions (Wrangel Island, Svalbard, Ellesmere Island and Axel Heiberg Island) have favoured directions in terms of numbers of glaciers (expressed by vector means) very different from their lowest mid-altitude directions. These anomalous results are unusual and require explanation.

Out of a total 685 glaciers, Novaya Zemlya has 395 valley, 158 mountain glaciers and 27 glacierets, giving 580 local glaciers, 574 of which have the aspect and altitude data for analysis of local asymmetry. The direction of minimum mid-altitude as predicted from the regression on latitude, longitude, sine and cosine of aspect is 098° ±18° (95% confidence limits), which is not far from the vector mean of 062° ±15° (vector strength 23%). Novaya Zemlya has extra glaciers facing northeast and southeast, giving a significant eastward component. This is explained by the effects of wind from westerly directions, mainly through drifting snow to lee slopes.

In Svalbard, 241 out of the 406 glaciers inventoried and classified are 'local'. Mid-altitude is predicted to be minimal at an azimuth of 109° ±46°, an eastward tendency that is significant only when position is included in the regression. However, the vector mean is 014° ±17° with a strength of 21%, so the two approaches are inconsistent. This may relate to the unusually low average gradient (5.8°) and great length (average 8.86 km) of these glaciers, differing by a factor of at least two from those of all other regions, and reducing the importance of aspect. The 'non-local' glaciers have no significant resultant vector, and their lowest altitude is weakly eastward.

Franz Josef Land (81°N) has 995 glaciers, but most of them are ice caps and outlet glaciers and thus not suitable for these analyses. There are 153 glacierets and 6 other local glaciers, yet all but 7 lack either lowest or (mainly) highest altitude in the Inventory. Likewise in Severnaya Zemlya (79°N), highest altitude is normally missing. However, vector analyses of local glaciers show highly significant asymmetry, tending to 312° ±17° for Franz Josef Land and 022° ±24° for the 118 on Severnaya Zemlya, with vector strengths of 33 and 31% respectively.

Further east, all 101 glaciers on Wrangel I. (71°N) are local, and from 0.1 to 1.2 km long. Their vector mean is 335° ±13° but they are lowest when facing

* e-mail: i.s.evans@durham.ac.uk

Table 1. Arctic glaciers in the late twentieth century

Region	Area [km²]	Number	ELA [m]	Lat. [deg]	Long. [deg]
Alaska					
Brooks Ra.	722	995	1500–2100	69	146W
Canada					
Axel Heiberg I.	11735	1101	200–1200	80	92W
Ellesmere I	80500		200–1000	79	80W
Devon I.	16200	1835	400 ?	75	83W
Bylot I.	5000		500–850	73	78W
Baffin I.	37000		>11,364 300–1100	71	72W
other islands	1356		(150)	77	100W
GREENLAND	76200* excl. ice sheet		0–1600	73	42W
JAN MAYEN I.	115	>30	1000	71	8W
N. NORWAY & SWEDEN	1441*	1487	600–1600	68	17E
SVALBARD	36591	2128	100–800	79	19E
Russia					
Franz Josef Land	13734	988	100–400	81	57E
Novaya Zemlya	23600	685	300–700	76	61E
Ushakov I.	325	2	130	81	80E
Polar Urals	29*	143	600–1000	66	64E
Severnaya Zemlya	18300	287	350–600	79	98E
Byrranga Mts. (Taimyr)	30	96	700–1000	76	108E
Putorana Plateau	2	22	950	69	94E
Orulgan Mts.	18	74	1600–2000	68	128E
Chersk Mts.	155*	372	1900–2300	65	145E
De Longa I.	81	15	200	77	152E
Wrangel I.	3	101	400	71	179E

Mainly from World Glacier Monitoring Service (Haeberli et al. 1989, 1998), updated from Dowdeswell, Hambrey (2002). Note that numbers differ from those cited in the text, based on on-line World Glacier Inventory data
* includes some glaciers south of Arctic Circle

southeast: 143° ±34°. Disregarding position, the 25 north-facing glaciers average 511 m in mid-altitude; the 9 facing south, SE or SW average 292 m. Although small, the numbers are sufficient to provide significant Fourier coefficients and it must be admitted that on Wrangel, as in Svalbard, glacier numbers and altitude reflect aspect in different ways. Thus, in most of the Eurasian Arctic, more local glaciers face northward, but they are often higher than south-facing ones. Linear trends with position seem insufficient here to allow for the regional effects of moisture brought from southerly sources.

In Canada, Axel Heiberg was the subject of a major trial of the Inventory methodology, and Ommanney (1969) provided an early complete Inventory. Although local glaciers have an insignificant vector resultant, the lowest altitude is clearly northward. On the other hand nearby southeast Ellesmere Island, with only 80 local glaciers, has a significant southward resultant but no significant altitude variation.

Further investigation is under way of the effects of glacier type (Table 3) and other characteristics on these results. Anomalies in Svalbard and elsewhere may be due to tidewater glaciers. Calving gives a ‘premature’ glacier termination, such that averaging lowest and highest glacier altitudes does not give a good estimate of ELA – unlike the situation for

Table 2. Vector and regression results and average characteristics for ten Arctic regions, ordered by latitude

Region	number of glaciers	midalt [m]	length [m]	grad [°]	vmean [°]	mindir [°]	vs [%]	asymalt [m]	Lat [°]	Long [°]	abbrev
Suntar-Khayata& Chersk	381	2274	1440	24.8	4	351	78	86	64.4N	142.1E	SK
West Greenland > 64.8N	1775	939	1038	14.8	349	354	38	90	67.9N	52.7W	Gn
Northern Scandinavia	1146	1167	902	20.7	53	35	52	30	68.0N	17.5E	NS
Orulgan	74	1867	724	23.3	25	17	78	44	68.4N	128.5E	Og
Brooks Range	876	1790	1383	20.1	358	25	64	45	68.5N	148.8W	Br
Wrangel I.	101	428	162	18.8	335	144	55	102	71.1N	179.0E	WI
Novaya Zemlya	574	524	3514	11.2	62	98	23	43	74.0N	56.2E	No
SE Ellesmere I.	80	526	4379	15.1	185	323	23	20	77.3N	78.5W	El
Svalbard	241	470	8861	5.8	14	109	21	33	78.5N	15.1E	Sv
Axel Heiberg	289	844	4305	12.9	255	15	7	30	79.3N	91.5E	AH

The variables are: region – region name, number of glaciers – number of glaciers in the analyses, midalt – means of mid-altitude, length – length, grad – gradient, vmean – vector mean direction, mindir – direction of minimum glacier altitude, vs – vector strength, asymalt – degree of altitude asymmetry, Lat – mean latitude, Long – longitude, abbrev – abbreviation for Figs.

Table 3. World Glacier Inventory classification of glaciers in the Eurasian Arctic

Class	Region					
	Svalbard	Nov.Zem	Franz Josef	Sev.Zem.	*	Total
uncertain	27	1	0	0	0	28
field	3	0	0	0	0	3
cap	20	37	349	68	0	474
outlet	115	67	487	100	0	769
valley	237	395	1	17	0	650
mountain	0	158	5	44	22	229
glacieret	4	27	153	57	0	241
shelf	0	0	0	3	0	3
*	488	0	0	0	0	488
Total	894	685	995	289	22	2885

* = missing data

simple mid-latitude glaciers. Also the different mass balance components on Arctic glaciers (Dowdeswell and Hagen 2004; Etzelmüller and Sollid 1996; Hagen et al. 2003) may change the relationships between ELA and glacier geometry. Currently, glacier mid-altitude is available for many more glaciers than is ELA. In the Arctic, the effects of solar radiation on glacier mass balance are present, but are small and easily masked or overturned by other factors such as wind, and varying diurnal cycles of ablation.

Acknowledgement

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Landform Analysis, Vol. 5: 25–29 (2007)

Permafrost and periglacial environment of Western Tibet

Monique Fort*

UMR 8586 PRODIG, CNRS, UFR GHSS, Université Paris Diderot-Paris 7, Paris, France

Brigitte van Vliet-Lanoe

UMR 8110, CNRS Processus et Bilans des Domaines Sédimentaires, SN5, Université Sciences et Technologies de Lille, Lille, France

Northwestern Tibet (=*Qiantang*) (78°–81°E; 33°–36°N), located between the Kunlun (7178 m a.s.l., West Kunlun Peak) and the Karakoram ranges, is the highest (mean elevation 5500 m a.s.l.), the coldest (MAT > 0°C) and the driest part (P >50 mm/y) of the Tibetan Plateau. This remote region, a mountainous desert with lacustrine depressions surrounded by patches of steppe, has rarely been visited so far. Except for a few early explorers, the first detailed records dealing with the glaciers and the periglacial environment prevailing on the plateau were provided in the early eighties by the scientific expeditions led by Chinese and Sino-Japonese teams. During summer 1989, the Sino-French Kunlun-Karakoram Geotraverse (Fig. 1a) carried out multidisciplinary research aimed at studying tectonics, Quaternary evolution and present environment of Qiangtang (Fort and Dollfus, 1992). We describe here the conditions and types of periglacial features observed and their controls (elevation, lithology, geomorphology). Evidence of fossil, periglacial features eventually suggest that periglacial environments prevailed here during most of the Upper Pleistocene.

Physical conditions favouring periglacial environment

Climatically, Qiantang is known as the driest part of the central Asiatic mountains (Fig. 1a). The annual amount of precipitation, which occurs mostly during summer as convective rain/snow falls (Flohn 1968, 1981), is very low. On the plateau, the mean annual precipitation is estimated to be only 20–50 mm (Chang, 1981), although it is probable that more precipitation occurs at higher altitudes (Flohn 1981). For instance, Ohata et al. (1989) have estimated the precipitation to be < 200 mm/yr at 5200 m a.s.l. (Aqsay Qin lacustrine plain) and < 350 mm at 6300 m a.s.l. (flank of West Kunlun Peak). At Tianshuihai (4860 m a.s.l.), our one-year record data (Aug. 1989 – Aug. 1990) (Fig. 1b and c) suggest that annual precipitation can be as low as 23 mm/yr, and show that the moisture content of air never exceeds 90%. At the same site, the total potential evaporation calculated for the same one-year period is 1607 mm (Dobremez, unpubl. data). This extreme aridity of Northwestern Tibet makes this area certainly the most severe, alpine desert on earth.

The mean annual temperature is negative (-2.1°C measured during our one-year record at Tianshuihai), yet this value is 8–9°C higher than what would be expected at these altitudes compared to adjacent areas of the same latitude, due to plateau effect (radiative budget in excess during summer). At Tianshuihai also, the contrasts recorded between the coldest (January) and warmest (July) months (mean annual amplitude: 20°C, for a maximal range of temperature of 45°C), and between the coldest and warmest hours of each day, show that there are many freeze-thaw cycles during a year (153 cycles during the period Aug. 1989 – Aug. 1990) (Fig. 1d). This number is certainly much higher if ground temperature, instead of air temperature, is considered, because of the heat concentrating on ground surface.

* e-mail: fort@paris7.jussieu.fr

In the adjacent Eastern Pamir, there are no fewer than 300 freeze-thaw cycles annually in soil at a depth of 2 cm (Gorbunov, 1983). From this we may also infer that the south-facing slopes can potentially undergo a freeze-thaw cycle every day of the year. On north facing slopes or in mountain shadow, the radiative budget is negative, promoting permafrost extension.

At these altitudes, the winds are strong and frequent. At the Tianshuihai site, the windy days (wind velocity > 4.5 m/s) exceed 80–100 days/year (Li et al. 1990). The wind may interfere subtedly on the development of permafrost, especially in flat areas: by lowering significantly the air temperatures thus increasing the ground-air thermic gradient, by increasing the dryness of the ground surface by exportation of fine particles, or/and by increasing its water retention by importation of salt particles. Occasionally, the wind may drift away the rare snow cover, which anyway is never thick enough to act as a thermal protection.

Geomorphologically, Western Tibet is not "simply" a plateau, a flat elevated terrane, but an alternation of mountain ranges (6200 m–6800/7000 m a.s.l.) and discrete, lower (5500–6200 m a.s.l.), flattened ridges, separated by 80–100 km^2 wide, endorheic

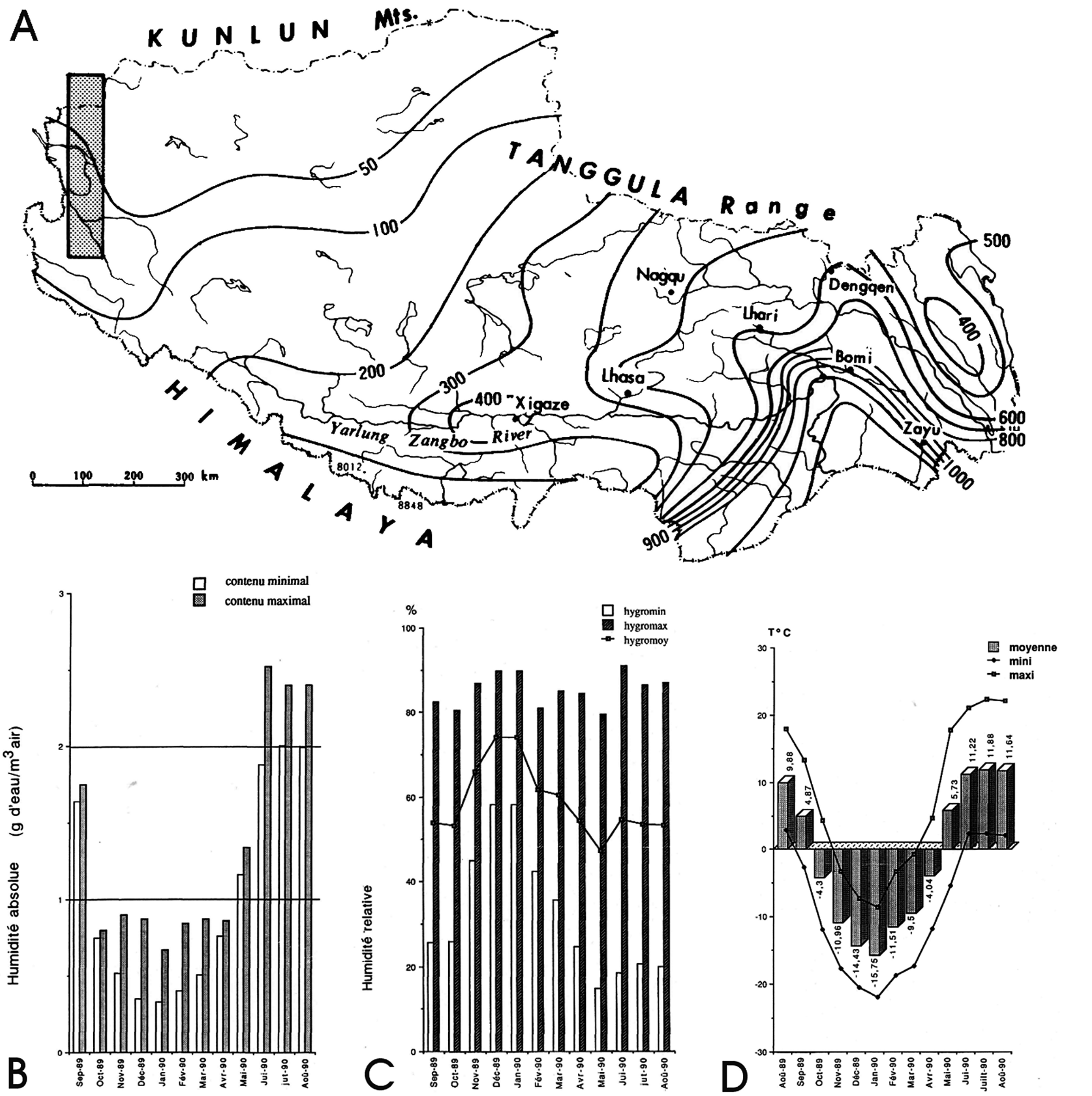

Fig. 1. Map (A) of mean annual precipitation distribution in Tibet (after Li, Zheng 1981) – the studied area corresponds to the gray rectangle (left); Hydrometric data (B, C) and thermal data (D) of Tianshuihai station (4860 m a.s.l.) annual cycle (Aug. 1989 – Aug. 1990), recorded by the Sino-French Kunlun-Karakoram Geotraverse (after Fort, Dollfus 1992)

plains (4400–5000 m a.s.l.), where fluvial, fanglomeratic and lacustrine sediments have accumulated during Pleistocene and Holocene (Fort and Dollfus, 1989; Gasse et al. 1991). The mountain ridges receive most of the precipitation, predominantly as snow. The modern glaciation is limited to ice-caps on summits >6400–6500 m. The snowline altitude varies from north to south between 5800–6200 m (north-facing slopes) and 6000–6400 m (south-facing slopes), thus reflecting a dryness gradient as a result of the general decrease of the mean elevation of valleys southwards.

The contrasts between slopes and nearly flat areas cause an unequal distribution of water in ground, which controls the occurrence of periglacial features. Zones of moisture concentration are mainly found along slopes high enough to get ice/snow meltwaters, at the foot of slopes or at the issue of gullies collecting melt/outspring waters, along the river floodplains and low terraces, and close to the lake shores and meadows developed on former lacustrine sediments. These latter are the source area of most salt particles blown away by wind and deposited on the surrounding slopes.

Under these climatic, altitudinal and topographic conditions, high-cold desertic and high-cold steppic landscapes prevail. In the northern part, the very short or non-existent growing season results in a very sparse vegetation with adapted species, like the cryophytic-xeric cussionlike Ceratoides compacta (>8% soil cover). Steppic communities are locally developed on the piedmont slopes (Carex moorcroftii, Stipa purpurea) and around the lake depressions (Stipa glareosa, Stipa subsessiliforma), whereas the only woody plants are found along some river beds (Myricaria hedini) (Chang 1981). In the southern part of Western Tibet, extensive Caragana versicolor and Juniperus communities are not rare in the lower (4300 m a.s.l.), warmer valleys, whereas the slopes are often covered by a Stipa purpurea steppe (Chang 1981).

From this rapid presentation of the Qiangtang environment, it appears that frost is acting everywhere. In the ground, the cryogenic activity, induced by the great number of freeze-thaw cycles, is directly controlled by moisture availability. Because of the extreme aridity of the climate, the edaphic conditions (controlled by slope and site) become a predominant factor in the distribution and type of periglacial features, which are also locally influenced by salt occurrences.

Present permafrost and associated cryogenic landforms and figures

Located quite above the 0°C isotherm (estimated to be about 3800 m a.s.l.; Li and He, 1989), North-western Tibet is potentially a zone of high altitudinal, continuous permafrost. Evidence for frost-shattering processes are everywhere in the landscapes. However, we found that the zone of detectable, continuous permafrost is not as widespread as formerly mapped (Li and He, 1989), because of limited moisture content in ground. Instead, we observed a rather extensive zone of discontinuous detectable permafrost, the patches of which are developed only on sites where water is available. Thus it can be questioned whether the absence of water in soil is a limiting factor to the development of features generally associated to permafrost. The summer dryness, by lowering the thermal conductivity, and increasing the albedo of the dry sediments and salt efflorescences, limits the heat transmissivity and the depth of the active layer.

Because the slopes represent dry, well drained areas, they are mostly shaped by frost shattering processes. Typical mountains slopes of Western Tibet are debris mantled, and/or are shaped as rectilinear, Richter slopes. Yet, additional forms of solifluxion also occur, limited in their extent by favourable factors: lithology, aspect, position in the hillslope profile, all related to the moisture content.

The rectilinear profiles correspond to unvegetated, rocky, denudation (or Richter) slopes, covered with a thin (few decimetre thick) sheet of debris, with a slope angle corresponding to the angle of stability (varying between 27–28° up to 34°) related with bedrock parameters such as jointing, bedding. The best examples of Richter slopes are encountered on monzonites (northwestern edge of the plateau), where they reach their final stage without any rocky spur but on their very top. Other examples are also found on shaly substrate.

In the composite slopes, the upper section may develop as a steep, rocky cliff, yet, most of the time, it exhibits an alternation of frost weathered pinnacles and frost gullies controlled by joints. The frost debris accumulates downwards as talus screes or cones (slope angle between 30–34°), built up by a combination of debris falls, avalanches and debris-flow. Solifluction features usually develop along the lower part of the slopes, resulting in a noticeable lowering of the slope angle (down to 20°).

Although frost shattering (and probably salt shattering as well) is the dominant process, solifluction –in fact seemingly mostly frost creep- plays an active part in the slope evolution. Typically, the talus or cone profiles are ondulated, and soli-gelifluction lobes occur in the lower part of slopes, where moisture may generally concentrate. Their size varies from 1–10 m in width, depending on rocky material. Their lower limit observed is on south facing slopes at about 4800 m, and may be as low as 4500 m a.s.l. on north-facing slopes. Even on Richter slopes, the thin sheet of debris displays a flat girdle pattern on surface, outlined by coarser material. This feature seems to be characteristic of arid alpine hillslopes, as also observed in the Kunlun Range (Iwata and Zheng, 1989) and in the Ladakh-Gandise range (Fort 1981), this being probably a result of dry slumping.

When the ground moisture content increases, the conditions are at best for the rock-glaciers to develop. This is particularly true when there is an upper, rocky slope that concentrates the snow or glacial meltwaters, or along avalanche tracks and/or small ravines. The lower altitudinal limit of active rock-glaciers varies between about 4800 m a.s.l. (south aspect) and 4600 m a.s.l. (north aspect). It can exceptionally descend lower, when rock-glaciers develop on a terminal moraine (Iwata and Zheng, 1989), hence probably revealing a MAT <-2°C, as noted in the Alps (Haeberli, 1985).

If patterned grounds are not absent from mountain slopes, their best yet limited occurrences are in the valleys or depressions, mostly at altitudes ranging between 4500 m a.s.l. and 5200 m a.s.l. They are good indicators of the existence of permafrost. Li and He (1989) have studied the most widespread continuous permafrost encountered in this region: the vicinity of the West Kunlun range, near and in the Tianshuihai lacustrine depression (4900–5000 m a.s.l.). There, the active layer is 1.0 to 1.5 m thick, the mean annual ground temperature is –3.2°C. The depth of no annual temperature amplitude is 13–15 m. From geothermal gradient and geophysical methods, Li and He (1989) estimated the permafrost thickness to be 117,9 m and 77,0 m respectively. They also indicate that this permafrost contains thick horizontal ice-rich layers and ice masses near the permafrost table. In association with it, masses of segregated ice, ice wedges and ice veins (>0.5 m high), thermokarst lakes, have also been reported by the same authors, together with ice-core mounds (pingos) about 1m high, with fissured summits and basal diameters of 3–5 m. Only one large pingo (basal diameter of 100 m), surrounded by slumped layers, has been observed (Li, 1987). This ice rich layer could be a relict of Holocene, more humid period, as suggested by the extension of lacustrine layers. In such arid and cold environment, the presence of shallow lakes seems to be a reinforcing factor for the surficial expression of permafrost.

In fact, observations made along our 500 km long transect led us to think that the Tianshuihai area cannot be considered as fully representative of the northwestern part of Tibet, because of the proximity of large and flat lacustrine water bodies susceptible to provide the moisture necessary for ground ice to develop. In other places, the nature and thickness of permafrost is unknown and, in fact, cryogenic features are more subtle and more spatially confined. Patterned ground, non sorted circles, cryogenic mounds (palsas; Fig. 2), upheaved stones, and icing along the river beds are the most typical landforms we encountered, with some characteristics that reflect the overall aridity of the environment, yet are not necessarily significant of the presence of permafrost (palsas excepted).

The best examples of patterned grounds are circles observed north of the Sumxi lake (5350 m a.s.l.). These figures developed at the base of a 5500 m a.s.l. high ridge underlain by Jurassic, north-dipping limestones, with dip slope flanks (24–28° slope angle) mantled by decimetric blocks dislodged by widely open, frost cracks. From the lower north-facing slope, we successively observed from top to foot solifluction lobes (17–7° slope angle), passing downward to non sorted, elongated stone circles developing on a colluvial piedmont (5350–5200 m a.s.l.), sloping (4–2°) in a northwest direction. The colluvial material includes frost-shattered, limestones clasts, calcareous, silty particles derived from frost shattering and dissolution, and from wind blown lacustrine silts (winnowed from the Holocene lacustrine deposits of lakes Sumxi and Longmu Co), and salt particles, also derived from the same lacustrine outcrops. On surface, fine particles are protected by a thin veneer of clasts (deflation pavement). The ground patterns observed are non sorted and very shallow, due to the low soil moisture content which favours an active layer without mid-portion desiccation (Van Vliet-Lanoe 1985). The ground temperatures measured (20/07/89, early afternoon) indicate that the active layer is 1.5 m thick at 5100 m a.s.l. (i.e. close to the limit of continuous permafrost), and only 0.4 m thick at 5350 m a.s.l. Yet, the excavations performed in the still frozen active layer (segregation ice) have shown that this layer varies in thickness.

Mineral palses (Fig. 2) are certainly the most striking periglacial features that typify this part of Tibet. These 5-to-10 m high mounds may extend laterally over tens of metres, and their surface is affected by desiccation cracks. Their development is related to segregation ice-lenses, very much dependant upon

Fig. 2. Cryogenic mound (mineral palse) associated to discontinuous permafrost affecting early-middle Holocene lacustrine sediments. East of Domar, Bangong lake watershed (photo M. Fort)

Fig. 3. Several generations of rock glaciers: inherited rock glaciers mantled with a thin veneer of sands and silts in the lower part; active rock glaciers (>4800 m a.s.l.) in the upper part. Chanthang valley, west of Rutog (photo M. Fort)

site-specific conditions. Indeed, these palses are only found in the vicinity of lacustrine bodies and/or in the valley bottoms, their growth being favoured by the presence of early-middle Holocene, laminated, lacustrine silts (Fort and Dollfus, 1992) and illimited supply of water.

Eventually, permafrost and periglacial features of Western Tibet appear as sensitive to and good indicators of climate change. On the one hand, inactive, silt-blown mantled rock-glaciers (Fig. 3) and rectilinear hillslope observed at lower (<4800 m a.s.l.) altitudes suggest that periglacial environment prevailed here during most of the Upper Pleistocene, hence refuting the idea of a generalized glaciation over Tibet. On the other hand, the warming trend in air temperature recently detected East of the Tibetan Plateau (Wang and French, 1994) might possibly cause progressive permafrost degradation, increasing dryness and potential risk of desertification in the near future.

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Glacial Quaternary of the Levant from speleothems, lakes and loess

Amos Frumkin*, Sorin Lisker

Department of Geography, Hebrew University of Jerusalem, Jerusalem, Israel

Anton Vaks

Institute of Earth Sciences, Hebrew University of Jerusalem, Jerusalem, Israel

Miryam Bar-Matthews

Geological Survey of Israel, Jerusalem, Israel

Speleothems, lake levels and loess data have been collected and radiometrically dated, providing a comprehensive insight of climate change in the Levant during the last 300,000 years.

In the Mediterranean climatic belt, glacial periods were generally cool, wet and dusty. Lake levels were high; speleothems were deposited in presently Mediterranean and semi-arid zones; thick loess beds were deposited. Interglacial periods in the Mediterranean climatic belt were mostly drier, warmer, with less dust deposition. Lake levels were low.

In the arid zone of the northern Saharan belt, conditions were mostly dry, during both glacial and interglacial periods. However, short wetter episodes occurred mainly during previous interglacial periods, at marine isotope stages 5, 7, and 9. Climatic variability of the Levant is highest during interglacial periods, especially at the onset of these periods.

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* e-mail: msamos@mscc.huji.ac.il

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The heavy metals in water of select Spitsbergen and Iceland glaciers

Marek Jóźwiak*, Małgorzata Jóźwiak

Institute of Geography, Saint Cross Academy, Kielce, Poland

In the summer time, as a result of ice melting, which takes part inside of the glaciers, melted water often floats down. It floats trough subglacier tunnels, as supraglacier streams and in coast areas of glacier. Melted water is usually poorly mineralized. However it contains some substances which composition and concentration is various according to the localization of the glacier (Jóźwiak, Kozłowski 2005). The chemical composition of glacier water is formed by the chemicals from snow, chemical composition of air contamination, in wet and dry form, including sea aerosols, chemical composition of rainwater, chemical composition of water which flows from outside of the glacier and chemical reaction inside the glacier area (Jania 1997). One of the component in supraglacier water are heavy metals. Their source are dusts placed in snow which come from Europe (Pacyna et al. 1985), especially from North Europe (Paatero et al. 1993) and industrial areas of Siberia (Mielnikov 1991). The pollution transport, including heavy metals, takes place also in Arctic Sea area (Pfirman et al. 1995). On Iceland, the source of heavy metals in water, which flows over the glacier are also volcanic dusts.

The purpose of this article is to present the content of heavy metals in supraglacier water streams of the selected glaciers of Svalbard and Iceland.

In july 2003 was taken water samples from one of the glaciers of Spitsbergen. The chosen glaciers were: Waldemar (NW Spitsbergen), Ebba (Middle Spitsbergen) and Hans (South Spitsbergen). On Iceland the glaciers which were chosen to examine were placed in south Solhejmajökull and south-west Fláajölull part of the Iceland, from where the water samples were taken in August 2005 (Fig. 1, 2). One liter of water was taken into the BRAND containers and transported in temperature of 4°C to Poland to

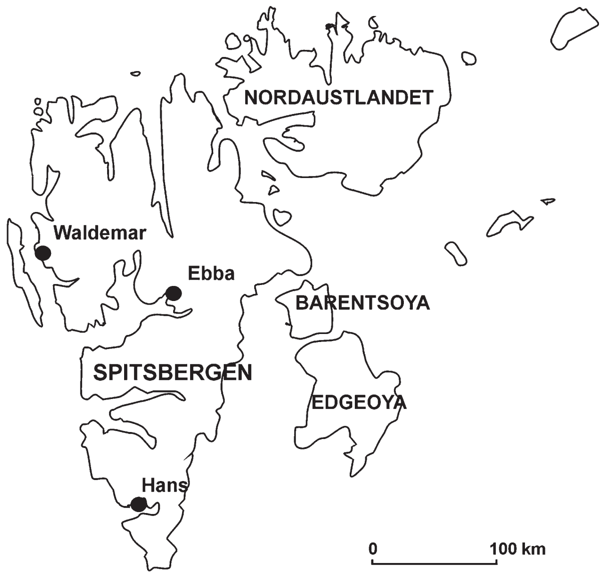

Fig. 1. Spitsbergen glaciers localizations

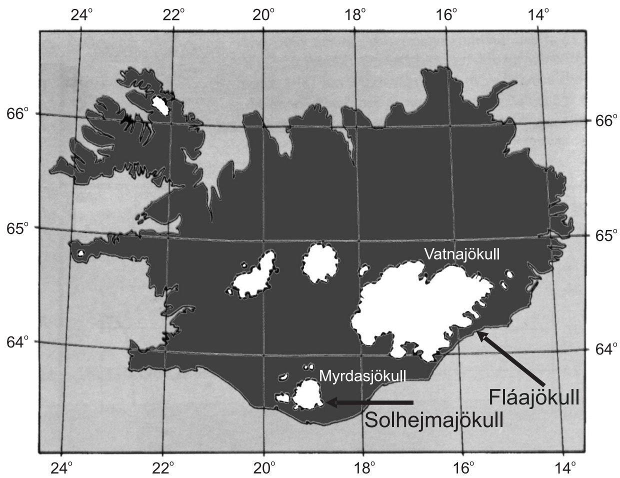

Fig. 2. Island glaciers localizations

* e- mail: marjo@pu.kielce.pl

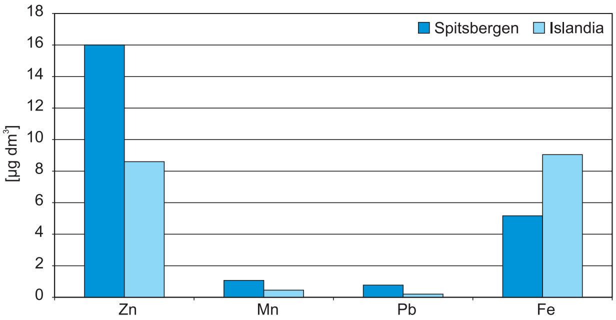

Fig. 3. Participate heavy metals in the water of supraglacial streams Spitsbergen and Iceland glaciers

Laboratory of Province Inspector Environmental Protection in Kielce (the certificate of Polish Center of Research and Certification AB 106) and in Laboratory Environmental Monitoring Institute Environmental Protection in Warsaw (the certificate of Polish Accreditation Center AB 337) where was defined content of zinc, lead, manganese and iron according to the norm adequate to each analyze.

The Waldemar glacier lays on the north-west Spitsbergen in Kaffiøyry region (78°33' ÷ 78°44'N and 11°43' ÷12°13'E). It is the valley glacier, alpine type, area 2,68 km^2 (Sobota 2003), It lays between 130 m a.s.l. and 490 m a.s.l. and consists of two parts (2,25 km^2, 0,43 km^2) separated by the middle moraine. The main, north part gradually falls into south-west and its surface is quite flat, with poorly marked faults on the sly surface (Lankauf 2002). The supraglacier rivers on the front are numerous but shallow.

The Ebba glacier is edging-valley (Rachlewicz 2003). It lays on middle Spitzbergen in Petuniabukta region (78°40' ÷ 78°50'N and 11°43' ÷ 12°13'E). It lays between 700 m a.s.l. and 1000 m a.s.l. The area was set on 25 km^2 (Hagen et al. 1993).

The Hans glacier – area 58,4 km^2 – lays in south part of Wedela Jarslberga Land (77º05'N and 15º38'E) on 500 m a.s.l. Its' length is 16 km, average surface slope under 2^o. It has meridian cost and goes into Hornsund fiord as a glacier cliff (Jania et al. 2003).

The Solhejmajökull is one of the edging glacier in south part of Mýrdalsjökull, in area of 596 km^2 and lays on vollcano-mountain massif over 1200 m a.s.l. high (Karasiewicz 2005). The face is on 110 m a.s.l. The length of tongue is about 15 km, and its area is about 45 km^2 (Sigurdsson 1998). The width of the glacier is about 2 km in the middle and it narrows into 1 km in the south part (Eiriksson et al. 1994). The capacity of ice in the glacier is estimated on about 12,3 km^3 and its thickness is 268 m average. Under the glacier is caldera of the volcanic Kalt system. The whole area lays on the south-east part of icelandic neovolcanic-ryft and that is why the earth-quakes and subice volcanic eruption are often there (Einarsson, Brandsdóttir, 2000).

The Fláajökull is the edging glacier of Vatnajökull and it lays in south-east Iceland. The length of the glacier is 15 km, and the average width is 2,5 km (Dąbski et al. 1998).

The researches show the differences in participation of marked heavy metals in supraglacier water between examined glaciers of Spitsbergen and Iceland. On Spitsbergen, except the iron, dominated all analyzed metals (ryc. 3). The concentration of zinc was about 46,3% higher, with average for tree glaciers 16,00 μg dm^{-3} , manganese 57,9% higher, with average 1,07 μg dm^{-3} and lead 74% with average 0,77 μg dm^{-3}.

On Iceland, iron has dominated in water of the examined glaciers with average 9,05 μg dm^{-3} and it was about 42,9% higher than average for the glaciers of Spitsbergen.

The analyze of the content of heavy metals in water floating from Spitsbergen glaciers shows, that the amount of heavy metals decreases from the south (Hans glacier) to north (Waldemar glacier). That confirms the results of Pacyna's (and others 1985) and Paatero's (and others 2003) researches about the sources of heavy metals on Svalbard.

On Iceland, the higher contet on heavy metals in water on glacier was marked on south-east part of the Iceland. In this case the sources of heavy metals should be recognized in local conditioning.

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Application of remote sensing and mathematical morphology of landscape for studying thermo-karst processes

Veronika Kapralova

Institute of Environmental Geosciences, Russian Academy of Sciences, Moscow, Russia

Climatic and technogenic changes render diverse effect on the environment, and it in turn affects economic activities. The area of development of permafrost rocks is one of the most sensitive to changes.

More than 60% of Russia is a permafrost zone. Accordingly problems of permafrost rocks and integrated with them exogenous geological processes are very actual for our country, especially for Siberia, which greater part is within the permafrost area and where the most of mineral deposits of Russia are placed.

As a rule, the frozen ground processes including those in a stage of stabilization or attenuation become more active under technogenic intervention and climatic changes, receiving a new impulse. And even more, they can reach higher degrees of intensity in their development. Besides other processes arise, which were not developed earlier at this territory.

Thermo-karst is one of geocryological processes especially sensitive to climatic changes. Thermo-karst is a process of formation of the closed negative landforms as a result of degradation of the soils containing ice.

Thermo-karst originates when the following conditions are satisfied:

a) Soils contain ice in the form of beds or schlieres;
b) Depth of seasonal thawing exceeds a depth of occurrence of underground ice or the soils containing schlieres of ice;
c) The water formed after ice thawing is filtered off in a thawed zone and due to it sinking of soil occurs.

Thermo-karst depressions much depend on types of underground ice and ground that thaw and on conditions of water flow.

Thermo-karst is mainly associated with loams, sandy loams especially containing big amount of particles of 0.05–0.002 mm, sandy silt and it is frequent within peatland.

Presence of close water-bearing horizons promotes thermo-karst processes, therefore the significant amount of thermo-karst forms is associated with alluvial sediments of ancient and actual river systems (Kachurin 1961).

Rate of thermo-karst processes depends on ice content of sediments: the higher it is, the faster ground destroys. Rate of destruction of walls of an exposure depends on latitude of locality, composition of deposits overlapping ice, ice thickness, exposure of slopes and depth of erosion base level. Persistent thermo-karst destruction and self-development of this process is possible at water body depth of 1,5 m and more. The stable water regime of thermo-karst lake is provided under condition that thawing stocks of underground ice in an ice complex make more than 35% of total volume of ground of an ice complex. At thawing of such amount of underground ice there is a self-development of thermo-karst lake irrespective of weather conditions of the year. In this case the further evolution is limited by underground ice amount and drainage conditions of locality. In case of formation of a drainage canal along which water from thermo-karst lake outflows, the water level drops sharply, leading to temporary stabilization of shores of the lake. Growth of thermo-karst lakes slows down sharply (Are et al. 1974).

One of the important problems is to find out principles of distribution and dynamics of thermo-karst development with the purpose of forecast of environment changes.

Studying regularity of distribution of thermo-karst forms, Romanovsky (1977) specifies their difference in different frost-temperature zones. These differences are caused by:

- genesis, scale of evolution of underground ice and ice content of the ground, which have zone features;
- the beginning of evolution of thermo-karst process, which in the north is connected mainly with alteration of seasonally thawed layer, and in the south in the greater extent with degradation of a permafrost;
- the probability and extent of thawing of underground ice increases in a direction from the north to the south.

Near to southern boundary of permafrost and especially outside its limits all types of ice and grounds containing ice thaw through.

Territories of distribution of sediments with low ice content appear to be hypsometrically above sunk sections (Romanovsky 1977).

In our work we use a method of mathematical morphology of a landscape – a branch of landscape science, investigating quantitative laws of construction of mosaics which are formed on an earth surface by natural units, and methods of the mathematical analysis of these mosaics – landscape patterns (Viktorov 1998).

Canonical initial mathematical models play a special role in mathematical morphology of a landscape. They deal with the patterns developed in uniform conditions, that is, at a constancy of major factors of landscape differentiation. The further combination of such models, in view of interaction of processes allows us to describe all variety of the morphological patterns developed in the diversified combinations of natural conditions. The fact of existence of such possibility to construct the model capable to describe complete variety of geometrical features of morphological patterns of the given genetic type with several equations is quite a real fact, though it is not obvious at first sight. Also, it should be noted, that basic equations do not depend on a lot of particular conditions, for example, a material structure of surface sediments, annual sum of precipitations, etc. Thus, the model allows us to examine the problems in general, i.e., obtaining a solution fair for a broad spectrum of natural geographical conditions.

We have investigated principles of surface change in connection with dynamics of thermo-karst processes. Researches were conducted on several reference sites. The section within West Siberian plain on the Pjakupur River has been taken as the basic, reference one. Additional sections are on Alaska, on Yamal peninsula and in Western Siberia. Aerial photographs and cartographical materials of different time dates were used: 1969, 1980, 1990 and 2000 and of different resolution from the one of more than 1 meter, up to 30 meters. The maps also were of different scale 1:100000 and 1:200000.

Measurement of size, area, and parameters of distribution of thermo-karst lakes was carried out. Various software products have been applied for the measurement: MapInfo Professional 7.5, ERDAS, Vektorizator, GST3, Excel and others.

The equations of the mathematical model of a morphological pattern for thermo-karst lake plains were used for the analysis of data and forecast constructions. They represent combination of the probabilistic mathematical relations reflecting the most essential geometrical features of the pattern. The equations include:

- probabilistic distribution of a number of thermo-karst lakes, which have appeared within a specified site during the given time interval (Poisson process).
- probabilistic distribution of changes of thermo-karst lake diameters (Winer random process relative to logarithms of diameters).

The analysis has shown:

- conformity on the whole of distribution of the centers of lakes to Poisson distribution with certain deviations;
- conformity on the whole of distribution of sizes of lakes to lognormal distribution with certain deviations;
- complex dynamics of similar territories.

The analysis of deviations of data distributions from theoretical model allows to outline influence of the varying factors acting over thermo-karst processes (climate change and the anthropogenous factor).

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Ice-cored moraines in the Petunia Bukta area – examples from Ragnar marginal zone

Leszek Kasprzak, Marek Ewertowski*

Institute of Paleogeography and Geoecology, Adam Mickiewicz University, Poznań, Poland

Ragnar glacier is localated in central Spitsbergen, in the area of the north part of Billefjorden – Petuniabukta (fig. 1 & 2). Marginal zone of studied glacier consists from four morphogenetical zones (fig. 3):

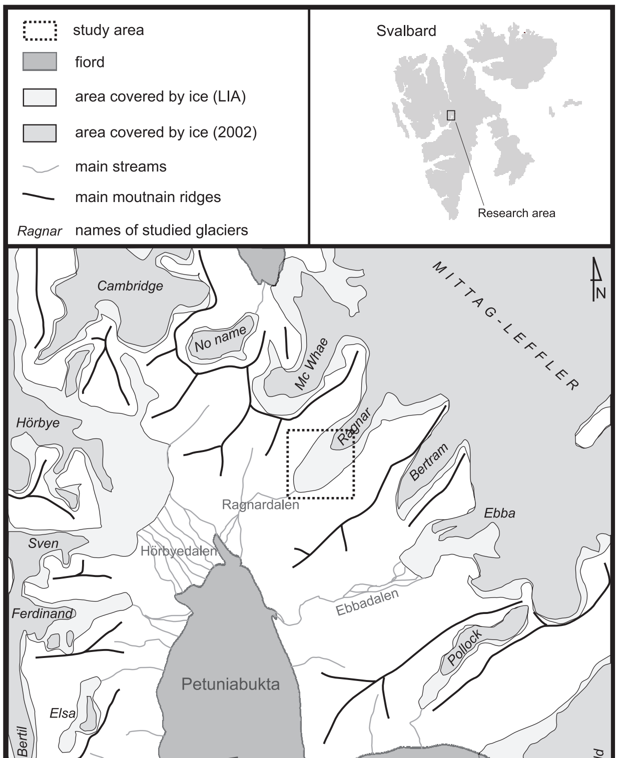

Fig.1. Localization of study area

1) Outer ice-cored moraine ridge (fig. 3 – A) – It has arcuate shape, length about 1,2 km and relative height about 30 m (fig. 4A). Its width in the central part is about 130 m, decreasing to 80 m in the wings. It was probably shaped during Last Glacial Maximum. The outer ice-cored moraine is built from lithofacies of coarse- to medium-grained, angular gravel (fig. 4B).

2) Inner ice-cored moraine ridges (fig. 3 – B) – This zone is stretched between outer moraine ridge and proglacial lake and is about 250 m width. It is comprised of four chains of hillocks probably with ice-cores (fig. 4C). Between them there are kettle-holes remnants after decay of dead-ice blocks, some of them are filled with water (fig. 5A). The longer axis of hillocks and kettle-holes imitate the shape of the outer ice-cored moraine ridge (fig. 5B). Hillocks are built usually from fine-grained, rounded

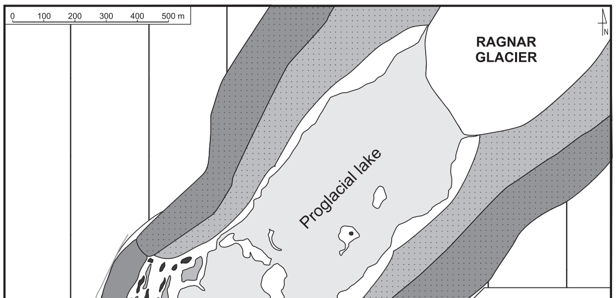

Fig. 2. Elements of marginal zone on Ragnar glacier

* e-mail: evert@amu.edu.pl

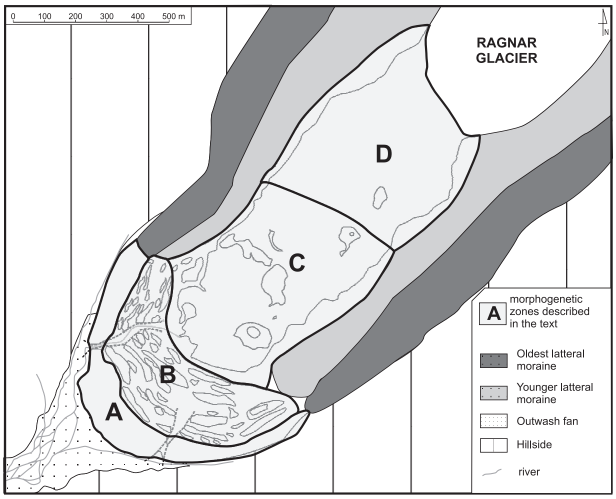

Fig. 3. Morphogenetic zones (A-D) on the Ragnar glacier forefield

down of dead-ice blocks. Here, lithofacies of sand and silt are dominant (fig. 6B & 6C)

4) Proximal zone of proglacial lake (fig. 3 – D) – this zone has a direct contact with ice-margin. There are no landforms above water level except for one island built completely from ice with very thin (3–5 cm) debris cover.

Interpretation

Connection between shape, orientation and lithology of ice-cored moraines and rate of Ragnar glacier recession (fig. 7):

A – outer ice-cored moraine ridge was formed during the equilibrium state of ice-front at the maximum extent (MEL). Majority of debris were delivered by supraglacial transport so it is coarse-grained and angular. During relatively long time of ice-front stay a large amount of debris were delivered so the large-size ice-cores could be preserved.

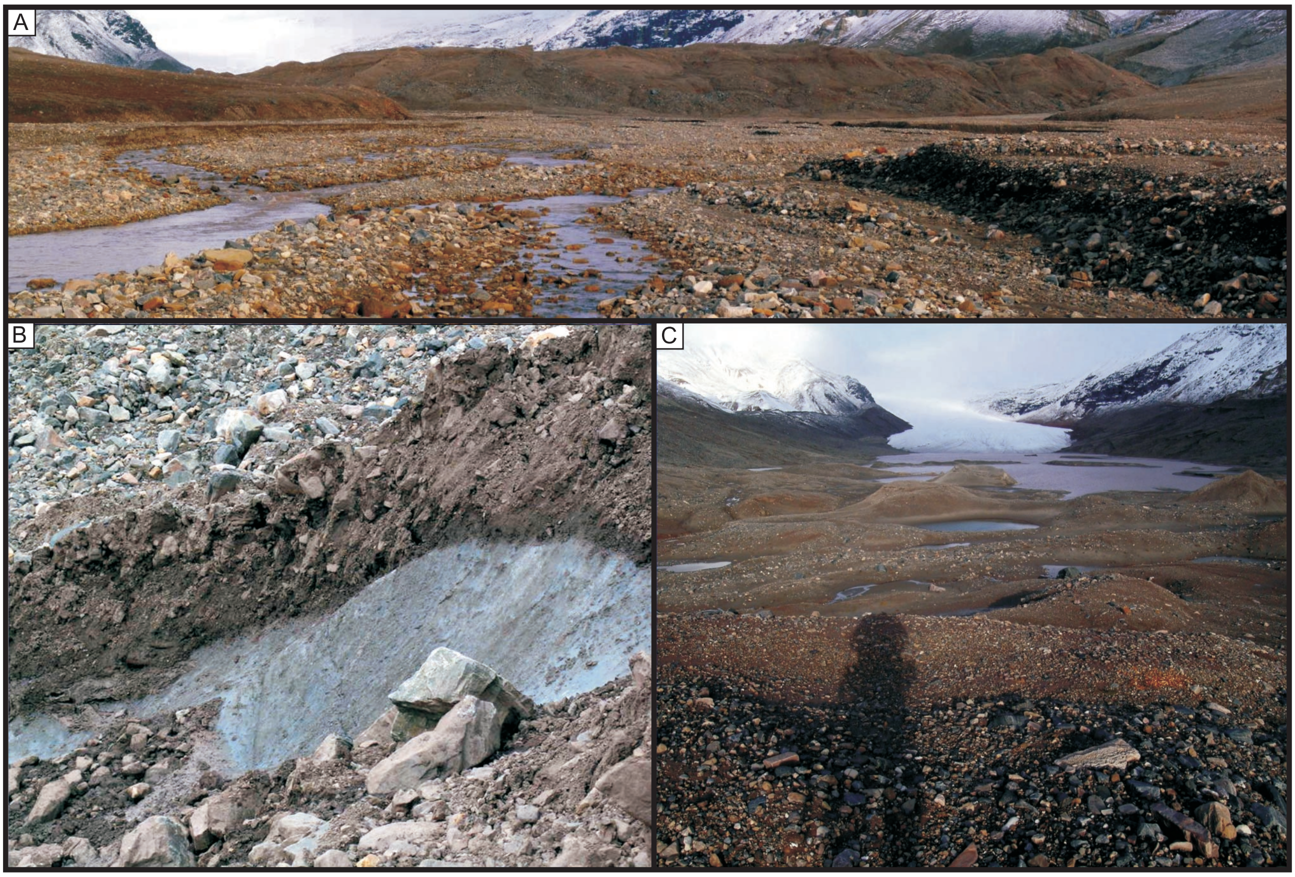

Fig. 4. A – Outer (oldest) ice-cored moraine; B – Exposure of ice cored, Outer ice-cored moraine; C – Zone of inner ice-cored moraines

gravel and sand. Lithofacies of silt and sand occur in the holes.

3) Distal zone of proglacial lake (fig.3 – C) – one of the most characteristic features of this zone are small islands (fig. 6A). They are the upper parts of ice-cored ridges situated 1–3 m above water level. They are built from fine-grained gravel and sand, usually massive or horizontal bedded. Some of the islands have inner water basins, created after melt-

B – after long period of ice-front stays Ragnar glacier started to retreat. At first, recession was relatively slow and interrupted by short-term stays. The chains of hillocks in the inner zone of ice-cored moraine are interpreted as a record of this short-time

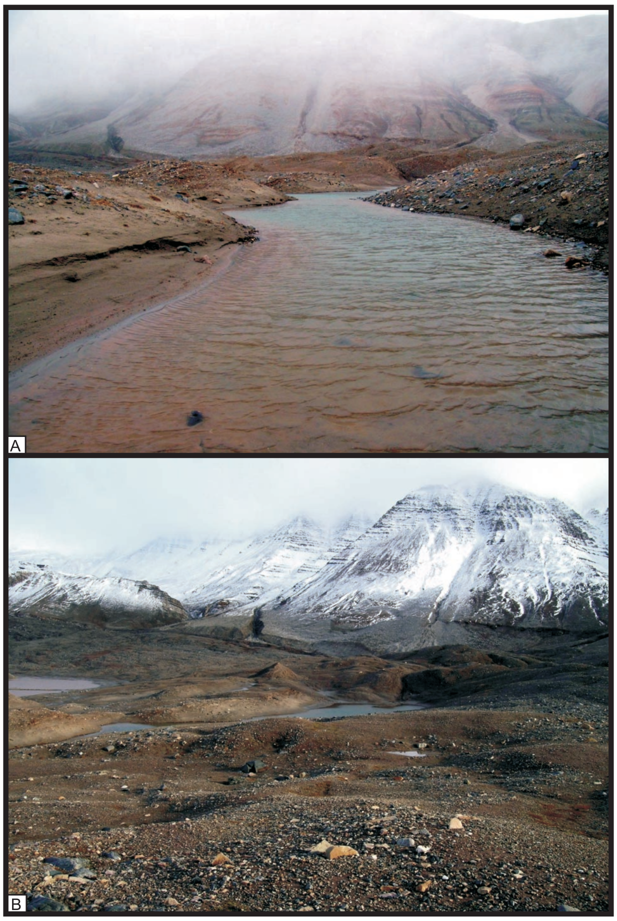

Fig. 5. A – Joined kettle-holes, zone of inner ice-cored moraines; B – Zone of inner ice-cored moraines

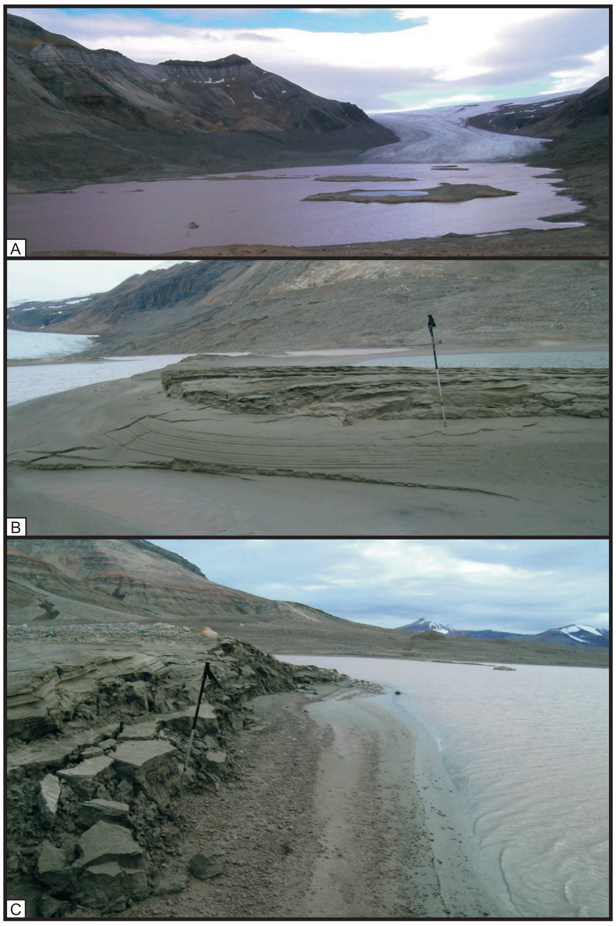

Fig. 6. A – Distal zone of proglacial lake; B – Lithofacies of horizontally laminated sand, distal zone of proglacial lake; C – Lithofacies of sands and gravels in distal zone of proglacial lake

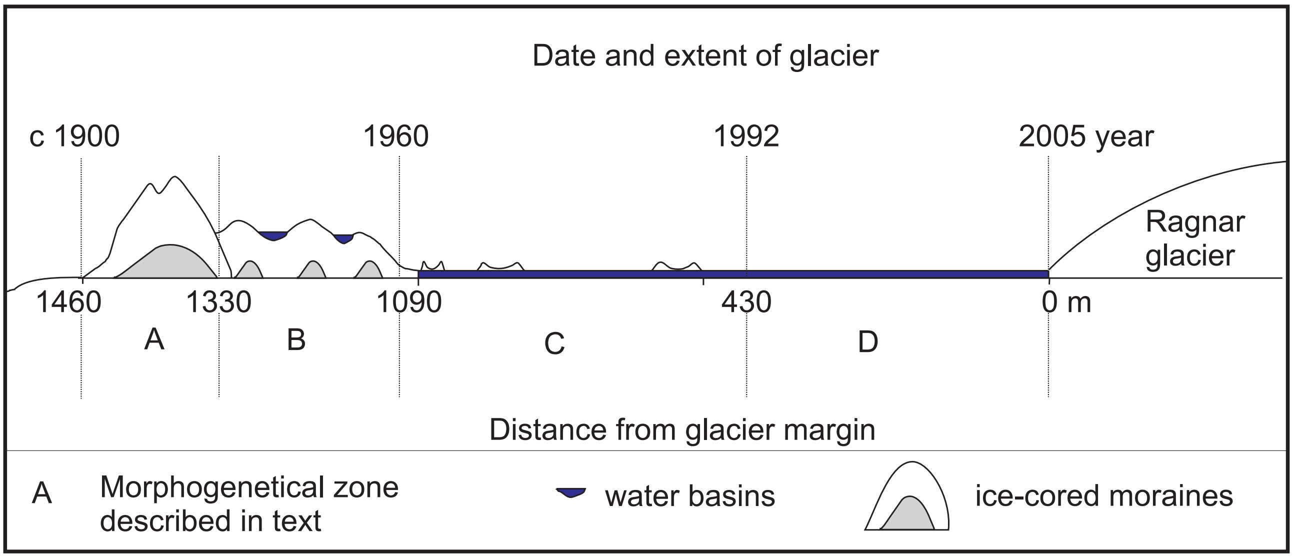

Fig. 7. Morphogenetic zones and rate of recession of Ragnar glacier

events. Quantity of debris was sufficient to preserve only some of the ice-blocks, the rest melted and the kettle-holes were created, still followed, however the shape of the ice-margin.

C – further increase of recession rate had two consequences. The first one was larger amount of meltwater and formation of proglacial lake. The second one was increasing quantity of debris depositioned at ice margin and consequently poorer conditions for creation and preservation of ice-cores. Only isolated ice-blocks survived and formed characteristic island with inner water basins.

D – During the last 20–30 years rate of recession was very high, so depositional effectiveness of glacier was very low. Debris delivered contemporary to ice-margin is immediately removed by meltwater so there is no possibility of any accumulation on ice surface.

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Present-day geomorphological activity in the Arctic

Andrzej Kostrzewski, Grzegorz Rachlewicz*, Zbigniew Zwoliński
Institute of Paleogeography and Geoecology, Adam Mickiewicz University, Poznań, Poland

Introduction

The landscape uniqueness of the Arctic polar zone manifests itself in morphological traces of older glaciations and marine transgressions, areas of present-day glaciations, multi-year permafrost, multi-year snow covers, deglaciation processes variable in time and space and resulting in an expansion of ice-free areas, multi-directional geosuccession, and finally in the various responses of the Arctic peoples to landscape changes and the growing human impact. The present study rests on the following assumptions:

– the state of geoecosystems in the Arctic polar zone is the product of their former and current development varying over space and time;
– the operation of present-day geoecosystems is affected by climatic variability and a growing human impact;
– the seasonal rhythm of polar geoecosystems is disturbed by processes of above-average and extreme nature which bring about changes in their internal structure or lead to the disappearance of the existing geoecosystems and the emergence of new ones; and
– a research on the present-day polar geoecosystems of the Arctic should be organised, integrated, and based on comprehensive projects, both national and international, involving the participation of the Northern peoples.

The abrupt landscape changes taking place over a period shorter than the life span of a single generation have currently become readily visible in the Arctic region. They can be due to a wide variety of natural causes, whether endogenous or exogenous, or to the increasing, multi-directed human activity. At present, however, their principal cause is believed to be climate change at a variety of spatial scales. The paper presents examples of contemporary changes of morphologic surfaces from all the territory of the Arctic, with special attention paid to Spitsbergen.

Spatial extent of the environmental changes in the Arctic

Irrespective of the adopted criterion of delimiting the boundary of the Arctic (Fig. 1, Kostrzewski et al. 2006), one can note ever greater changes in its southern course after the Little Ice Age has ended (Halsey et al. 1995; Laberge, Payette 1995; Osterkamp, Romanovsky 1999). They are first of all regional in nature (Burgess et al. 2001; Osterkamp, Romanovsky 1999; Romanovsky 2006). Thus, a distinct shrinkage in the area of permafrost can be observed in Alaska, the valley and delta of the Mackenzie, Spitsbergen, and the Dvina and Pechora Plains; the decay rate is somewhat slower in the Canadian High Arctic, northern Norway, the Kola Peninsula, and the northern regions of West Siberia, while the extent of permafrost stays the same or tends to grow slightly in the middle reaches of the Yukon, the eastern Canadian Arctic (in the late 1980s and early '90s), and the eastern margins of East Siberia[1]. The cause of those changes is the northward succession of boreal forests pushing far-

* e-mail: grzera@amu.edu.pl

1 Detailed data are supplied by the international research programmes, Circumpolar Active Layer Monitoring – CALM, and the Global Terrestrial Network for Permafrost – GTN-P.

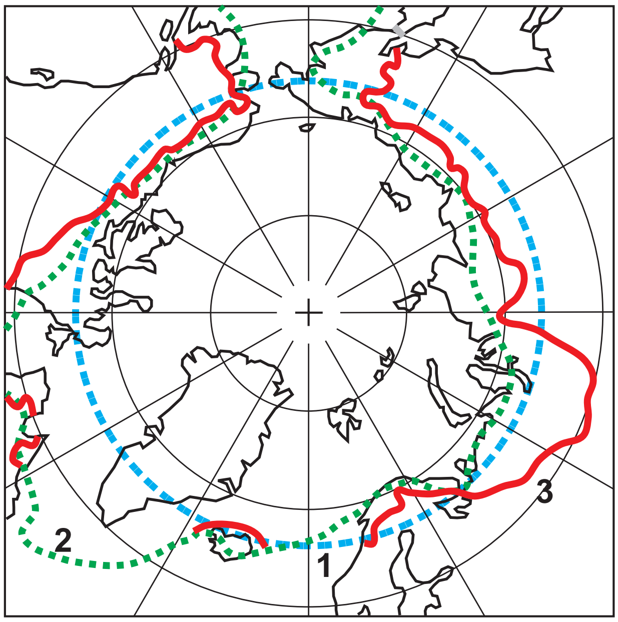

Fig. 1. Different boundaries of the Arctic
1 – northern polar circle, 2 – course of July isotherm equal to 10°C, 3 – northern limit of boreal forest

Fig. 2. Comparison of Hans Glacier' front extent in the 50-years interval
A – year 1957 (Archive of the Institute of Geophisics), B – year 2007 (photo A. Nawrot)

ther north their usual limits in many places. This development has been brought about by several types of feedback, e.g. warming, an increase in the thickness of the active layer, increased evaporation, a decline in the area of wetlands, etc. Irrespective of the current trends in the shifting of the Arctic limits, one should note that the prepared scenarios anticipate their northward shift by hundreds of kilometres, and this means changes in the functioning of terrestrial geoecosystems that will embrace an area of thousands of square kilometres.

Environmental changes in the Arctic

The changes observed in the polar environment, especially in the Arctic terrestrial geoecosystems, are connected with the many signals of global and regional changes that conform to the global change pattern. The latter has been particularly discernible over the last half-century (Fig. 2), which has also been characterised by an ever-intensifying human impact (IPCC 2002, Macdonald et al. 2003). It is assumed (ACIA 2004) that in the Arctic the mean annual temperature grew in the second half of the 20th century by 2-3°C in Alaska and Siberia, while it dropped by about 1°C in the southern regions of Greenland. The extent of floating ice has shrunk by 8% over the last three decades, but the changes are most conspicuous during summer when the marine ice area dwindles by 15–20%. Among the most significant environmental changes noted in the polar regions, Zwoliński (2007) lists the following:

– air temperatures frequently exceeding the hitherto absolute maxima,
– an increase in annual precipitation totals, first of all in the form of rain, also during the cold period,
– cold periods becoming milder and shorter,
– transitional periods becoming longer: spring coming earlier and autumn ending later,
– a decrease in the thickness, persistence and area of the sea-ice cover,
– an increase in the number of icebergs from intensively calving glaciers,
– an increase in the temperature and a decrease in the salinity and density of ocean waters; changes in the thermohaline circulation,
– an increase in the level of the world ocean,
– intense ablation and rapid recession of the majority of polar glaciers,
– a decrease in the area of nival covers,
– intensive thawing of multi-year permafrost, mainly in continental parts,
– changes in the water cycle manifested by an increase in the surface runoff in streams and a shortening of the period of freezing of streams and lakes,
– an increase in the area of some wetlands and a decrease in others,

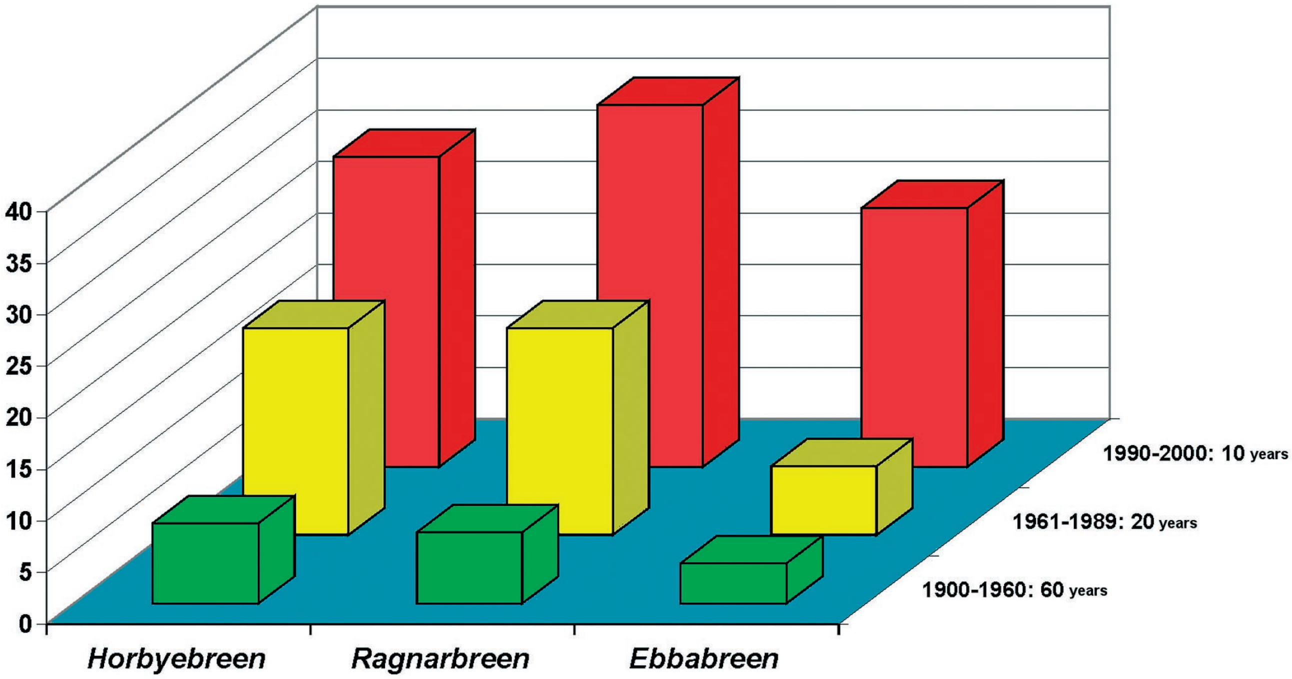

Fig. 3. Rates of glacier front recession for glaciers in the vicinity of Petuniabukta, Central Spitsbergen (acc. Rachlewicz, Szczuciński (2002), changed)

– a northward shift of geoecological, including vegetation, zones,
– changes in the carbon cycle in the geoecosystems manifesting themselves in an increase in biogenic carbon dioxide and methane, and
– an increase in the frequency and magnitude of forest fires.

All those symptoms of climate change affect the terrestrial geoecosystems of the Arctic to a greater or lesser degree. The ever-growing role of rock geoecosystems, crucially dependent on glacier and nival geoecosystems, results from intensive glacier recession (Fig. 3) and the melting of permafrost and snow covers. It has been estimated (ACIA 2004; Haeberli et al. 1989) that losses in the cumulative volume of glaciers in the other half of the 20th century amounted to nearly 500 km^3 in the North American Arctic and more than 100 km^3 in the Russian Arctic, and it was only Eurasia, mostly Scandinavia, which recorded an increase of some 200 km^3.

Geomorphological changes in the Arctic terrestrial geoecosystems

The observed geosuccession phenomena are especially conspicuous in paraglacial areas. Throughout the Arctic, on each of the continents one can find numerous examples of geosuccession changes. One of the most striking is the foreland of the Breidamerkur ice-cap (Iceland), whose margin has retreated several kilometres over the last 30 years leaving behind a diversified morphological surface with a variety of morphogenetic and sedimentary environments (Fig. 4). The stabilisation of such areas is a rather slow process and each year one can note

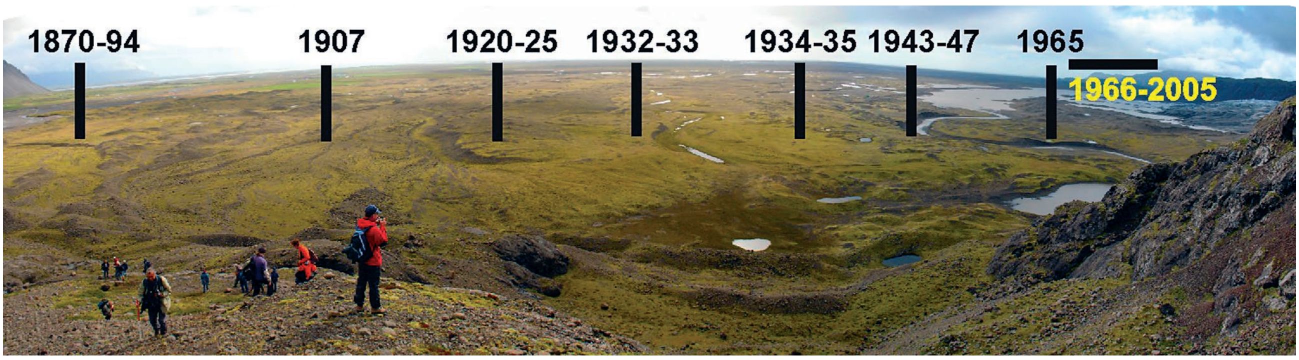

Fig. 4. Geosuccession traces of fast recession of the Fláa Glacier (Iceland); black labels depict moraine-recessional stages (acc. Dąbski 2002) during last 130 years (photo Zb. Zwoliński)

Fig. 5. Surface of alluvial fan modified by "desertification" processes, southern side of the Ebba Valley, Central Spitsbergen (photo Zb. Zwoliński)

morphological changes in the freshly forming topographic surface.

The intensive glacier retreat, downwasting of nival covers, thawing of permafrost and an increase in precipitation over land areas have altered the water cycle: the amount of water in streams, rivers and surface bodies has grown. The six principal Eurasian rivers: the Dvina, Pechora, Ob, Yenisey, Lena and Kolyma, discharge a total of 2,000 km^3 of fresh water (ACIA 2004) to the Arctic Sea, up 7% from the late 1930s (Peterson et al. 2002). In the 21st century, in winter European rivers carry some 100 km^3 of water more than they did in the mid-20th century (ACIA 2004). The increase in the discharge of water by Arctic rivers accelerates thermal erosion of their bank scarps with their underground glaciation (Siberia) and reduces the salinity of sea waters in the Arctic Ocean, especially in the coastal zone (Arctic Change 2006).

The rising level of the world ocean makes abrasion processes more intensive in the Arctic coastal zones. It is especially readily visible on high coasts, cliffs built of loose rocks of glacial origin, where frequent landslides occur. Abrasion is intensive e.g. in the coastal morainic zones of Spitsbergen or along the distributaries of the Mackenzie delta. Abrasion processes on the Arctic coasts and increased discharges of the pan-Arctic rivers enrich the coastal sea waters with sediment, but such developments as the building up of spits or beaches are rare and occur at a more significant rate only locally.

One can also observe an increase in the proportion of high winds, which reinforce abrasion processes through an increase in waving which affects the far-from-stable, fresh shores in the immediate neighbourhood of polar oases, as well as deflation and accumulation processes of eolian and niveo-eolian deposits. The deposits come from supraglacial moraines and marginal zones, but also from areas drying as a result of permafrost melt-out. An unfavourable development is the blowing out of fine-grained sediment covers in tundra-supporting areas, especially in the case of pioneer tundra moving onto land newly available for colonisation. Sometimes this process can even resemble desertification as an effect of the morphogenetic sequence of the thawing of the active layer, permafrost melt-out, drainage of meltwater, and the drying of land through evaporation and deflation, i.e. a typical example of a geosuccession (Fig. 5).

A result of the increase in precipitation in the Arctic regions has been a higher frequency of ephemeral snow covers during summer. They do not last long and usually form in the highest parts of elevated areas, but even so they do affect the summer water cycle. Such occurrences have been observed in Spitsbergen, among other places. In 2002 the melting of snow and ice covers in Greenland was recorded up to an altitude of 2,000 m a.s.l. Also the earlier coming of spring and the later ending of autumn changes the duration, extent and thickness of nival covers, both on glaciers and their surrounding areas.

The Spitsbergen area (77–80°N), because of its location within a very dynamic maritime influences, is especially susceptible to processes of intensified activity of energy circulation and matter transfer. It is favoured by general tendency of climate warming, with the elongation of the period of morphogenetic processes activity, as well as the rise of frequency of above-average phenomena. A number of signs of environmental changes on areas of Polish research activity on Spitsbergen were denoted in extensive monographs edited by Kostrzewski, Pulina, Zwoliński (2004) and Kostrzewski, Zwoliński (2003).

Contemporary changes of Arctic climate are influencing a number of geomorphologic processes leading to essential structural and external aspects transformations of the landscape. These changes are visible in the functioning of contemporary terrestrial geoecosystems of Spitsbergen, among which one can rank (e.g. taking into account Ebba Valley, the area of investigations of Poznań University – fig. 6):

Fig. 6. Spatial pattern of selected morphogenetic domains in the Ebba Valley, Central Spitsbergen (photo Zb. Zwoliński)
Morphogenetic domains: 1 – glacial, 2 – marginal, 3 – braided proglacial, 4 – meandering fluvial without vegetation cover, 5 – meandering fluvial with vegetation cover, 6 – littoral of raised marine terraces, 7 – estuarial, 8 – littoral of bay/fjord, 9 – fluvial of alluvial fans, 10 – slope with talus cones, 11 – rock weathering, 12 – weathering within waste cover, 13 – permafrost (periglacial), 14 – eolian.

A – in morphogenetic (hierarchical) depiction:
- weathering sub-system,
- nival sub-system,
- glacial sub-system,
- glacifluvial, glacilimnic, glacimarine sub-systems,
- fluvial sub-system of proglacial rivers: braided and meandering,
- denudational-fluvial sub-system,
- limnic sub-system,
- eolian sub-system,
- litoral sub-system,
- permafrost related sub-system,

B – in spatial (cascade) depiction
- field sub-system,
- slope sub-system,
- valley sub-system,
- piedmont sub-system,
- coastal sub-system.

Presented sub-systems are not depleting the full inventory of morphogenetic and sedimentary environments of polar areas but are realizing about two facts: a large diversification of quality of factors and morphogenetic processes in an apparently monotonous polar environment and a very fast spatial migration of these sub-systems observed over past 25 years, referring to changes of their extents – some of them growing and some shrinking.

Geomorphological trends within the Arctic terrestrial geoecosystems

The polar research to date and scenarios of development of the Arctic polar regions indicate that the Arctic landscapes have been undergoing rapid changes recently. Their pace and intensity over the last 100 years have varied owing to climate change connected with the ever-growing human impact. The Arctic is an especially interesting area whose transformation brings about global changes in the surface of the Earth.

According to the models and scenarios of ACIA (2004), for the Arctic regions the increase in air temperature should be 5°C (scenario B2) or 7°C (scenario A2) as compared with the 1981–2000 figures. It should be the steepest in Siberia and the eastern Canadian Arctic. Precipitation in the form of rain should grow by some 20%, mainly in summer. The anticipated areas of increased rainfall are the North American Arctic and northern Russia, while northern Scandinavia is expected to receive lower rainfall. Sea ice may be expected to dwindle by as much as 50%. According to some models, in the summer of 2100 there may be no ice cover whatsoever on the Arctic Ocean. The rise in the level of the world ocean is expected to range from 10 cm to 70 cm in both scenarios, A2 and B2. The extent of the terrestrial ice cover should shrink by about 20%, mainly in spring, causing an earlier start of flow of the pan-Arctic rivers, whose discharges may grow by 10–25% in winter and spring to decline in summer owing to increased evaporation. It should be emphasised that all the analysed scenarios anticipate roughly the same increase or drop in the parameters in question over the 21st century. However, observation to date has shown that their pattern may be different.

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Landform Analysis, Vol. 5: 47–48 (2007)

The impact of frost action and nivation on relief formation of Velebit Mt. (Croatia)

Nina Lončar*, Perica Dra en

Departmant of Geography, University of Zadar, Zadar, Croatia

Although the Velebit is a relatively low mountain, situated in the moderate climate zone, there exist periglacial processes in relief modelling in its highest part. Their appearance is caused by interdependance of geological, geomorphological, climatic, vegetational and pedological influences, but also long antropogenic and zoogenic influences which accelerate the influence of periglacial processes in the relief modelling. Among periglacial forms the features which originated from the activity of nival and frost processes can be singled out.

The Velebit Mountain has been attracting scientist since the 18th century (Hacquet 1785) who research its geomorphological characteristics, only in the middle of 20th century Poljak (1947) and Rogić (1958) in their works paid attention to the problems of periglacial modelling of relief of the Velebit Mountain. In respect of the problematic of this work, the main way of researching has been connected with the terrain investigations.

At the highest parts of the Velebit mountain (above 1400 m a.s.l.) besides the karst and derasion processes there are periglacial processes which are of great importance for relief modelling. In respect of intensity and lasting of periglacial processes, the climatic elements are very important.

The border position of the Velebit Mountain between the coast and inland area is expressed by the climatic characteristics of the mountain ridge. During the cold half of the year in the higher parts of the Velebit (above 900 m a.s.l.) there is a frequent appearance of cold, ice-cold and chilling days, which cause the freezing of water in the rock fissures and appearance of the cryogenic process. Although the daily oscillation of the air temperature during the winter months at Zavižan (1594 m a.s.l.), and espe-

Fig. 1. The avalanche accumulation (South Velebit)

Fig. 2. Stone rivers (stream) in the sinkhole (South Velebit)

* e-mail: loncarn@yahoo.com

cially Baške Oštarije (924 m a.s.l.) is only about 2°C, because of frequent oscillation of the temperature around 0°C there is dissolving and refreezing and, due to that, the strong mechanical wearing out of the rocks appears.

There are two main winds which dominate in the Velebit area. Bora, dry, stroking and cold wind, most frequent on the SW slope and warm and moisture Jugo on the SE slope. Despite of large amount of snow, due to influence of those winds many parts of Velebit Mt. stay uncovered (without termic protection). These uncovered parts are more exposed to the frost action.

In periglacial modelling of relief the structural characteristics of the rock complex are of great importance, especially a density of appearance of the primary and secondary fissures and holes, and also the inclination of the layers. The domination of layers (mostly of carbonate sediments) which incline toward sea with general direction of NW-SE is at the highest parts (above 1200 m a.s.l.) of the mountain.

The snow moving in the shape of avalanche is mostly expressed on the slopes with inclination from 30° to 60°. On the slopes covered by wood vegetation, as well as on the ones without it where slope inclination is mostly less than 30°, there is a slow creeping of the snow cover. On the slopes overgrown by wood vegetation, because of the snow cover weight which creeps down the slope and because of the suffosional activity of the water (snow-water, but also the other precipitation) there is bending of the trees at their basic part. On the slopes without wood cover with the slope inclination over 30° the appearance of the snow avalanches is frequent (Perica et al. 2002). Their appearance can be found at the highest range on the SE slope of the South Velebit. There can be found the avalanche accumulations – cone, at the foot of slopes which, can be detect by the mixture of unsorted karst and wood fragments and stone blocks (Fig. 1).

Cryofraction and solifluction are the most expressive of all periglacial processes. The cryogenic process is most frequently connected with escarpments, sinkholes, hollows and uvalas of karst polje at the highest parts of the Velebit mountain. Due to the ice contractions along the fissure there is breaking of the stone complex the extracted parts of which and the influence of gravitation are being accumulated at the footslopes, at the bottom of sinkholes, hollows and uvalas of karst poljas, in the shape of talus cones, colluvial cones and colluvial fans (Fig. 2).

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Weathering rates, natural organic matter and global climate change: Are they related?

Donald L. Macalady*

Professor of Chemistry and Geochemistry, Colorado School of Mines, Golden, CO 80401 USA

James F. Ranville

Associate Professor, Department of Chemistry and Geochemistry, Colorado School of Mines, Golden, CO 80401 USA

Ola Magne Sæther

Research Scientist, Norwegian Geological Survey (NGU), Trondheim, Norway

Some things concerning the title of this presentation are certain. Global warming is happening (Hinzman 2005). As polar ice cover recedes, senescent organic matter will be exposed and partially degraded. Primary productivity in polar regions will increase, producing increased loads of detrital organic matter (Striegl et al. 2005, 2007). Natural organic matter (NOM) and concomitant small chain organic acids will be produced in increasing quantities (Michaelson et al. 1998, Benner et al. 2004, Kawahigashi et al. 2004, Frey, Smith 2005). However whether or not this will have dramatic effect on mineral weathering rates is far from certain (Antweiler, Drever 1983, Ranville, Macalady 1997, Rauland-Rasmussen et al. 1998, Anderson, Drever 2000). Higher temperatures may have an effect (Veibel 1983)? Increased levels of carbon dioxide and, probably, lower freshwater pH values, will almost certainly increase weathering rates (Raymond, Cole 2003). Will NOM exacerbate or mollify such weathering rate increases? Will effects be different depending on soil cover and soil type (Jardine et al. 1989a, b, 1990, McCarthy et al. 1993, 1996). This presentation will provide data, both new and from the published literature, to support arguments on both sides of the issue.

* e-mail: dmacalad@mines.edu

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Measurements of selected water balance components in Ebbaelva catchments, Svalbard – pilot study

Marek Marciniak*, **Anna Szczucińska**
Institute of Physical Geography and Environmental Planning, Adam Mickiewicz University, Poznań, Poland

Krzysztof Dragon
Institute of Geology, Adam Mickiewicz University, Poznań, Poland

Introduction

The research of the water balance in polar catchments are relative unique. The water balance components are often estimated approximately or calculated using conceptual models. Killingtveit et al. (2003) reviews and summarizes all known previous water balance studies in Svalbald and give only several examples of water balance computation in unglaciated catchments. The water balance research was also undertaken in region of Bellsund (Bartoszewski 1988) and Horsund (Pulina et al. 1984). Even more difficult to recognize is water balance calculated for glaciated catchments, where the intensive water circulation take places only in summer months. In the remaining part of hydrologic year water circulation is connected with solid phase of water (sublimation, glacial retention, etc.). Moreover, in case of examination of water balance in polar catchments the complexity of this type of environment should be taken into consideration (e.g.: retention of water connected with permafrost, the seasonal effects connected with polar day and night, snow driffing between catchments, irregular locations of polar meteorological stations and its seasonal functioning in summer period.

The water balance formula includes the following water circulation components:

$$P = E + Qr + Qp + Qg + \Delta R \qquad (1)$$

where:
P – precipitation [mm],
E – evaporation [mm],
Qr – river runoff calculated for catchments area [mm],
Qp – superficial runoff calculated for catchments area [mm],
Qg – groundwater flow calculated for catchments area [mm],
ΔR – retention changes [mm].

The aim of this study is computation of the water balance components which are the most unique estimated in polar environments: actual evaporation and groundwater flow. The methods of research are illustrated by the result of pilot studies performed in Ebbaelva catchments in Svarbald.

The Study area

For the pilot studies (in 2005 and 2006) and for research planned in the 2007 summer season a glaciated catchment of Ebba River (Ebbaelva) was selected. The river enters Petunia bay (Petuniabukta) – part of Billefjorden, in central Spitsbergen (Fig. 1). The catchment (51.5 km^2) consists of Ebba valley (Ebbadalen), which is covered in more than 50% by Ebba and Betram glaciers (Ebbabreen and Bertrambreen). Their meltwaters are the main supply of the Ebba River, which also collects waters from several streams flowing down from mountain ridges in the north and south. The average summer discharges of the Ebba River are between several and 20 m^{3}s^{-1}. The Ebba glacier belongs to polythermal type, so water outflows are active also in

* e-mail: mmarc@amu.edu.pl

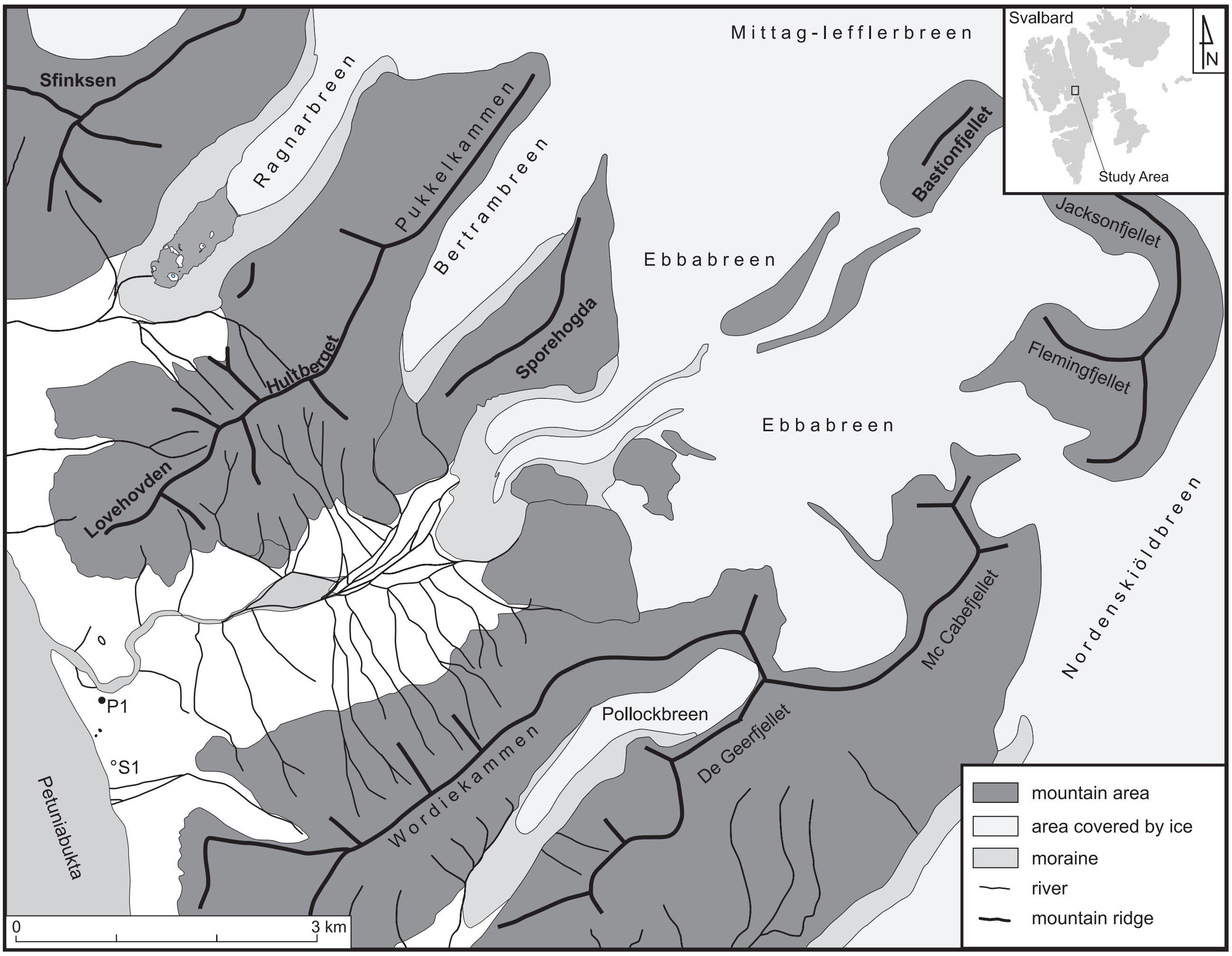

Fig. 1. The study area

winter time (Rachlewicz 2003a; Gibas et al. 2005). The catchment was previously often investigated, for instance, in regard to water discharge (Kostrzewski et al. 1989; Choiński 1989; Rachlewicz 2007) and meteorological conditions (Kostrzewski et al. 1989; Rachlewicz 2003b, c). The most of the catchment is in permafrost zone. The maximum active layer thickness is at the end of summer and may reach even 2 m (Gibas et al. 2005) – it makes existence of groundwater component of the Ebba River supply being very likely.

Methods and examples of their applications

Measurements of evaporation from free water surface are taken with an evaporation pan fixed within a measuring station (Figs. 2, 3). The main element of the measuring station is a pan (A) filled with water, of surface area about 1 m^2. Its upper rim should be installed slightly above land surface. A detail measurement of water surface elevation in the pan at the beginning and end of the period of observation allows assessment of the evaporation. The measurement is taken in a limnometric well (B). Its main part is a 4 cm in diameter pipe (B3), which is placed vertically on a base disc (B1) in the pan. For levelling of the base disk is used a level (B2) installed on it. The well is submerged in water, in such a way that free water surface is more or less in a half of its height. The main purpose of using the well is to eliminate water surface waving. In the well is a vertically installed sharp ended micrometric screw (B4). A male screw of the micrometric screw has vertical distance between spiral ridges of 1 mm. In consequence, screw rotation of 3.6 o gives vertical displacement of 0.01 mm. On the upper rim of the well is mounted a turn buckle (B6) and scale (B5). A task for an investigator is to recognize moment when the sharpened end of the screw (B4) touches the water surface and to read its position on the scale (B5).

It is very difficult to take precise daily measurements of the evaporation due to very small changes in water surface position in the pan. It is the reason for reporting and analyzing at least 10 days or two weeks sums of the evaporation. Additional difficulty during the interpretation of the results is a necessity of taking into account a correction for rainfall. It requires to take measurements of the rainfall with precision of at least the same accuracy as the measurements of evaporation.

In polar conditions the evaporation is smaller so its measurements are even more difficult. It is the reason for proposed below four major modifications of the evaporation pan for polar conditions. These modifications are results of experience gained during a pilot measurement series taken in Ebbadalen in summer 2006.

- First of all it is necessary to eliminate possibility of refilling water in the evaporation pan (A) through rainfall or snowfall. Correction for precipitation of the evaporation measurements requires to take measurements of the precipitation with precision of at least the same accuracy as the measurements of evaporation, which is unrealistic taking into account commonly used rain gauges. It is suggested to install a kind of tarpaulin roof (E) of area significantly bigger than the evaporation pan (A) on at height of 1 m above it. The roof will be installed on stakes (F) and fixed with ropes (G). It will unable rain water to reach the evaporation pan even during oblique direction of rainfall but will allow free air movement above water in the evaporation pan.
- Secondly, during taking the measurements in the evaporation pan a cover (C) should be placed on it to eliminate waving of the water surface due to wind. During the pilot studies such a waving, caused by even very fair wind, made the correct reading of the water surface elevation impossible. In the cover a window made of plexy and small opening will be left to make possible the observation and operation of the micrometric screw. Except of the time of the measurements the cover should be displaced.
- Thirdly, the sharp end of the micrometric screw should be equipped in electronic system for optical and acoustical signalization (D) of the moment of contact with water surface. For the confidence of an observer the system should be mounted on the cover (C). It will help to eliminate difficulties in determination of the moment of reaching the water surface by the screw, which exists in classical evaporation pan.
- Fourth, the evaporation pan (A) should be covered with a net, which will protect the pan from birds and other animals (damage, drinking water etc.). The net shall not influence the evaporation.

All of the above listed improvements should make the field measurements of daily sums of evaporation possible in polar conditions during summer. Laboratory tests proved that evaporation pan has sufficient measurement resolution for such a purpose.

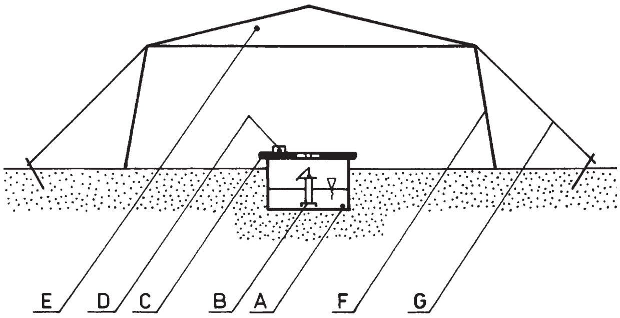

Fig. 2. Polar evaporymeter in the field position
A – evaporation pan, B – limnometric well with a micrometric screw, C – cover, D – electronic contact indicator, E – tarpaulin roof, F – stakes of roof, G – tighten rope

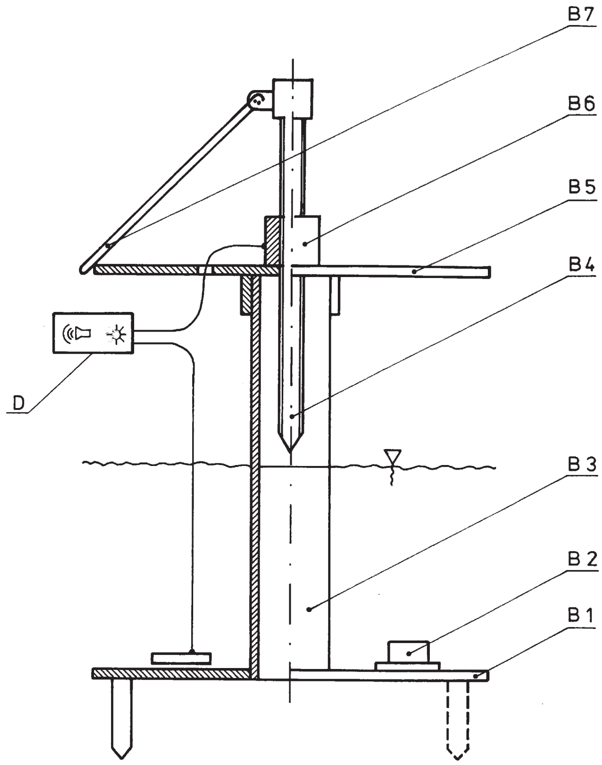

Fig. 3. Limnometric well of the polar evaporymeter
B1 – the base disc with tree leveling screws, B2 – spirit level, B3 – pipe of limnometric well with a linear scale, B4 – micrometric screw, B5 – angle scale, B6 – a nut of micrometric screw, B7 – pointer, D – electronic contact indicator

The groundwater flow measurements

For estimation of groundwater level fluctuations in the Ebbaelva River catchment the 0,7 m deep piezometer P1 was installed. This piezometer was located about 10 m from stream channel near the river mouth (Fig. 1). In this region sands and gravels filling the river valley during drilling were detected. The daily measurements of water level in the period between 15.08.2005 and 20.09.2005 were performed. Also the water level fluctuations in the river and run-

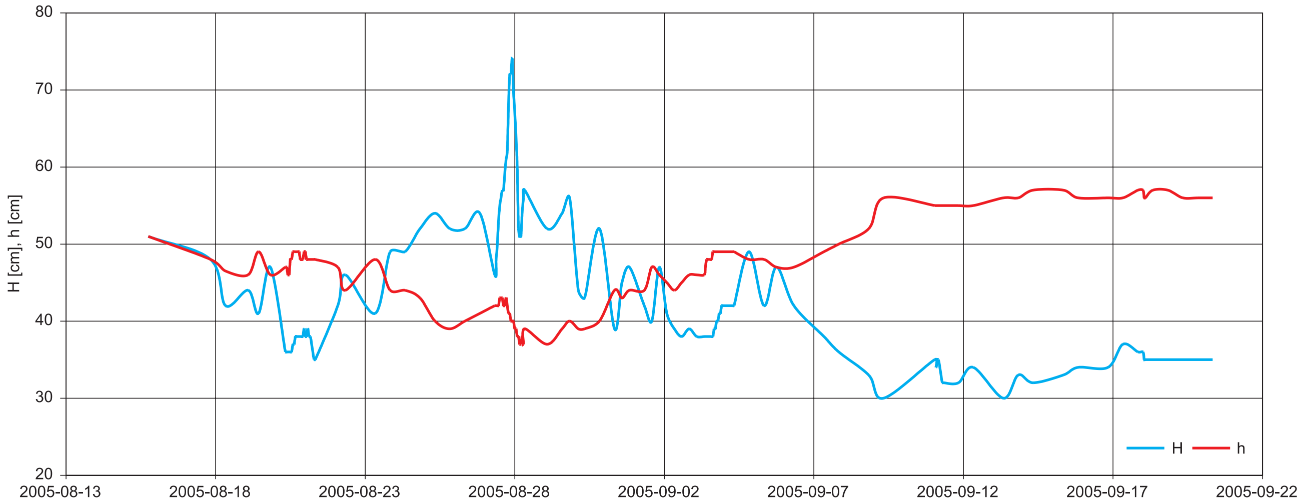

Fig. 4. Variations of free groundwater surface depth (h) and water level in Ebba river (H) during summer season 2005

off in this some period were measured. The result of this research on Fig. 4 is presented.

The strong relationship between subsurface water level fluctuations and stream regime are observable. The fluctuations of groundwater level indicate that groundwater flow play important rule in water circulation in Ebbaelva catchments especially in the stream channel vicinity.

Conclusions

The specific nature of polar environments cause that the measurements of water balance components are scarce. It is relating mainly to actual evaporation and groundwater flow. The pilot study performed in Ebbaelva catchments (Spitsbergen) show distinct groundwater water level fluctuations. These fluctuations prove that groundwater flow play important role in water circulation in this catchments. The test of evaporation with use of basin evaporometer shows that this apparatus should be adapted to specific polar environment. The detailed research of groundwater flow and actual evaporation will be performed in Ebbaelva catchments in summer 2007.

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Relief forms on a place of retreated glaciers, Spitsbergen

Bulat R. Mavlyudov*

Institute of Geography, Russian Academy of Sciences, Moscow, Russia

Relief development on areas recently free from glaciers is determined by following factors: relief features existing before glacier, conditions under glaciers (rocks and sediments temperature, features of glacial exsaration and moraine accumulation, washing up again of sediments by water streams), features of area clearing from ice, modern processes relief forming. In Spitsbergen conditions permafrost with all accompanying processes is added. We shall consider these features of relief formation separately.

Previous relief on the place of existing modern glaciers is unknown. It is possible to study it on areas, which now are not covered by glaciers. As a rule it is wide valleys modeled by glaciers during the previous epoch and in which for the climatic reasons there are no modern glaciers. Analyzing features of such valleys structure it is possible to speak that for previous relief wide trough valleys with the smoothed forms of surfaces, with washed up moraine sediments, with wide development of congelifraction on exposed rocks surfaces were typical.

There is specific question about time of glaciers formation on archipelago. This question till now is not solved completely. The structure of a relief shows that modern glaciation was imposed on areas which long time was free from ice. Approximately 6000 years ago on archipelago there was a wood vegetation (Lavrushin 1969). Researches of peats in different parts of island West Spitsbergen show that 3–4 thousands years ago on archipelago prevailed enough warm conditions not favorable for formation of glaciers (Surova et al. 1988). On available data still about 2000 years ago on a place of Longier Glacier the moss could grow (Hormes 2003). It means that influence of a modern glaciation on a relief was not long.

Glaciers formation occurred in favorable climatic conditions of the recent past when winter snow accumulation essentially exceeded its summer melting. It is clear that at the first stage were formed extensive snowfields, which gradually transformed into glaciers. Therefore all over again snow fields have only covered rocks, having protected them from sharp temperatures fluctuations. When thickness of snow-firn-ice thickness has increased ice movement was accompanied by transportation of moraine material in ice thickness from upper parts of glaciers to lower and by increasing of exaration activity. Exsaration depends on glacier bottom conditions, degree of it preparation by congelifraction and glacier bottom temperature. Researches have shown that if glaciers had ice thickness less then 100 m they were cold i.e. under them there is permafrost (Vasilenko et al. 2001). If glaciers thickness is more than 100 m they will became two-layers (polythermal) with upper cold ice layer and lower temperate layer and presence of talic under ice thickness. It is clear that such glaciers have different influence on the bottoms. Temperate ice layer located at the basis of polythermal glacier is favourable for exaration but exposed permafrost at glacier bottom was not favorable for exaration. Thus in all cases exaration more efficiently effects on rocky ledges (riegels) while at depressions between them there was first of all an accumulation of moraine sediments. As prior to the glaciation beginning rocky ledges have been well prepared by congelifraction glaciers easy strip off this layer of strongly weathered rocks while the further exaration has been complicated.

As each glacier passed the certain stages of development its influence on glacier bottom changed in time. At the first stage of glacier formation cold ice moves upon frozen rocks and exaration was minimal. In this case glacier influence on its bottom was insignificant. When ice thickness has increased and in the

* e-mail: bulatrm@bk.ru

lower part of glacier there was a layer of temperate ice the glacier begin to move more quickly, and influence of glacier on its bottom increased. It corresponds to glaciers increasing stage. On Spitsbergen such stage occurred at the end 19 – the beginning of 20 centuries (Koryakin 1988). At this particular time big part of mountain-valley glaciers have grown in sizes. For example, Aldegonda Glacier at the beginning of 20 centuries ended in the sea (now its tongue situated as far as 2 km from seashore), and on Western Grøfjord Glacier there was surge at that time. At this particular time there was a serious reorganization of subglacial and close to glacier relief with common tendency to less billowy areas. In this period there was a formation of powerful lateral moraines of Aldegonda Glacier and end moraine of Western Grønfjord Glacier. Apparently at this stage of glaciers development prevailed exaration of subglacial relief.

At the beginning of 20 centuries glaciers increasing change into their degradation that was well visible by steady positive glaciers mass balance during long time (Hagen, Liestøl 1990, Mavlyudov, Solovyanova 2005), and also by essential displacement of glaciers tongues (Mavlyudov 2004, 2007a). At this time processes of exaration at glaciers bottoms decreased. At this time begin processes connected with development of internal drainage systems of glaciers (Mavlyudov 2006a). Thus in ice thickness and under the glacier enough steady drainage channels are formed. Channels provide fast water movement to glaciers bottoms. It led to involving of moraine sediments on glaciers bottoms in washing up again by water streams and in partial sediments carrying-out of glaciers limits. Despite of great significance of this process influence of water on glaciers bottoms is localized within the limits of channels beds of subglacial streams, which occupy only insignificant areas under glaciers. Therefore only insignificant part of moraine sediments at glaciers bottoms can be rewashing by subglacial water streams.

Glaciers degradation led to reduction of ice thickness and when it decreased up to 100 m talics under glaciers disappeared and glaciers became cold (Gokhman, Khodakov 1983). It led to braking of all processes under glaciers. But as have shown recent researches on Tavle Glacier (Mavlyudov 2007b), even in cold glaciers internal drainage system disappears not at once so rewashing of moraine sediments by water streams under glacier can proceed at this stage of glacier development.

Reduction of ice thickness of retreating glaciers, which besides have lost the accumulations areas (glaciers occurred below ELA) as it is observed on glaciers of Nordenskiold Land now, translates degrading glaciers in the category of passive (Bol'shiyanov 2006). In this case there is no influence of ice on underlying relief.

The important influence on underlying relief occurred by character of glaciers degradation. If glaciers retreating occurs gradually and is accompanied by steady lowering of ice surface and very slow ice movement in this case glacier bottom usually completely become free from ice and projective moraines sediments cover rocks surfaces (Lavrushin 1969). As a rule thickness of such moraine cover varies from 0.3 up to 1 m (less often up to 2 m) as it was observed on Aldegonda and Western Grønfjord Glaciers. These moraines remind the fine ridges stretched by parallel strips from glacier tongue.

When some part of glacier ice is separated from retreating glacier by water stream such area of dead ice can be kept long time under moraine sediments cover. Time of such ice existence considerably increases when water streams do not influence it. Influence of water streams on dead ice leads to cavities formation inside of its thickness because of in dead ice collapses, subsidences and temporary lakes are typical. As a rule after dead ice disappearance on a place of its distribution arises original hummocky relief. Especially intensively such relief develops at degradation of dead glaciers tongues after surges. To formation of such relief conducts intensive development of a glacial karst to which development favour presence of thick moraine cover atop of dead ice, presence of wandering water streams or their partial damming (Mavlyudov 2006b).

As a result of influence of all above mention processes on a place of the retreated glaciers the original relief is formed. The analysis of existing data allows to subdivide areas that was open after retreating mountain-valley glaciers tongues into some types:

1. weak rebuilding of left moraine deposits when from glacier tongue we can find traces of moraines that come from ice to surrounding area (so-called projective moraines) which testify absence of ice movement (Glaciers Aldegonda, Western Grønfjord, Werenskiold etc.);
2. strong rebuilding of left moraine deposits which is accompanied by strong sediments washout by water streams (Glaciers Tavle, Drøn etc.);
3. complex of final moraines (Glacier Longier);
4. complex of dead ice (Glaciers Tunge, Lars);
5. formation of moraine-dammed lakes (Iren, Eastern and Western Grøfjord, Hanna etc);
6. formation of glacier-dammed lakes (Glacier Elfenbein).

Only in rare cases it is possible to observe each of these allocated types in the pure state. It is possible to see partial complication of one type by another or a combination of these types more often. For example on Werenskiold Glacier the most part of moraine sediments is poorly rebuild (type 1), but in the right part of valley rebuilding of moraine sediments is significant (type 2), and by traces of lake sediments it is possible to say that there was moraine-dammed lake some years ago (type 5). Retreating of glaciers tongues ended in the sea, which are known only in a northwest part of archipelago, are accompanied basically by change of position of glaciers tongues and by lowering of their surfaces.

If to look on archipelago maps it is not difficult to see that much houses, which now are used by hunters, fishermen and tourists are usually situated on moraine sediments where at the end of 19th – the beginning of 20th centuries there were glaciers. It also concerns and to almost all modern existing (and earlier existed) settlements on archipelago: Longyearbyen, Pyramiden, Sveagruve, Nu-Alesund. It means that if in the future there will be glacier growing on archipelago which on the scales will be compared with glaciers increasing at the end of 19th – the beginning of 20th centuries, big quantity of houses and settlements on archipelago can occur under destruction threat.

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Landform Analysis, Vol. 5: 58–59 (2007)

Concept of geomorphological analysis of previously glaciated areas (based on analysis of the surroundings of Prášilské jezero lake and Jezero Laka lake, Šumava Mts., Czech Republic)

Pavel Mentlík*

Department of Geography, University of West Bohemia, Pilsen, Czech Republic

Lenka Lisá

Department of Quaternary Research, Czech Geological Survey, Prague, Czech Republic

Jozef Minár

Department of Physical Geography and Geoecology, Faculty of Natural Science, Comenius University, Bratislava, Slovak Republic

Geomorphological analysis is an important part of geomorphological investigation. However, the process of this analysis itself is in fact still unclear, even though geomorphologists use it very often.

Therefore, a firm concept of geomorphological analysis has been postulated according to a proposal published by Urbánek (2000a, b) and applied to geomorphological research of previously glaciated areas in the Šumava Mts. (the Czech Republic).

Geomorphological information system (GmIS) has been suggested as the environment and the tool of the analysis (Minár et al. 2005, Mentlík et al. 2006). A layer of elementary forms of relief comprises the main part of this system – all other information is connected with a particular elementary form.

Generally, the process of geomorphological analysis has seven steps. Delimitation of the area under consideration is made in the first step – called identification. The second step – differentiation (which is separated into three substeps) deals with demarcation of elementary forms of relief in the particular area. Spatial analysis, when the positions of elementary forms are compared and identified genetically, is done in the third step. The next two steps are independent from the previous parts. They deal with research of current geomorphological processes and analysis of mophochronology (the main steps of the development of the particular area are postulated). The data obtained by all the steps of the analysis are summarized and the hypothesis of the genesis of the area of interest is postulated in the next step. The verification of this hypothesis is done in the last part of the analysis by means of independent (non geomorphological) methods.

The presented approach was applied in two areas (see above). The analysis of quartz grain surfaces and other sedimentological methods (analysis of orientation of clasts and analysis of clast shape and roundness) were used in this step there. Finally, hypotheses of development of relief were postulated for each area.

Two main further divided phases of glaciations were investigated in both areas. The older glaciation was more extensive (suggested TPW-ELA ~1080 m a.s.l., TP-ELA 1200-1300 m a.s.l.). The glaciers were ~1500 m long, ~600 m wide and ~ 50 m thick. Although the glaciation was less extensive during the

* e-mail: pment@kge.zcu.cz

second phase, various glacial environments existed – particularly in the surroundings of Prášilské jezero lake. Firstly a glacier rock glacier (comp. Benn, Evans 1996) developed and secondly the cirque glacier existed there. The method of analysis of quartz grains was especially useful for identification of the remnants of the glacier rock glacier.

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Pavel Mentlík*

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Assumption and realization of Arie catchment measuring system, Spitsbergen

Adam Nawrot*
Institute of Paleogeography and Geoecology, Adam Mickiewicz University, Poznań, Poland

Michał Petlicki
Department of Polar and Marine Research, Institute of Geophysics, Polish Academy of Sciences, Warszawa, Poland

Introduction

Geoecosystems of polar areas are particularly sensitive to climate change. The fluctuations of mean annual temperatures of Svalbard areas recorded for many years and progressing glacier recession indicate unequivocally that the Arctic areas strongly react to long-term weather and climate changes. Energy released in the form of proglacial waters contributes considerably to the transformation of the biotic and abiotic systems. However, it is necessary to note that once the polar environment has changed, it is a very hard and long process to restore it to its former state. In this way natural processes accelerate for the next environmental changes (Zwoliński 2007). As suggested by Kostrzewski et al. (2006), of great significance in the terrestrial geoecosystems of the Arctic is the recession of glaciers and the decline of permafrost and snow covers.

The monitoring of the polar environment in the Svalbard area included periodic research in glaciated basins with areas considerably larger than that of the suggested Arie catchment. This means that the glaciated meso- and micro-basins omitted by the researchers may prove particularly important geoindicators of periodic weather changes for the polar zone. The dynamics of energy flow and matter cycle in a small-sized basin differs definitely from that in big catchments.

It is reasonable for balance investigations to use a river basin as the reference spatial unit. In many areas of the Arctic, glaciated and non-glaciated geoecosystems are basic spatial units in environmental investigations (Kostrzewski, Zwoliński 2003, Kostrzewski et al. 2004).

The measuring system of the glaciated Arie basin has been geared to the needs of the doctoral projects undertaken:

– Operation of the geoecosystem of the glaciated Arie catchment, and
– Impact of superimposed ice on runoff from a glacial catchment.

In the research, the Arie catchment is treated as an independent structure with an autonomous flow of energy and matter reflecting the mechanism of processes operating in the polar zone.

Location of the Ariedalen

The drainage basin of the Arie glacier is located in the proximity of the Hornsund fjord, in the south-western part of West Spitsbergen Island belonging to the Svalbard Archipelago situated in the border zone between the Eurasian and American Arctic (Fig. 1). Glaciers and ice caps cover 36,600 km^2, or about 61% of the Svalbard Archipelago area (Hagen et al. 2003). Situated at the mouth of the Rev valley, between two massifs, Skoddefjellet and Ariekammen, and a short distance from Polish Polar Station, is a parallel firn field of the Ariebreen cirque glacier (Fig. 3). Small cirque glaciers are numerous

* e-mail: nawrot@amu.edu.pl

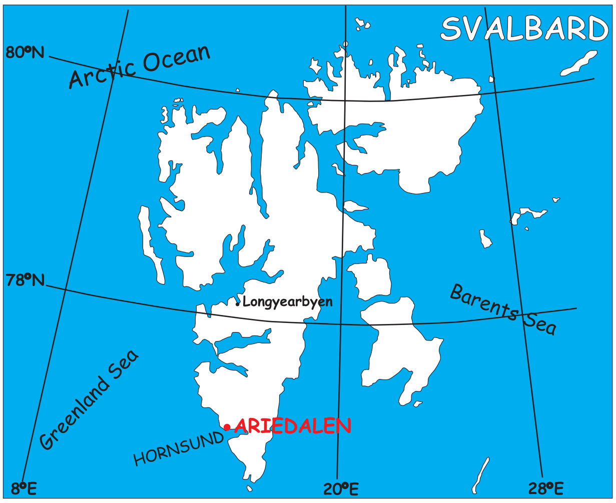

Fig. 1. Location Ariedalen in Svalbard Archipelago

and typical, especially in the alpine mountain regions of western Spitsbergen (Hagen et al. 2003).

The Arie cirque glacier belongs, in thermal terms, to the type of cold body glaciers. Its thickness does not surpass 75 metres, and its area amounts to 0.48 km^2. Hagen et al. (2003) found that small glaciers (A < 10 km^2) are often not much more than 100 m thick and, therefore, the main body of those glaciers is cold. This is the case with the Arie area. Below the firn field the glacier surface is steep and narrows towards the valley mouth where the foreland of the glacier has the form of a regularly drained water body. Proglacial streams drain the Arie glacier catchment along channels inside front ice-moraine ridges with relative heights of up to 15 metres. Ice cores are covered by a 1- to 1.8-metre-thick stone ablation moraine layer with big blocks of rock. The marginal zone is drained by an external, high-hanging outwash trail which ends at the exit of the Arie valley with a steep rock bar (Karczewski 1984). A proglacial stream cuts into the bedrock and flows through a gorge creating a distinct alluvial fan on the foreland.

The Ariedalen, according Pulina's classification (2004), belongs to glaciated basins in a residual stage which include valleys of coastal mountains. In such an area water circulates exclusively in summer, but it stops in the polar winter time entirely. Because of its small area (2.3 km^2), relief (a valley surrounded by tall rock walls) and the rapid recession of the Arie glacier (0.48 km^2 in area), the Arie drainage basin is a model research field with only a minimal human impact.

Polar monitoring of the Arie catchment

The organization and implementation of the monitoring of polar geoecosystems should rest on a comprehensive concept of the operation of glaciated and non-glaciated geoecosystems (Zwoliński 2005).

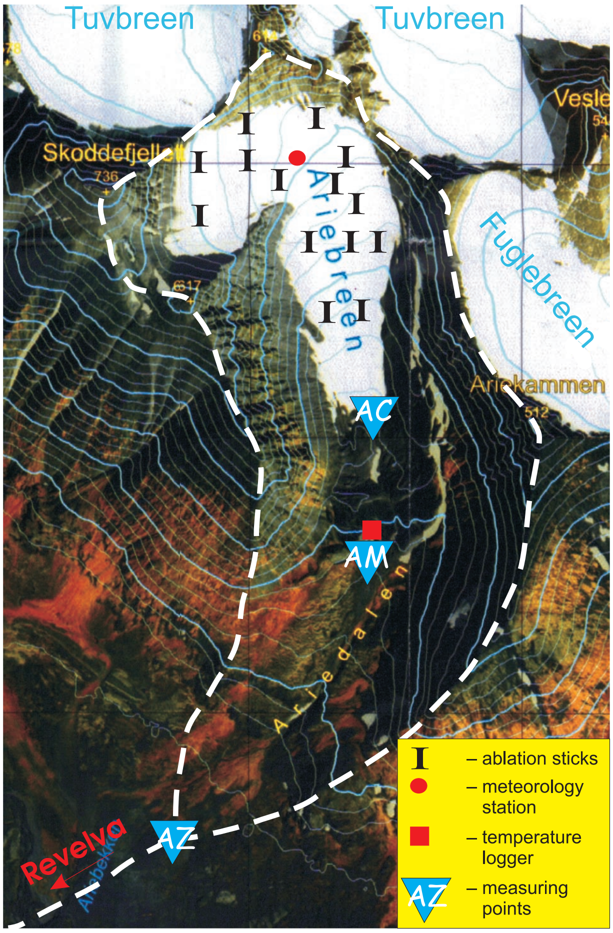

Fig. 2. Monitoring system of Arie basin. Werenskioldbreen Orthophotomap (Kolondra 2002) change by authors

Thus understood, the research procedure can be defined as an integrated monitoring of the natural environment of polar geoecosystems (Kostrzewski et al. 2006).

The Ariebreen terminates on land, with the outlet stream, Arieelva, collecting the whole runoff and flowing out through a lateral moraine. The Arieelva has a closed catchment with only one outflow channel, which offers the possibility of calculating its complete water balance. It is a rare situation in the Hornsund area, and therefore gives the researcher an opportunity for a more precise estimation of water cycle parameters unavailable for other glaciated catchments.

Because of the great diversification of the fluvial system caused by the geomorphological structure of the valley, the Arie catchment was divided into two sub-catchments. Important information about mechanical and chemical denudation is provided by three measuring points (Fig. 2). Samples are taken from them twice a day (AZ) and once a day (AM, AC). In the laboratory of the Polish Polar Station water samples undergo chemical analyses and their content of suspended material is determined. Major ions are analysed for cations and anions by ion chromatography on a Methrom 761 Compact IC. The temperature, pH and specific electric conductivity are measured in situ near the measuring points.

The water level in the Arieelva is measured with a POLON L-04p/01 automatic gauge at 10-minute intervals. The gauge is situated on an old marine terrace, just upstream of a bifurcation. Actual runoff is calculated with a rating curve, calibrated by tracer methods and current meter measurements. A staff gauge has also been installed at the site to provide independent information about the Arieelva water level. Runoff measurements are also taken in a few locations on the Arieelva between the glacier snout and the automatic gauge.

The elevation of ablation sticks is measured once a week. Fourteen ablation sticks are inserted nine metres deep in the ice (Fig. 2). The data from the ablation sticks provide information about the ablation rate, snow accumulation, and the dynamics of ice flow. Once a month the limits of the glacier are measured with a differential GPS Leica System 1200.

Radon ^{222}Rn concentrations in the outlet stream are measured once a week. Samples are taken to the laboratory and ^{222}Rn activity is measured with a liquid scintillator. It provides information about the origin of water - whether it comes directly from precipitation or from the glacier bed, originating from the melting of glacier ice.

A weather station situated on the Ariebreen at an altitude of 400 metres above sea level and another automatic temperature logger on a moraine provide information about temperature, wind, humidity, pressure, and rain. Data from the meteorological station of the Polish Polar Station and weather stations situated on the Hansbreen near ablation sticks number IV (200 metres a.s.l.) and IX (400 metres a.s.l.) are very important for making comparisons.

Conclusions

In answer to the "Rapid Landscape Change and Human Response in the Arctic" declaration of 17 July 2005 signed in Whitehorse, a systematic integrated monitoring of the Arie catchment has been set up. A correct identification of the natural processes in polar geoecosystems and determination of corresponding geoindicators will provide a basis for an estimate of the condition of and direction of change in selected High Arctic areas. In the light of the anticipated climate change, it will be important to gain a better understanding of the effect of refreezing of meltwater in snow and firn, and the extent and amount of superimposed ice formation (Hagen et al. 2003). The research undertaken by the authors in the glaciated Arie catchment will fill a scientific niche and allow comparing the glaciated areas of Svalbard as to energy change in the context of climate change.

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Landform Analysis, Vol. 5: 63–65 (2007)

Geomorphological map of the surroundings of Cortina d'Ampezzo (Dolomites, Italy)

Alessandro Pasuto
IRPI -CNR, Padova, Italy

Mauro Soldati*
Dipartimento di Scienze della Terra, Università di Modena e Reggio Emilia, Modena, Italy

Geomorphological investigations in the area of Cortina d'Ampezzo (Dolomites, Italy) have been carried out since the 1980s mainly within national and European research projects. This has enabled the researchers to define, on the one hand, the geomorphological evolution of the area and, on the other hand, the spatial and temporal occurrence of landslides, that are the most spread geomorphological feature of the studied area.

A detailed geological and geomorphological survey was carried out at a scale of 1:10000 and a geomorphological map at a scale of 1:20000 was produced following to the Italian geomorphological mapping methodology (Gruppo …. 1994). The survey was combined with multitemporal aerial and ground photograph analysis. The examination of archive photographs (late XIX and early XX century) was especially significant for the slope evolution to be evaluated thanks to the low degree of human activity in the area, together with the scarce extension of woodland.

The Cortina d'Ampezzo valley, situated in the eastern Dolomites, is surrounded by high mountain groups such as Tofane, Lastoni di Formin, Croda da Lago, Faloria, Cristallo and Pomagagnon. The valley is crossed in a N-S direction by the Boite torrent, a right tributary of the Piave River.

From a climatic point of view the Cortina d'Ampezzo area is quite varied. This is mainly the result of the wide altitudinal range. The climate corresponds to the Alpine type, from cold to temperate, with variably cold winters and mild summers. The pluviometric regime reflects the typical pattern of an Alpine climate with two maxima of rainfall, of which the main one is late spring-summer and the other in autumn. As regards snowfall, the period of permanent or almost permanent snow normally begins in December and lasts until April. In the higher zones naturally this period lasts for a couple of months longer.

The geological structure of the area, characterised by an alternation of dolomitic rocks and successions of prevalently pelitic components, has markedly conditioned the morphological evolution of the slopes after the retreat of the LGM glaciers (Pasuto et al. 1997). The stratigraphical sequence outcropping in the area of Cortina d'Ampezzo covers a period of time ranging from Middle and Upper Triassic to Lias. The Quaternary deposits, mainly deriving from landslide phenomena, are widespread inside the valley, masking the substratum and making the recognition of tectonic elements along the valley bottom particularly difficult. The Triassic rocks outcrop especially in the peripheral parts where the highest mountain groups are located

The slope morphology is softly degrading in the medium and lower parts where pelitic formations outcrop, while at higher altitudes subvertical dolomitic walls rise up, eventually interrupted by typical ledges, thick scree slopes, located in correspondence with more erodible formations. The whole area has often been affected by landslide phenomena of various types and of sometimes notable dimensions, some of which are still active today. As

* e-mail: soldati@unimore.it

a result of the favourable morphological conditions, the area has witnessed progressive urbanisation, which has also been tied to an intensive tourist development. Because of this intense urbanisation and the interest which this region holds for tourism, the presence of some active landslides and of a large number of dormant landslides makes this area particularly vulnerable and subject to a high geomorphological risk (Soldati 1999; Pasuto, Soldati 2004).

The valley of Cortina d'Ampezzo during the Würm glaciation was almost completely covered by ice masses reaching thicknesses of 1000 m; only the highest dolomitic peaks emerged above the glaciers. At that time several glacial tongues converged in the valley from the surrounding mountain groups, forming a thick glacial tongue which met the Piave glacier near Pieve di Cadore (about 35 km valleyward of Cortina d'Ampezzo). Thus the slope evolution and the setting of the surficial deposits of the studied area mostly date from the post-Würmian period. As regards glacial deposits, only a few outcrops datable to late glacial stadial phases were found. The scarcity of such deposits, in contrast with other dolomitic areas, is related to the numerous and extensive landslide movements which occurred after the retreat of the glacial masses (Panizza et al. 1996; Soldati et al. 2004).

The Cortina d'Ampezzo area appears to have always been prone to instability phenomena for the following different reasons. First of all the structural conditions of the valley must be taken into account. The stratigraphic succession is, in fact, characterised by an alternation of dolomitic rocks showing a brittle mechanical behaviour (Dolomia Cassiana, Dürrenstein Formation and Dolomia Principale) and rocks with a ductile mechanical behaviour (San Cassiano Formation and Raibl Formation). This situation has favoured the development of mass movements and deep-seated gravitational slope deformations; the latter, which were widely recognised in the area, may have favoured themselves or induced the occurrence of several landslides (Soldati and Pasuto 1991).

Furthermore the incidence of tectonic activity is also significant as regards landsliding; in fact, the dolomites were affected by an intense jointing in correspondence with the principal faults, thus creating discontinuities which became potential sliding surfaces and preferential seepage zones for water which could reach and moisten the underlying marly and clayey formations. In addition, tectonic activity might have favoured landsliding giving a diffuse presence of the San Cassiano Formation in the middle-lower part of the slopes as a result of partial doublings of the stratigraphic sequence due to overthrusting (Pasuto et al. 1997).

The effects of glacier retreat on the slopes must not be overlooked. It is likely that the pressure of ice on the valley sides determined rock deformations in correspondence with surfaces of structural discontinuity, favouring the formation of sliding surfaces.

An effort has been made to assess which of the numerous radiocarbon dates collected in the study areas of Cortina d'Ampezzo are related to events of a type or magnitude that might be indicators of Holocene climatic changes (Soldati et al. 2004). The first phase of marked slope instability is observed in the Preboreal and Boreal (about 11500 to 8500 cal BP) and includes, on the one hand, large translational rock slides, which affected the dolomite slopes following the withdrawal of the Lateglacial glaciers and the consequent decompression of slopes and, on the other hand, complex movements (rotational slides and flows) which affected the underlying pelitic formations and were probably favoured by high groundwater levels resulting from an increase of precipitation and/or permafrost meltdown. A second concentration of landslide events is found during the Subboreal (about 5800 to 2000 cal BP), when slope processes, mainly rotational slides and/or flows, took place in both the study areas. These slides may be considered as reactivations of older events linked to the phase of precipitation increase, which has been documented in several European regions during the mid-Holocene period. On the other hand, during the Little Ice Age, the scarce number of landslides dated in the study areas does not enable an increased frequency of landslides to be detected.

The recurrence in time of landslide activity since the Lateglacial was certainly also influenced by non-climatic factors. Among these, the influence of human activity, which is proved to have been crucial for other Alpine areas at least since the mid-Holocene, seems to be almost irrelevant in the study areas where only seasonal or isolated settlements were present before the Middle Ages. Nevertheless, the present state of the investigations does not allow the assessment of the degree of influence of this and other possible non-climatic causes, which might have "disturbed" the climatic signal identifiable in the landslides studied.

In the geomorphological map of the area surrounding Cortina d'Ampezzo landforms are depicted according to the Italian mapping methodology. This methodology pays particular attention to the genesis of landforms (glacial, periglacial, gravitational, alluvial etc.), that are distinguished using symbols of different colours, and their degree of activity, marked with more or less intense colouring. Morphometric aspects are also represented on the map by specific symbols. The outcropping of geological formations and tectonic elements are mapped, too. In the specific case of the Cortina d'Ampezzo area, this is of fundamental importance to understand the past, present and future evolution of the relief, because the structural control on landforms is relevant.

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Floods in High Arctic valley systems and their geomorphologic effects (examples from Billefjorden, Central Spitsbergen)

Grzegorz Rachlewicz*

Institute of Paleogeography and Geoecology, Adam Mickiewicz University, Poznań, Poland

Introduction

The supply of European rivers is mainly an effect of water delivered from rainfall and the melting of snow cover. These phenomena are strongly diversified seasonally and regionally determining not only the outflow regime, but also natural and economic relations in river valleys (Knapp 1979). In polar and high-mountains glaciated areas the main volume of water, reaching 80% of the total, is transferred during the short melting season from ablating ice covers (Singh, Singh 2001). These regions are characterized by limited possibilities of water storage in the ground because of thin sedimentary covers and permafrost occurrence, temporarily unfreezing to the depth of 1 m below the surface or more. These areas are also under strong and distinct influence of the global warming and hydrological changes (Nelson 2003). Main media of water storage in glaciated basins in different time scales are snow and ice covers (Jansson et al. 2003). Water is released from them with different intensity during the short period of summer positive temperatures. Even some slight environmental changes may have the influence on abrupt release of considerable amounts of water. It is deciding about short-term rhythm and seasonality of processes run in glaciated catchments, triggering significant floods, rebuilding valley floors, not stabilized with plant covers. The aims of the present paper are to describe types of floods in glaciated catchments of the High Arctic with some examples of their geomorphologic effects.

Study area

Billefjorden, the NE branch of Isfjorden system in the central part of Spitsbergen (Fig. 1) is ending in the North with a comparatively shallow bay – Petuniabukta (bukta = *norw.* bay). The shallowness of the bay is a consequence of its lateral position to the main stream of ice during the Pleistocene, shaping the fjord bottom from the East (Karczewski 1995). Only some smaller ice tongues, flowing from the Lomonosov Plateau and valley glaciers, founded its outlet there. Such a setting created also conditions to develop, in the inner part, a large accumulation plane with overlapping glacio-fluvial and tidal influences (Borówka 1989). Contemporary glaciation around the Petuniabukta is reduced to the inner valley parts in the phase of continuous retreat from the position of the maximum of Little Ice Age advance (600–100 BP) marked with distinct frontal moraines (Rachlewicz et al. 2007.). The valleys, affected by strong slope and mass movement processes, in conditions of continuous permafrost occurrence and weak plant cover, they are the background for the operation of proglacial outflow, magnified by the decay of seasonal and perennial snow covers. The whole catchment of Petuniabukta has an area of 162.133 km^2, about 36% of which is covered with glaciers. The largest glaciers (breen = *norw.* glacier) are Ebbabreen (25 km^2), Hörbyebreen (20 km^2) and Ragnarbreen (7 km^2).

The climate of the central part of Spitsbergen is of quasi-continental type. Precipitation slightly exceeds

* e-mail: grzera@amu.edu.pl

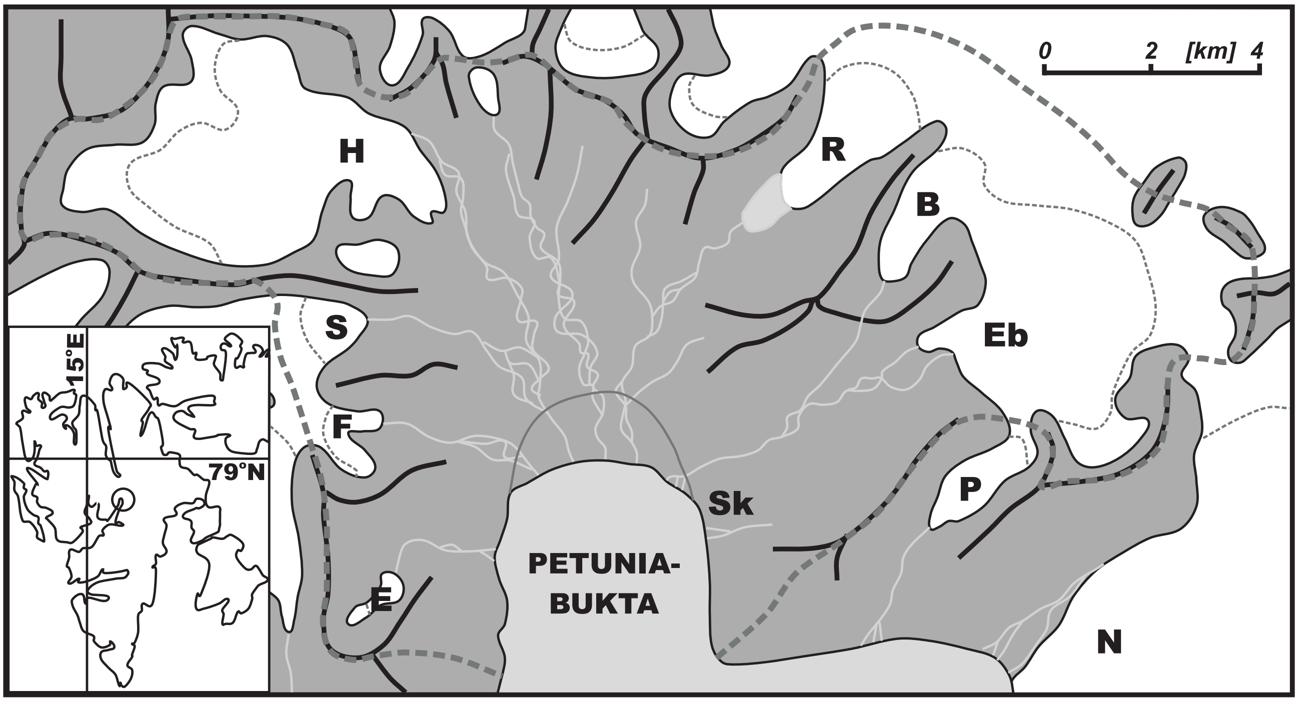

Fig. 1. Location of the study area
Black lines – main mountain crests; bold dashed lines – Petuniabukta watershed; thin dashed line – equilibrium line altitude (about 500 m a.s.l.); E – Elsabreen; F – Ferdinandbreen; S – Svenbreen; H – Hörbyebreen; R – Ragnarbreen; B – Bertrambreen; Eb – Ebbabreen; P – Pollockbreen; N – Nordenskjöldbreen; Sk – Skottehytta

200 mmy^{-1} and average summer month (June-August) temperatures are about 5°C, rarely reaching more than 10°C (Rachlewicz 2003a). Meteorological conditions were monitored at the shoreline in the base Skottehytta and by several automatic stations along valleys and on glaciers Ebba and Hörbye. The outflow is mainly of glacial origin, appearing periodically during positive temperatures occurrence and partly continued in autumn and winter in the case of larger, polithermal glaciers, in front of which icing fields are present (Bukowska-Jania, Szafraniec 2005). Episodic streams appear also on slopes, in cuts and gullies, alimented with melting snow cover and rainfalls. Problems of outflow and floods of the surveyed area and its neighbourhood, was earlier fragmentary studied by Gokhman, Khodakov (1986), Kostrzewski et al. (1989) and Rachlewicz (2003b, 2004).

Floods in glaciated valleys – examples and effects

Floods in glaciated valleys are generated through water stored in ice and snow covers in liquid or solid state. Main factors triggering flood waves are meteorological conditions in the form of rainfalls or bringing snow and ice to melt. Other features of glaciers and their surroundings like geology, endogenic activity, relief, availability of sediments, thermal state of ice etc. could not be neglected. These phenomena are a subject of studies of many specialists, also practical like planning of water supply, hydro-technical devices management or prediction of catastrophic events (Hock 2005).

Main criteria of identification of various types of floods, also with participation of melting ice masses, are their seasonality linked to the supply of water from various levels of its storage. Four types of floods generated in glaciated valleys were distinguished:

- snow-melt floods in spring,
- summer ice-ablation floods,
- summer rainfall generated floods,
- föhn-like floods.

Apart of this a separate group of short flooding waves is represented through singular outbursts of water, incurred according to the opening of englacial channels chopped with ice and snow or the draining of supraglacial or terminal lakes. On Spitsbergen such events are known for example as the single outflow of 1.0×10^6 m^3 from Tillberg ice-plateau during winter (Liestøl 1977) or the subglacial lake drainage from Kongsvegen of about 40×10^6 m^3 of water (Hagen 1987). However the most spectacular phenomena are connected with endogenic activity. Enormous jökullhlaups known recently from Island, was observed in 1995, when the lake Grimsvotn from the Vatna ice-cap released at once 1.9×10^9 m^3 (Björnsson 1998). For comparison, the total yearly outflow from the Ebba glacier is estimated to about 58×10^6 m^3 (Rachlewicz, unpubl.).

Fig. 2. Alluvial cone built during a single flood event 2003-07-06 at the outcome of small non-glaciated catchment in Ragnar glacier valley

Atmospheric conditions of the summer season (usually between June 20th and September 5th) in the Northern part of Billefjorden, generating the above mentioned four types of floods are observed with various frequency from year to year. During their occurrence the discharge of water is rising from the seasonal average of about 10 $m^3 s^{-1}$ to even more than 22 $m^3 s^{-1}$ (Rachlewicz 2003c). According to the thermal gradient measured in longitudinal profiles of valleys from 0 to 500 m a.s.l. (equilibrium line altitude [ELA] for this region, after Hagen et al. 1993) is equal to 0.6°C 100 m^{-1} (Rachlewicz 2004, Górska-Zabielska et al. 2007). The first threshold activating snow melt on slopes and glaciers is the rise and remaining of temperature above 5°C, at the level of the sea. Nivation processes reveal unequal distribution and certain amount of water is stored in the snow and firn covers. Thus intensive floods are observed in lower located small catchments on valleys slopes (Fig. 2). These processes are the most actively transforming the relief of these systems (Rachlewicz et al. in prep.).

Highest water stages are connected with the rise of air temperature above 15°C and the average daily temperature above 10°C. During the four observation seasons such situations occurrence varied from 2 to 7 days. Characteristic reaction of glacifluvial outflow is shown on Fig. 3.

Rainfall induced floods are rare for this part of Spitsbergen. It is an effect of low total precipitation and the intensity usually not exceeding 2.0 mm h^{-1}. Besides that rainfalls are associated with flushes of cold and wet air masses, transforming at the level of 200–300 m a.s.l. into snowfalls, giving weaker effects of immediate outflow. If either rainfall floods generate big changes on the glaciers and in valleys (concentration of the outflow from large surfaces, thermal effect etc.) its participation in the total outflow is small, reaching at least 4% (Rachlewicz 2003b). Time of the reaction of rivers to big intensity rainfall is very fast (Fig. 4), with its cumulating in lower parts due to tributary supply of sheet-wash and concentrated outflow. As quick is the return to the average, ablation controlled, state.

The location of the study area is favouring occurrence of föhn phenomena. The nature of föhn-like weather conditions is the flush of warm air masses over large areas, including highly situated glacier surfaces. The orographic obstacle forcing rapid air flow is the watershed between Wijdefjorden (on the North) and Billefjorden, exceeding 1000 m a.s.l.

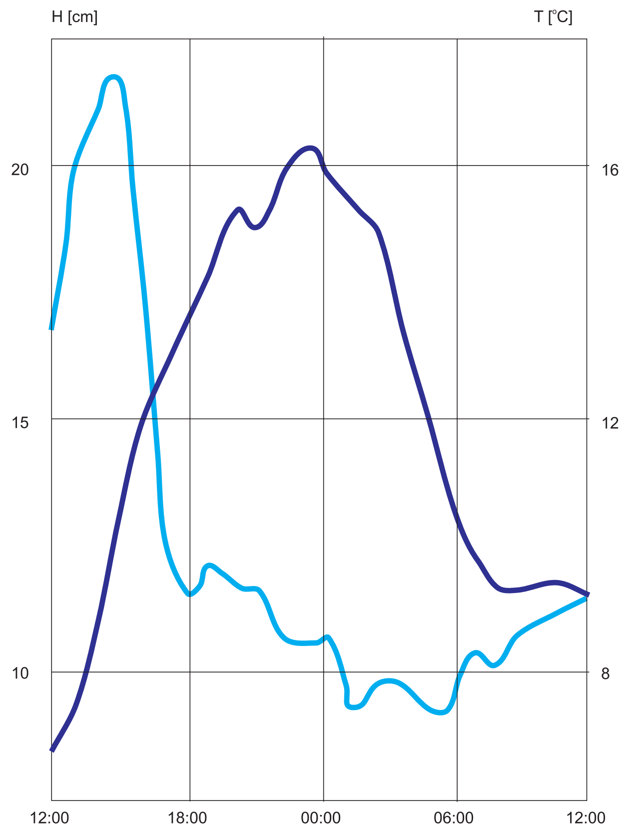

Fig. 3. A plot of daily course of Ebbaelva water level at the catchment closing point (black) and air temperature from Skottehytta (gray) from 2003–07–28/29, showing an additional peak of the flooding wave from Bertram Glacier

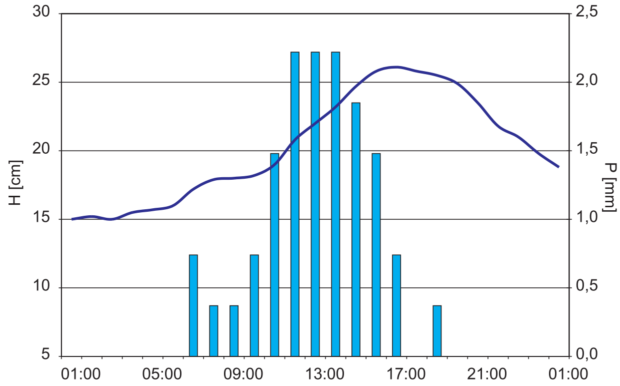

Fig. 4. Hydrogram of the flooding wave on Ebba river the 2003–08–11, on the background of hourly sum of precipitation in Skottehytta

Fig. 5. Sandy-gravel deposition of flood facies above the actual river channel in front of Hörbye glacier. Measuring rule is 1 m long

These events are disturbing thermal stratification before air masses re-cool over upper parts of glaciers. Föhn phenomena are usually followed by intensive rainfalls magnifying flood events.

Upper stages of water in river channels lead to intensive mobilization of the material stored on glaciers, beneath their covers and on their forefields: in marginal (morainic) and outwash zones. At glacier edges, where subglacial channels often operate in conditions of increased hydraulic pressure, sets of boulder layers are observed. Gravely and sandy material is transported along the whole valleys length and deposited up to 0.5 m above the average water level (Fig. 5). The intensity of glacifluvial processes transporting big amount of sediments in traction and suspension is reconstructing the layout of flat-bed braided channels. In terms of the lack of vegetation, there is also observed a lateral supply of fine material, washed out to the foot of valley slopes, where extreme floods are trimming lower parts of alluvial fans.

Conclusion

Floods and their effects are also dependent on actual state of the environment, i.e. freezing of the ground, snow cover occurrence and coverage by ice (also icing). Big dynamics of outflow increases possibility of sediments transportation and in consequence enlarge area and grain-size of deposited covers. Particularly in this parts of valley floors, where braided channels are common, their arrangement undergo distinct changes. In the gorge segments processes of lateral and bed erosion are dominating. A variety and frequency of floods is deciding about remodelling of the landscape architecture at the bottom of post-glacial valleys. It is one of the most active zones of transformations of paraglacial environment, treated as areas of fresh glacial retreat, with an unstable relief configuration and predominance of sediment transit processes (Ballantyne 2001).

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Strontium isotope systematics in the Oppstryn drainage basin, western Norway

Ola Magne Sæther*, Achim Beylich
Geological Survey of Norway, Trondheim, Norway

Göran Åberg
Institute for Energy Research, Kjeller, Norway

The strontium (Sr) isotope composition of runoff and particulate material can be used in catchments to determine the proportion of weathering products originating in areas with different bedrock as a supplement to major ion geochemistry. The Sr budget of a catchment is determined by the relative contributions of erosion and weathering of carbonate rocks versus silicate rocks, but also the preferential weathering of carbonate minerals versus silicate minerals and the contributions from different silicate minerals within the same rock unit. As an example it has been suggested that the importance of carbonate relative to plagioclase weathering could be exaggerated in cases where only the plagioclase-to-kaolinite dissolution reaction is considered (Pretti, Stewart 2002).

Samples of filtered river water and suspended particulate material collected on the filters are collected to estimate the particulate and dissolved loads of runoff from the northwest end of the Jostedalen glacier, western Norway. Strontium isotopic fingerprinting will be carried out by analysing the same samples for Sr-87 and Sr-86 isotopes in an attempt to delineate the relative contribution the two to three different major types of bedrock in the area under and adjacent to the glacier.

Strontium isotope systematics could enhance the precision of the more general mass balance which is performed with respect to major in geochemistry in the Oppstryn drainage basin and contribute to understanding the contribution of different types of minerals, rock types and sub-catchments when estimating the overall erosion and weathering in this part of Europe.

* e-mail: Ola.Sather@ngu.no

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Criteria to discriminate between proglacial and paraglacial environments

Olav Slaymaker*

Department of Geography, University of British Columbia, Vancouver, Canada

The definition of "periglacial" has drifted from the original meaning of "areas peripheral to Pleistocene ice sheets and glaciers" (Lozinski 1909) to one which stresses the distinctive processes of freeze-thaw and permafrost formation. In spite of the comment by Worsley (2004) that the term periglacial has no universally accepted definition, there are really only two options that are in common use:

a) an environment of frequent freeze-thaw cycles and deep seasonal freezing (encompassing about 35% of the earth's continental surface and/or
b) a permafrost environment (only 20%).

The international journal Permafrost and Periglacial Processes implies the broader definition. We therefore conclude that the definition of periglacial is not contentious. Not only are we comfortable with the idea of periglacial processes, but equally there is agreement over the approximate extent of past and present periglacial environments, depending on whether the narrower or broader definition above is used.

The situation is not so clear in the context of the terms "proglacial" and "paraglacial", especially since the publication of a major paper on paraglacial geomorphology (Ballantyne 2002a). French (2007) has a nice discussion on this issue in which he notes that the proglacial environment, which refers specifically to ice-marginal conditions, is a periglacial environment in the original sense of Lozinski.

Proglacial systems, sediment-landform associations and landform assemblages

There is no debate about the literal meaning of proglacial, which is "in front of the glacier" (Penck, Bruckner 1909) but its use has not been consistent in the literature. As pointed out by Embleton-Hamann (2004) there is a transition between ice contact, proglacial and paraglacial environments, processes and forms in space and over time (see Warburton, 1990; Hasholt et al. 2000).

It is also helpful to consider a scale hierarchy of proglacial environments (Benn, Evans 1998):

a) proglacial fluvial systems;
b) proglacial associations of sediment and landforms and
c) proglacial landform assemblages.

They further sub-divide proglacial landform assemblages into

a) ice sheet systems,
b) mountain valley systems and
c) subaquatic landsystems.

The question can be formulated in two parts "how far in front of the glacier does the proglacial environment extend?" and, when interpreting landforms and sedimentary facies "how does the proglacial signal differ from that of the paraglacial signal?".

* e-mail: olav.slaymaker@ubc.ca

Paraglacial systems, sediment-landform associations and landform assemblages

Neither is there any debate about the literal meaning of paraglacial which is "beyond the glacier" but the use of this term has also evolved (Slaymaker 2004). The term was introduced by Ryder (1971) to describe alluvial and colluvial fans that had accumulated through the reworking of glacial sediments by rivers and debris flows following late Wisconsinan deglaciation in the interior of British Columbia. She showed that fan accumulation had been initiated soon after valley floors became ice free and continued until shortly after the deposition of Mazama tephra (6600 yrs. BP). The paraglacial concept was formalized by Church, Ryder (1972). They defined the paraglacial environment as one that is characterized by non-glacial processes that are directly conditioned by glaciation. They identified three aspects of the influence of paraglacial sediment supply on fluvial transport:

a) the dominant component of reworked sediment may shift from till to secondary sources, such as alluvial fans and valley fills;
b) regional uplift patterns will condition the timing of changes in the balance between fluvial deposition and erosion and
c) consequently the total period of paraglacial effect is prolonged beyond the period of initial reworking of glacigenic sediments.

Clague (1986), Slaymaker (1987), Church, Slaymaker (1989) and Muller (1999) refined the concept further.

Benn, Evans (1998) consider paraglacial activity under

a) terrestrial ice-marginal environments;
b) paraglacial associations of sediment and landforms and
c) the paraglacial land system.

They make the case that because there are no processes unique to paraglacial environments it would be better to think of paraglacial as referring to a period of time.

Ballantyne (2002a) points out that between 1971 and 1985 the paraglacial concept was largely ignored outside North America. Since 1985 he sees four trends:

a) an extension of the geomorphic contexts in which the paraglacial concept has been explicitly used;
b) a focusing of research on present-day paraglacial processes and land systems;
c) use of the paraglacial concept as a framework for research across a wide range of contrasting deglacial environments; and
d) a growing awareness of the palaeo-environmental significance of paraglacial facies in Quaternary stratigraphic studies.

He proposed a working definition of paraglacial as "non-glacial earth surface processes, sediment accumulations, landforms, land systems and landscapes that are directly conditioned by glaciation and deglaciation". Geomorphic contexts in which the term paraglacial is now being used include, in addition to the original debris cone, alluvial fan and valley fill deposits

a) rock slopes;
b) sediment-mantled slopes;
c) glacier forefields;
d) glacilacustrine systems and
e) coastal systems.

The problem which is becoming evident in the literature is that the term "paraglacial" is now being widely used without careful distinction between it and the long established traditional term "proglacial". Indeed, and in part as a result of Ballantyne's magisterial papers (Ballantyne 2002a, b), the term paraglacial is now being used to cover a bewilderingly large variety of circumstances, almost to the point of making the word redundant.

It is the confusion between proglacial and paraglacial and the lack of clarity of the use of the term paraglacial that forms the motivation for this paper. We review recent usage of the terms proglacial and paraglacial processes, landforms and environments.

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Mass balance of Kaffiøyra glaciers, Svalbard

Ireneusz Sobota*

Department of Cryology and Polar Research, Institute of Geography, N. Copernicus University, Toruń, Poland

The studies of the structure of mass balance of Kaffiøyra glaciers refer to the Waldemarbreen, Irenebreen and Elisebreen. The data on the structure of the mass balance of Waldemarbreen were based on the direct field measurements conducted from 1996 to 2006 (Sobota 1999, 2000, 2004, 2005, Grześ, Sobota 2000, Sobota, Grześ 2006). The studies of the mass balance of Irenebreen were taken between 2001 and 2006. In 2005 the studies of the mass balance of Elisebreen began. This research is continued. At the same time geodetic and cartographic measurements were carried out (Lankauf 2002, Bartkowiak et al. 2004).

Glaciers are located in the northern part of the Oscar II Land, Kaffiøyra, north-western Spitsbergen (Fig. 1). Waldemarbreen is about 3.5 km long and has an area of 2.6 km^2. The ice originates in one cirque and flows from an elevation of more than 500 m to the present terminus at 130 m a.s.l. Irenebreen, a valley glacier located to the south of Waldemarbreen, flows down towards the Kaffiøyra plain. The area of Irenebreen amounts to 4.2 km^2. Elisebreen area is 11.9 km^2. Its length is about 7 km, while its width is up to 1.8 km. To the north the glacier borders Agnorbreen which is often treated as part of Elisebreen.

Fig. 1. Location of analysed glaciers on Kaffiøyra (photo A. Tretyn)

* e-mail: irso@geo.uni.torun.pl

In order to estimate the mass balance of Kaffiøyra glaciers the method of direct measurements was used. It was based on a set of ablation poles completed with the studies of the snow cover in the snow profiles. This method belongs to the most precise and most often used (Østrem, Brugman 1991; Kaser et al. 2003, Hubbard, Glasser 2005). Twenty-two poles were placed on Waldemarbreen; Irenebreen and Elisebreen had ten poles installed each.

All the ablation poles were drilled 10 m deep with a steam driven Heucke Ice Drill (Heucke 1999). Snow, firn and ice ablation were converted into water equivalent (w.e.). The ice density of 0.9 g cm^{-3} was used to convert ablation thickness to water equivalent. Where snow was found on glacier the appropriate snow density was applied to the computations.

The studies of winter mass balance mainly referred to the estimation of the size of the snow accumulation on glaciers, as well as its selected properties. Soundings of the snow depth on both Waldemarbreen and Irenebreen were carried out in about 150 measurement points. Location of the measurement points was based on both geodesic and the GPS measurements. The measurements also were made in the selected snow profiles in accordance with the International Commission on Snow and Ice (ICSI) standards.

The studies on the summer and winter balance enabled the author to estimate the net mass balance of the studied glaciers in the analysed period.

Time changeability of ablation processes of Waldemarbreen and Irenebreen at various latitudes was significantly diverse. The greatest changeability was observed in the lowest parts of glacier. With the growing altitude the fluctuations decrease. The average summer balance of Waldemarbreen amounted to -104 cm w.e. for the period of 1996–2006. The average summer balance of Irenebreen amounted to -124 cm w.e. for the period of 2002–2006. In 2006 the summer balance of Elisebreen was -135 cm w.e.

Spatial distribution of winter snow accumulation on Waldemarbreen shows some regularity. The largest accumulation is found in the accumulation part and at the foot of the mountain slopes. The smallest accumulation, however, is observed in the front part of glacier up to the altitude of 220 m and at the foot of the medial moraine. In the case of Irenebreen snow accumulation increases significantly from the front part of glacier towards the accumulation fields. In 2005 and 2006 measurements of snow accumulation on Elisebreen were taken as well.

The measurements of structure and graining of the snow cover were not undertaken during all of the analysed periods. When undertaken, the studies included making a few snow profiles in the selected parts of both Waldemarbreen and Irenebreen. Snow cover shows some specific physico-chemical properties. Its vertical profile shows a variety of snow types of diverse level of metamorphosis, hardness and wetting. Snow structure reflects prevailing weather conditions at the time when the snow cover formed.

The lowest snow accumulation on Waldemarbreen was recorded in 2000; it amounted to 32 cm w.e. on average. The highest snow accumulation of 75 cm w.e. was recorded in 1996. From 1996 to 2006 the mean snow accumulation on Waldemarbreen was 47 cm w.e. From 2002 to 2006 the mean snow accumulation value for Irenebreen was 52 cm w.e. In 2005 the snow accumulation for Elisebreen was 59 cm w.e., while in 2006 it was 63 cm w.e. These values are similar to those estimated for other studied Svalbard glaciers.

Mass balances of glaciers based on a network of poles are burdened with measuring errors. They result from reading errors of the poles, as well as from the influence of the morphological conditions of a given glacier, which often make field-readings difficult. Additionally, mass balance estimations are influenced by a complicated process of glacier feeding and its outflow. As numerous papers indicate, a standard measuring error can be calculated: its value is similar for most glaciers. It was also found out that the direct measurements of mass balance of Waldemarbreen, Irenebreen and Elisebreen are burdened with a small error. This mainly results from reading errors of the measurement ablation poles. Errors connected with the inner accumulation, condensation and evaporation do not play a significant role in estimating the size of glacier's mass balance. The error of the annual mass balance for Waldemarbreen, Irenebreen and Elisebreen, based on various methods and formulas used by various authors, as well as on direct field measurements, was estimated at about ±10–20 cm w.e.

Spatial diversity of mass balance of Waldemarbreen, Irenebreen and Elisebreen is mainly influenced by the weather conditions in a specific part of glacier and by local morphological conditions. The areas of the glaciers may be generally divided into the part of the negative mass balance and the part of the positive mass balance. In the case of Waldemarbreen the year 1998 was exceptional, as the entire glacier showed negative mass balance. Irenebreen shows more positive mass balance in its both accumulation parts. The accumulation part of Elisebreen also shows positive mass balance. This results from the fact that they both are located at higher altitude than Waldemarbreen.

According to Dyurgerov (1986), those glaciers whose mass balance altitude profiles show the least year to year variability are more useful for assessing the changes of climate. Waldemarbreen, Irenebreen and Elisebreen are one of such glaciers (Fig. 2). Thanks to the direct measurements, the average location of the equilibrium line on Waldemarbreen

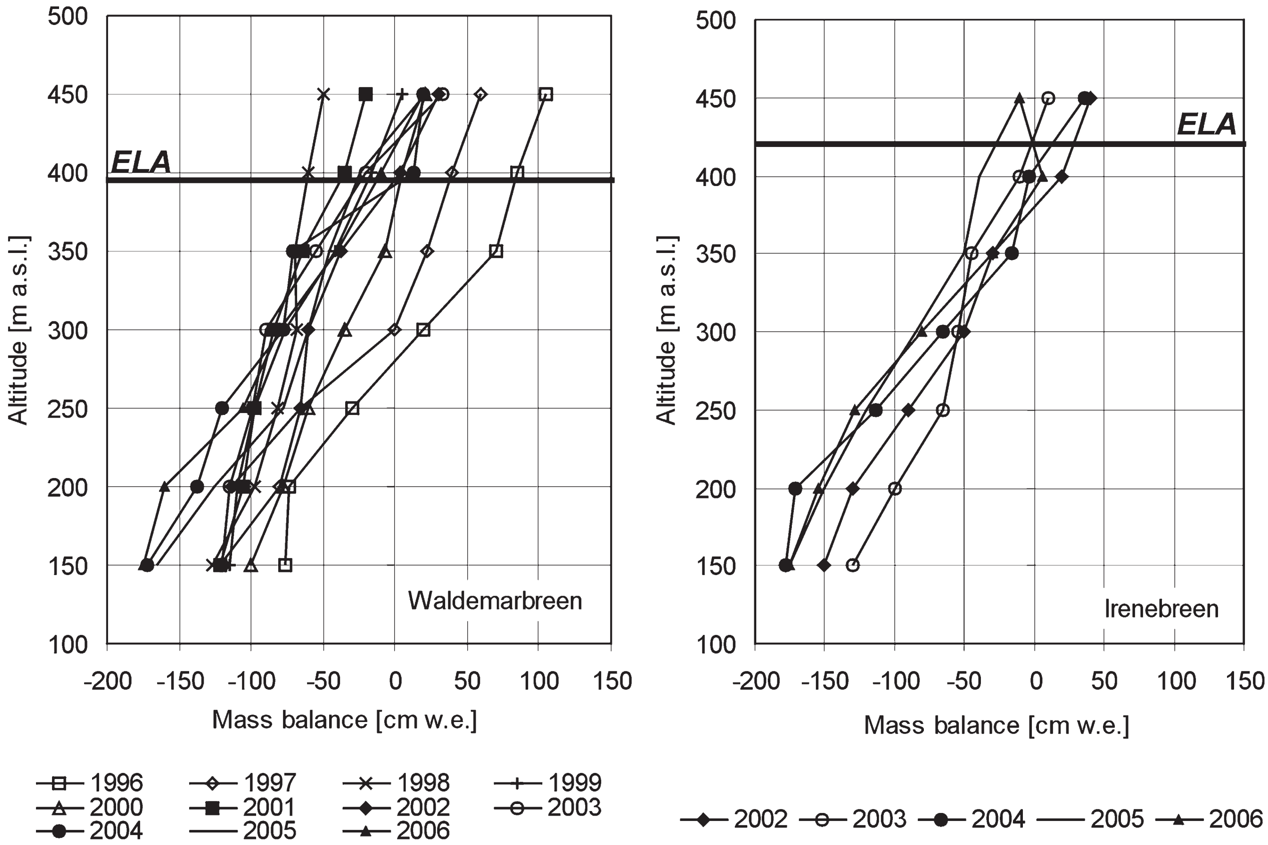

Fig. 2. Mass balance as a function of elevation for the period 1996–2006

was estimated at the altitude of 397 m in 1996–2006. From 2002 to 2006 the annual equilibrium line altitude was 421 m a.s.l. for Irenebreen, while for Elisebreen it was 365 m a.s.l. (Fig. 2).

The average mass balance of Waldemarbreen amounted to -57 cm w.e. in 1996–2006. Between 2002 and 2006 the mean annual mass balance of Irenebreen was -71 cm w.e. In 2006 the mass balance of Elisebreen was -73 cm w.e., and was similar to other glaciers of the region, even though its winter mass balance was larger. In this case the important factor was the low altitude of the frontal part of the glacier which results in intensive ablation.

Mean annual net mass balance of Kaffiøyra glaciers (Waldemarbreen and Irenebreen) is close to other Svalbard glaciers of similar size. These glaciers have negative long-term mass balance.

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Selected climatic and geodetic methods for estimating the mass balance of Waldemarbreen, Svalbard

Ireneusz Sobota*

Department of Cryology and Polar Research, Institute of Geography, N. Copernicus University, Toruń, Poland

The net mass balance of Waldemarbreen, Svalbard, has been measured with a glaciological method since 1996 (Sobota 1999, 2000, 2003, 2004, 2005). Several indirect methods, however, were also used for estimating its mass balance. The comparison of a current map with that of 1978, and climatic records enable us to calculate the change in the mass balance of Waldemarbreen which has taken place for over 34 years.

Direct glaciological investigations of the glacier mass balance are not always possible. Therefore, an attempt to assess the mass balance of Waldemarbreen through indirect methods was undertaken. The obtained results prove both validity and accuracy of these methods. They made it possible to define and recognise the variable elements of the glacier balance over dozens of years preceding the study period. These methods were divided into two major groups: climatic and geodetic.

PDD-model, PT-model and mapping metod were comapred with direct mass balance observations of standard glaciological method on Waldemarbreen. The results were correlated with observations and made a mass balance reconstruction back to 1970.

Waldemarbreen is located in the northern part of the Oscar II Land, Kaffiøyra, north-western Spitsbergen (Fig. 1). Kaffiøyra is a coastal lowland situated on the Forlandsundet. In the north it is bordered by Aavatsmarkbreen, which terminates in Hornbaek Bay, and in the south by Dahlbreen and the bay of the same name. Waldemarbreen is about 3.5 km long and has an area of 2.6 km^2.

Climatic methods are widely applied all over the world for glaciers investigations. They are based on the correlation of mass balance elements with meteorological parameters. The influence of such elements as air temperature or precipitation on the values of summer ablation and winter accumulation

Fig. 1. Waldemarbreen during summer time (photo I. Sobota)

* e-mail: irso@geo.uni.torun.pl

can clearly illustrate accuracy and representation of these methods.

The glacier mass balance depends on weather conditions prevailing in a particular balance year. Climatic methods use meteorological input data to estimate the mass balance. A method to estimate an empirical mass balance was the least-squares regression of the measured series data with two climatologic parameters: PDD – the sum of the positive daily average values of air temperature (positive degree-days, PDD-model), T – the daily average air temperature (PT-model), (method called 1 and 2).

The glacier mass balance was also estimated through the comparison of winter snow accumulation and summer ablation (method 3). The winter accumulation was calculated as the sum of precipitation in a given balance year, in a period with daily average air temperature less and equal 0°C. On the other hand, ablation was computed on the grounds of the relationship with the sum of the positive degree-days (PDD) in the period from June to September.

Geodetic methods involve detailed analysis of topographic materials, and aerial and satellite photographs of the studied area (Bukowska-Jania, Jania 1988, Echelmeyer et al. 1996; Finsterwalder 1954; Jania 1988; Jania et al. 2006; Krimmel 1999). They are based on the comparison of the accurate topographic maps and the determination of the volume change for the period between the photogrammetric surveys. They enable to compare the glacier mass balance in a given region and estimate its entire size over a certain period of time. A method similar to the one proposed by Finsterwalder (1954) was applied to estimate the mass changes of Waldemarbreen over a period longer than the direct field measurements. Maps of Waldemarbreen were made in 1978 and 2000 (Lankauf 2002).

The average mass balance of Waldemarbreen, computed by climatic methods, amounted to -0.42 m a^{-1} of water equivalent (w.e.) for the period of 1970-2004, and to -0.51 m w.e. for the period of 1996–2004. These balances were compared with the glaciological balance over the same period, which amounted to -0.54 m w.e. between 1996 and 2004. The geodetic balance was also computed, giving -0.52 m w.e. from 1978 to 2000. Waldemarbreen mass balance was found to have been negative over the years, yet there were years with a positive balance. The formulas used for Waldemarbreen may be also useful for estimating the mass balance of other Svalbard small glaciers which which terminate on land.

The methods which were used enabled the author to define successfully the variability of the cumulative mass balance of Waldemarbreen. It was found that the cumulative mass balance in the years 1978–2000 was -9.48 m w. e. (method 1), -7.82 m w.e. (method 2), and -9.12 m w. e. (method 3). Additionally, the geodetic method gave a similar value of the mass balance in the multi-annual period of 1978–2000, i.e. -11.60 m w.e. (Fig. 2). The value received by the geodetic method is larger, which is the consequence of the estimated error of ±0.09 m w.e., the cumulated value of which is ±2 m w.e. This method, however, shows the mass balance changes of the glacier during a multi-annual period of time if based on the analysis of just two maps drawn at different times. As a result, the geodetic method may be a very good way to estimate the mass balance of the glaciers for multi-annual periods of time when direct glaciological measurements are not conducted. It must be remembered, though, that climatic methods also carry errors which result in differences in the cumulated values.

The conclusion which can be drawn from the used models, is that the mass balance of Waldemarbreen is primarily dependent on the mean temperature of the summer season and ablation, and, secondly, on the winter snow accumulation. The exceptions are the years of significant snowfall and snow accumulation. Such a high dependence on the weather conditions confirms the idea that the mass balance of Waldemarbreen, similarly to other glaciers, can be an indicator of climatic changes (IAHS(ICSI)/UNEP/UNESCO 2003, 2005; Haeberli 1995; Klok, Oerlemans 2004).

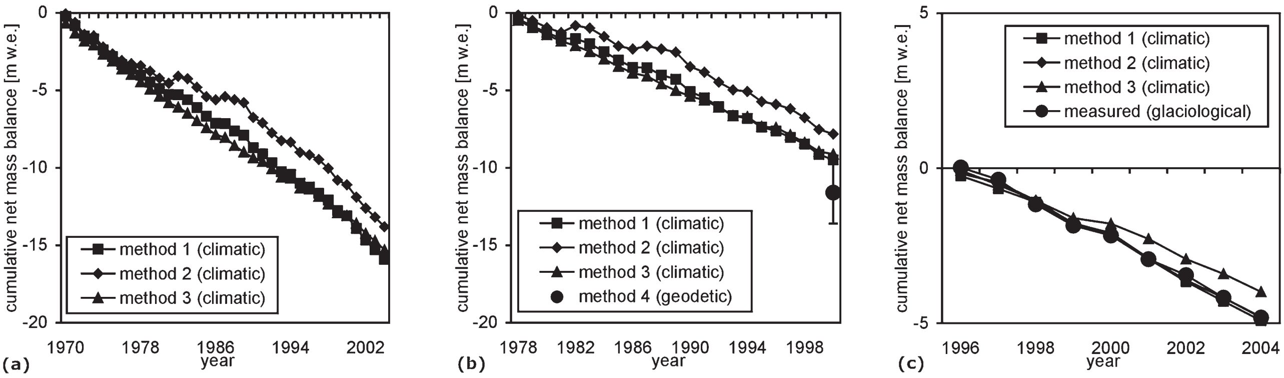

Fig. 2. Comparison of cumulative glaciological, climatic and geodetic mass balances
(a) 1970-2004 (b) 1978–2000 (c) 1996–2004

Mean glaciological mass balance value of glacier Waldemarbreen for the years 1996–2004 can be treated as a representative one for small valley glaciers of the north-western part of Svalbard. This conclusion is supported by the received values of mass balance for a long-term of time, which were based on the chosen climatic and geodetic methods. These methods enabled to select the years of the lowest mass balance, as well as the highest. Strict conformity of the results received with the use of indirect (climatic and geodetic) and direct (glaciological) methods means that they can be alternative ways of estimating glaciers' mass balance in the years where undertaking direct field glaciological research is impossible. The accuracy of the mass balance estimation may be increased significantly by using all the suggested methods, especially for longer periods of time.

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The dynamics of suspended and dissolved transport in a High-Arctic glaciated catchment in ablation seasons 2005 and 2006, Bertram River, Central Spitsbergen

Mateusz Strzelecki*

Faculty of Geographical and Geological Sciences, Adam Mickiewicz University, Poznań, Poland

According to Warburton (1999) proglacial river geoecosystems provide a key link between glacial processes and the wider (paraglacial) environment. One of the most sensitive indicators of changes in glacial environments which inform about interaction between glacier, climate and landscape change are suspended sediment yields (Hodgkins et al. 2003). Many studies of sediment storage were made in alpine regions, however relatively few researches have been carried out in the High-Arctic locations, what caused a gap in the understanding of fluvial matter transport in glaciated catchments.

This paper presents data from investigations on denudational processes carried out on the proglacial Bertram River located in Petuniabukta region in Central Spitsbergen. Major research data and conclusions are also summarized on the poster prepared for the conference poster session.

Catchment of the Bertram River exhibits many interesting geomorphological characteristics when compared to other proglacial river geoecosystems. Bertram River is distinguished by a waterfall system, which divide 4.9 km^2 catchment into two parts: upper – glaciated part located on mountainous plateau, and lower – where braided planform was formed within the bottom of the Ebba River valley. About 60% (2.9 km^2) of the total area of the Bertram catchment is occupied by small cold-type Bertrambreen, which since LIA is in a continuous retreat (Kłysz et al. 1989)). The lithology of glaciated part of the catchment includes Precambrian metamorphic rocks, Paleozoic dolomites, limestones, shales and sandstones (Szczuciński 2003). Sandstones contain iron compounds, which turns red water colour. The

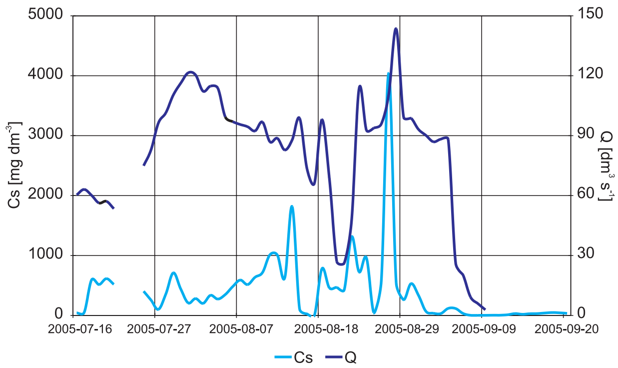

Fig. 1. Suspended sediment concentration Cs and river discharge Q in melt season 2005

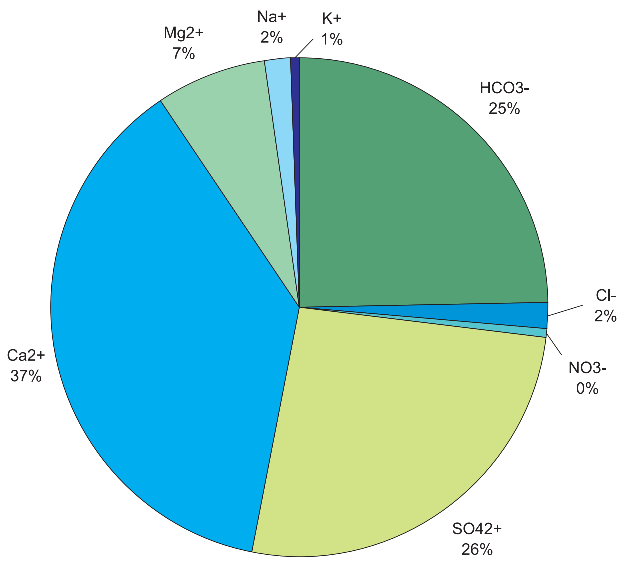

Fig. 2. Average chemical composition of Bertram River water in 2005 ablation season

* e-mail: mstrzelecki@mediasynergia.pl

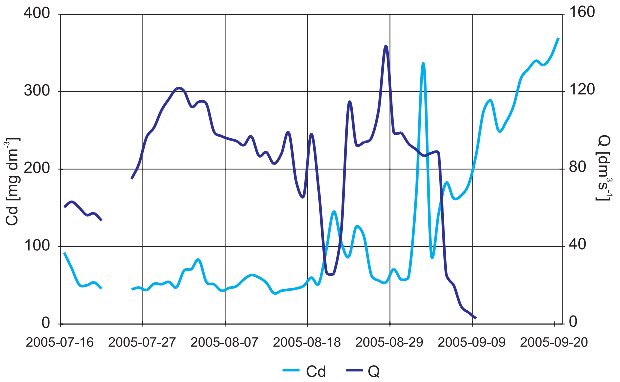

Fig. 3. Solute matter concentration Cd and river discharge Q in melt season 2005

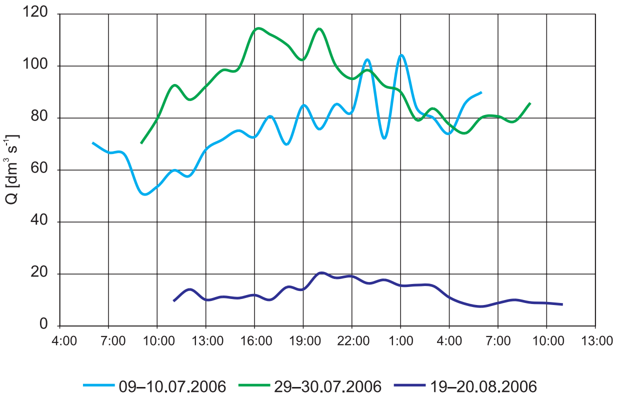

Fig. 4. Changes of water discharge Q during three days of 24-hour hydrochemical monitoring

lower part of the catchment is covered with unconsolidated and poorly sorted glaciofluvial deposits.

During the first expedition, which took place in melt season 2005, the main purpose was to observe the seasonal changes in discharge and concentration of suspended and dissolved matter (Fig. 1, Fig. 2) Investigations on chemical composition of glacial meltwaters (Fig. 3) were carried out simultaneously with observations of suspended sediments to estimate the participation of chemical denudation in the total rate of catchment denudation.

Daily hydrological and hydrochemical data and samples, collected between 16th July and 20th September 2005, were analyzed together with geomorphological surveys and meteorological observations. One of the most interesting events happened between the 27th and 28th of August when after few days of precipitation, strong and warm foehn occurred and intensified ablation causing bankfull discharge and extensive flood in the Bertram River and next in the Ebba Valley also. Extreme event which disturbed the seasonal distribution of fluvial transport served as a basis for 2006 detailed measurements.

The main purpose of the second campaign of measurements was to observe the diurnal fluctuations of matter concentration and stream discharge (Fig. 4) during three selected days during ablation

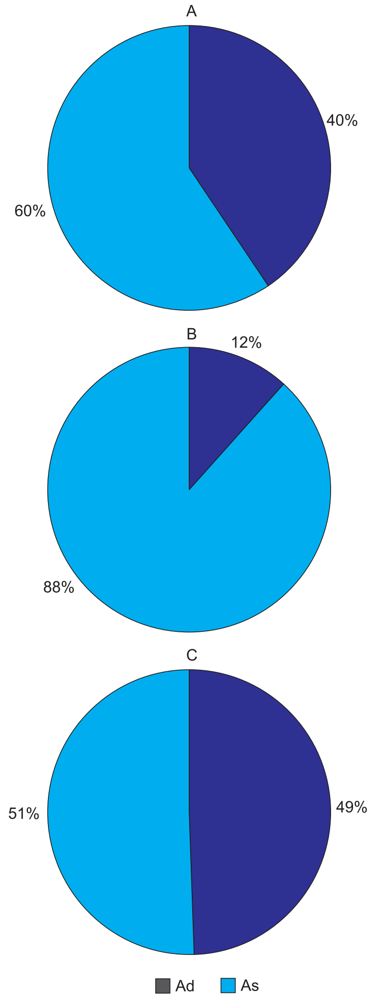

Fig. 5. Proportional participation of suspended sediment fluxes As and solute matter fluxes Ad in the total matter flux form the Bertram River catchment during 24-hour monitoring

A: 09–10.07.2006, B: 29–30.07.2006, C: 19–20.08.2006

season 2006. 24-hour hydrological and hydrochemical surveys were carried out:

a) at the beginning of the July (09/10.07.2006) in the phase of increasing discharge,
b) at the end of the July (29/30.07.2006) during the phase of the highest summer discharges,
c) at the end of August (19/20.08.2006) in early phase of the decay of water flow in the river.

The final results of the measurements revealed the compact dependence between the matter fluxes and water discharges. During the first observing season about 60% of denudated matter were suspended sediments (Fig. 5A). At the peak of the ablation season 2005 the amount of suspended matter in the total load of the fluvial transport increased to 88% (Fig. 5B). The results form the last survey from the second decade of August showed that during period of lower discharge the amount of suspended and soluted matter fluxes was almost equal (Fig. 5C).

Diurnal observations from 2006 season indicate different rate of reaction of fluvial system on weather

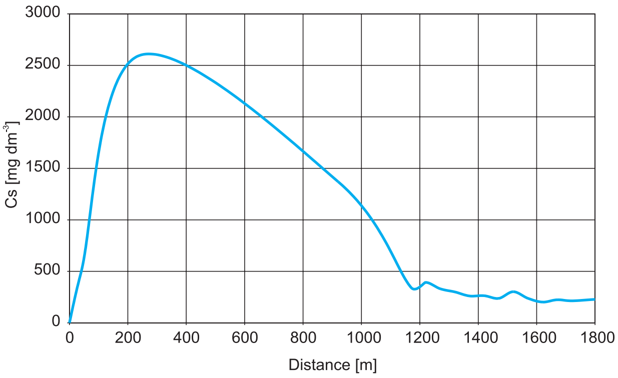

Fig. 6. Changes of suspended sediment concentration Cs along longitudinal profile of the Bertram River between the glacier source and river mouth

conditions. Maximum discharges were registered from 4 p.m. till 1 a.m. on the following day. Although different rate of transport and loads of suspended and solute matter during 24 hours demonstrate their significant dependence on meteorological conditions and definitely weaker dependence on bed rock lithology.

The last part of the observation in melt season 2006 was hydrochemical mapping carried out on the main channel of Bertram River. The research assumed observing the changes in water chemistry and channel morphology from the glacier front, through the waterfall system to the stream channel, where Bertram River joins the Ebba River. The strong connection between waterfall system and the change of suspended sediment concentration were found (Fig. 6) as well as the changes of solute matter concentration in the lower part of the catchment. Waterfall system located between the 920–1170 meter of the river length caused the dispersion of the suspended sediments and slow down the process of braided-channel pattern formation. The huge difference between chemical composition of meltwaters from the upper and lower part of the catchment will be also discussed.

Both researches from 2005 and 2006 observation seasons enable to present the detailed image of the geomorphological and hydrological conditions of seasonal and diurnal variations in suspended sediments and solute transport in High-Arctic catchment.

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Landform Analysis, Vol. 5: 85–86 (2007)

Geomorphological mapping in high mountain watersheds: the contribution of geomorphology to the evaluation of sediment transfer processes

David Theler*, Emmanuel Reynard

Institute of Geography, University of Lausanne, Lausanne, Swiss

We will present the first results of a PhD thesis carried out at the Institute of Geography of Lausanne University (IGUL, under the supervision of Prof. Emmanuel Reynard) and with the collaboration of the Roads and Watercourse Service and the Forests and Landscape Service of the canton of Valais, Switzerland.

Introduction

A huge range of geomorphological legend systems have been developed since the sixties all over the world and currently, geomorphologic mapping is one of the main research interests of the Institute of Geography at the University of Lausanne (IGUL) that has developed its own legend, based on various European legend systems (Schoeneich et al. 1998). The legend represents landforms by their genesis more than by their dynamics and has been used in the Swiss Alps for twenty years. In fact, the IGUL legend is mostly used for inventories and the management of landforms or landscape protection and it is insufficient to appreciate dynamic processes like debris flows.

Problematic

After severe floods in Switzerland in 1987, the Swiss federal laws and ordinances on river engineering and forests impose the responsibility of establishing hydrological hazard maps, which should become an obligatory tool for land planning. The method applied by the federal authorities consists of three steps: the first one consists in establishing a "phenomena" map produced by using field geomorphological evidence (Kienholz, Krummenacher 1995) ; then, based on this evidence, intensity maps are produced, either by numerical modelling and/or expert-system mapping; the last step, called hazard map, is a much more synthetic map, which shows the different degrees of danger and is based on two main parameters: intensity and probability of hazard. The hazard map allows the representation of five degrees of danger.

This methodology is also used for snow avalanche and rockfall danger. The variety of tools available leads to some inconsistencies in the field. In fact, the recommended legend for mapping the phenomenon only gives a momentary vision of one single event. In theory, the phenomena map should be redrawn after each new event and all the maps should finally be superposed to have a global view of the flooded area. Indeed, experiences in debris flow mapping (Bonnet-Staub 2001; Bardou 2002) have shown that landforms related to theses fluvial phenomena are very active and may change very quickly over a short time and space scale. Fore example, some important characteristics of an event / torrential system are not considered, like the integration of past events / history of the stream, distinction between punctual and potential sediment alimentation of a debris flow, distinction of the different processes in the deposition zone or distinction of different deposition landforms.

There is, therefore, a need for more detailed information about volumes of potentially mobilised sediments, especially in densely populated mountain regions with a high potential of natural hazards. A better cartographic recognition of the slope system

* e-mail: david.theler@unil.ch

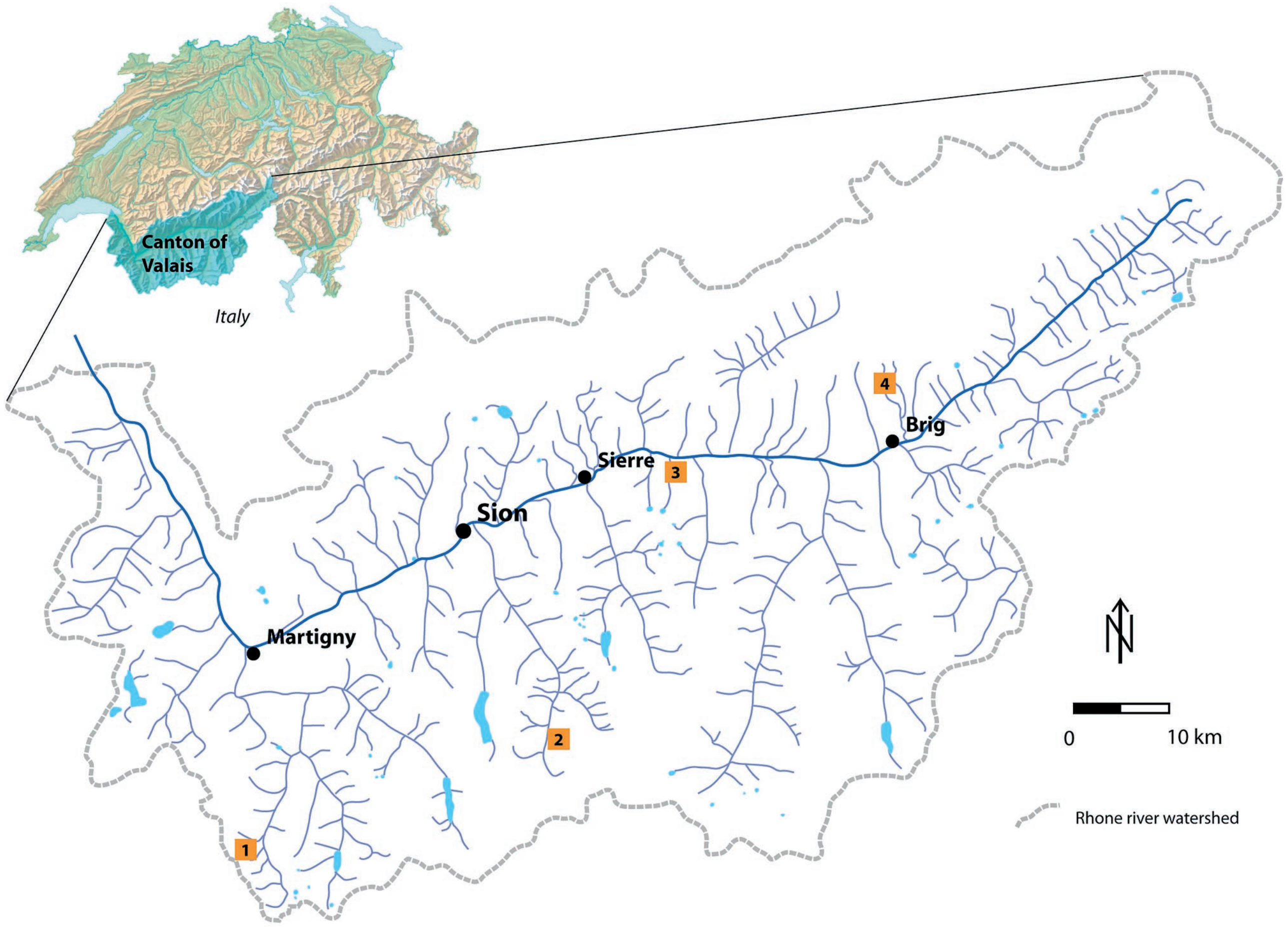

Fig. 1. Localization of the different study sites of the project within Rhone River watershed

and especially the sediment transfer processes, linked with a specific legend, could improve the knowledge on hydrological hazards in the studied area.

Project

Thus, since 2006, a new symbol-and-GIS based detailed geomorphological mapping system is in development. We have developed a conceptual geomorphic model, based on an "erosion system" from the top (rock escarpments, free faces in high altitude) downwards (alluvial fans, flood plain and scree cones, etc.). The aim is to consider the slope as a succession of connected reservoir subsystems varying in storage periods and emptying velocity. These reservoirs depict glacial processes and landforms (till accumulation, morainic bastions, etc.), periglacial processes (permafrost creeping, rock glaciers, solifluction, etc.), gravitational processes (landslides, rock falls, etc.), fluvial processes (debris flows, alluvial fans, etc.) and snow processes (snow avalanche deposits). This methodological approach may be used to quantify the postglacial sediment filling of alpine valleys (e.g. Schrott et al. 2003). The legend system should be able to consider all the factors governing sediment transfers, to produce susceptibility erosion maps and quantify the potential volume of sediment that may be mobilize.

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Chemical weathering on the glacial foreland of Storbreen, Jotunheimen Mountains, Norway

C.E. Thorn*

Department of Geography, University of Illinois, Urbana, USA

R.G. Darmody

Department of Natural Resources and Environmental Sciences, University of Illinois, Urbana, USA

C.E. Allen, S.W. Campbell

Independent Scholar

Storbreen is a cirque glacier in the Jotunheimen Mountains of Norway. Its Little Ice Age maximum occurred about 1750 when it deposited a well-defined end moraine. Since then it has retreated leaving a series of recessional moraines. All of these moraines have been dated lichenometrically by Matthews (see Matthews, 1992 for a comprehensive summary). We used this series of dated moraines to investigate the early stages of pedogenesis, the early stages of the chemical weathering of cobbles, and the impact of variation between 1750 moraine crest and moraine proximal base positions upon soil chemistry and mineralogy.

The overall pedological research design (Darmody et al. 2005) embraced three distinct elements. First, there was a constant elevation sequence undertaken on the southern flank of glacier foreland with triplicate soil pits excavated at the present glacier snout, on the crests of the AD 1750, 1810, 1870, and 1928 moraines, and finally upon an approximately 10 000 year-old surface beyond the glacier foreland. Second, there was a vertical sequence investigated up and down the 1750 moraine (low = approximately 1165 to 1180 m a.s.l.; middle = approximately 1310 to 1330 m a.s.l.; high = approximately 1400 to 1465 m a.s.l.). This sequence included matched triplicate pits on the 10 000 year-old surface, the 1750 moraine crest, and the 1750 moraine proximal base at all three elevations. Third, there was a skeletal matching of 1750 moraine crest pits on the southern and northern flanks of the foreland at the three aforementioned elevation levels. Within the overall pedological study a reduced set of the moraine-crest sites was used to study the development of chemical weathering within surficial lichen-free-free, surficial lichen-covered, and buried cobbles. Porosity within feldspar minerals was determined with backscatter electron microscopy. The third element of the study compared and contrasted clay mineralogy and soil chemistry up and down the 1750 moraine to determine if the published identification of a 'green zone' had measurable significance beyond the lichenometric one already demonstrated in the literature.

Soils were mostly frigid or isofrigid, coarse textured, and poorly developed Cryorthents (American Soil Taxonomy). There were differences between same-age soils at different elevations, same-elevation soils of differing ages, and soils from moraine crest and base pairs. Primary minerals, quartz, mica, feldspar, and amphibolites dominated soil mineralogy. However, secondary minerals, in particular hydrobiotite, increased with age and elevation. Despite the generally poor soil development, detectable topochronosequence differences in soil and associated weathering trends emerge in this young, cold environment.

* e-mail: cthorn@uiuc.edu

The initial stages of cobble weathering, measured as increasing porosity, were calculated for sets of cobbles taken from in front of the 1998 glacier snout, the 1928, 1870, 1810, and 1750 moraine crests, and from the 10 000 year-old land surface beyond the Neoglacial foreland limit. Findings indicate that cobbles close to the glacier snout are largely unweathered, also weathering is generally weak in the 1928, 1870, and 1810 positions, but statistically significantly higher in the 1750 and 10 000 year-old positions. Weathering of buried cobbles always exceeds weathering of exposed cobbles and may possibly reach a value beyond which it cannot progress while retaining surface cohesion. The degree of weathering on lichen-free and lichen-covered cobble surfaces is not initially distinguishable, but diverges sharply after ~250 years when lichen-covered surfaces experience significantly higher totals. Overall, the weathering trends in cobbles match those found in soils at the same sites.

Haines-Young (1983) demonstrated that there was a clear statistically occurrence of larger lichen thalli, Rhizocarpon geographicum spp., along the base of the 1750 moraine at Storbreen when compared to matching moraine-crest positions. We investigated crest-proximal base pairings to determine if this lichenological phenomenon had a matching soil and/or chemical weathering pattern. Our results were variable, in many instances, e.g., loss on ignition, organic matter, and carbon content, there was no statistically distinguishable results. However, in a limited number of instances, e.g., percentage of the secondary mineral hydrobiotite, statistically different results emerged between matching pairs. We take such differenced to reflect the fact that positional differences that have prevailed for approximately 250 years are sufficient to establish microenvironmental differences in chemical weathering. Such differences are presumed to be associated with ground moisture and temperature variability derived from different, but repetitive, seasonal snowcover differences between moraine crests and bases.

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Pressuremeter test in glaciated valley sediments (Andorra, Southern Pyrenees) Part one: An improved approach to their geomechanical behaviour

Valenti Turu*

Igeotest SL, Marcel Chevalier Foundation Project, Andorra la Vella, Principality of Andorra

The Andorra Glaciated Valley

Setting

The Principality of Andorra is a little county (465 km^2) located between France and Spain (42°30'N, 1°30'E), at the foothills of the SE Pyrenees. The tributary valley system has an "Y" shape and at the Upper Pleistocene different glaciers came together (Turu et al. 2007) to the main valley, where actually is locate the biggest city (Andorra la Vella) of Andorra. Understanding the stratigraphy of the glacial loaded sediments of Andorra is particularly important for civil engineers (Turu 2000). Glacial sediments produced during Quaternary glacial periods are widespread in both mountainous and lowland zones and influence many construction projects. One of the characteristics of such sediments is the great variability and unpredictability of the consolidation state and accurately geotechnical and geophysical surveys are needed.

Geomechanical data, pressuremeter tests

Intensive investigations of the architecture and character of valley floor sediments have been undertaken in the main Valley, in association with site investigations for major constructions until 1995 (see Turu et al. 2007) with up to 900 geotechnical surveys in the country.

The conclusion of all those surveying years is that the best geotechnical data to obtain the stress/strain behaviour of glaciated sediments are pressuremeter tests data.

The theoretical basis for this test was provided by Ménard and Baguelin et al. (1978) who also created a commercial design. Interpretation procedures are described by AFNOR (1999, 2000). In this test, a pneumatic cell, with flexible walls in a metallic slotted-tube is pushed into a pre-existing bore-hole. This push-in technique (Reid et al. 1982; Fiffle et al. 1985) reduces possible soil disturbances. A hydro-pneumatic system controls cell pressure, and expanding cell walls exert a horizontal stress on the bore-hole walls, whose deformation is concurrently measured by the expansion of the cell wall. Once the test is ended the pneumatic cell and the slotted-tube are extracted, cleaned, eventually repaired and calibrated.

Basically, when a certain pressure threshold is exceeded, volume expansion of the pneumatic cell increases rapidly, marking the change from elastic to plastic soil behaviour.

Rheological interpretation is based on the assumption of radial expansion of a cylindrical form in an isotopic elasto-plastic medium (Cassan 1982), and the test also yields the Young's modulus of the soil for a given value of Poisson ratio.

Stress/Strain analysis, the pressuremeter data

The most relevant data obtained will be synthesised in this paper without taking into account their geological setting, specifically data obtained from pressuremeter tests.

* e-mail: vturu@andorra.ad

As previously stated, this test has been performed in bore-holes, introducing the cell at depths between 5 and 25 meters which, in the best case scenario, implies ground pressures acquired according to a gravitational gradient between 0.1 to 0.5 MPa. However, with pressuremeter tests, overconsolidation pressures up to ten times greater than these have been obtained, implying that glacial sediments may be strongly consolidated.

Stress/strain data (pressuremeter P/V data) obtained permit us distinguish basically three types of charts (see Fig. 1):

- Type 1: P/V evolution with a single yield point
- Type 2: P/V evolution with various yield point
- Type 3: P/V evolution without any apparent yield point and strain rebounds are observed (ratcheting)

Type 1 P/V evolution is that which is most commonly described in the literature, a linear stress/strain behaviour from elastic domain is observed until a yield point is reached where start non-linear stress/strain behaviour from the plastic domain. Type 2 P/V evolution may appear in the literature, but is generally interpreted in the same manner as type 1, and in certain cases this type of curve is attributed to poor e xecution of the test, perturbation of the ground tested or the influence of large boulders near the pressuremeter testing cell; but since the same kind of diagrams in widespread glacial sediments is observed (subglacial tills, melt-out tills, glaciotectonites, lateral tills), we should think as inherent to those sediments, only in soft rocks with penetrative cleavage had been also observed (Devinzenci, Turu 1999). Type 3 P/V evolution is generally interpreted in the literature as corresponding to very compact ground, ratcheting is observed by strain rebounds on that type 3 diagrams, but no notice is known int the specialised literature about that phenomenon. The exact value of the critical state or yield point being usually unknown.

Discussion

The discussion deals with the purpose of this paper, the rheological interpretation of type 2 and type 3 pressuremetric curves. I will begin by explaining type 1 and continue with the subsequent types.

Type 1 P/V curves

These present a unique yield pressure which may correspond to pressure that is gravitational (normally consolidated), or perhaps greater than gravitational (overconsolidated). Commonly type 1 diagrams are interpreted using elasto-plastic models (i.e. Modified Cam-Clay), where the elastic behaviour is equivalent to those obtained in oedometric tests (one dimensional compression tests with lateral constraint), and the plastic behaviour is interpreted using the Coulomb failure criterion (Fig. 1a).

In Andorra this curve can be obtained if the effective pressure in the system has always been increasing or constant, with no load or unload cycles due to an ancient subglacial drainage. Usually the sediments showing type 1 diagrams had not shear strain structures, so the consolidation of those sediments were acquired in a low subglacial shear stress context.

Type 2 P/V curves

More than one yield point is observed in that type of diagrams (Fig. 1b). We can attempt to interpret that behaviour by continuous hyperplastic constitutive model in which continuous stress/strain memory (Einav et al. 2003) is related. So in type 2 diagrams the tensional history of the sediment is archived.

Usually the sediments showing type 2 diagrams have shear strain structures, like most of the subglacial tills (Evans et al. 2006). Hyperplasticity is based in the modified cam-clay constitutive model (Einav et al. 2003), and some particularities should be taking in account when pervasive subglacial shear stress is present.

The zone of till where the available shear strength is less than the constant pervasive subglacial shear stress imposed by the overlying glacier ice, undergoes critical state consolidation (Quan, 2005). That can be explained by modified Cam-Clay constitutive model, where small load-unload hydrological cycles produce that the stress state of the subglacial sediment moves away or close from the critical state line (Fig. 2a). Such consolidation is known as critical state consolidation (Quan, 2005) and can be more than 1.8 times greater than the isotropic consolidation. In other hand if the available shear strength is beyond the constant pervasive shear stress, the effective stress path goes away from the critical state at constant shear stress. Such consolidation acquired with constant shearing (Quan 2005) is lesser than the isotropic consolidation, especially for low effective pressure increments.

From geomorphology data (high position of lateral moraines) is possible to say that preconsolidations obtained from type 2 diagrams in Andorra, are always lesser or quite equal to the gravitational ice consolidation. In that sense something happen that inhibit the critical state consolidation in Andorra.

In that sense is known that for temperate glaciers, meltwaters drainage is subjected to climatic, annual and even diurnal cycles (see i.e. Boulton et al. 2001). All the subglacial hydrology is ruled to those melting cycles, the load and unload cycles transmit pore water pressure variations in the subglacial aquifer. Crit-

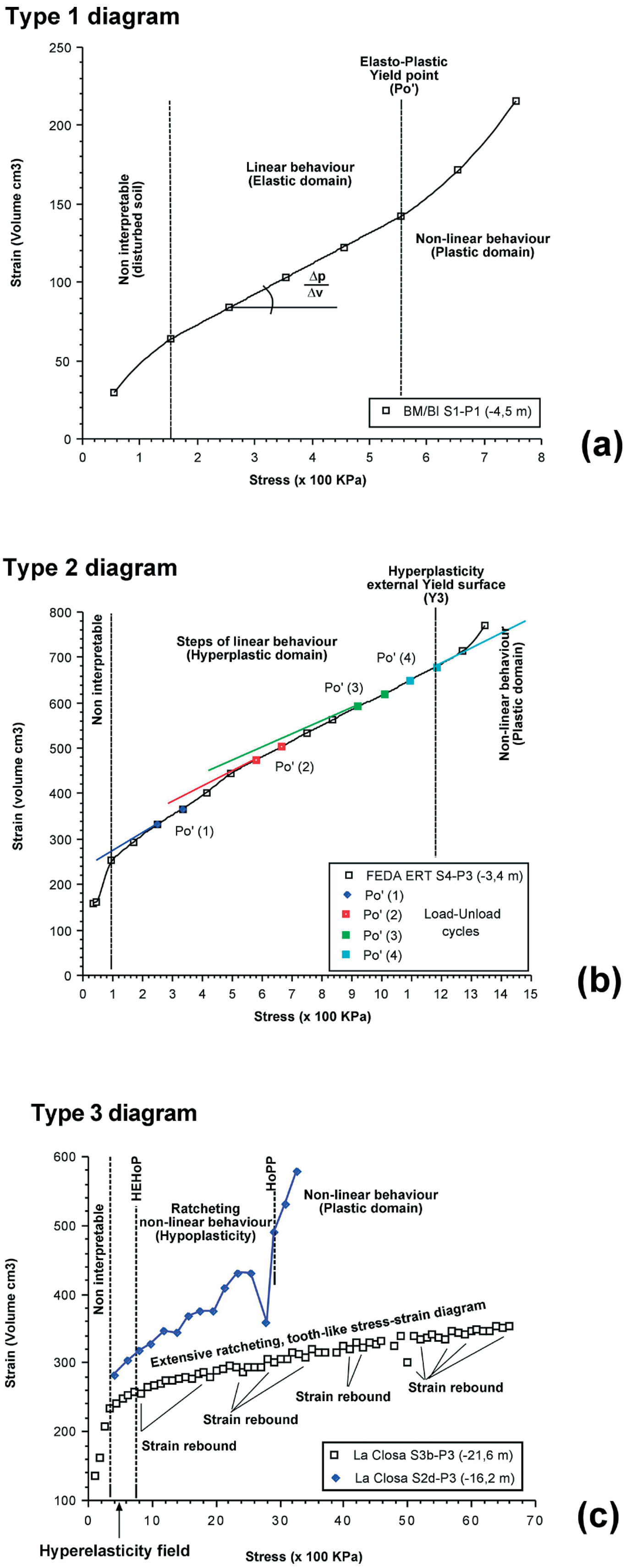

Fig. 1. Most representative type of stress/strain diagrams from pressuremeter data

a) Type 1 diagram showing an elasto-plastic behaviour. We can distinguish an elastic domain where deformation modulus is obtained by G = k ?p/?v (k is a pressuremeter constant). b) Type 2 diagram with a hyperplastic behaviour, showing the ability to record the stress/strain history of soil. Four pressure steps with a growing stiffness (less slope) of linear behaviour. c) Type 3 diagram showing strain rebound, called ratcheting as it is similar to those described in hypoplastic models. Two yield points can be distinguished were ratcheting happen between both, an hyperelastic yield point (HEHoP) and a Hypoplastic yield point (HoPP), over which failure criterion is reached

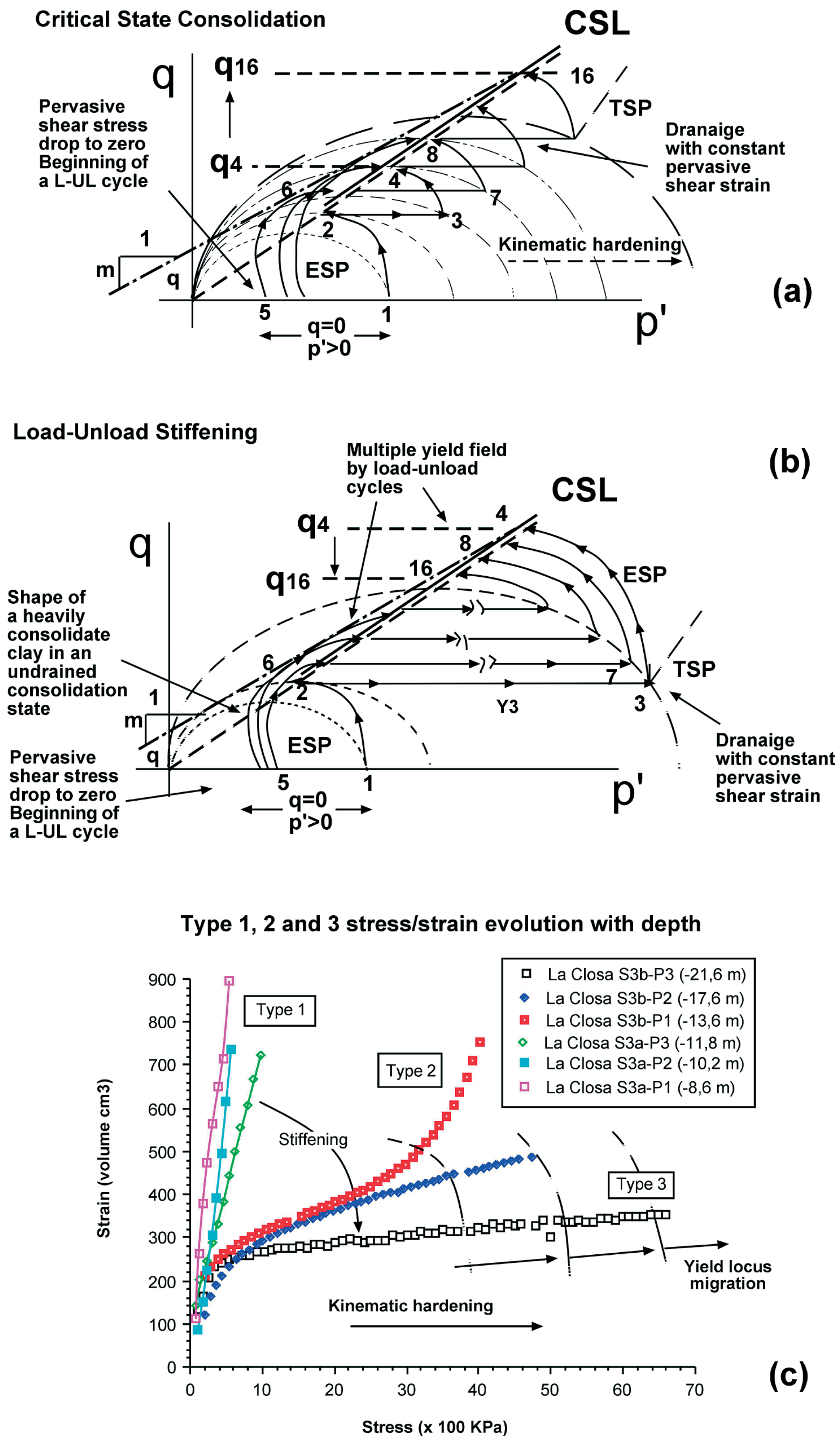

Fig. 2. Idealised behaviour of the glaciated sediments
a) Using a Modified Cam-Clay diagram of increasing or decreasing load-unload (L-UL) cycles. In a increasing strain/stress evolution, load and unload cycles with a constant pervasive shear stress can produce a critical state of consolidation. b) Using a MCC diagram in a decreasing stress/strain evolution the soil can show more than one preconsolidat,ions. c) Behaviour evolution from the pressuremeters diagrams in depth. Stiffening diminishes the slope of the stress/strain diagram and kinematic hardening produce the migration of the locus yield

ical state consolidation can be inhibit after a load event with pervasive constant shear stress if the unload event is associated with a pervasive shear stress drop; or in other words, if the unload event is associated with a very bad drainage of the subglacial hydrologic system and the glacier lost contact (décollement) with its sole by flotation uplift.

If the net evolution of subglacial effective pressure over different cycles has been decreasing (Fig. 2b), it is known from constitutive models (elasto-plastic, hyperplastic, hypoplastic, hyperelastic, ...) that load and unload cycles stiffen consolidated sediments; and is manifested in the elastic field of the pressuremetric curve by a decrease in its slope (greater stiffness) by steps (Fig. 1b), with each step corresponding to a range of effective pressures of the load-unload cycles. The greater consolidation state can be rheologically assimilated to the expansion of the yield curve due to plastic hardening. If we follow the continuous hyperplasticity model (Einav et al. 2003) the outer most yield surface should be the Y3 hyperplastic yield surface (Fig. 2b). In the other hand, if the net evolution of subglacial effective pressure over different cycles has been increasing, the pervasive shear stress field consolidation can undergo the soil to critical state consolidation (Quan 2005). If pervasive shear stress is not negligible, increasing evolution of subglacial effective pressure over the different cycles will show preconsolidations greater than the decreasing evolution.

In hyperplasticity constitutive models three yield surfaces are used, a inner yield surface (Y1) were stress-strain answer is purely elastic, an outer surface (Y2) representing the outer boundary of non-linear behaviour, and both yield surfaces inside of a third one from modified cam-clay large-scale yield surface (Y3) that is the outer boundary of plastic behaviour. Type 2 diagrams multiple yield zone should be interpreted as a multiple elastic soil behaviour below the Y3 yield point (Fig. 1b).

Type 3 P/V curves

These curves have lost their tensional history and I think that those diagrams correspond to an evolution toward the hyperelasticity and hypoplasticity of type 2 curves, let me explain:

The consolidation of the subglacial sediments situated near hydraulically singular points (subglacial tunnel drainage), is subject to an intense flow of water due to being situated near the place of drainage where there is a high hydraulic drop, and therefore also subject to greater high pervasive shear stress. If high water flow through porous media produce fine grain cleaning (supported by soil analysis and geophysical data in Andorra), subglacial shear stress can rearranges the sediment grains. The soil will appear to be undergoing consolidation when its stress state is close to critical state (Quan 2005), reflecting a consolidation pressure greater than the isotropic one (Fig. 2a).

The different load-unload cycles of subglacial drainage not only lend greater stiffness to the sediment in the elastic stage, but the progressive fine grain cleaning, together with the rearrangement of the grains, also provides denser packing leading the soil to reduce its void ratio to such a degree that granular contact does not permit it to consolidate further.

Dense packing of glaciated sediment grains was detected by Turu (2000) in Andorra comparing seismic shear modulus with the pressuremeter shear modulus.

Hyperelasticity can explain easily the behaviour of dense packing soils for small strains (see Niemunis 1996; Niemunis, Cudny 1998), where the stress is transferred through the porous media and small intergranular strain occurs without new rearrangement of grains, so the strain can be considered as reversible. Nevertheless different behaviour is expected for large deformations.

For extreme stress ubiquitous ratcheting effects may be possible (Niemunis com. pers. 2007) and has been observed in type 3 stress/strain diagrams (Fig. 1c). Typical saw-tooth-like stress-strain diagrams are obtained in the vicinity of yield stress predicted by the hypopasticity models, but since now not observed experimentally because the performance of the model in comparison to experiment were evidently poor (Niemunis, Triantafyllidis 2003).

So in type 3 diagrams different stress/strain behaviours can be observed. Hyperelasticity behaviour for intergranular small-strains (Niemunis 1996; Niemunis, Cudny 1998), while for larger strains extensive accumulation of deformation by load cycles leads toward an hypoplasticity behaviour (see Niemunis, Triantafyllidis 2003). Upon the hypoplastic yield stress more larger strains are obtained for small stress increments, leading towards a failure criterion behaviour.

Separation between hyperelastic and hypoplastic behaviours should corresponds to a inner yield surface (like Y1 hyperplastic yield surface) that we will call HEHoP; while an external yield surface (like Y3 hyperplastic yield surface), formed near the critical state, should corresponding to the separation between hypoplastic and failure behaviours that we will call HoPP (Fig. 1c).

Conclusions

The hyperelastic and hypoplastic behaviour of type 3 curves derive from previous hyperplastic behaviour from type 2 curves, while hyperplasticity of type 2 in turn derive from the elastic behaviour of type 1 curves. The principal mechanism to that evolution is due to load-unload (L-UL) cycles, producing stiffening and kinematic hardening of the subglacial sediment (Fig. 2c).

The evolution from type 2 to type 3 soil behaviour should start with a critical state consolidation (HoPP yield), wile the HEHoP yield point appear when the soil is led to a dense packing by further fine grain cleaning and rearrangement of grains. Between both, type 2 expansion of the yield curve due to plastic hardening by load-unload cycles derive to ratcheting in type 3 diagrams by extensive accumulation of deformation by those cycles.

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Landform Analysis, Vol. 5: 95–101 (2007)

Pressuremeter test in glaciated valley sediments (Andorra, Southern Pyrenees) Part two: Fossil subglacial drainage patterns, dynamics and rheology

Valenti Turu*

Igeotest SL, Marcel Chevalier Foundation Project, Andorra la Vella, Principality of Andorra

Introduction

At the Upper Peistocene in Andorra, as the almost glacial valleys, several glacial tongues join to form a significant accumulation of ice at the end of their trajectory, in the ablation zone. In the same manner, glacial fusion waters were carried from secondary valleys to the main glaciated valley.

Meltwaters generally follow various paths until arriving at the snout, but significant amount enters through glacier crevasses and moulins as well as through lateral moraines, until saturating the subglacial aquifer (Menzies 1995). Eventually poor drainage of the system may accumulate water under the ice until a certain piezometrical height resulting from the balance between ice fusion and water drainage. If the glaciostatic pressure is exceeded the glacier follows the Archimede's law, basal contact is lost and a surge event can be produced (Nielsen 1969). Once subglacial drainage is again established efficiently, by one or several subglacial tunnels (see i.e. Boulton et al. 2001), the entire system is conditioned: glacial flow, aquifer drainage, subglacial shearing, subglacial sedimentation and erosion, and consolidation and dilation of the subglacial sediments.

In this sense, pre-existing morphologies may condition the position of these channels or tunnels beneath the glacier (Menzies 1995), such as subglacial gorges or the confluence of glacial tongues. Subglacial gorges constitute entryways for subglacial water from tributary valleys, while confluence between glaciers constitutes a lineal anisotropy from which, if conditions are favourable, a subglacial drainage tunnel may be formed. This is the case that appears to have occurred in the Andorra valley.

Following a profile parallel to the main axis of the valley, overconsolidation has been observed to increase upstream (Turu et al. 2007), that mean that the effective pressures where greater upstream rather than on the snout zone. That can be easily explained because upstream the glacier thickness is greater rather than in the snout zone, also greater meltwater is present at the ablation zone near to the snout for temperate glaciers.

The magnitude of the preconsolidations observed in Andorra should be taken as an indicative value of ancient effective pressure beneath the andorran valley glacier. The value of these preconsolidations are compatible with the presence of R and C subglacial drainage channels beneath the valley glacier (Menzies 1995). Following a profile perpendicular to the main axis of the Andorra valley, overconsolidation pressure has been observed to vary, being greater in the centre, so in ancient times effective pressures where greater in the mean valley, and a major tunnel or drainage channel might existed there.

Stress/strain data obtained in pressuremeter tests not only have been observed to vary regarding the location in the glaciated valley, but also in depth at the valley floor. As noticed in a parallel communication here, stress/strain evolution named Type 1, Type 2 and Type 3 P/V diagrams are observed in Andorra and discussed here taking into account their geological setting.

* e-mail: vturu@andorra.ad

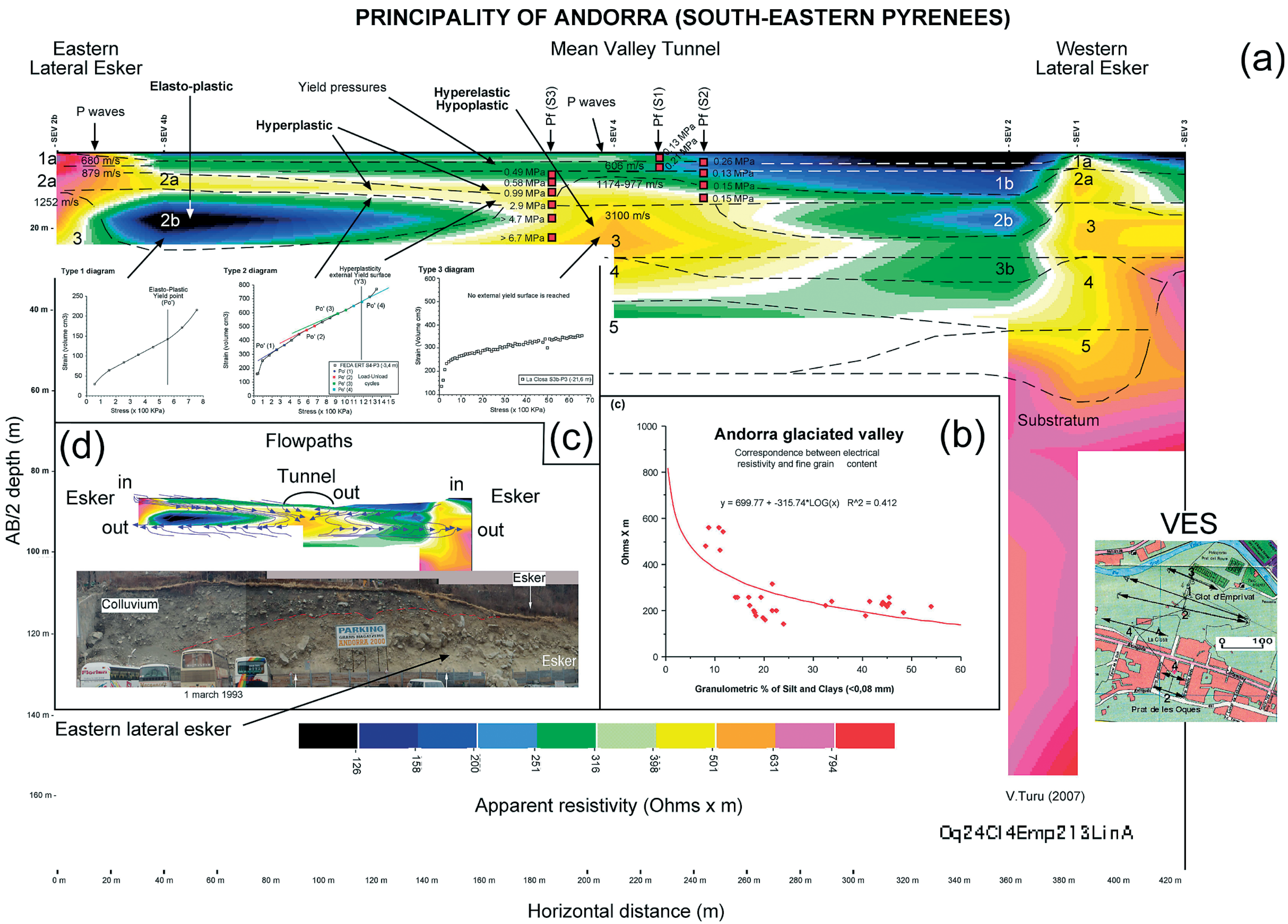
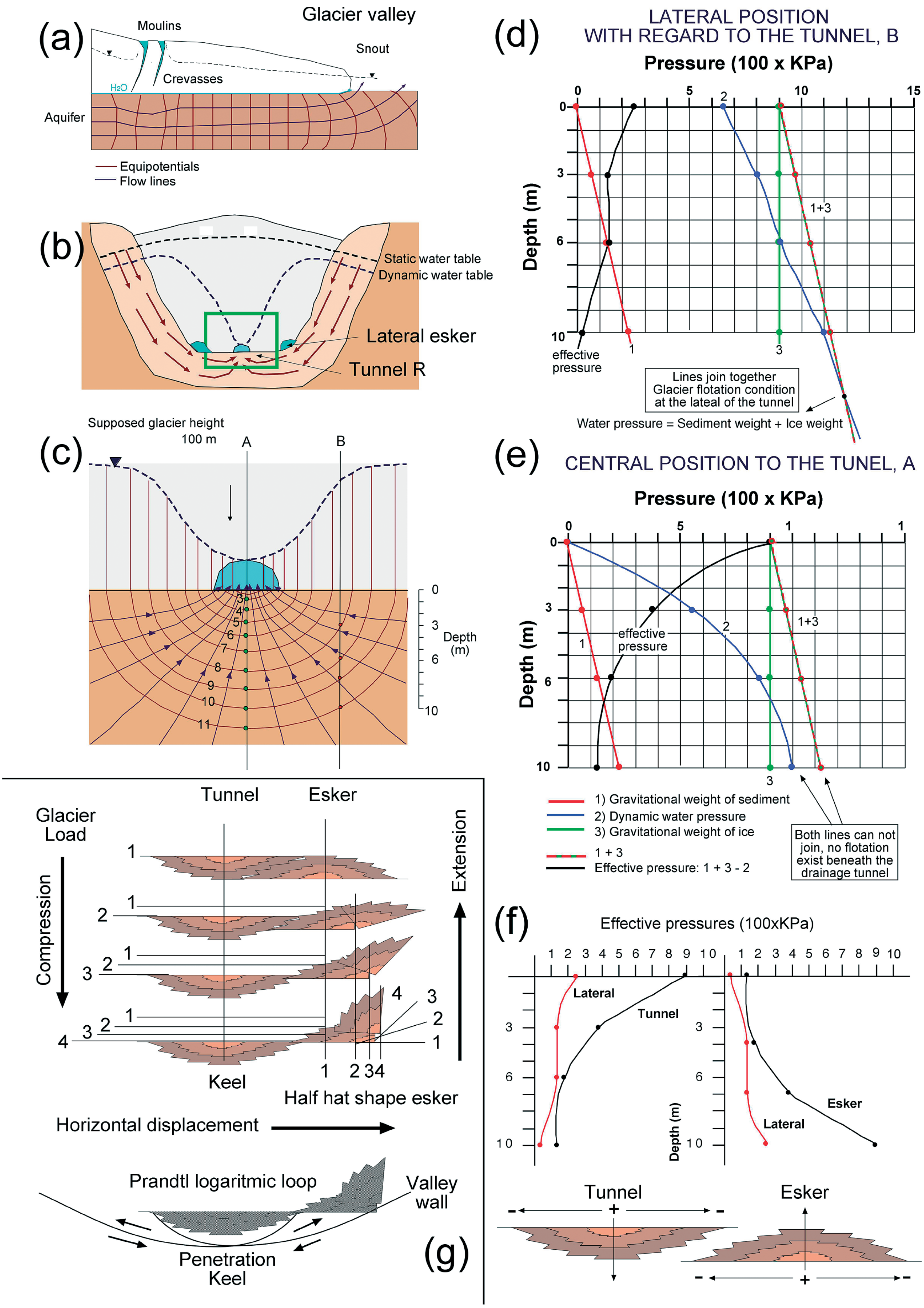


Fig. 1. Principality of Andorra (south-eastern Pyrenees
a: Resistivity profile from vertical electrical soundings (VES) using the Bovachev et al. (2003) software utilities. Apparent resistivity are plotted at an equivalent depth of AB/2 (half VES distance); b: Correspondence between electrical resistivity and fine grain content in the sediment. Note that under a 15% of fine grain particles (under 0.08 mm diameter) the resistivity changes quickly; c: Position of the representative stress/strain diagrams at the ancient subglacial aquifer. Note a close correspondence between high resistivity and high stiffness of the pressuremeter diagrams. The main rehological behaviours are also located; d: General flowpaths from an ancient subglacial drainage are represented. Central tunnel drain out the water from the subglacial system. Lateral water contributions came laterally from throughout the lateral eskers (in). Preconsolidation data (ancient effective pressures, Turu 2003a, b) show that ancient lateral eskers could act also as a drainage conduits (out)



(a)
Glacier valley
Moulins
Snout
Crevasses
H_2O
Aquifer
Equipotentials
Flow lines
(b)
Static water table
Dynamic water table
Lateral esker
Tunnel R
(c)
Supposed glacier height
100 m
A
B
Depth (m)
(d)
LATERAL POSITION
WITH REGARD TO THE TUNNEL, B
Pressure (100 x KPa)
Depth (m)
effective pressure
1+3
Lines join together
Glacier flotation condition
at the lateral of the tunnel
Water pressure = Sediment weight + Ice weight
(e)
CENTRAL POSITION TO THE TUNEL, A
Pressure (100 x KPa)
Depth (m)
effective pressure
1) Gravitational weight of sediment
2) Dynamic water pressure
3) Gravitational weight of ice
1 + 3
Effective pressure: 1 + 3 - 2
Both lines can not join, no flotation exist beneath the drainage tunnel
(f)
Effective pressures (100xKPa)
Lateral
Tunnel
Esker
Depth (m)
Tunnel
Esker
Glacier Load
Compression
Tunnel
Esker
Extension
Keel
Half hat shape esker
Horizontal displacement
Prandtl logaritmic loop
Valley wall
Penetration
Keel
(g)

Fig. 2.

a: Temperate glacier at its ablation zone can be assimilated as a karstic aquifer with englacial conduits. Metwaters can be infiltrate through moulins and crevasses to deeper levels until reaching the subglacial porous media. In turn the aquifer can be drained by channels present at the subglacial floor (i.e. R channels); b: The aquifer drainage generate water flow through the porous media and the flowpath follow the piezometric gradient. Lateral water inputs can be present, specially from lateral valleys lateral eskers and by lateral moraines. The subglacial drainage is lead by a tunnel between the lateral inputs; c: For a supposed saturated glacier height of 100 m aquifer drainage net is drawn. Water flow came from the aquifer to the drainage tunnel. Near to the tunnel (section A) water pressure drop quickly (equipotentials) but lesser in a lateral position from the tunnel (section B); d & e: From figure 2c example, evolution of effective pressure in the aquifer (line 2) with depth, in a lateral position with regard to the drainage tunnel (Fig. 2d) an beneath the tunnel (Fig. 2e). At the same time glacier load and glacier flotation can coexist beneath the subglacial floor; f: Aquifer effective pressures from figures 2d and 2e beneath the drainage tunnel and the opposite happen beneath a lateral esker water input. Beneath the tunnel preconsolidations might be bigger at the top of the strata and a "bicouche" can be formed. Beneath the esker preconsolidations might be bigger at the bottom of the strata and a inverted "bicouche" can be formed; g: High effective pressures at the top of the strata imply that glacier load is transmitted to the valley floor (compression), while high effective pressures at the bottom of the strata only imply consolidation of sediments by the Bernoulli effect, because high water pressures uplift the glacier and traction stress (extension) happen at the top of the esker. Such 2–D stress configuration will promote the collapse of the mean subglacial valley floor by uplifting the side subglacial valley floor, following the known Prandtl logarithmic loop failure criterion in common civil engineering. Side valley margins confine the subglacial sediments laterally, so the only way to generate more space for faulting is deforming the lateral sediments (eskers), and the result of that process is the pile-up of eskers related materials producing an half hat shape of the lateral eskers. Lateral eskers showing an half hat shape is quite common in Andorra (Turu 2003b). If there is further glacier load in the mean valley floor the double Prandtl logarithmic loop will generate a penetration keel and plastic hardening might happen for sediments inside the keel, also efficient glacier coupling can promote further consolidation by pervasive shearing and progressively reaching an hyperelastic-hypoplastic penetration keel under the glacier at the mean valley.

Stratigraphical architecture of the glaciated valley

Geoelectrical survey data represented in a transverse profile to the main axis of the Valley (Fig. 1a) shows a symmetrical distribution of electrical resistivity. The resistivity symmetry consists of the existence of three highly resistive cores, two of which are located in the sides of the valley and one in the centre. The position of the lateral high resistive bodies coincides with the position of the subglacial gorges of the tributary valleys, while the resistive body located in the centre of the valley coincides with the position of the confluence of the two largest glaciers. Between them low resistivity sediments are present. The group is stratified showing almost five geoelectrical units and are interpret as sedimentary starts.

On the other hand, has been empirically determined in Andorra that there is a strong relationship between fine grain content (grains less than 0.08 mm in size) and the resistivity (Fig. 1b). It has also been observed that lateral high resistive bodies are primarily formed by boulders, while no boulders have been detected by bore-holes in the central resistive body. The origin of the boulders must be attributed to lateral moraine erosion and to sedimentary contribution channelled by subglacial gorges. From these descriptions, those high resistive lateral cores could by assimilate into eskers. The high resistivity results from the scarcity of fine grains (< 0.08%) due to a cleaning of the matrix produced by significant channelled subglacial water flows (R or C channels, or tunnels).

Rehological architecture of the glaciated valley

The rheology of the sediments are related with its stress/strain behaviour. From parallel communication it is known that from pressuremeter P/V diagrams rehological behaviour from tested soils are obtained. In that sense if we plot the most representative P/V diagrams on the resistivity profile (Fig. 1c) some conclusions can be done:

1. Type 1 – diagrams are mostly located in the less resistive layers
2. Type 3 – diagrams exclusively are located in the resistive bodies
3. Type 2 – diagrams are widespread located, close to the others

Sediments showing Type 1 diagrams will present an elasto-plastic stress/strain behaviour (see Fig. 1c).

The sediments with Type 3 diagrams are restricted to the high resistivity core at the mean valley; hyperelastic behaviour for small strains (seismic waves) is expected, hypoplastic behaviour for larger strains is also expected, and finally for very large strains a failure criterion can be obtained.

Between them Type 2 diagrams domain, with sediment showing continuous stress/strain memory until hyperplastic yield is exceed, then a classical plastic behaviour is expected.

Noticed in a parallel communication Typer 2 diagrams, which are quite widespread in the glaciated valley, are related with ancient subglacial load and unload (L-UL) cycles related with ancient subglacial drainage.

Subglacial drainage pathways of the glaciated valley

It is also acknowledged in the literature (Boulton, Zatsepin 2001) that the glacial ablation process is not continuous through time and is subject to seasonal, daily and climatic cycles. Thus the subglacial sediments have been subjected to various load and unload (L-UL) cycles and generated the consolidation of subglacial materials, with the particularities mentioned in a parallel communication.

In Andorra Type 2 diagrams will show us the sites where the L-UL cycles have been recorded. Type 2 diagrams present more stiffness with depth but also laterally close to the high resistivity bodies. Also Type 2 diagrams present less stiffness in the low resistivity bodies, there where sediments with Type 1 diagrams also exist.

Two main subglacial drainage pathways can be distinguished regarding its valley position.

Drainage in the central part of the valley

Type 3 diagrams are the stiffest one, only present at the high resistivity body in the mean valley, and its presence is related with the most important piezometric drop in the glaciated valley (Fig. 2a-c). In that sense the resistivity data and the stress/strain data show us roughly an important drainage flowpath in the mean valley for the ancient subglacial aquifer.

Drainage in the lateral part of the valley

In essence, lateral eskers would basically correspond to zones of meltwater entry in the subglacial system, with the water being drained out of the system by the underlying granular aquifer as well as by the central tunnel (Fig. 2a-c).

The stratification observed in the valley by geophysical data clearly show a sedimentary accretion, closely related to the drainage process beneath the ancient glacier. The subglacial sedimentary accretion can be interpreted as a constructional process (Hart, Boulton 1991) and some consequences of that architecture in the subglacial dynamics are expected:

a) Abandoned eskers went no more directly connected with the lateral valleys drainage, but is expected that they could act as pipe conduits keeping hydraulically connected distant subglacial regions with different water levels.
b) Subglacial sedimentary accretion implies that the deepest layers have been subjected to more hydraulic cycles than the shallow ones. Also the layers close to the principal drainage pathways (central tunnel and the lateral eskers).

Valley glacier subglacial drainage pathways

From outcrops, bore-holes sedimentological data, and pressuremeter tests point out that, at the high resistive cores strata accretion is also present. Layers showing light stiffness Type 2 diagrams were detected in silty-gravely layers. In the high resistivity cores these layers have less thickness than the layers showing heavy stiffness, while at the low resistivity bodies these layers have greater thickness than the stiffen Type 2 diagrams.

The presence of these layers showing small stiffness, lightly consolidated, below layers with great stiffness (heavily consolidated), was firstly indicate and explained by Turu (2000) in Andorra (Fig. 2 d-f). At the mean valley both layers are always present together, named as “bicouches” by Turu et al. (2007). The heavily consolidated layer and the lightly consolidated layer from theses “bicouches” were named as “a” and “b” respectively by Turu (2000) and it’s geometry across the valley has be studied by Turu et al. (2007).

Type “b” layers were of great significance for the aquifer drainage, acting as a important drainway for the ancient subglacial system, so was not possible for those layers to consolidate further. Since that kind of layers are present in the aquifer, many of the drainage might go through keeping hydraulically connected the central tunnel and the lateral eskers.

Taking into account these particularities and the general behaviour of the subglacial drainage, the ancient flowpaths in the glaciated valley aquifer are drawn (Fig. 1d).

Subglacial dynamics of the glaciated valley

Subglacial pervasive shear stress should be also archived in the subglacial sediments, there where water pore pressures were low, specially at the mean valley where the central tunnel was present.

Subglacial coupling might happen at the mean valley position, at the same places where pervasive shearing was greater and best transmitted.

Should be noted that only the materials present at the mean valley show hyperelastic and hypoplastic behaviours for small and large strains respectively. Those materials show Type 3 diagrams and are the most consolidated in the valley. In a parallel communication is noticed that these consolidation can be easily 1.8 greater than those reflected in Type 2 diagrams from hyperplastic materials. Dense packing of the porous skeleton (Turu 2000) was expected for that kind of terrain.

It is known from the literature (Menzies 1995; Evans et al. 2006) that an efficient glacier coupling leads ploughing over unconsolidated sediments. In Andorra the ancient glacier might not have an efficient coupling at those strata showing Type 1 diagrams or lightly consolidated Type 2 diagrams, but coupling might be largely done at the mean valley position where heavily consolidated materials are present. Is expected that ploughing happen at the beginning of the consolidation process but might diminish for further consolidation.

If we take into account the "bicouche" structure of the strata from the ancient subglacial aquifer, pervasive shearing might not being transmitted to further depth, because the "b" type layer of the "bicouche" will significant reduce the pervasive shear stress transfer to further depth by its weakness, but ploughing of the whole "bicouche" could happen and substantial pile-up of "bicouche" can result (see Turu et al. 2007). That pile-up only could happen at the mean valley subglacial floor, there where was the subglacial tunnel. If the amass entails a drainage decrease in the tunnel, subglacial water pressure could grow submitting the glacier in a flotation condition toward a decoupling from its bed. Subglacial sedimentation can then happen and subaquaceus facies can be deposited (specially turbidites), as is explained by Brennand (2000) for subglacial meltwater drainage. When subglacial drainage becomes again efficient enough to permit a new coupling between the glacier and its bed, the new subglacial sediment undergo to consolidate following the Type 1, Type 2 and eventually Type 3 stress/strain behaviours and a new "bicouche" is formed.

Tunnel subglacial drainage did permit the glacier weight transmission to the mean valley floor, while at the lateral valley floor low subglacial effective pressures were present. These stress patterns at the valley floor could derive to an overload faulting following the suhoritzontal structure of the "bicouches", similar happen to shallow foundations when the bearing capacity is exceeded. Here hyperelastic and hypoplastic terrain will act as a shallow foundation, the glacier weight as the load, and the elasto-plastic & light stiffen hyperplastic materials (the "b" layer of a "bicouche") could only impose a low bearing capacity, so a pile-up is expected at the valley sides by the penetration keel of hyperelatic-hypoplastic (HEHoP) material under the glacier (Fig. 2g). It is noted here that if HEHoP keel produce further penetration ancient sediments will be preserved at the valley sides and it has been observed in Andorra by Turu et al. 2007) but also in many valley glaciers in the Alps (Nicoud et al. 2002).

Conclusions

Any subglacial sediment subjected to drainage with load and unload hydrological cycles should present consolidation patters similar to those here described. Without the use of pressuremeter tests might be impossible to obtain a significant number of strain/stress data to permit the rheology study of glacial sediments at Andorra. However similar rehological behaviours to those of type 2 curves have been obtained from oedometric tests, but the lack of data inherent to the granulometry of the glaciated sediments did not permit to get further data by that way. Without representative data of the whole family of subglacial sediments (Evans et al. 2006) the rheological study of them is almost impossible, but much research is still needed to be able to completely explain the rheological characteristics of subglacial sediments, especially comparative studies all over the glaciated areas.

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The conditioning of the evolution of NW part of the coast of Wedel Jarlsberg Land (Spitsbergen) during the last century

Piotr Zagórski*

Department of Geomorphology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland

In the area of NW part of Wedel Jarlsberg Land (southern Bellsund, Spitsbergen) the studies of formation of coastal zone began during the 1st Polar Expedition of Maria Curie-Skłodowska University to Spitsbergen in 1986 and continued the following years. They covered about 70 km long coast from Dunderdalen to the eastern coast of Rrecherchefjorden (Harasimiuk 1987, Harasimiuk, Jezierski 1991, Zagórski 2002, 2004). It is characterized by alternate abrasive and accumulative parts. Their spatial location and development depend mainly on geological structure of background as well as exposition to waving.

In this article, in reference to the last century, a special attention was paid to the evolution of a coast in the section from Skilvika to Josephbukta. In the Skilvika region the coast is in the form of cliff developed within the Proterozoic rocks (western part) and Tertiary rocks with the additional series of Quaternary sediments (eastern part) (Dallmann et al. 1990, Birkenmajer 2004, Landvik et al. 1992, Pękala, Repelewska-Pękalowa 1990). In the vicinity of Renardodden, due to the intensive accumulation, there were a few storm ridges formed, at present – fossil, on the surface of which numerous stations of XVII and XIX centuries settlements are located (Krawczyk, Reder 1989). In the section between Renardodden and Josephbukta, the coast is of accumulative character with a full profile beach. It is formed by the marine terrace, 2–8 m a.s.l. (terrace I) and 40-180 m wide, separated with a section of cliff shore in the zone of marginal moraine of Renardbreen (Harasimiuk 1987, Zagórski 2002). The terrace is built of sands and gravels transported to the shore zone by streams from the tundra area and a river flowing from the Scott and Renard Glaciers (Fig. 1).

During the last century the coast of NW part of Wedel Jarlsberg Land was and is still affected by various morphogenetic factors including littoral, glacial, fluvioglacial and fluvial processes. They reflect internal dynamics, feedbacks of atmosphere, cryosphere and hydrosphere. The Little Ice Age was a remarkable glacial episode in this area. Its end is dated on XIXth and XXth centuries (Isaksson et al. 2005). Large glaciers getting into the sea like Renardbreen and Rrecherchebreen largely affected coast transformation. Their marginal zones invaded partly the terrace I level (Fig. 1). There, among others, exarative redeposition of sediments and fossil flora took place, e.g. in the case of Renardbreen forefield – fossil flora dated with the radiocarbon method as 660±80, 1040±80 and 1130±80 BP (Dzierżek et al. 1990). Additional finding is enriched with an occupation layer (Renardbreen 1) glacially remodelled (Jasinski, Starkov 1993). Based on the archaeological-geomorphological work, there were found small changes of sea level that were probably due to glacio-isostasy (Jasinski et al. 1997). Another effect could be also caused by intensification of abrasion processes.

During the last century the most important factors that affect transformation of coasts were marine processes (waving, tides, longshore current). Within the coast zone their actions are intensified by fluvial, glacial and mass movement processes. Destructive effect of waving is particularly evident in the case of cliff shores (Fig. 1). The example can be the Skilvika region where the cliff evolution is additionally predisposed by occurrence of Tertiary carbonaceous

* e-mail: pzagorsk@biotop.umcs.lublin.pl

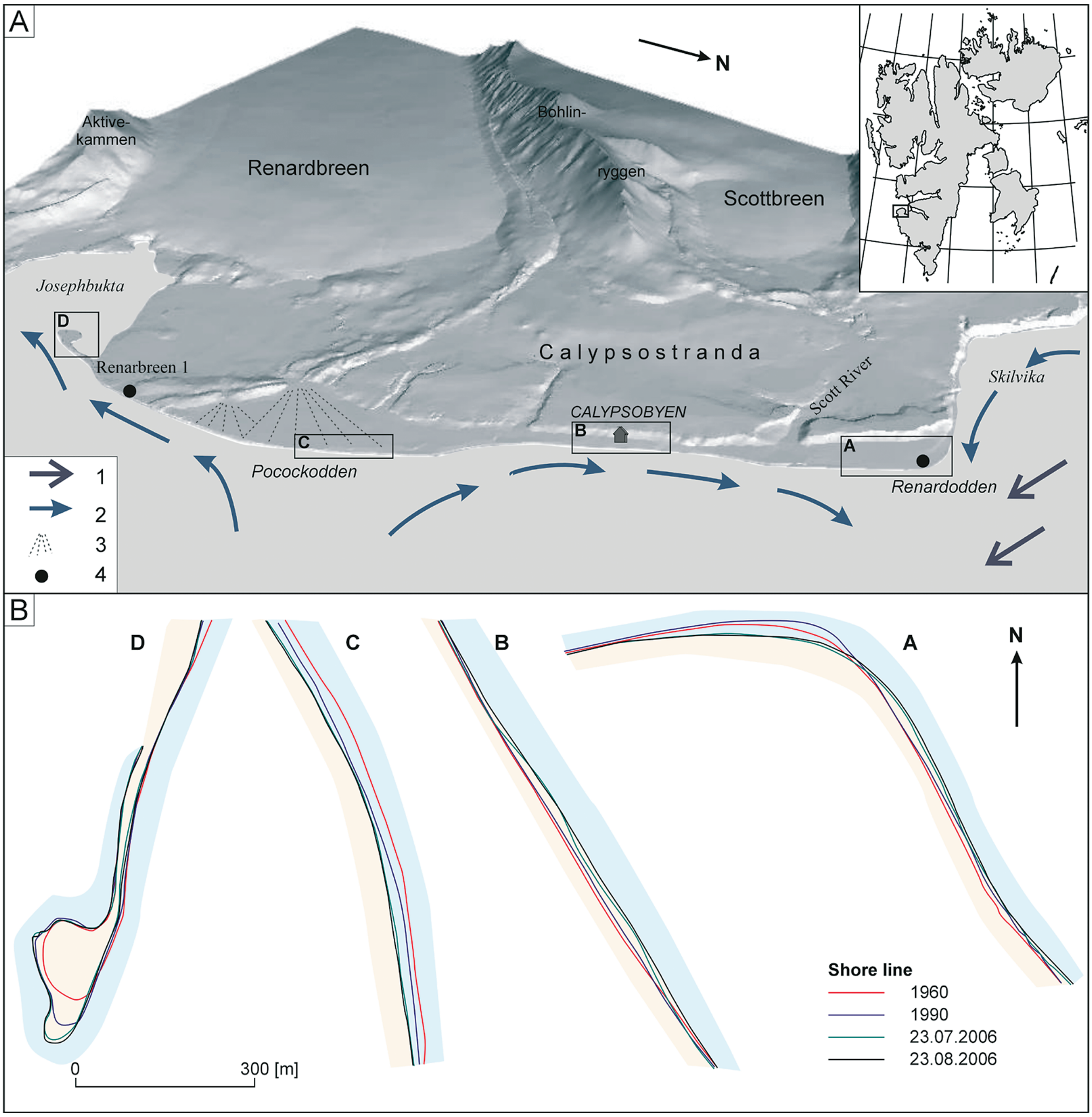

Fig. 1. A: 3D model of study area: 1 – oceanic drift (high wave energy), 2 – longshore current (after: Harasimiuk, Jezierski 1991), 3 – sandur cones, 4 – location of archaeological sites. B: Changes of shoreline

formation filling up the tectonic graben (Birkenmajer 2004). However, at the cliff foot there is formed an abrasive platform, cleared out of sediments by waving and broadening with cliff recession (Harasimiuk 1987, Zagórski 2004).

The delivery of material increased during the glacier recession (mainly: Scottbreen and Renardbreen) when the Little Ice Age was over and the zone of longshore currents convergence diminished the abrasion rate in the Calypsostranda region and made accumulation predominant e.g. in the vicinity of Renardodden. The archaeological date and geomorphological works carried out in this region indicate intensive evolution of Renardodden from the XVIIth century (Jasinski, Zagórski 1996). The closest to the present coastal zone (about 60 m from the coastal line) is the site Renardodden 1 which is the survival of the Russian station of walrus hunter from the first half of the XIXth century (Jasinski, Zavyalov 1995). Originally the hunter station building was out of the reach of storm waving. However, due to increase of abrasive processes activity, the old storm ridge was destroyed and the waves dragged pieces of brick and organic remains over the tidal zone (Fig. 1, 2). This condition was still maintained up to the beginning of the 60s i.e. when quick reces-

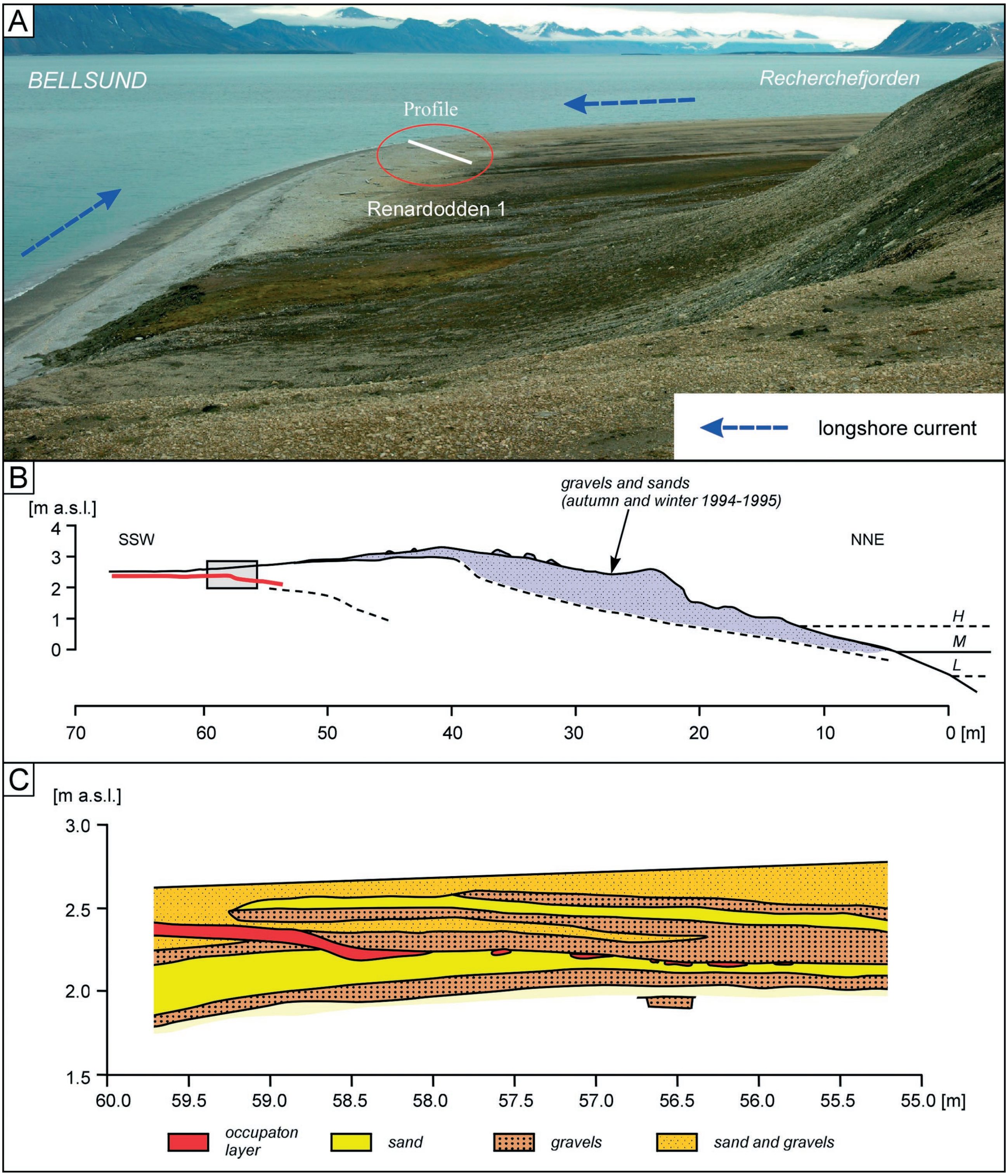

Fig. 2. Archaeological site Renardodden 1
A: General view, B: Geological profile across the storm ridge, C: Geological profile across the fragment of storm ridge with dragged occupation layer (after Jasinski, Zagórski 1996)

sion of the Scottbreen began (Reder 1996, Zagórski, Bartoszewski 2004). Till 1990 intensification of material supply resulted in the extension of the cape by almost 20 m (Fig. 1). However, lately there some developing changes of cape geometry has been observed due to poorer material supply from the marginal zone of the Scottbreen to the coastal zone and increasing role of marine processes (waving, longshore current). The part from Skilvika was largely sheared and that towards the Scott River estuary was aggradated (Harasimiuk 1987, Zagórski 2004).

From the works carried out in the 80s of the last century the area of accumulative coast situated between the Scott River and Pocockodden is considered to be relatively stable where the north-west coastal current is of saturated character and its whole energy makes dislocation of sediments along the shore (Harasimiuk 1987) (Fig. 1). Observations and measurements of coastal line changes performed using the GPS receiver during the polar expeditions in 2000, 2005 and 2006 under quiet meteorological conditions indicate gradual building up of a new gravelly ridge. However, strong storm conditions caused removal of the coast by a few meters and return to the previous state. Comparison of many years' observations of the coastal line was based on benchmark point and GPS measurements point to gradual change of coast geometry (Zagórski 2002, 2004). The autumn-winter storms of extreme sizes contribute largely to these changes (particularly in 1992/1993) (Rodzik, Wiktorowicz 1996, Zagórski 1996). Over ten metre removal of coastal line and addition of gravelly, gravelly-sandy and vegetable covers on the storm ridge area took place at that time in the vicinity of the station in Calypsobyen (Zagórski 1996). In successive years such rapid changes were not observed but gradual reconstruction of the devastated area took place.

At the beginning of the XXth century up to the sixties fluvial and fluvioglacial processes combined with marine ones had a significant effect on the coast shape. Their role was to deliver terrigenous material to the coastal zone (Harasimiuk, Król 1992). Such situation was observed, among others, in the case of vast sandur cones on the distal side of moraine ridges of Renardbreen (Fig. 1, 2). Slightly oblique area of semicircular shape was formed. At present due to Renardbreen recession and fluvioglacial supply disappearance they have become dead forms. Disappearance of land material supply resulted in increase of marine processes activity which, turn in, caused formation of a gravelly ridge inhibiting cone destruction (Harasimiuk 1987, Zagórski 2004). However, local longshore currents of which one flows towards north-west and the other south are of significant importance for evolution (Harasimiuk, Jezierski 1991). The latter supplied with the material from destruction of fluvioglacial cones of the Renardbreen affected the formation of sand spit bordering Josephbukta in the east and at present plays an important role in its transformation. Its evolution was also predisposed by occurrence of glacial sediments of Renardbreen marginal zone (Harasimiuk 1987, Zagórski 2004).

Occurrence of coastal ice is also an essential factor (for example: Jahn 1977, Jezierski 1992, Rodzik, Wiktorowicz 1996). Its quick accumulation provides effective protection of the shore against destructive activity of waving. Similarly, its long existence in the spring-summer season inhibits transformation of the coastal zone (Zagórski 1996, 2004).

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The geoecosystem of polar oases within the ice drainage basin of Admiralty Bay, King George Island, Antarctica

Zbigniew Zwoliński*

Institute of Paleogeography and Geoecology, Adam Mickiewicz University, Poznań, Poland

Introduction

The paper presents the results of a research into the geoecosystems of polar oases located on King George Island in the South Shetlands in the Maritime Antarctic. The recession of glaciers brings about an expansion of ice-free areas, landform metamorphoses, changes in land cover, redeposition of sedimentary covers, changes in the water cycle and mineral circulation, and finally a transformation of landscapes in the ice drainage basin of Admiralty Bay. The increase in ice-free areas observed in the 20th century is variable but quite rapid (Zwoliński 2007).

Corresponding to ice-free areas, the geoecosystems of polar oases have a relatively narrow range of geomorphic functioning since they are largely determined by the duration of snow and glacier covers and the accessibility and amount of solar energy reaching the Earth surface. The great significance of the geoecosystems of polar oases in the modern sub-Antarctic zone is due to the following:

– areas emerging from under ice, which are sites of a fast landform evolution, fast changes in the land cover and fast landscape transformations, and consequently of a rapid geosuccession due to the rebuilding of the internal structure of those geoecosystems,
– areas with distinct qualitative changes, mainly involving energy and matter,
– extensive, exposed areas susceptible to all morphogenetic factors such as sunshine, wind, water, snow, ice, gravitation and, increasingly, man,
– degraded periglacial landforms, highly sensitive to even the slightest climate change,
– manifestations of succession and biological colonisation, including communities of lichens, mosses and lower plants, and colonies of birds and pinnipeds as well as communities of the freshwater fauna,
– the potential space for settlement (at present mainly in the form of scientific stations and research sites), conducting of various types of economic activity (mainly mining), and a growing penetration by tourists.

Study area

The South Shetland Islands are located between two continents: Antarctica (the Antarctic Peninsula) in the south and South America (Tierra del Fuego) in the north, and between two ocean basins: the Pacific (the Bellingshausen Sea) in the west and the Atlantic (the Weddell Sea) in the east. This makes the waters surrounding the archipelago highly dynamic: they form a peculiar, internally mobile oceanographic junction which exerts a great influence on the dynamics of atmospheric and oceanic processes, and thus on the geomorphic processes within the terrestrial geoecosystems of polar oases.

In the drainage basin of Admiralty Bay within King George Island (Fig. 1), four morphogenetic microchores (types of terrain) were distinguished: 1. glacial, 2. postglacial and periglacial, 3. periglacial and 4. non-glacial, which belong to the category of paraglacial areas.

* e-mail: zbzw@amu.edu.pl

Table 1. Area dynamic of polar oases within drainage basin of Admiralty Bay in years 1979–1988

No.	Name of polar oasis	Area [km^2]					Rates [km^{2}a^{-1}]	
		1979		1988	Difference			
		(A)	(B)	(C)	(C-A)	(C-B)	for (C-A)	for (C-B)
1	Red Hill	0,38750	0,29251	0,43315	0,04565	0,14064	0,00457	0,01406
2	Patelnia (Telephone Point)	0,13750	0,22959	0,22880	0,09130	-0,00079	0,00913	-0,00008
3	Blue Dyke	0,27875	0,41434	0,46547	0,18672	0,05113	0,01867	0,00511
4	Demay Point	1,48125	1,64770	1,62575	0,14450	-0,02195	0,01445	-0,00220
5	Bastion	0,08530	0,12383	0,13538	0,05008	0,01154	0,00501	0,00115
6	The Tower	0,04560	0,07556	0,08373	0,03813	0,00817	0,00381	0,00082
7	Brama	0,20000	0,10817	0,12763	-0,07237	0,01946	-0,00724	0,00195
8	Siodło	0,04125	0,02989	0,04296	0,00171	0,01306	0,00017	0,00131
9	Zamek	0,13125	0,24717	0,37457	0,24332	0,12740	0,02433	0,01274
10	Sphinx Hill + Błaszyk Moraine	0,58000	0,67765	0,73803	0,15803	0,06038	0,01580	0,00604
11	Rescuers Hills	0,46875	0,52326	0,62168	0,15293	0,09842	0,01529	0,00984
12	Arctowski Oasis	4,19375	4,48843	4,69146	0,49771	0,20303	0,04977	0,02030
13	Breccia Crag	0,26000	0,21332	0,24321	-0,01679	0,02989	-0,00168	0,00299
14	Cytadela	1,19375	1,13419	1,21059	0,01684	0,07640	0,00168	0,00764
15	Belweder	0,17708	0,28272	0,29423	0,11715	0,01151	0,01171	0,00115
16	Scalpel Point	0,04560	0,04719	0,08168	0,03608	0,03450	0,00361	0,00345
17	Pond Hill	0,41000	0,49514	0,51996	0,10996	0,02482	0,01100	0,00248
18	Dufayel Island	0,44380	0,45625	0,47801	0,03421	0,02176	0,00342	0,00218
19	Nunataki Emerald Icefall		0,61951	0,72125	0,72125	0,10174	0,07213	0,01017
20	Klekowski Crag	0,26250	0,30622	0,32367	0,06117	0,01746	0,00612	0,00175
21	Admiralen Peak	0,02060	0,06887	0,05050	0,02990	-0,01837	0,00299	-0,00184
22	Komandor Peak	0,11000	0,13457	0,14745	0,03745	0,01289	0,00375	0,00129
23	Crepin Point	0,44125	0,70141	0,75251	0,31126	0,05109	0,03113	0,00511
24	Cockscomb Hill	0,02500	0,01701	0,01434	-0,01066	-0,00267	-0,00107	-0,00027
25	Garnuszewski Peak	0,02680		0,03297	0,00617	0,03297	0,00062	0,00330
26	Keller Peninsula	4,18750	2,67146	3,15485	-1,03265	0,48339	-0,10327	0,04834

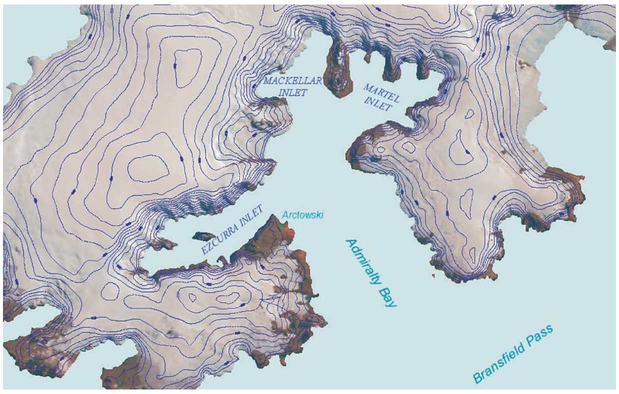

Fig. 1. Polar oases within drainage basin of Admiralty Bay on the background of ice caps' hypsometry (Project KGIS 2005, changed)

Overview of polar oases

The studies performed during a 4-year campaign corroborated the following regularities observed in polar areas (Zwoliński 2007):

– exceeding of the hitherto absolute maxima of air temperature at the Arctowski Station,
– an increase in annual precipitation totals at the Arctowski Station, first of all in the form of rain, also during the cold period,
– cold periods on King George Island becoming shorter and climatically milder,

No.	Name of polar oasis	Area [km^2]					Rates [km^{2}a^{-1}]	
		1979		1988	Difference			
		(A)	(B)	(C)	(C-A)	(C-B)	for (C-A)	for (C-B)
27	Shark Finn	0,01875	0,01830	0,04520	0,02645	0,02690	0,00264	0,00269
28	Stenhouse Bluff	0,08312	0,11265	0,09337	0,01025	-0,01928	0,00102	-0,00193
29	Ullman Spur	1,30375	1,22184	1,28726	-0,01649	0,06541	-0,00165	0,00654
30	Precious Peaks	0,63125	0,75084	0,74226	0,11101	-0,00858	0,01110	-0,00086
31	Ternyck Needle		0,03045	0,04183	0,04183	0,01138	0,00418	0,00114
32	Szafer Ridge	0,33250	0,31961	0,36199	0,02949	0,04238	0,00295	0,00424
33	Tern Nunatak	0,02280	0,01751	0,02194	-0,00086	0,00443	-0,00009	0,00044
34	Warkocz	0,20000	0,28403	0,30464	0,10464	0,02061	0,01046	0,00206
35	Smok Hill	0,50180	0,46810	0,44932	-0,05248	-0,01878	-0,00525	-0,00188
36	Mount Wawel (Hennequin Pt)	1,29750	1,02270	1,13405	-0,16345	0,11134	-0,01635	0,01113
37	Bell Zygmunt		0,05262	0,12136	0,12136	0,06875	0,01214	0,00687
38	Manczarski Point		0,06905		0,00000	-0,06905	0,00000	-0,00691
39	Rembiszewski Nunataks		0,03556	0,03658	0,03658	0,00102	0,00366	0,00010
40	Puchalski Peak		0,02056	0,02568	0,02568	0,00513	0,00257	0,00051
41	Vauréal Peak	0,20190	0,25391	0,28158	0,07968	0,02767	0,00797	0,00277
42	Harnasie	0,36620	0,51857	0,38814	0,02194	-0,13043	0,00219	-0,01304
43	Czajkowski Needle		0,03283	0,02430				
44	Northern Sphinx Hill		0,01697	0,10624				
45	Table Hill		0,00611	0,00984				
46	Three Musketers		0,02211	0,03496				
47	Krak Glacier		0,03726	0,07878				
48	Chabrier Rock		0,05548	0,06477				
49	Polar Club Glacier			0,02885				
50	Northern Blue Dyke			0,01844				
51	Baranowski Glacier			0,05901				
52	Ecology Glacier			0,06247				
53	Doctors Icefall			0,02799				
54	Dobrowolski Glacier			0,02145				
	Area of oases:	20,59435	21,37302	23,46612	Average rates:		0,00556	0,00411

- transitional periods in the South Shetland Islands becoming longer: spring comes earlier and autumn ends later,
- a decrease in the thickness, duration and spatial range of the sea-ice cover in Admiralty Bay,
- an increase in the temperature and a decrease in the salinity and density of the bay waters,
- an increase in the number of ice-floes and growlers from glaciers in the Admiralty Bay catchment,
- intense ablation and rapid recession of the majority of glaciers in the vicinity of Admiralty Bay,
- a decrease in the area of nival covers in the Admiralty Bay drainage basin,
- intensive thawing of the multi-year permafrost in the Admiralty Bay drainage basin,
- changes in the land water cycle manifested by an increase in the surface runoff in streams and a shortening of the freezing period for streams and lakes in the Admiralty Bay drainage basin,

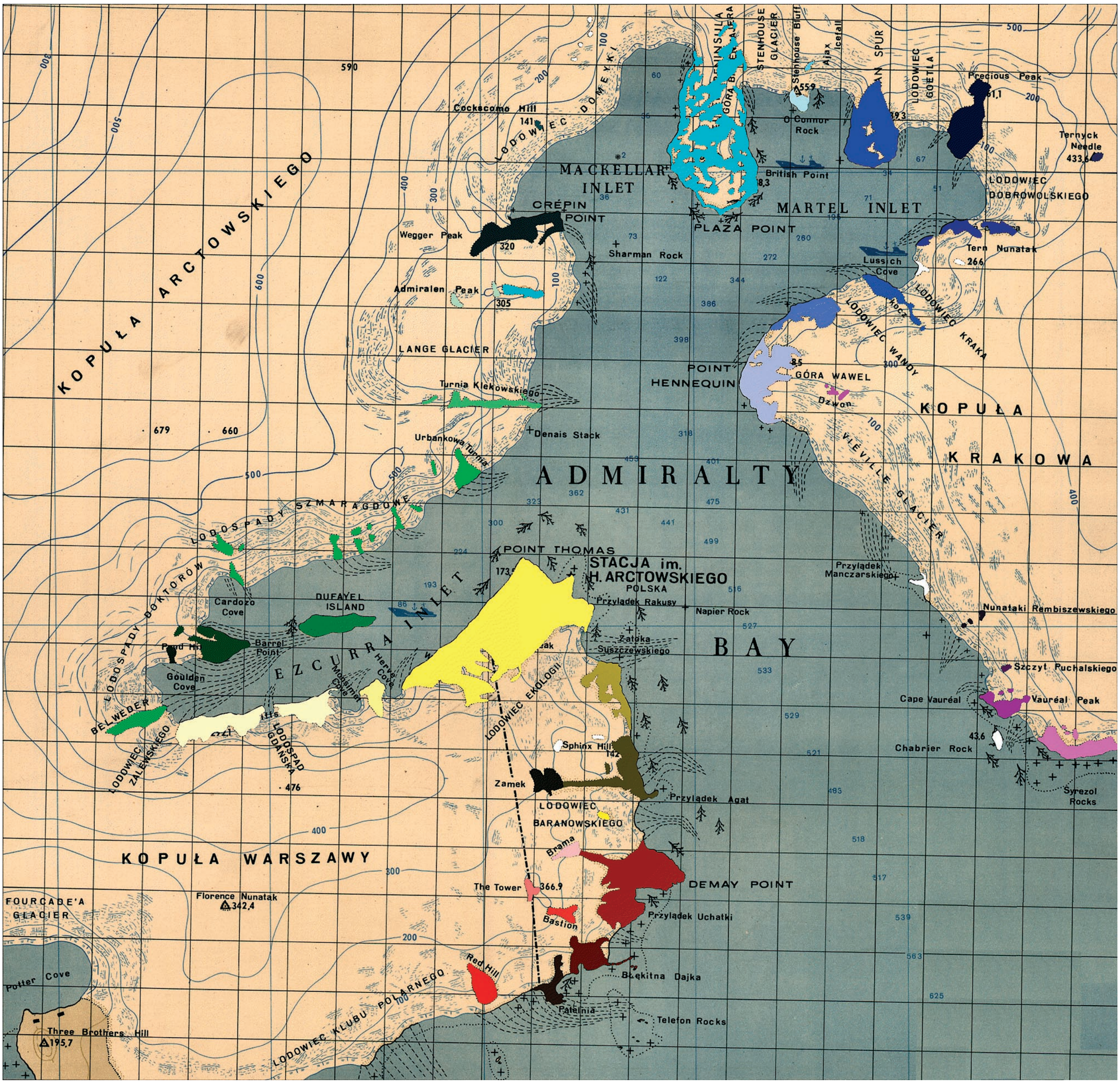

Fig. 2. Spatial pattern of polar oases on the topographic map „Zatoka Admiralicji" (ZBP IE PAN) in scale 1:50 000 according to data from January 1979

- a growing area of surfaces saturated with meltwater from degraded buried ice-cores in marginal zones of glaciers in the vicinity of the Admiralty Bay, and
- a shift of geoecological zones in the Admiralty Bay drainage basin.

The effect of the above set of causal and resultant tendencies is that an ever-growing amount of mineral matter (rock waste, morainic and diamicton covers) is transferred within areas newly exposed from beneath glaciers and ice caps. As a consequence, there is a rapid change in the initial topographic surface, the inclusion of mineral matter into the transport, a transformation of various types of cover deposits, and the formation of new landscapes, mainly postglacial/non-glacial hybrids. This makes landscape transformations a common phenomenon in the paraglacial zones.

Spatial pattern of polar oases

In the distribution of oases in the surroundings of Admiralty Bay (Fig. 2 and 3) several regularities can be observed (Zwoliński 2002, 2007):

- most of the oases are located in maritime situations in which at least one of their sides meets the bay waters; only a few oases or nunataks occur in places surrounded on all sides by glacier ice;
- the greatest number of oases occur on north-facing slopes, e.g. the southern shores of the Ezcurra

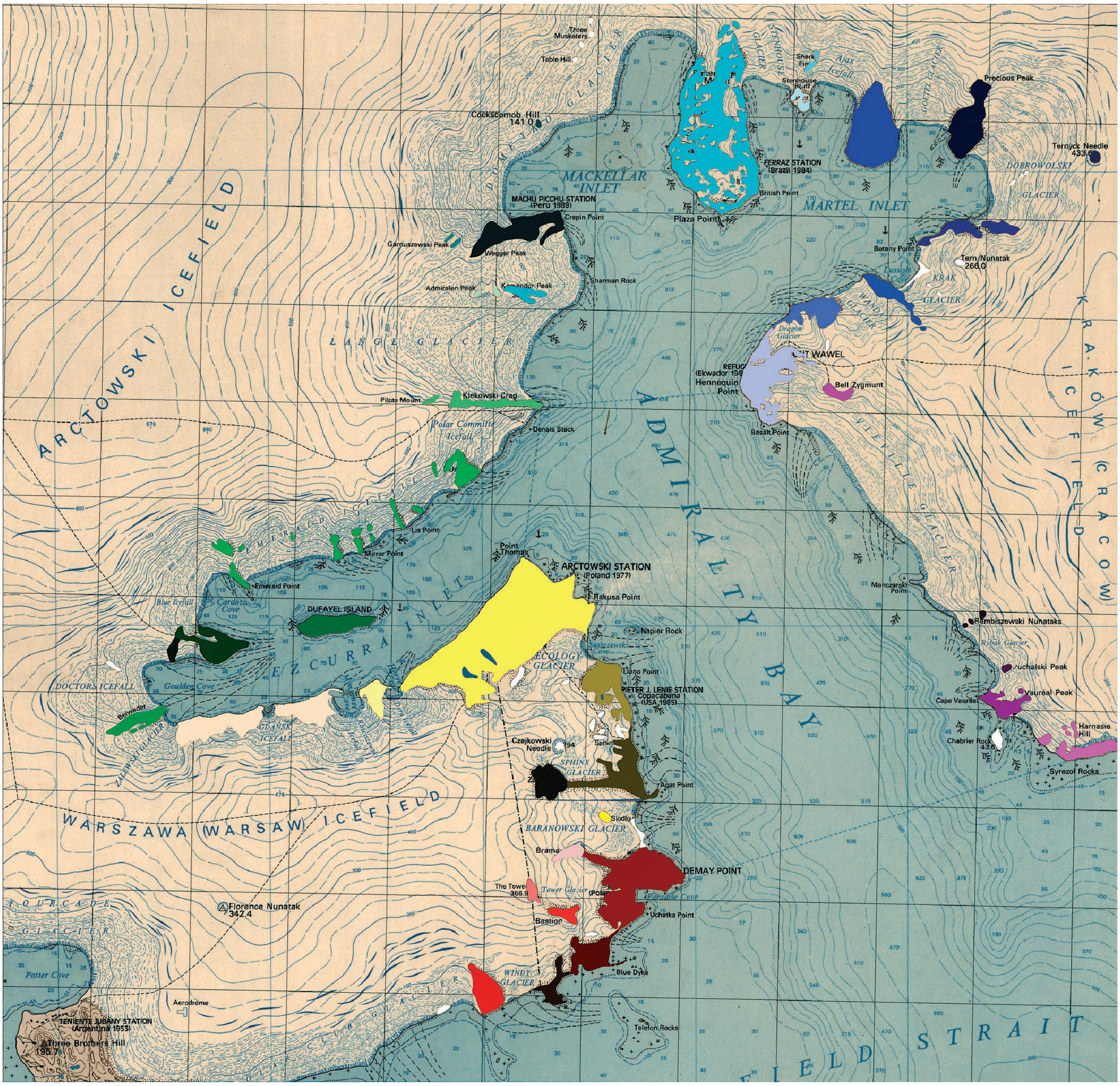

Fig. 3. Spatial pattern of polar oases on the topographic map „Admiralty Bay" (Battke 1990) in scale 1:50 000 according to data from 1988

and Martel Inlets; this is connected with the prevailing direction of incoming solar radiation irrespective of the season of the year;

- on south-facing slopes the oases are sporadic and small because solar radiation either does not reach them at all or its access is limited, e.g. the northern shores of the Ezcurra and MacKellar Inlets;
- generally, on the eastern shore of Admiralty Bay there are no oases due to intensive glacier alimentation from the Cracow Cap; and
- on the western shore numerous small oases have developed as an effect of the dying out of the Warsaw Cap, which is not directly supplied from the centrally located Arctowski Cap.

The above distribution of oases clearly indicates that it is dependent on the climatic and topoclimatic conditions, primarily on the exposure and ice balance of the icefields. Superimposed on these characteristic features of oases throughout the whole area is its geological-morphological aspect, i.e. the occurrence on the bedrock of rock ranges and massifs resistant to denudation processes and hence being natural places for oases to develop. The spatial pattern of oases in the Admiralty Bay drainage basin can be considered a regularity in the polar sub-Antarctic zone because similar locations can also be observed in the whole of King George Island as well as the other islands of the South Shetland archipelago.

Matter flux in polar oases

From the point of view of the mobility of mineral matter, the following properties of the geoecosystems of polar oases in sub-Antarctic islands should be emphasised:

- the occurrence of a debris, morainic or waste cover under which glacier ice may be buried,
- frequent changes in the morphological surface, including hypsometric changes,
- a climate distinct from that of the surroundings,
- air and ground temperatures above 0°C, at least at the height of the warm period,
- the occurrence of water flow in the systems of streams and lakes which thaw during summer, but also the occurrence of water as ground moisture,
- the formation of initial soil covers, and
- a gradual appearance of conditions for the succession and development of biological life.

The system of mineral matter cycle within modern terrestrial sub-Antarctic geoecosystems is third in importance in the Admiralty Bay drainage basin after the climatic and hydrological systems. The boundaries of the matter cycle system are fuzzy, zonal and permeable in any direction and through any medium. This makes it susceptible to a dynamic exchange of energy and matter with the adjacent systems. That is why the thickness of a 3–dimensional solid representing the matter cycle varies spatially and in the case of the Admiralty Bay drainage basin depends on at least three factors:

- the altitude of the bedrock, which is a resultant of the geological history of the Scotia Arc and the previous and modern history of volcanism, tectonic and isostatic movements,
- the thickness of the ice caps, which is the effect of the snow-ice mass balance, and hence of the alternate advances and recessions of glaciers,
- the occurrence of coastal and inland oases and their areal expansion.

Concluding remarks

The glacier recession, expansion of ice-free areas, and paraglacial transformations of the land cover, water cycle, mineral matter circulation, sedimentary covers, landforms, and finally of the landscapes in the Admiralty Bay drainage basin, became an inspiration for the formulation and development of the conception of geosuccession and its justification in theoretical and application terms. The result of a geosuccession is an overlap of qualitative and quantitative changes in the weathering, denudation, transport and deposition processes occurring at any spatial-temporal scale and leading to transformations in the style of functioning of morphogenetic domains. The change in morphogenetic domains expresses changes in the nature of the relief and its forms, and as a consequence, changes in the landscape caused by the interchangeability of dominant and secondary processes. The sensitivity of these domains is evidence of an enrichment or impoverishment of the geodiversity of the given polar area.

Among the key features of the presently forming paraglacial areas within the geoecosystems of polar oases are high-energy, fast morphological and depositional changes as well as the freshness and youth of landforms and sedimentary covers, hence great geomorphic dynamics. The modern environments of polar oases are highly tender geoecosystems sensitive to changes in and variability of geographical conditions, from the local through regional to the global scale. Therefore features of areas in the sub-Antarctic zone may be treated as very precise geoindicators and bioindicators of present-day environmental changes, showing the direction and rate of transformations in the abiotic (Table 1) and biotic environments.

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The geoecological model for small tundra lakes, Spitsbergen

Zbigniew Zwoliński*, Grzegorz Rachlewicz, Małgorzata Mazurek, Renata Paluszkiewicz
Institute of Paleogeography and Geoecology, Adam Mickiewicz University, Poznań, Poland

Introduction

Arctic wetlands are an important geoecosystem in an arid and cold tundra environment of polar oasis on the high-latitude of northern hemisphere (Kostrzewski et al. 2006a), which over the past few years have been subjected to significant environmental changes. Global climate changes have contributed to the significant development of wetlands in areas of glacial retreat and where the deeper summer thawing of permafrost has taken place. However, at the same time the patchy wetlands with a small area tend to disappear as a result of the gradual degradation of the permafrost (Woo & Young 2006, Kostrzewski et al. 2006b). High-arctic lakes are prevalent and highly sensitive types of polar wetlands affected by climate change and variations of geomorphic processes (Yoshikawa, Hinzman, 2003).

The remote location of wetlands in the High Arctic results in the fact that data presenting the current state of the geoecosystems of tundra lakes, supporting an analysis of the stages of evolution and forecasting of their future development is sparse. On the eastern coast of Petuniabukta there is a group of tundra lakes at various stages of geosuccession. Detailed studies of shallow tundra lakes located on raised marine terraces at Ebbadalen valley (Fig. 1, Mazurek et al. in print, Zwoliński et al. in print) discovered the influence of water that feeds into the lakes whose type depends on bedrocks, the intensity of biogenic processes and distance from the sea (Stankowska 1989).

The objective of this paper is to present the conditions and style of functioning of the geoecosystems of tundra lakes. An analysis of their functioning has been based on the diversification and variability of water chemistry and the sedimentological and geochemical properties of lacustrine deposits, which fill the basins of a number of lakes on the eastern coast of Petuniabukta. Hydrographical and lithological mapping covered 16 tundra lakes, and included the research of spatial and temporal variability (in the period 2001–2003 and in 2005–2006) of the chemical composition of water in selected lakes (Fig. 2). This analysis should point to the role of these basins in the paraglacial evolution of mineral matter in a post-glacial valley. In consequence, a model presentation of the development of arctic lake-like wetlands in the dry high-Arctic climate should be obtained.

Study area

The surveyed area is located on the north-eastern tip of Billefjorden, a branch of Isfjorden – the biggest fjord-system in the mid-western coast of Central Spitsbergen (Fig. 3). The eastern coast of Petuniabukta at the mouth of Ebbadalen (Ebba valley) forms a steplike system of raised marine terraces characteristic for the post-Pleistocene shoreline layout of Svalbard (Salvigsen, 1984; Landvik et al., 1987). Geologically, the eastern coast of Petuniabukta is dominated by the Ebbadalen Formation composed of Middle and Upper Carboniferous carbonates, anhydrites, gypsum, sandstone and conglomerates (Dallmann et al., 1994). The lower terraces comprise sandy and gravel sediments originating from local waste material and, to a smaller extent, in cryctallinic Hecla-Hoek rocks. The Ebbadalen tundra lakes occupy shallow depressions un-

* e-mail: zbzw@amu.edu.pl

Fig. 1. General view on raised marine terraces on the east coast of Petunibukta (photo Zb. Zwoliński)

derlain by initial soil or thin peat in permafrost terrain (Zwoliński et al., in prep.). According to Gulińska et al. (2003), the soil conditions on raised marine terraces suggest that they belong to the area with loose and poorly developed soils. Rare vegetation of dry deflational lichen-moss tundra ecosystems is comprised mainly of calciophylitic *Dryas octopetala*, *Salix polaris* and various species of *Saxifraga*, all partly interspersed with lichens.

Fig. 2. The closest lake to Skottehytta with water level gauge and measuremets of water temperature (comp. Fig. 4) (photo Zb. Zwoliński)

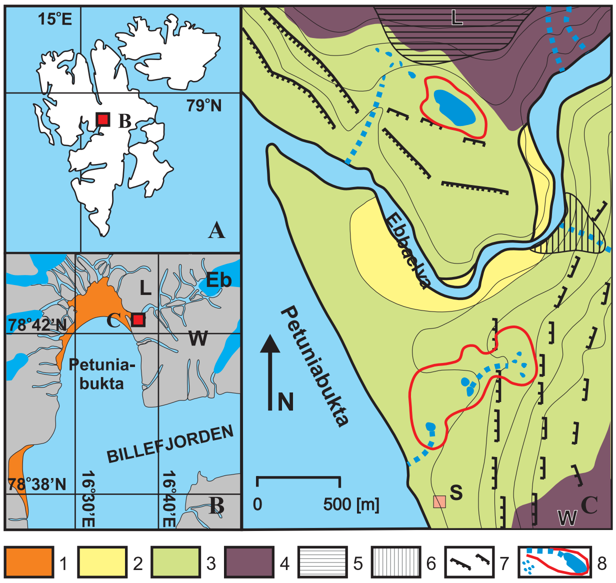

Fig. 3. Location of study area

1 – tidal flats; 2 – areas of glacifluvial accumulation; 3 – raised marine features; 4 – rock walls and slope deposits (undifferentiated); 5 – talus cones; 6 – alluvial fans; 7 – raised terrace level edges; 8 – lakes, episodic streams and borders of catchments under study (red lines); L – Lovehovden; W – Wordiekammen; Eb – Ebbabreen; S – Skottehytta. On the insert C – contour intervals marked with 5 m interval.

The functioning of tundra lake geoecosystem

The hydrological activity of non-glaciated catchments on the sea coast usually appear no longer than for 3–4 months in a year, from June to September (Fig. 4). Short period of functioning of tundra lakes induces that during summer they are one of the most dynamically changing morphogenetic and sedimentary environments, being at that time the highly valuable habitat for flora and fauna (Fig. 5).

During winter, the circulation of water in lake basins ceases, and is replaced by the accumulation of water in snow and permafrost. Shallow lakes and ponds can freeze to the bottom over winter. During warm season, the lakes are supplied initially by meltwater from the lake ice and snow cover melt, and then by the hillslope runoff and saturated overland flow from wetlands and creeks, and/or the seasonal thawing of permafrost. Processes of runoff trigger the inwash of fine particles, resulting in their seasonal decantation. Nevertheless the most efficient supply of sediments to ponds is connected with aeolian activity. Dust and fine-sand particles are delivered by wind either directly to the water, or deposited on the snow surface, mainly during early autumn, when the snow cover is not continuous, and pond depressions are privileged regarding its accumulation.

During the Arctic summer, the spread of tundra ponds changes, some of them disappear periodically or permanently, which results from the increase in air temperature and evaporation, and points among others to the gradual exhaustion of supply from the active layer of the permafrost. The magnitude of evaporation is also definitely influenced by the vegetation occurring on the margins of basins. The lakes usually froze slowly in early or mid-September.

The sources of water inflows reaching tundra lakes are reflected in the chemical composition of water which, in turn, affects the types of ecosystems, the geodiversity of bedrock, waste covers, soils and the biodiversity of the indigenous flora. The high total solute load and hydrochemistry in waters indicate the share of ions coming from weathering processes and supplied by water from the thawing of permafrost and circulation in the active layer (Mazurek et al. in print). Crustally derived ions are released from the permafrost and carried to basins in the active layer by suprapermafrost groundwater. Depending on the geological structure and the diverse weathering processes in the catchment area of lakes, the solutions fed into tundra lakes, with surface and ground water, may have varying contents of crustal components. Hydrochemical mapping showed that sulphates, which are connected with anhydrites and gypsum rocks, are a good geoindicator of terrestrial ions provenance. In quantitative terms, of lower importance for the chemical composition is the marine aerosol (Na^+, Cl^-) and atmospheric (Na^+, Cl^-, SO_4^{2-})

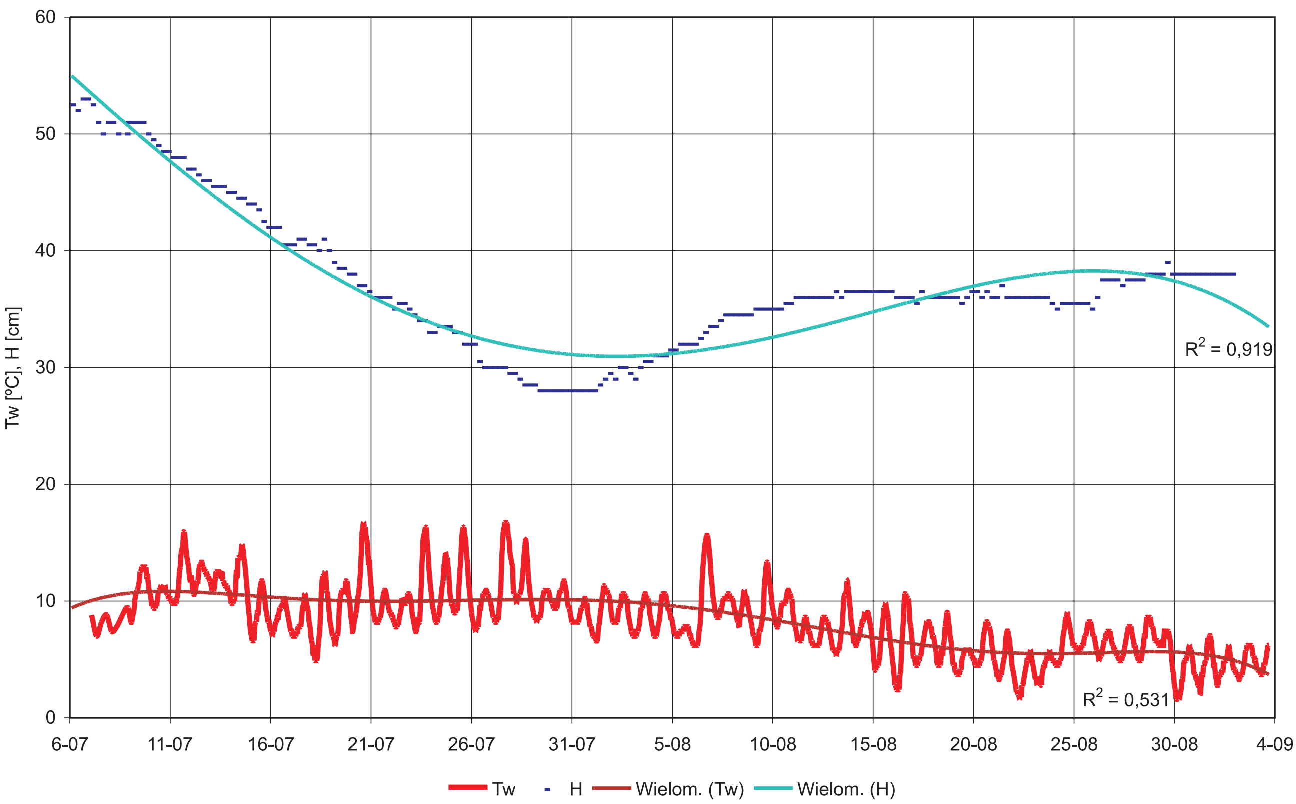

Fig. 4. Course of water level (H, four times per day) and water temperature (Tw, hourly intervals) in lake near Skottehytta during summer season 2006 (comp. Fig. 2)

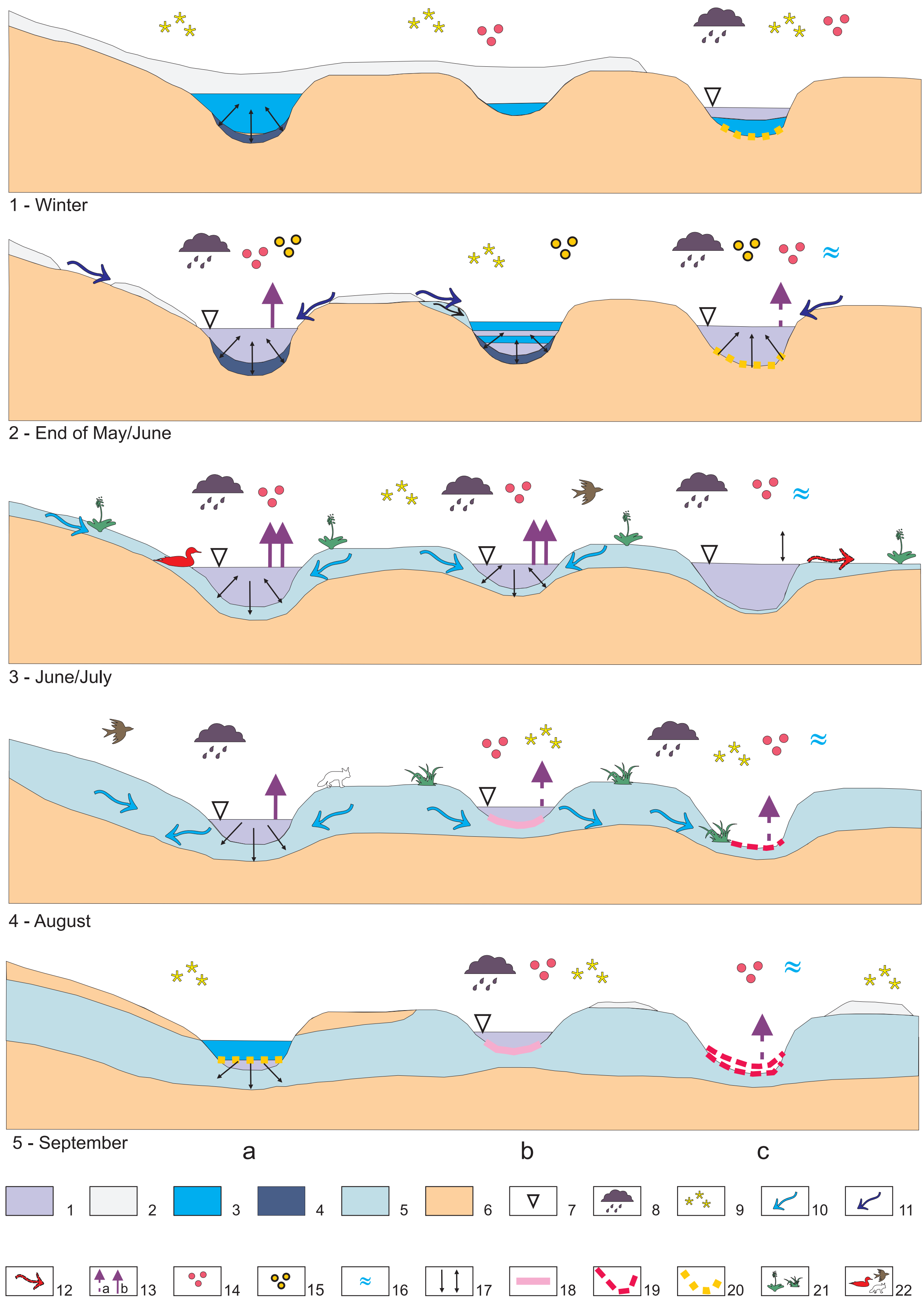

Fig. 5. Conceptualisation (1–5) and possible options (a-c) of input, transformation and output in context of matter flux within geoecosystems of tundra lake-like wetlands on raised marine terraces on Spitsbergen

1 – water, 2 – snow cover, 3 – ice, 4 – talik, 5 – active layer, 6 – frozen ground, 7 – water mark, 8 – rainfall, 9 – snowfall, 10 – subsurface runoff, 11 – surface runoff, 12 – episodic outflow, 13 – evaporation: a – low, b – high, 14 – wet and dry deposition, 15 – aeolian fallout, 16 – marine aerosol, 17 – matter exchange, 18 – mineral and organic sediments, 19 – precipitation of $CaCO_3$ and/or $CaSO_4$, 20 – cryochemical precipitation, 21 – tundra vegetation, 22 – animals.

supply, however these sources may modify the chemistry of waters in shallow terrace basins.

Saturated areas around lakes are often zones of richer tundra vegetation made up of various species of lichens, liverworts, moss, grass and herbs which are a source of components of biogenic origin. a factor that contributes to the eutrophication of soil and the resulting accelerated plant growth is the presence of birds around the surveyed lakes. Despite its slow rate, the decomposition of organic matter in the tundra ecosystem releases biogenic compounds, including sulphates, potassium and nitrates from waters feeding into lakes, in particular at the end of the polar summer.

The concentrations of most chemical components in the waters of the surveyed lakes display a slow increase from the beginning of the polar summer to the peak of the summer season. As the temperature rise, the water evaporates and the water table in the lakes is lowering, contributing to a gradual saturation of components. The soluble cations sequestered in near-surface permafrost comprise an important pool of solutes that may be released into the active layer during periods of deeper thaw (Kokelj, Burn 2003; Kokelj et al. 2005). As the permafrost table continues to degrade, the specific conductivity of lake water increases during the polar summer as did the share of calcium, magnesium, bicarbonate ions and sulphates. The highest specific conductivity values in water were observed at the time of decreases in air temperature and in the formation of the ice cover. The total loss of water in the basin results in a situation where the surface of lake deposits contains a large quantity of easily dissoluble constituents. During the precipitation period, these compounds may be washed inside deposits or may enrich waters in the succeeding phase of functioning of ponds.

The increase of air temperature leads to the melting of the first autumn ice cover and the relapse to the lower conductivity of water comparable with values preceding the formation of the ice cover. If the temperature falls below 0°C causing the lake freezing, the concentration of salt in the remaining bottom water strata increases due to cryochemical processes (Pulina 1990; Healy et al. 2006; Zwoliński 2007). The cryochemical processes lead to precipitation of calcite and gypsum crystals.

The sedimentation of autochthonous and allochthonous deposits, as well as the partial crystallisation of salts in lake basins is of particular significance for the creation of a present-day climax relief, aimed at levelling the topographical surface. The formation of such surface changes the hydrogeological conditions and reorganises the water runoff system from raised marine terraces. That occurs in spite of the fact that lakes are characterised by a relatively low dynamic and persistent morphogenetic and sedimentary systems.

Final remarks

Wetland geoecosystems on raised marine terraces operate only from June to August and occasionally to September each year. Factors such as the hydrological regime and water chemistry determine the development of Arctic wetlands on Petunia Bay coast in the dry climate zone of the High Arctic. Tundra pond levels usually fluctuate diurnally, seasonally and annually due to evaporation, groundwater recharge or seepage losses. They may evoke water level drawdowns which will result in partial drying and exposing the sediments. They may remain dry for several consecutive years, and consequently assume some characteristics of terrestrial geoecosystems until water levels are restored by above-average precipitation and runoff.

The gradual disappearance of lakes is pointed out by the morphological and sedimentological traces of their earlier extent. The bottoms of lakes, which are exposed to subaerial conditions, start to undergo other weathering processes and, with the participation of advancing tundra vegetation, develop under the influence of gradual pedogenic processes.

The chemistry of lake water on raised marine terraces of the eastern coast of Petuniabukta is influenced by the size of the surface catchment area, the type of substrate sediments (which include carbonate rocks and sediments) and the intensity of hydro-bio-geo-chemical processes occurring in the topsoil. Differences in the mineralisation and chemical composition of tundra lake water in catchments, with the significant thickness of permafrost active layer during the polar summer, result mainly from the supply of meltwater, from its degradation and from the thawing of snowcover. Due to low total of summer rainfall, the liquid precipitation in Arctic wetlands plays only a limited role.

The coastal wetlands of the Central Spitsbergen include, among others tundra lakes located in the zone of isostatically elevated coasts. a highly distinctive feature of the lakes is that they occur exclusively on the raised marine terraces. Due to their transitory location, the lakes are affected by the litoral and paraglacial geoecosystem and, on occasion, by the proglacial and glacial geoecosystem. Zwoliński (2002; 2007) classifies longer and shorter-term residence of minerals in the lakes as a part of the redeposition cascade found in the matter circulation system of the polar oasis. During its cycle, mineral matter undergoes various hydro-bio-geo-chemical transformations in solutions and diagenesis of muddy bottom sediments. The seasonal nature and the related hydrochemical dynamics of tundra lake water provide evidence of their role in the paraglacial evolution of mineral matter. Should be noted that in ecological terms, lake-like wetlands are the most lively systems from amongst geoecosystems distinguished by Kostrzewski et al. (2006a) in Ebbadalen.

Acknowledgements

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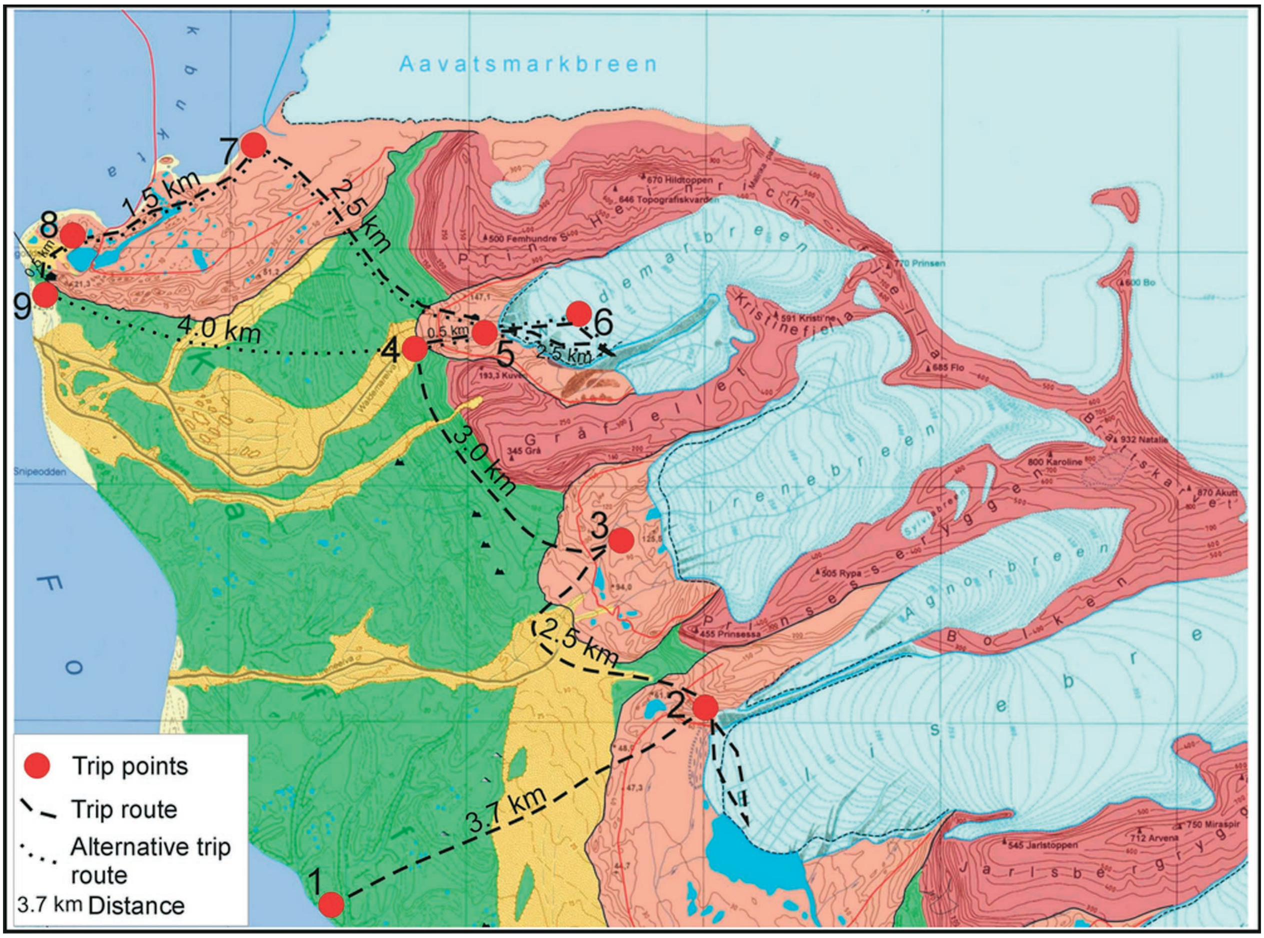
Glaciology, hydrology and geomorphology in the Kaffiøyra region – Leader Ireneusz Sobota



Point 1 – KAFFIØYRA 78° 37' 45" N 11° 56' 59" E
Marek Grześ, Ireneusz Sobota – Geography of Kaffiøyra

Point 2 – ELISEBREEN 78° 38' 44" N 12° 05' 38" E
Ireneusz Sobota – Summer balance of Elisebreen

Point 3 – IRENEBREEN 78° 39' 24" N 12° 03' 26" E
Ireneusz Sobota – Summer balance of Irenebreen

Point 4 – WALDEMAR RIVER 78° 40' 23" N 11° 58' 10" E
Point 4a – Marek Grześ – Naledi of the Kaffiøyra
Point 4b – Ireneusz Sobota – Discharge of Waldemar River and outflow from glacier
Point4c – Marek Grześ – Thickness of mineral covers on the ice-cored moraine and an active layer of permafrost on the western coast of the Oscar II Land (Svalbard)

Point 5 – WALDEMARBREEN 78° 40' 30" N 11° 59' 28" E
Point 5a – Ireneusz Sobota – Mass balance monitoring of Kaffiyra glaciers
Point 5b – Ireneusz Sobota – Summer balance of Waldemarbreen
Point 5c – Ireneusz Sobota, Marek Grześ – Snow accumulation of Kaffiøyra glaciers

Point 6 – ICE DRILLING 78° 40' 31" N 12° 01' 50" E
Erich Heucke, Ireneusz Sobota – The sample ice drilling on Waldemarbreen with Heucke Ice Drill – the demostration

Point 7 – AAVATSMARKBREEN 78° 41' 07" N 11° 53' 39" E
Marek Grześ, Michał Król, Ireneusz Sobota – Subaqual recordings of the changes in the range of glaciers in the Forlandsundet region (NW Spitsbergen)

Point 8 – GÓRNE LAKE 78° 40' 42" N 11° 50' 19" E
Ireneusz Sobota – Selected problems of changes in morphometry, bathymetry and thermal conditions in the lake complex at the forefield of Aavatsmarkbreen

Point 9 – NICOLAUS COPERNICUS POLAR STATION 78° 40' 33" N 11° 49' 36" E
Marek Grześ, Ireneusz Sobota – www.stacja.arktyka.com

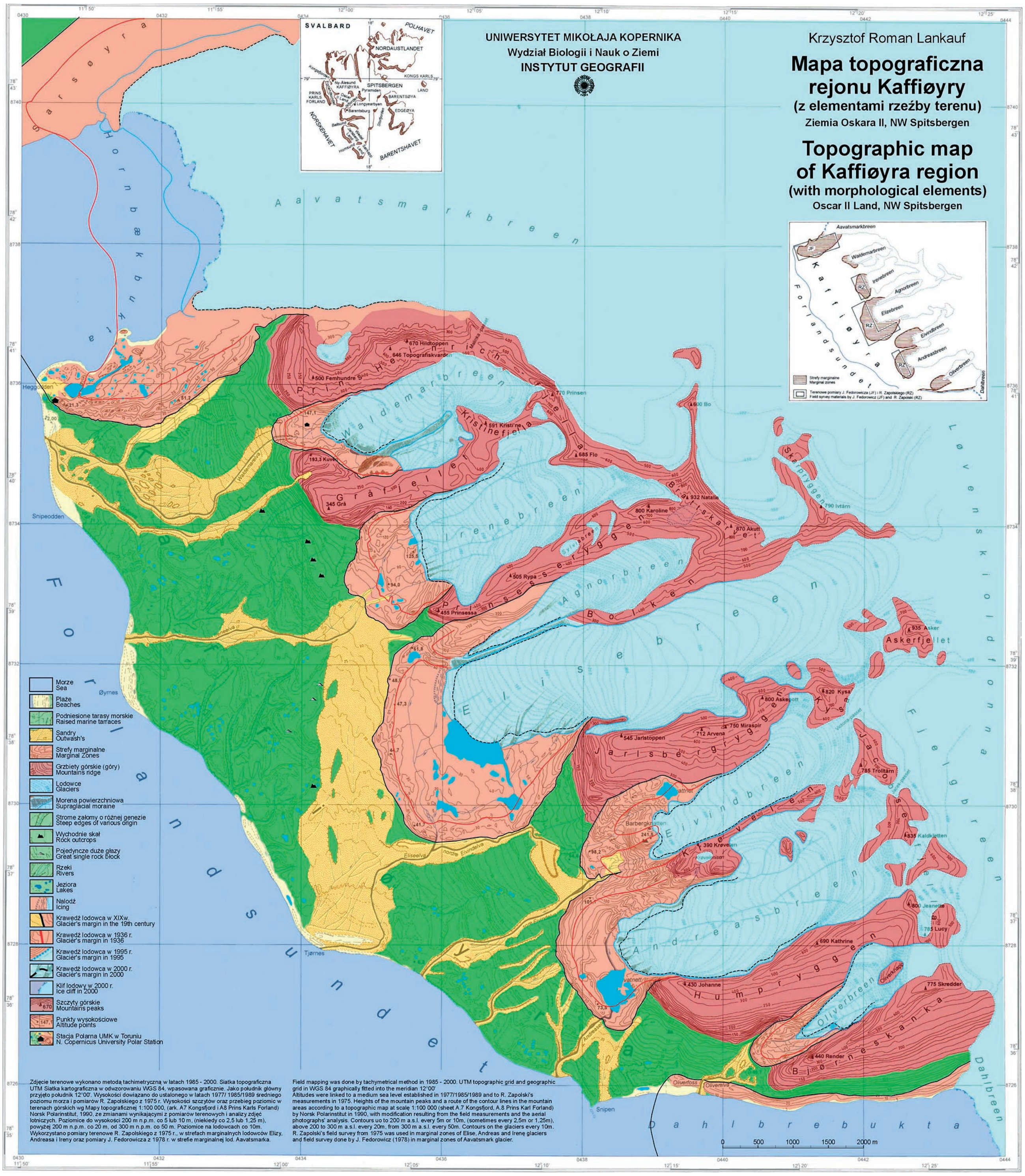


Fig. 1. Map of Kaffiøyra – K.R. Lankauf

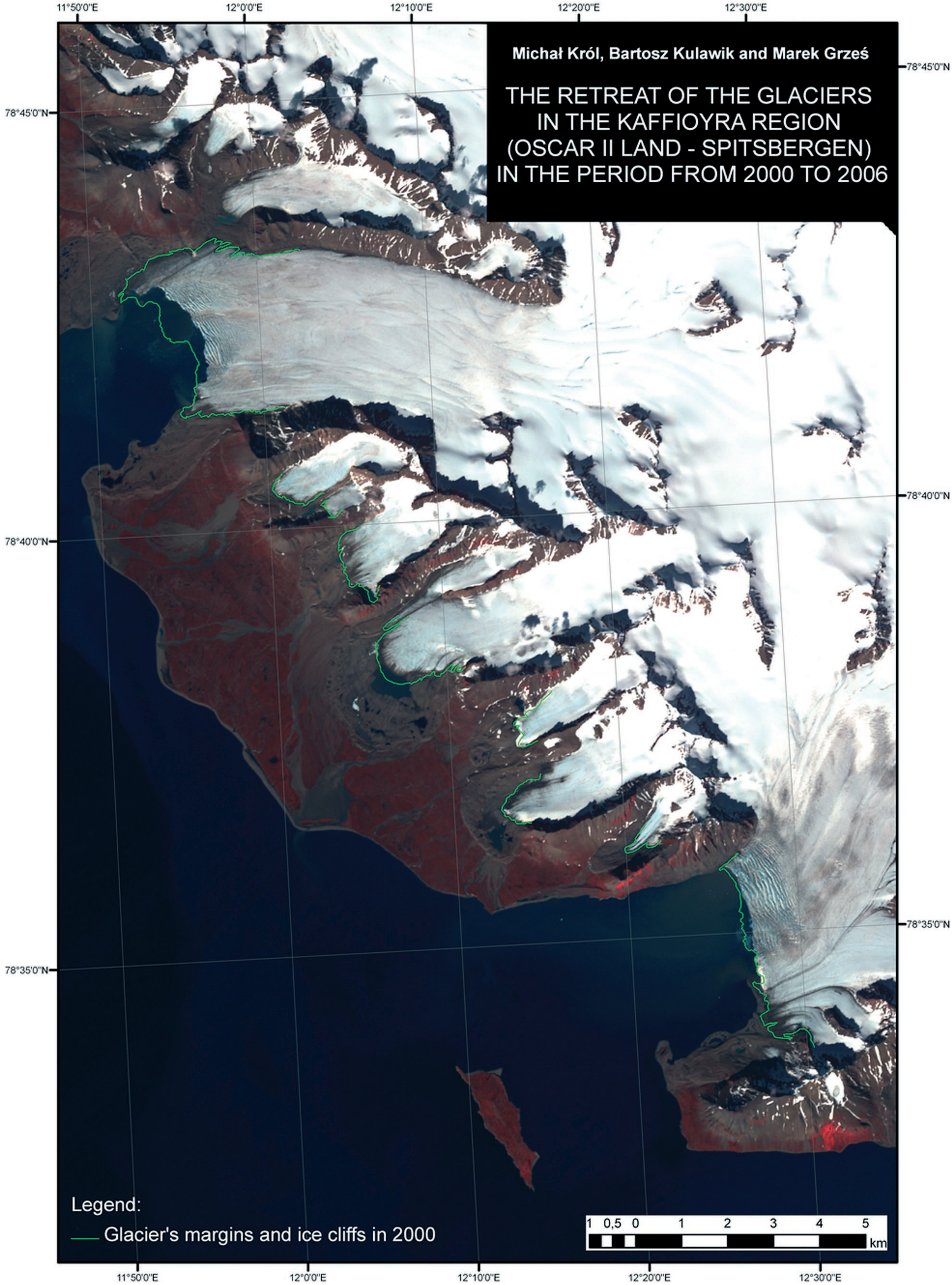


Fig. 2. The retreat of the glaciers in the Kaffiøyra region (Oscar II Land – Spitsbergen) in the period from 2000 to 2006 – M. Król, B. Kulawik and M. Grześ

Point 2 – Elisebreen *78° 38’ 44” N 12° 05’ 38” E*

Summer balance of Elisebreen

Ireneusz Sobota*

Department of Cryology and Polar Research, Institute of Geography, N. Copernicus University, Toruń, Poland

The studies of ablation of Kaffiøyra glaciers refer to Waldemarbreen, Irenebreen and Elisebreen. In 2006 the studies of ablation of Elisebreen began.

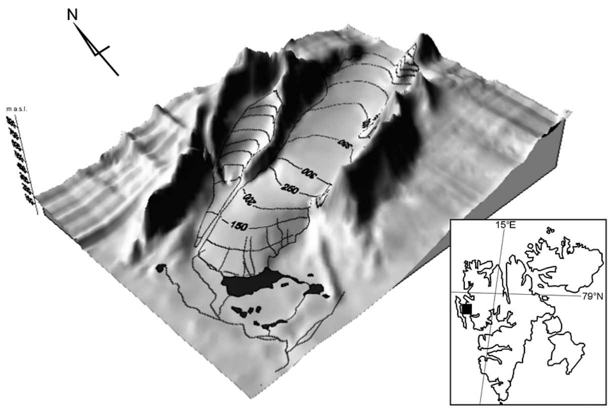


Fig. 1. Topographical draft of Elisebreen

These researches are continued (Sobota, 2004, 2005a). The measurements of surface ablation were made from July to September each year. All ablation poles were drilled 10 m deep with a steam driven Heucke Ice Drill (Heucke, 1999). Snow, firn and ice ablation were converted into water equivalent (w.e.).

Elisebreen area is 11.9 km^2. Its length is about 7 km, while its width is up to 1.8 km. To the north the glacier borders Agnorbreen which is often treated as part of Elisebreen. The northern border of the glacier is marked by the ranges of the Prinsesserygen and Prins Heinrichfjella, while the southern border by the ranges of thr Jarlsbergryggen, Kysa and Askerfjellet. In the east the glacier is connected with the Løvenskiold Plateau. The altitude of the frontal part of the Elisebreen is about 30–60 m above sea level (a.s.l).

Fig. 2. Frontal part of Elisebreen during summer time (photo I. Sobota)

* e-mail: irso@geo.uni.torun.pl

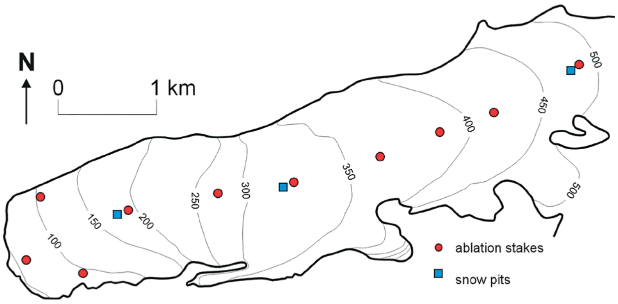


Fig. 3. Map of ablation stakes and snow pits on Elisebreen

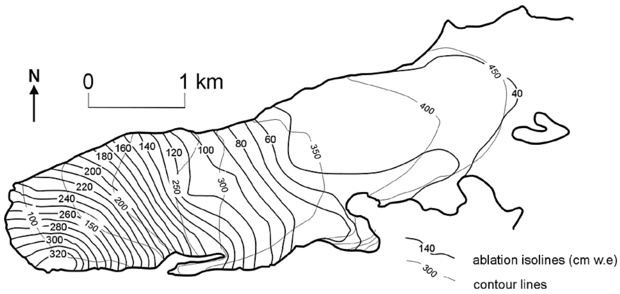


Fig. 4. Map of ablation of Elisebreen in 2006

Spatial diversity of ablation of Elisbreen shows clearly that the largest values were reached in the front part; they decreased towards the accumulation field, where snow cover was found throughout the entire summer season. The size of ablation in the frontal part of glacier (about 3 m of w.e.) was much higher than those of both Waldemarbreen and Irenebreen. This mainly resulted from the fact the altitude of this part of glacier is lower.

In 2006 the summer balance of Elisebreen was –135 cm w.e. Such large negative values during that period resulted from a very early beginning of the ablation season.

The weather conditions of the summer season of 2006 conditioned earlier, compared to previous years, ice- and snow-melting processes in lower parts of the glaciers. Thus, in spite of a large snow accumulation in winter, the mass balance of all the analysed s was negative.

Point 3 – Irenebreen   78° 39’ 24” N 12° 03’ 26” E

Summer balance of Irenebreen

Ireneusz Sobota*

Department of Cryology and Polar Research, Institute of Geography, N. Copernicus University, Toruń, Poland

The studies of the summer balance of Irenebreen were taken between 2001 and 2006. This researches are continued (Sobota, 2004, 2005b). The measurements of surface ablation were made every 5–7 days from July to September each year. All ablation poles were drilled 10 m deep with a steam driven Heucke Ice Drill (Heucke, 1999). Snow, firn and ice ablation were converted into water equivalent (w.e.).

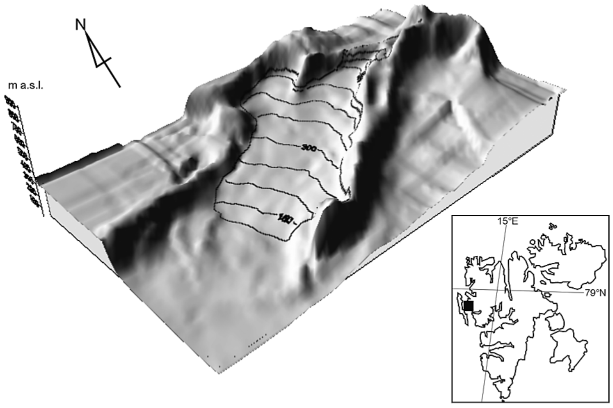


Fig. 1. Topographical draft of Irenebreen

Irenebreen is a valley glacier located to the south of Waldemarbreen, flows down towards the Kaffiøyra plain. In the north it borders the mountain chain of the Gråfjellet and Kristinefjell Range, in the east with the Prins Heinrichfjell Range, while in the south with the Prinsesserygen Range. The area of Irenebreen amounts to 4.2 km^2, its length to 4 km, while its width ranges from about 1 km in its frontal zone to about 1.5 km in the east section. Irenebreen has two significant accumulation zones. Ice masses flowing from them join into a glacier tongue which moves to the south west and ends on the Kaffiøyra.

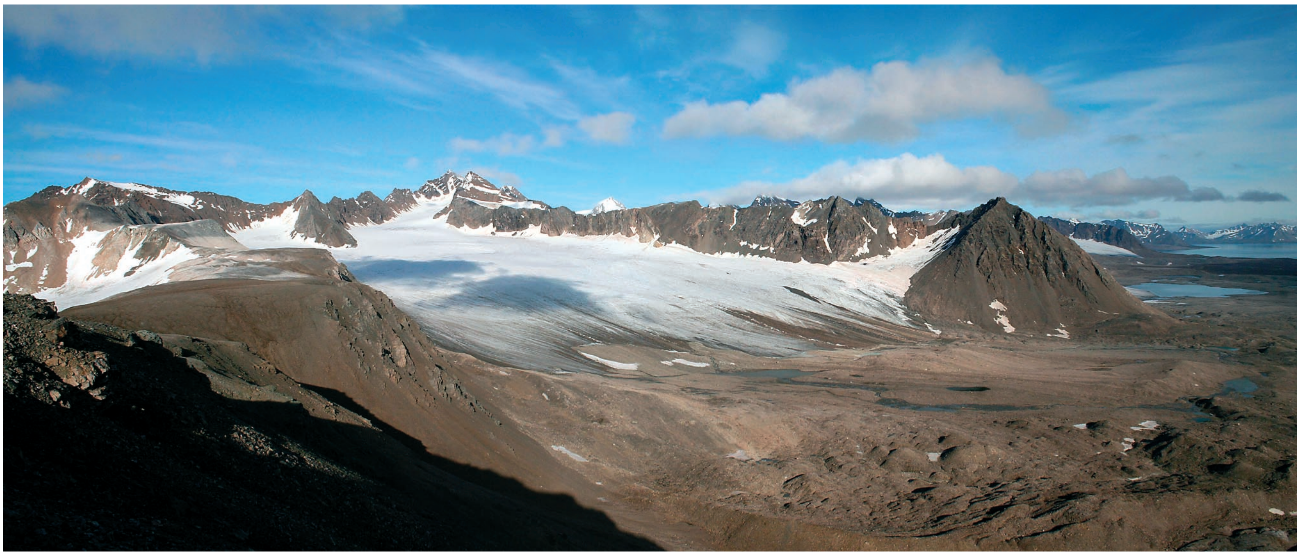


Fig. 2. Irenebreen during summer time (photo I. Sobota)

* e-mail: irso@geo.uni.torun.pl

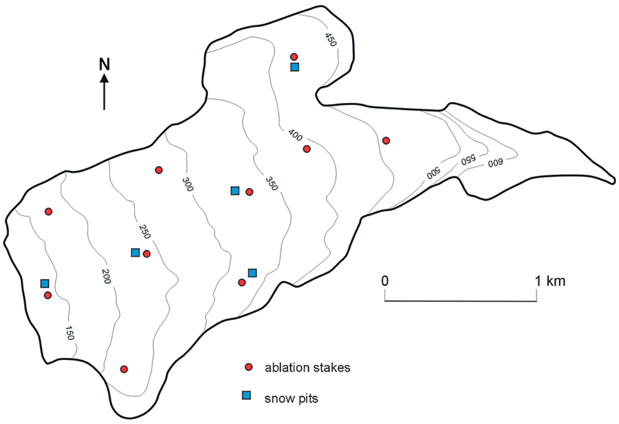


Fig. 3. Map of ablation stakes and snow pits on Irenebreen

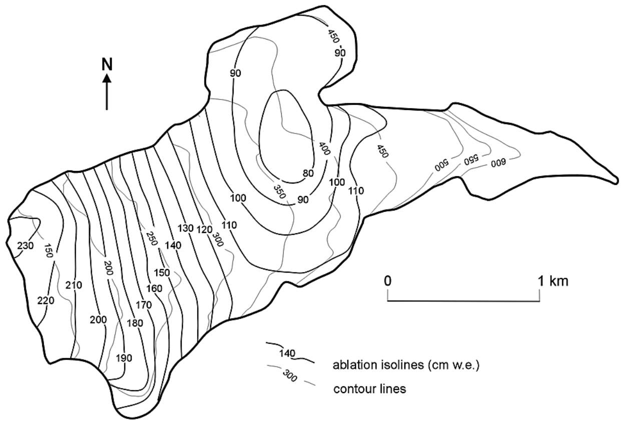


Fig. 4. Map of ablation of Irenebreen in 2006

Time changeability of ablation Irenebreen at various latitudes was significantly diverse. The greatest changeability was observed in the lowest parts of glacier. With the growing altitude the fluctuations decrease. There is a large difference in ablation intensity between the frontal part of the glacier and its accumulation part. This is mainly connected with the diverse weather conditions in these parts of glacier. The parts of glacier which are located high are influenced by lower air temperatures and thus ablation there are either much less intense or non-existent. The lowest part of glacier, however, is often located in the zone of much warmer air masses, and thus ablation are much more intensive there. Very often the altitude-related ablation is also influenced by local conditions of a given glacier, such as the slope aspect, its exposition, surrounding mountain slopes, the amount of morainic material on glacier and the system of supraglacial streams.

Spatial differentiation of ablation on Irenebreen also shows some regularity. The most intensive ablation was recorded in the northern part of the frontal section of glacier, while the least intensive was recorded on the accumulation field. Lower values of ablation intensity were also recorded in the south-central part of glacier. It is located at the foot of the mountains which means sunrays are blocked there. Such a situation influences the intensity of ablation.

The average summer balance of Irenebreen amounted to –124 cm w.e. for the period of 1996–2006. In the years 1996–2006 the cumulated total ablation of Irenebreen was about –740 cm w.e.

Point 4a – Waldemar River *78° 40’ 23” N 11° 58’ 10” E*

Naledi of the Kaffiøyra

Marek Grześ*

Department of Cryology and Polar Research, Institute of Geography, N. Copernicus University, Toruń, Poland

Most studies into the naledi of a glacial origin in the Svalbard were conducted to the summer season, that is when the naledi subject to greater disintegration or warnningg. The research focusec upon a morphogenetic role of the naledi.

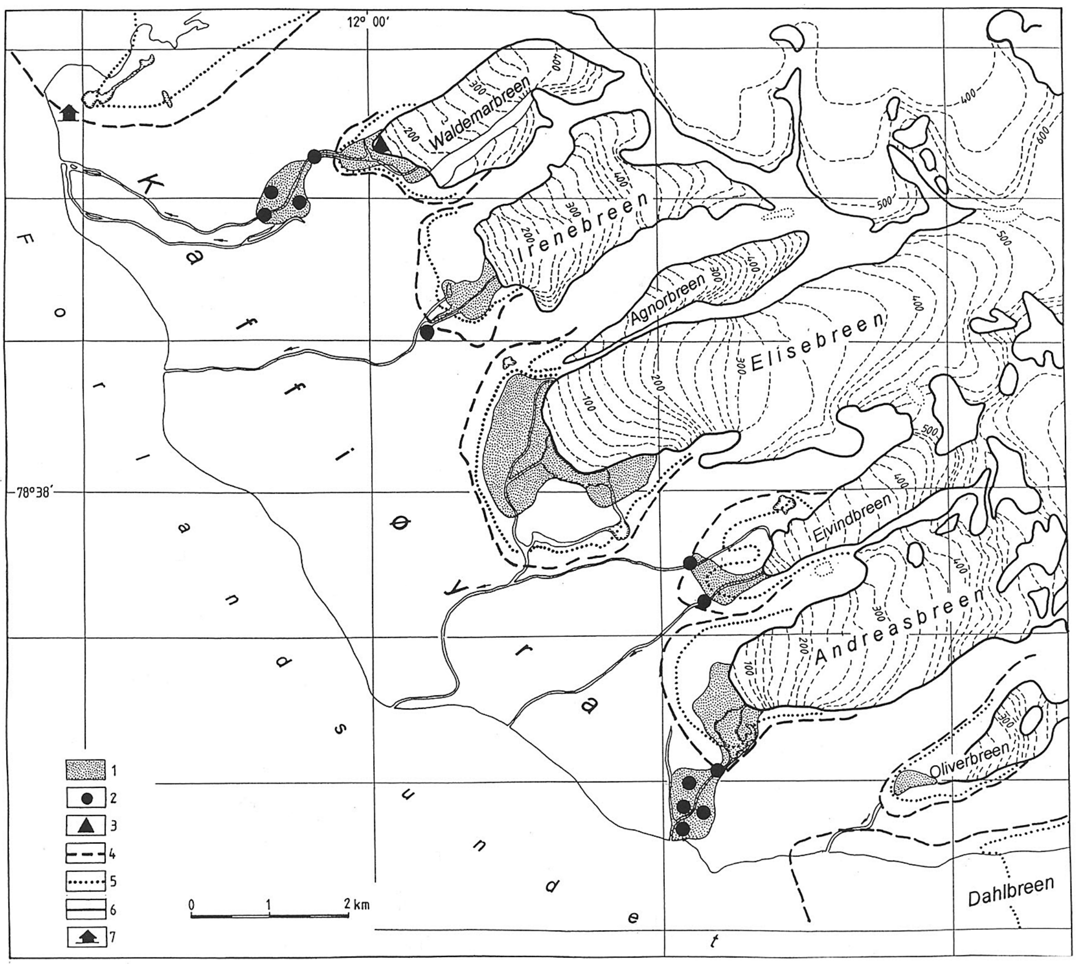


Fig. 1. Naledi of the Kaffiøyra
1 – areas occupied by naledi, 2 – icing mounds, 3 – supraglacial naled, 4 – extents of glaciers In the 18th/19th centuries, 5 – extents of glaciers in 1936, 6 – extents of glaciers in 1995, 7 – Nicholas Copernicus University Polar Station. The naled map was made on the grounds of the map of glacier extents by K.R. Lankauf (1999)

* e-mail: gmark@geo.uni.torun.pl

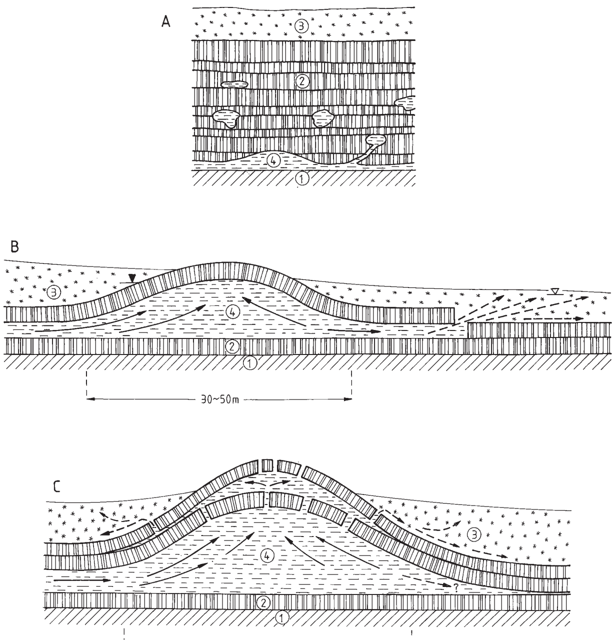


Fig. 2. Naled drainage schemes
A – Typical naled cross-section, B – Cross-section of a closed icing Mount, C – Cross-section of an open icing Mount; 1 – bedding, 2 – naled ice, 3 – snow, 4 – water

The present investigations were carried out in the forefields of six glaciers in the Kaffiøyra., in the the north-west of Spitsbergen.

The observation have been carried out since 1975, during spring time (April–May) and summer time (July–August). All the the glaciers were found to be accompanied by naledi.

However, it is only inthe case of the Waldemar and Andreas glaciers tht the naledi partially covered the outwash plains (see map). The studies explained this was related to the capasity of accumulating glacier waters in the marginal zone. The phenomenon was defiined as the marginal zone capacity and was was found to depend upon its configuraion. It must be assumed that at the time of the glacier progressing recession, naled cannot be formed in the extra-marginal zones.

It is very difficult to determine the area of the naledi precisely. They are mantled with a snow cover which plays decisive role in the migration of the outflowing water from the glacier. It was found that sesonal changes in the area of the naledi amounte to 15–20% front of the particular glaciers. The results of the investigations in the Kaffiøyra led to a conclusion that the naledi of a glacial origin reached the larges sizes in the winter and spring protected by a warmer and humid summer seasons

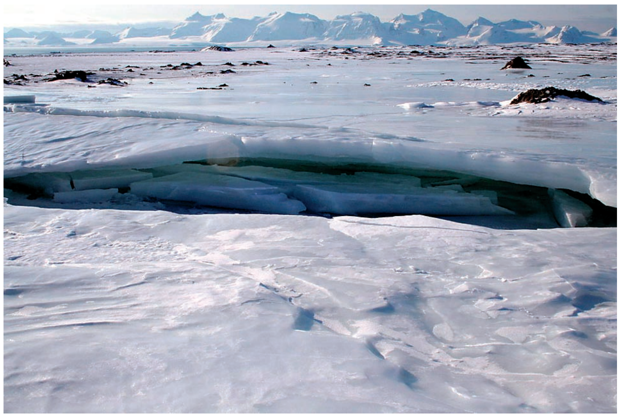


Fig. 3. Naledi (photo I. Sobota)

Every spring the naledi of the Kaffiøyra reaached the area of approximately 4.5 km^2 on average.

The comparison of values, that bigger galciers are accompanied by bigger naledi.

The volume of a naledi ice should be taken into consideration while making these estimates. It is complicated to establish the plume of a winter discharge from glaciers on the grounds of the naled ice.This results from a huge portion of water saturated snow (up to 80%). At the time of a naled-formation there are two systems of water migrations in them: free (gravitational) and forced (under pressure). Ice mounds are characteristic elements of a naled surface configuration.

One supraglacial naled was found on the Waldemar Glacier, but it was seasonal winter outflow only. It has had a consderable influuence upon the course of the glacier retreat. Very important problem concerning the volume of water trapped in naledies. In May 1998 it was equal to about 0.5 mln m^3. The average winter outflow from the Waldemar Glacier was estimated to 0.024 m^{-3} s i.e l km^{-2} (Grześ, Sobota 2000).

The wanning of the naledi was not the subject of the authors investigations. It was concluded, however, that the intensity of this phenomenon was determined by the disintegration of the naled sheets, division into smaller fields. This disintegration can occur in sub-naled and in-naled channels of winter drainage.

Point 4b – Waldemar River *78° 40' 23" N 11° 58' 10" E*

Discharge of Waldemar River and outflow from glacier

Ireneusz Sobota*

Department of Cryology and Polar Research, Institute of Geography, N. Copernicus University, Toruń, Poland

Catchment basins of six rivers can be distinguished on the Kaffiøyra plain. The catchment basin of Waldemar River is the smallest; its area takes 4.4 km^2, 62% out of which is taken by Waldemarbreen. The measurement site was located at the point where the river enters the outwash plain, about 500 m from the glacier frontal part. The length of Waldemar River from that place is about 1 kilometre. Below that point the river shows anastomosing character. The main factor shaping the catchment basin of Waldemar River is fluvioglacial water of the Waldemarbreen. Its area includes the streams fed both by the ablation water and precipitation water.

The largest intensity of the discharge corresponded with the period of highest ablation level. The closest correlation was visible when a few-day values were analysed. Additionally, there were periods when increased intensity of discharge was recorded later than the maximum of ablation. This mainly resulted from temporary retention of melted snow in the form of slush, large patches of which were found on glaciers. During every summer season ablation exerted distinct influence over the size of the discharge of Waldemar River. This is proved by the correlation between ablation and discharge.

Mean discharge of Waldemar River between 1996 and 2006 was 1.21 m^3s^{-1}, while in 2006 was 1.08 m^3s^{-1}.

In order to measure water stages and water temperatures at 5-minute intervals the HOBO logger was used. This enabled the author to estimate the discharge rate, both daily and mean for the entire summer season.

The mean outflow from Waldemarbreen between 1996 and 2006 was 4,578,641 m^3s^{-1} of water, which was carried away by Waldemar River. The share of the ablation within the outflow was, on average, 56%. The remaining part was made up by rainfall, outflow from the ice covers as well as other local sources of water (inter-glacier outflow, melting of snow from the mountain slopes).

Fig. 1. Waldemar River in 2006 summer (A) and 2007 spring (B), (photo I. Sobota)

* e-mail: irso@geo.uni.torun.pl

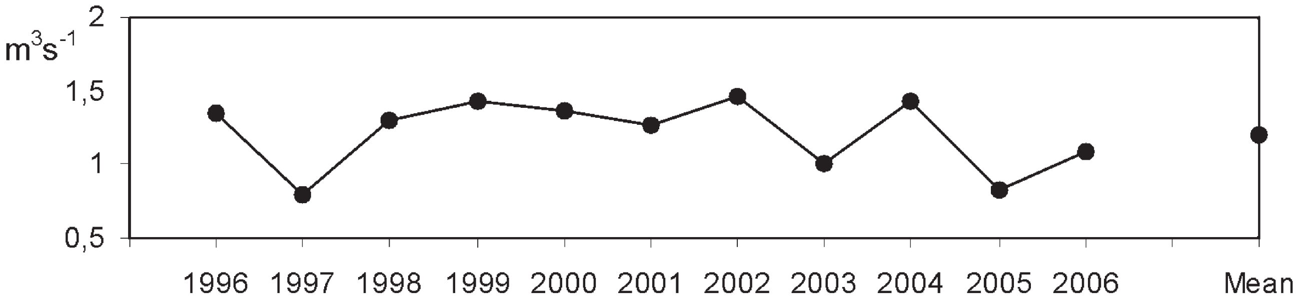


Fig. 2. Mean discharge of Waldemar River in 1996–2006

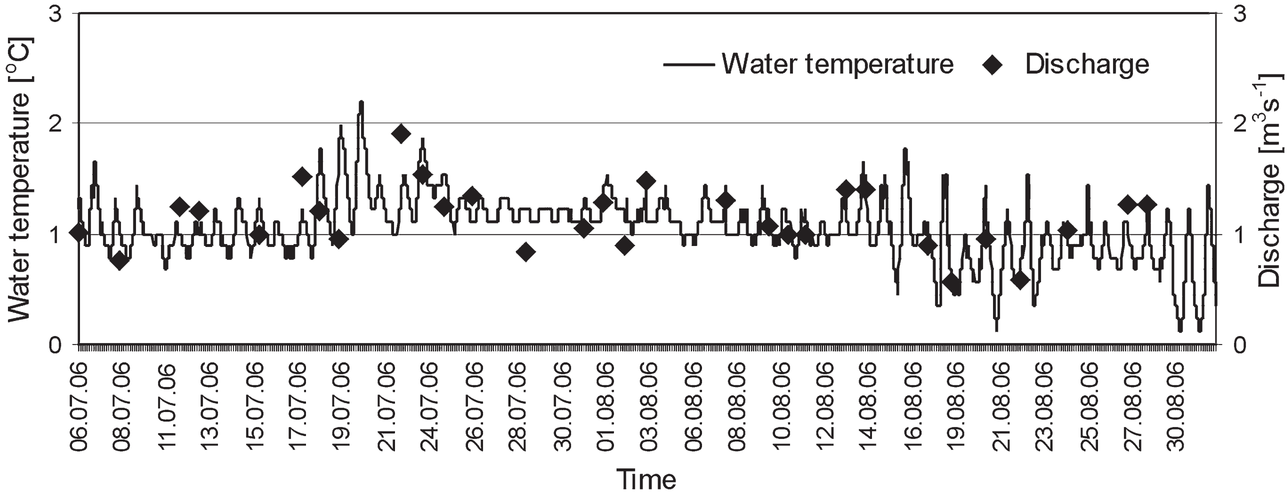


Fig. 3. Water temperature in Waldemar River against discharge in summer 2006

Point 4c – Waldemar River 78° 40' 23" N 11° 58' 10" E

Thickness of mineral covers on the ice-cored moraine and an active layer of permafrost on the western coast of the Oscar II Land (Svalbard)

Marek Grześ*

Department of Cryology and Polar Research, Institute of Geography, N. Copernicus University, Toruń, Poland

Ice-cored moraines belong to the dominating forms in the landscape of the glaciers marginal zones. The relict glacial ice existence in in the ice-covered moraines depends on climatic conditions, the thickness of mineral covers and their seasonal thawing. In the case of the active layers contacting the top of the relict ice, its disappearing is quicker. Mass movements uncovering ice play the greatest role then. That phenomenon is much more important at the initial phase of ice-cored moraines degradation. When the thickness of the mineral cover increases, the rate in which ie thaws decreases. As a result, slope processes slow down distinctly. At that moment ice-cored moraines reach the mature stage, which is indicated by numerous thermokarst hollows with no drainage.

The top parts of the ice-cored moraine lack snow covers throughout the whole year. The process of thawing last about 1.5 months longer as compare to the surrounding area. Its begins as early as in the second part April and gets to he depth of 0.3–0.4 m in first ten of May. Besides climatic conditions slope layers character play a significant role in relict ice degradation. Four characteristic phases distinguished in evolution of ice-cored moraine.

The questions arise: How long can the ice-core moraine exist? What geomorphological effects occur after the ice-core melting? Known that the outer series of ice-cored moraine were formed at the turn of the 19th century. These are "mature" forms which, however, contain ice inside.

The depth of summer thaw in various kind of ground has been presented as a scheme on the fig. 2. From the analysis of a set of estimates available, it follows that at least seven different environments which differ in the course and of thaw, i.e. the thickness of the active layer, my be recognized:

Ia, Ib) ground varying in particle, size distribution, with a blanketing continuous organic layer, more than 20 cm thick,

II) tills with high moisture contents which are colonized by luxuriant tundra vegetation,

III) gravels, sands and silts within depressions at marine terraces, which are occupied by luxuriant tundra vegetation,

IV) modeled (patterned) ground, the interior of which is build up of predominant earth particles and which is colonized by extremely luxuriant tundra vegetation,

V) sands, gravels of which marine terraces covered with sparse tundra vegetation are built up,

VI) gravels and sands forming present-day outwash fans,

VII) mineral mantle (boulders, silts, ...) over ice-core moraine.

The established pattern of seasonal thaw in various kind of ground at an 60 m a.s.l. provides the basis of plotting of eight (Ia, Ib, ... VII) empirical curves, the approximations of which are given by the formula:

$$h = a \lg (T \pm c) - b$$

where h is the depth of a thawed layer, a and b are constants coefficients defining a thawing layer, T is the duration of thaw in days, c is allowance for earlier (+) or a later (-) disappearance of snow at given locality. Index c is calculated in the following way: the actual number of the 24-hour of delay divided by 4.

* e-mail: gmark@geo.uni.torun.pl

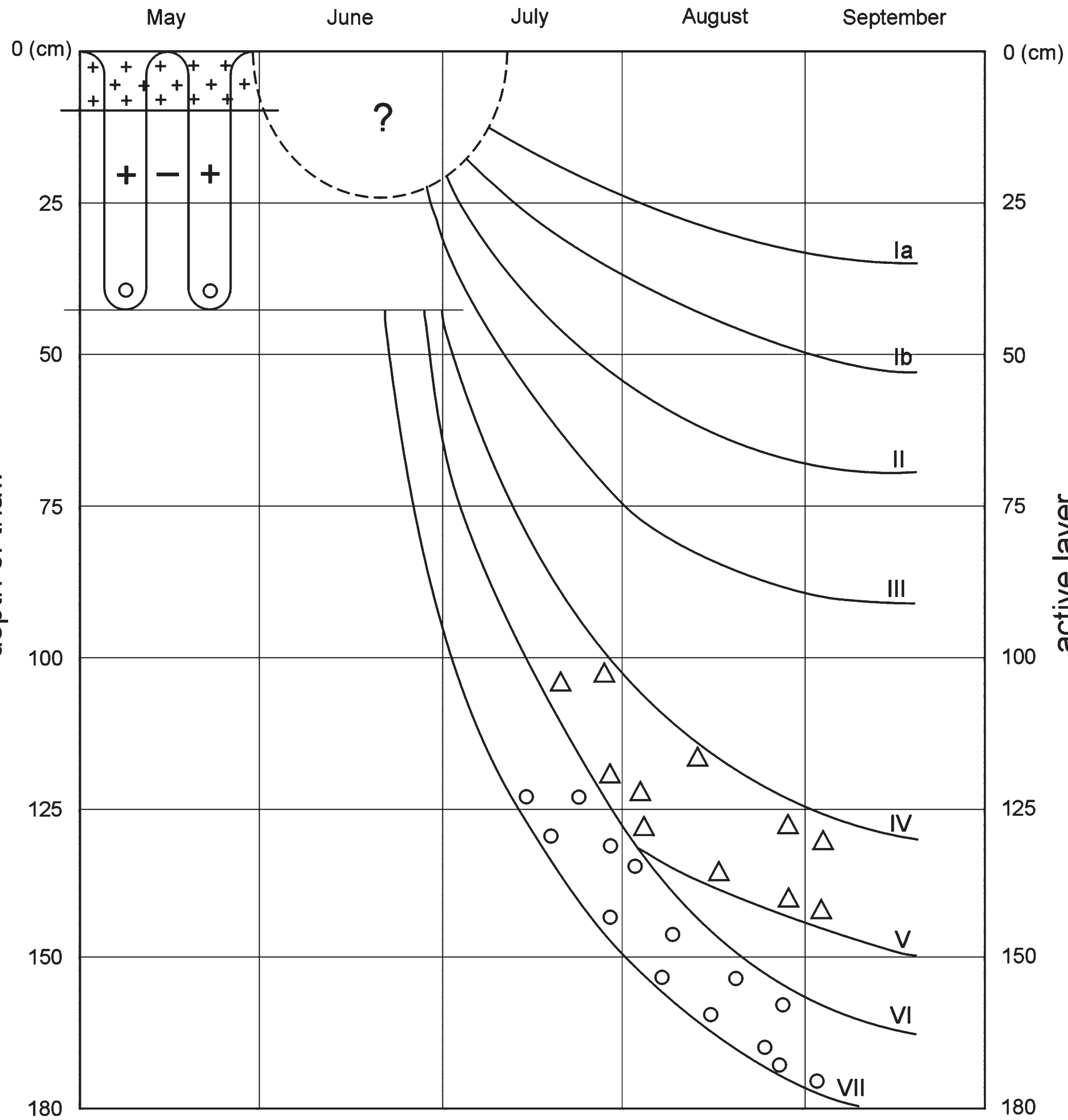


Fig. 1. Degradation scheme of ice-core moraine
1 – initial heiht (H1), 2 – formation of the initial thermokarst hollow with a small lake, 3 – formation of the complex of thermokarst hollows, 4 – formation of the therokarst hollows parallel to the ridge lines of the ice-core moraine, 4a – the final stage of the ice-core moraines degradation, the division of the ice core into two parts and their height H2, W1 and W2 – initial and final wi width of the form base

If the duration of thawing in a selected kind of environments is known, the above formula permit determination of the thickness of the active layer with high precision.

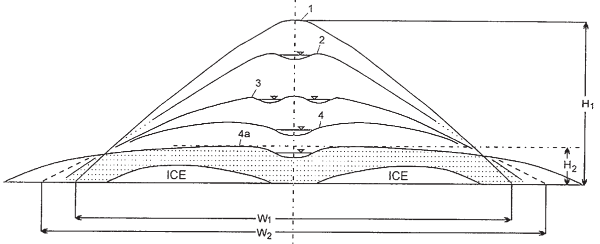


Fig. 2. Scheme of summer thaw (explanation in the text)

Point 5a – Waldemarbreen *78° 40' 30" N 11° 59' 28" E*

Mass balance monitoring of Kaffiøyra glaciers

Ireneusz Sobota*

Department of Cryology and Polar Research, Institute of Geography, N. Copernicus University, Toruń, Poland

The studies of the structure of mass balance of Kaffiøyra glaciers refer to the Waldemarbreen, Irenebreen and Elisebreen. The data on the structure of the mass balance of Waldemarbreen were based on the direct field measurements conducted from 1996 to 2006 (Sobota, 1999, 2000, 2004, 2005a, Sobota and Grześ, 2006). The studies of the mass balance of Irenebreen were taken between 2001 and 2006. In 2005 the studies of the mass balance of Elisebreen began. This research is continued. At the same time geodetic and cartographic measurements were carried out (Lankauf, 2002, Bartkowiak et al., 2004).

Glaciers are located in the northern part of the Oscar II Land, Kaffiøyra, north-western Spitsbergen. Waldemarbreen is about 3.5 km long and has an area of 2.6 km^2. The ice originates in one cirque and flows from an elevation of more than 500 m to the present terminus at 130 m a.s.l.. Irenebreen, a valley glacier located to the south of Waldemarbreen, flows down towards the Kaffiøyra plain. The area of Irenebreen amounts to 4.2 km^2. Elisebreen area is 11.9 km^2. Its length is about 7 km, while its width is up to 1.8 km. To the north the glacier borders Agnorbreen which is often treated as part of Elisebreen.

In order to estimate the mass balance of Kaffiøyra glaciers the method of direct measurements was used. It was based on a set of ablation poles completed with the studies of the snow cover in the snow profiles. Twenty-two poles were placed on Waldemarbreen; Irenebreen and Elisebreen had ten poles installed each.

Spatial diversity of mass balance of Waldemarbreen, Irenebreen and Elisebreen is mainly influenced by the weather conditions in a specific part of glacier and by local morphological conditions. The areas of the glaciers may be generally divided into the part of the negative mass balance and the part of the positive mass balance. In the case of Waldemarbreen the year 1998 was exceptional, as the entire glacier showed negative mass balance. Irenebreen shows more positive mass balance in its both accumulation parts. The accumulation part of Elisebreen also shows positive mass balance. This results from the fact that they both are located at higher altitude than Waldemarbreen.

Thanks to the direct measurements, the average location of the equilibrium line (ELA) on Waldemarbreen was estimated at the altitude of 397 m in 1996–2006. From 2002 to 2006 the annual equilibrium line altitude was 421 m a.s.l. for Irenebreen, while for Elisebreen it was 365 m a.s.l..

The average mass balance of Waldemarbreen amounted to –57 cm w.e. in 1996–2006. Between 2002 and 2006 the mean annual mass balance of Irenebreen was –71 cm w.e. In 2006 the mass balance of Elisebreen was –73 cm w.e. These values are close to other Svalbard glaciers of similar size.

* e-mail: irso@geo.uni.torun.pl

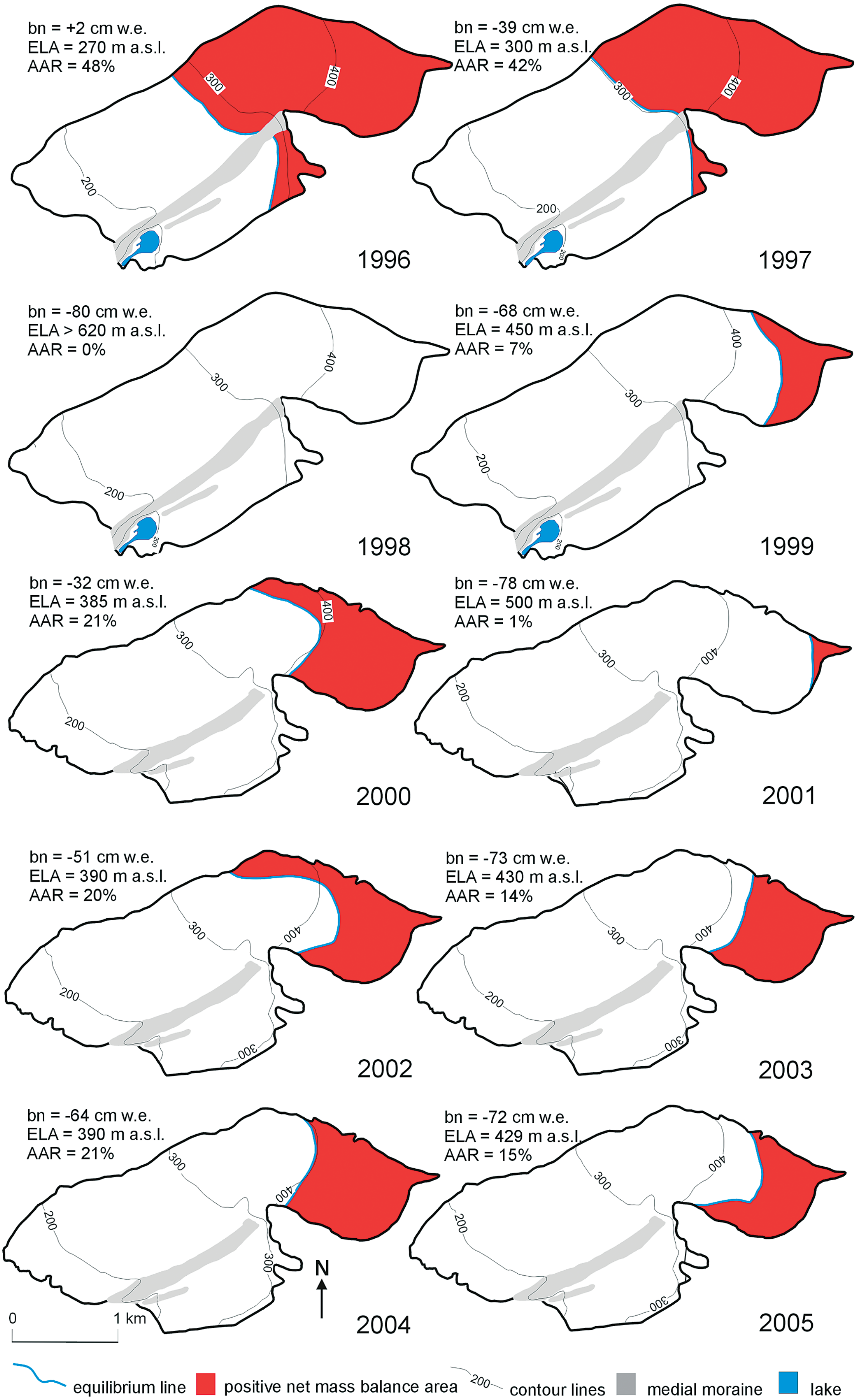


Fig. 1. Positive net mass balance area and equilibrium line (ELA) location maps of Waldemarbreen in 1996–2006

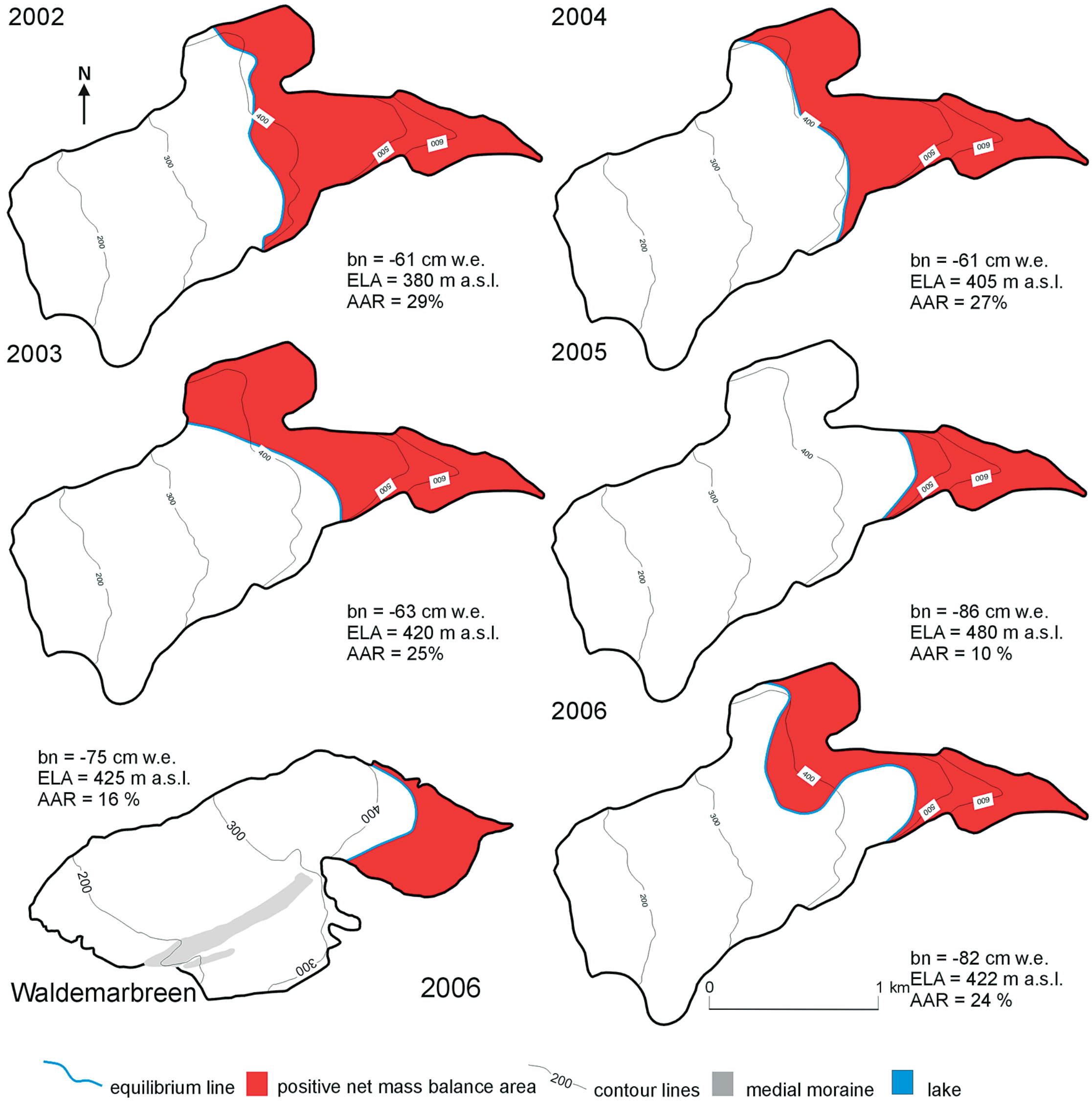


Fig. 2. Positive net mass balance area and equilibrium line (ELA) location maps of Irenebreen in 2002–2006

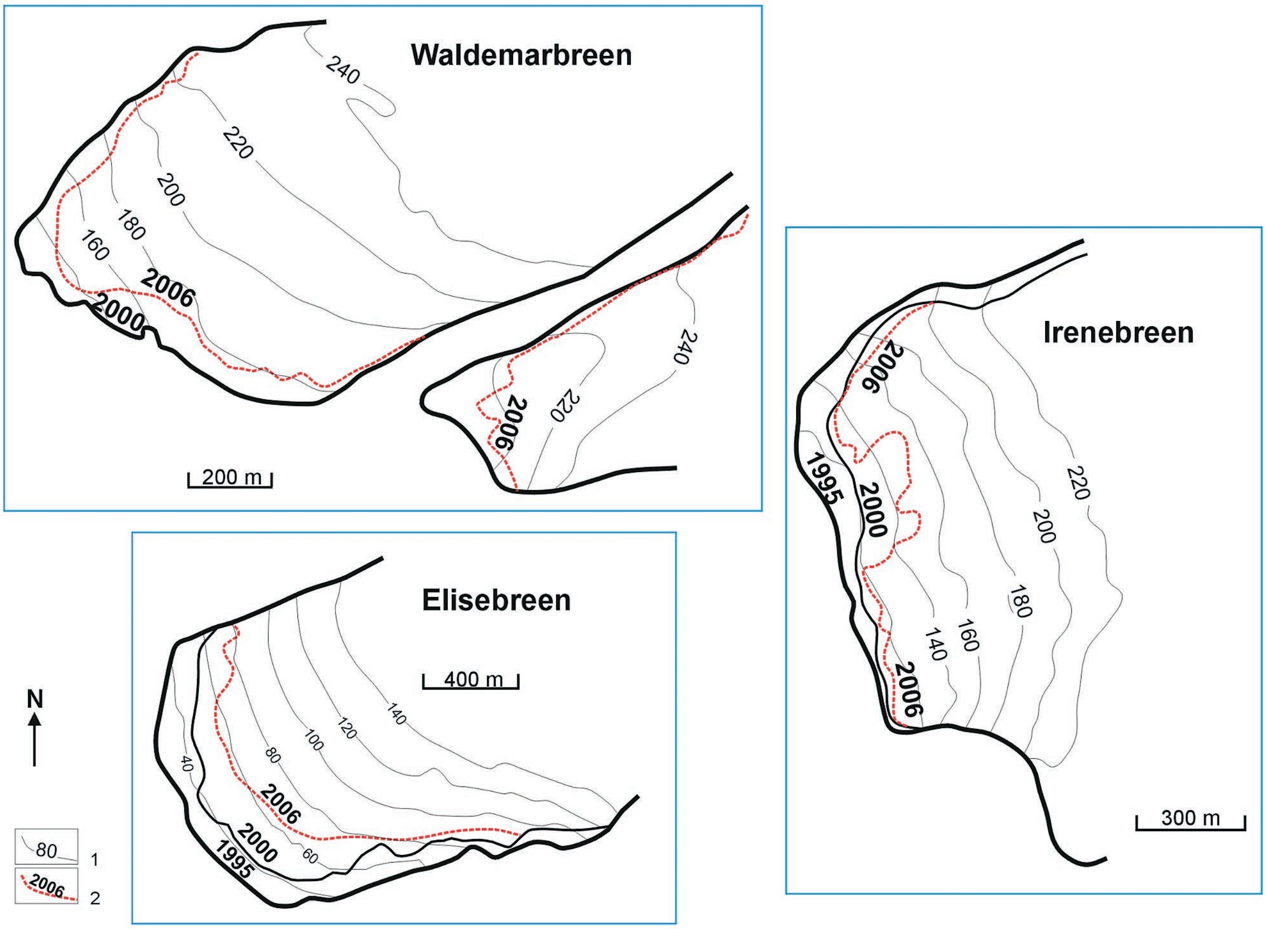


Fig. 3. The retreat of Kaffiøyra glaciers in the period from 2000 to 2006. Based on Lankauf's topographical map from 1995 (2002) and GPS measurements

1 – contour lines, 2 – location of glacier front in 2006

Point 5b – Waldemarbreen *78° 40' 30" N 11° 59' 28" E*

Summer balance of Waldemarbreen

Ireneusz Sobota*

Department of Cryology and Polar Research, Institute of Geography, N. Copernicus University, Toruń, Poland

Waldemarbreen is located in the northern part of the Oscar II Land, Kaffiøyra, north-western Spitsbergen. Waldemarbreen is about 3.5 km long and has an area of 2.6 km^2. The ice originates in one cirque and flows from an elevation of more than 500 m to the present terminus at 130 m a.s.l..

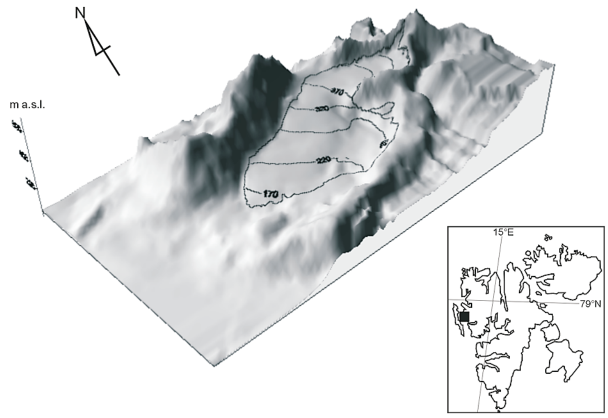


Fig. 1. Topographical draft of Waldemarbreen

The measurements of surface ablation of Waldemarbreen were made every 5 days from July to September each year for the period 1996–2006. The measurements were taken at 22 points of glacier. This is a large number if compared to the area of glacier. Such a dense network of the poles enabled us to estimate precisely the value of ablation at a given altitude, as well as the influence of the local conditions on its size. All the ablation poles were drilled 10 m deep with a steam driven Heucke Ice Drill . Snow, firn and ice ablation were converted into water equivalent (w.e.). The ice density of 0.9 g cm^{-3} was used to convert ablation thickness to water equivalent. Where snow was found on glacier the appropriate snow density was applied to the computations.

Time changeability of ablation processes of Waldemarbreen at various latitudes was significantly diverse. With the growing altitude the fluctuations decrease. There is a large difference in the ablation

Fig. 2. Waldemarbreen during summer time (photo I. Sobota)

* e-mail: irso@geo.uni.torun.pl

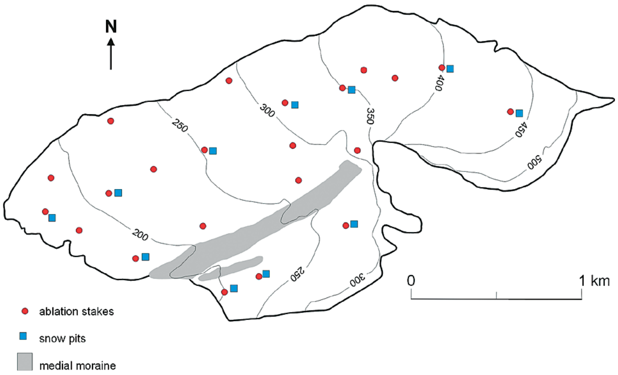


Fig. 3. Map of ablation stakes and snow pits on Waldemarbreen

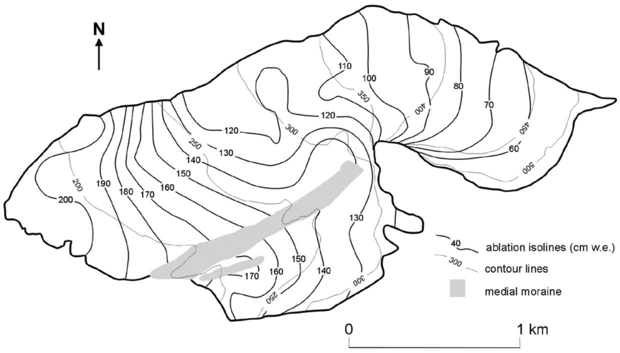


Fig. 4. Map of ablation of Waldemarbreen in 2006

intensity between the ablation part of the glacier and its accumulation part. This is mainly connected with the diverse weather conditions in these parts of the glacier. As far as Waldemarbreen is concerned, the highest ablation level throughout the studied period was found at the altitude of up to 250 m a.s.l. Above that level ablation decreases.

Spatial diversity of the ablation processes of Waldemarbreen was large. It was mainly caused by weather conditions in the individual parts of the glacier, as well as by the relief. Waldemarbreen is strongly inclined not only in its frontal part but also towards the medial moraine. Such a situation means a larger area of the glacier has southern exposition; additionally, the system of supraglacial streams develops and the amount of the moraine material on the glacier's surface increases. As a result, ablation processes in this part of glacier intensify.

The most negative mean summer balance of the glacier was –120 cm w.e. in 1998 and –130 cm w.e. in 2006, while the least negative was –63 cm w.e. in 2000. The average ablation of Waldemarbreen amounted to –104 cm w.e. for the period of 1996–2006. In the years 1996–2006 the cumulated total ablation of Waldemarbreen was about –1148 cm w.e.

Point 5c – Waldemarbreen *78° 40' 30" N 11° 59' 28" E*

Snow accumulation on Kaffiøyra glaciers

Ireneusz Sobota*, Marek Grześ

Department of Cryology and Polar Research, Institute of Geography, N. Copernicus University, Toruń, Poland

Studies of winter mass balance mainly referred to the estimation of the size of the snow accumulation on glaciers, as well as its selected physico-chemical properties. Soundings of the snow depth on Waldemarbreen, Irenebreen and Elisebreen were carried out in about 150 measurement points. They gave a very detailed picture of the spatial diversity of the winter snow accumulation at about 50 measurement points per 1 km^2. Measurement points were located relatively close to one another as the differences in the snow depth is often significant, which mainly results from topography and anemometric conditions. Location of the measurement points was based on both geodesic and the GPS measurements. The measurements were made in the selected snow profiles in accordance with the International Commission on Snow and Ice (ICSI) standards. Additionally, according to the above standards, the selected physical and chemical properties of the snow cover were measured. This mainly referred to the snow structure, graining, hardness and density.

Spatial distribution of winter snow accumulation on Waldemarbreen shows some regularity. The largest accumulation is found in the accumulation part and at the foot of the mountain slopes. The smallest accumulation, however, is observed in the front part of glacier up to the altitude of 220 m and at the foot of the medial moraine. Such a distribution is conditioned by anemometric situation and a larger inclination of this part of glacier. Some asymmetry in the snow cover depth was recorded. In the accumulation part of glacier the main factor influencing the depth of the snow cover was precipitation, while in the lower parts of glacier – local conditions (aspect) as well as wind directions and velocity (snow redeposition). The depth of the snow cover lowers from north east towards south west, i.e. in the direction of the medial moraine. Next in grows again towards the Gråfjellet Range. In the case of Irenebreen snow accumulation increases significantly from the front

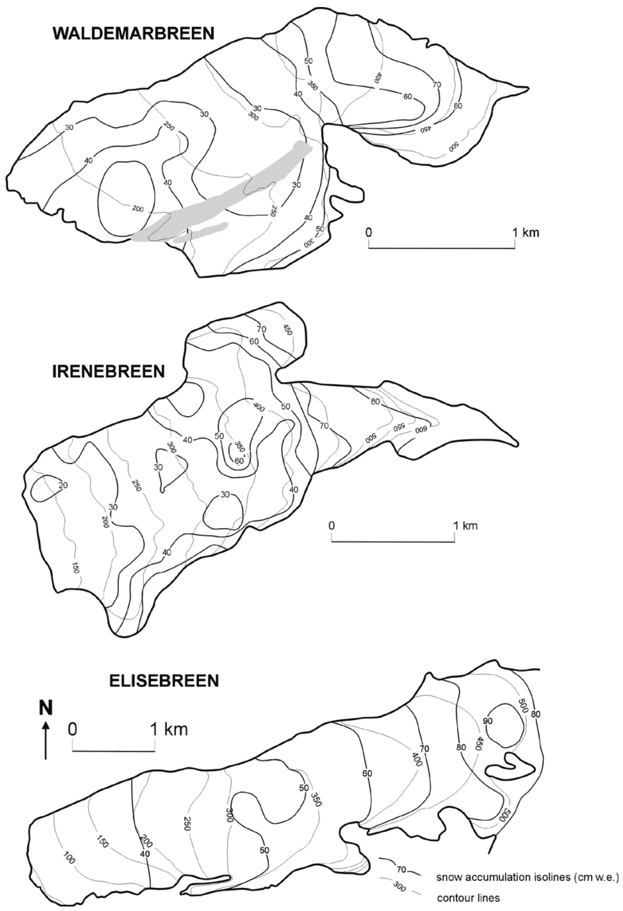


Fig. 1. Snow accumulations maps of Kaffiøyra glaciers in 2005

* e-mail: irso@geo.uni.torun.pl

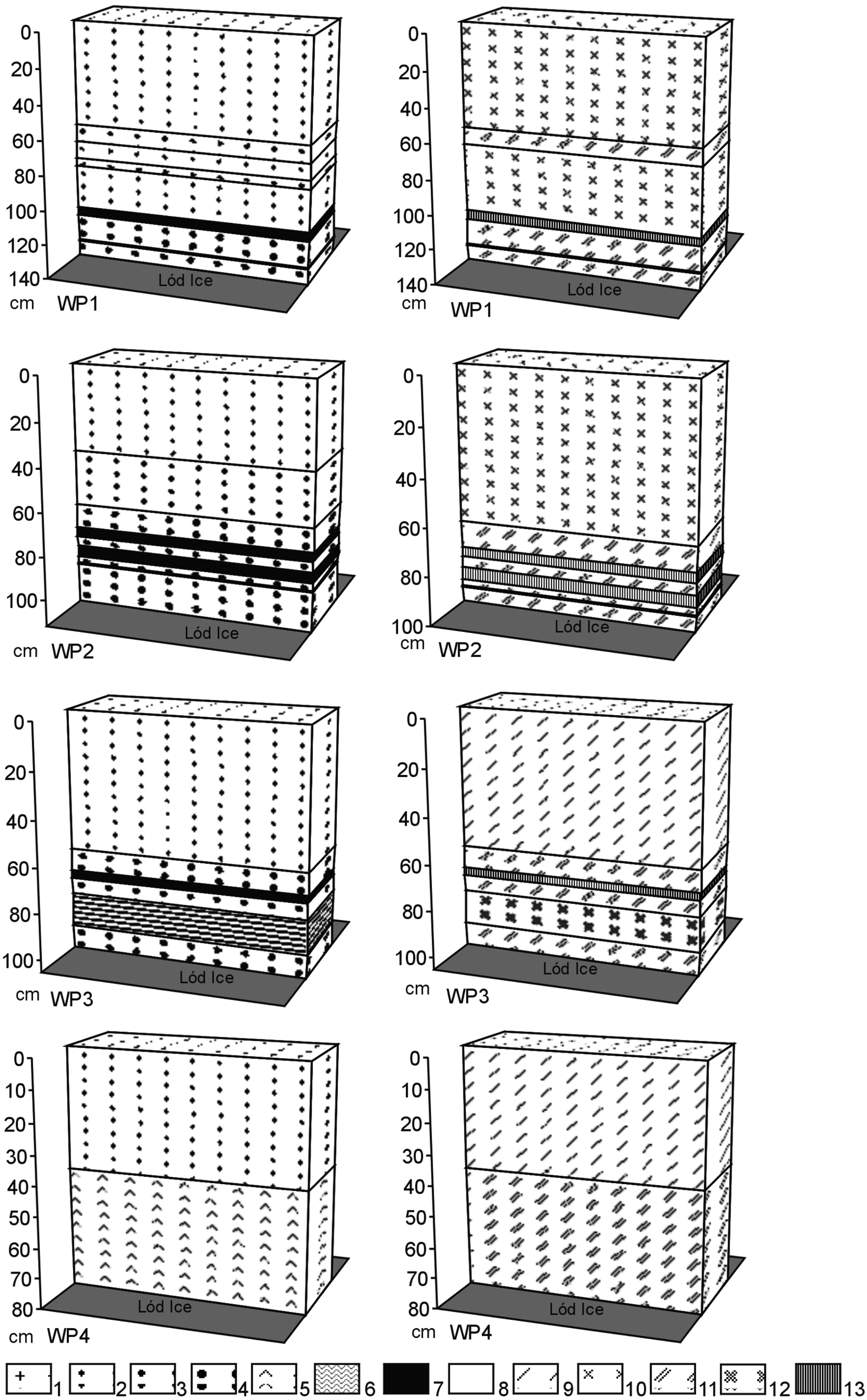


Fig. 2. Snow profiles at selected parts of the Waldemarbreen according to ICSI in May of 2005

Snow grain sizes and types: 1 – fresh snow, 2 – fine grained snow, 3 – medium grained snow, 4 – coarse grained snow, 5 – coarse grained snow, intesively matamorphosed (hoar snow), 6 – frozen snow with ice layers, 7 – ice layer. Hardness of deposited snow: 8 – very low (R1), 9 – low (R2), 10 – medium (R3), 11 – high (R4), 12 – very high (R5), 13 – ice (R6)

Fig. 3. Mesurements in snow pit (photo M. Grześ)

part of glacier towards the accumulation fields. On Elisebreen snow accumulation increases considerably with altitude until the ice-shed, i.e. from 40 cm w.e. to 150 cm w.e.

The measurements of structure and graining of the snow cover were not undertaken during all of the analysed periods. When undertaken, the studies included making a few snow profiles in the selected parts of both Waldemarbreen and Irenebreen. Snow cover shows some specific physico-chemical properties. Its vertical profile shows a variety of snow types of diverse level of metamorphosis, hardness and wetting. Snow structure reflects prevailing weather conditions at the time when the snow cover formed.

Snow density on Waldemarbreen ranged from 310 kg m^{-3} to 520 kg m^{-3} maximum. The mean snow density on both Waldemarbreen, Irenebreen and Elisebreen is similar and it amounts to about 400 kg m^{-3} on average. In the individual years the snow cover of the studied glaciers was dominated by fine-grained and medium-grained snow, while the layer above ice contained coarse-grained snow. Numerous ice layers were also found.

From 1996 to 2006 the mean snow accumulation on Waldemarbreen was 47 cm w.e. Cumulated value of accumulation for the entire glacier was 521 cm w.e. Between 2002 and 2006 the mean snow accumulation value for Irenebreen was 52 cm w.e., while the cumulated value of the winter balance for this period for the entire Irenebreen was 262 cm w.e. In 2005 the snow accumulation for Elisebreen was 59 cm w.e., while in 2006 it was 63 cm w.e. These values are similar to those estimated for other studied Svalbard glaciers.

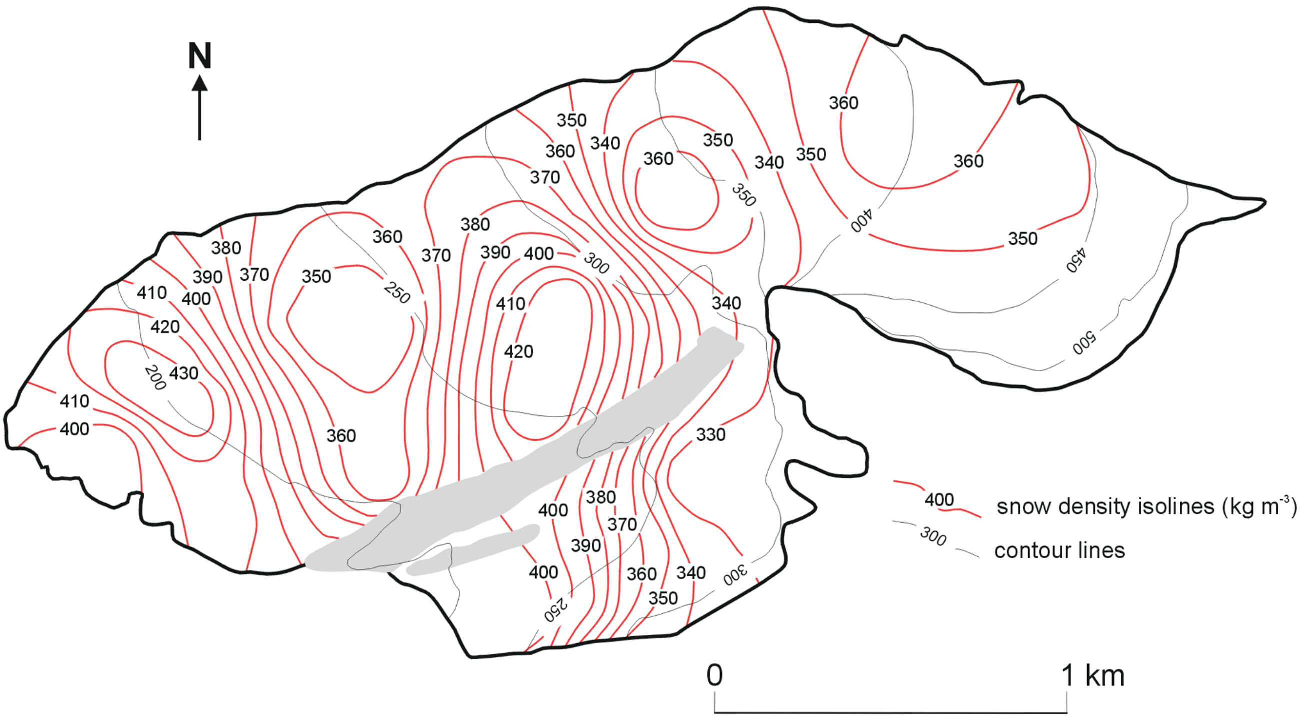


Fig. 4. Surface snow density map of Waldemarbreen, April 2007

The sample ice drilling on Waldemarbreen with Heucke Ice Drill – the demostration

Erich Heucke

Commission for Glaciology, Bavarian Academy of Sciences, Munich, Grafing,Germany

Ireneusz Sobota*

Department of Cryology and Polar Research, Institute of Geography, N. Copernicus University, Toruń, Poland

All the ablation poles on Kaffiøyra glaciers were drilled 10 m deep with a steam driven Heucke Ice Drill . Thanks it is possibility to install mesurements points for the long time period. It is necessary for this kind of investigations.

The purpose of this contribution is to explain the characteristics of a newly developed ice drill which is particularly geared to the needs of glaciologists. It is primarily designed fordrilling holes for ablation stakes and for measuring water levels or temperatures in fim areas. Its distinguishing features are its light weight, making it easy to cany even over long distances, and the variety of tasks to which it can be adjusted. Furthermore, it is easy to operate even by one person.

Water is heated in a boiler by two gas flames to produce steam, which flows through an insulated hose to a nozzle. When the valve is opened the issuing steam condenses, and the heat released in the process melts the ice. The heater is constructed in such a way that it caneasily be adapted to any formof gas supply locally available. It can be used for drilling in ice as well as in fun. The maximum drilling depth is 13 m in ice and 30 m in firn; hole diameters range from 25 to 45 rnrn.Mean drilling time is 16 min for 6 m, 35 min for 12 m in ice. The total weight is somewhat less than 16 kg, including all parts needed for drilling holes of 10 m in depth as well as the gas supply for one day. In recent years, devices of this type have been used successfully by scientists in various glaciated regions.

Fig. 1. Ice drilling with Heucke Ice Drill on glacier (photo M. Marciniak)

* e-mail: irso@geo.uni.torun.pl

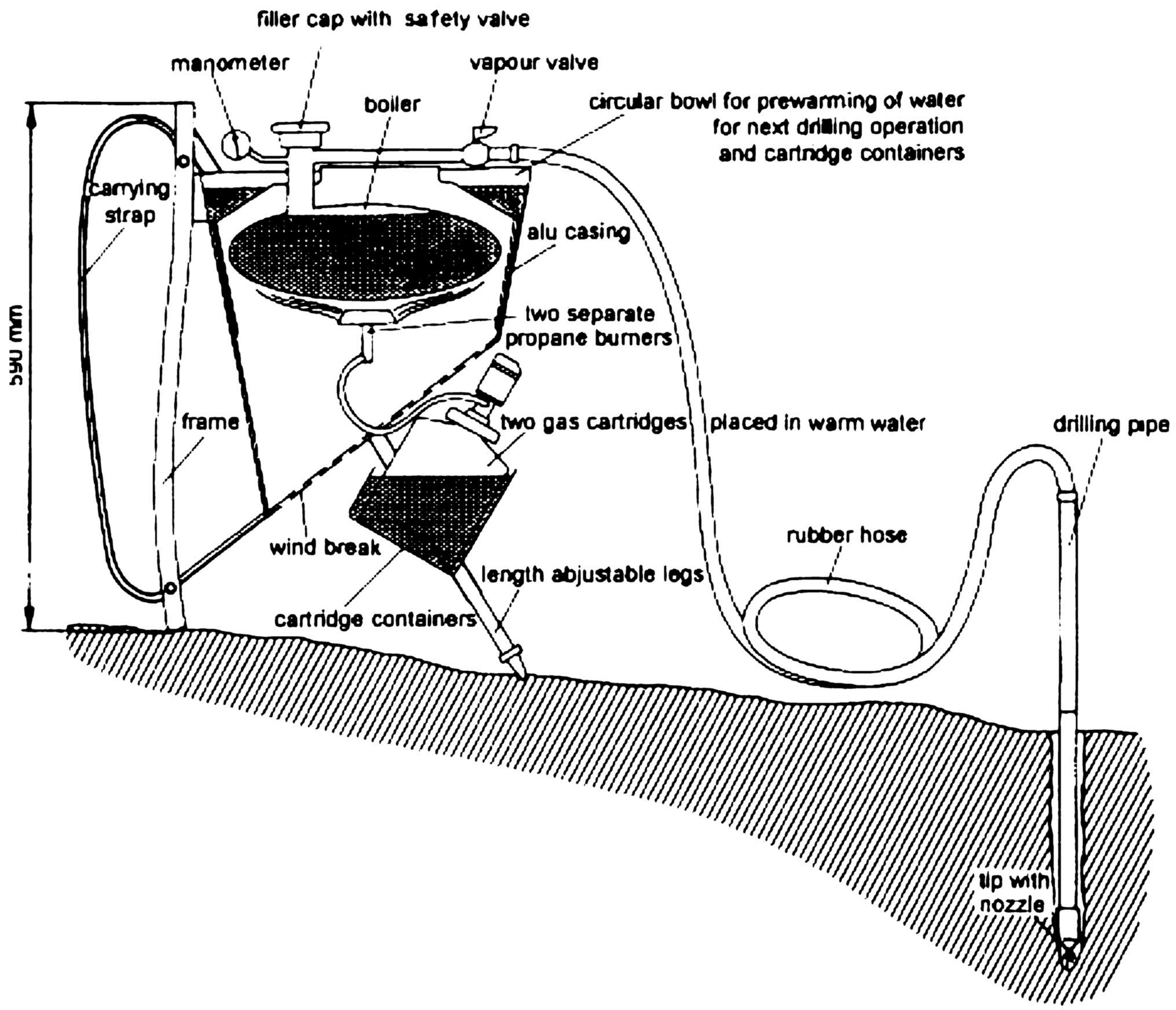


Fig. 2. Diagram of the entire drilling device

Point 7 – Aavatsmarkbreen 78° 41’ 07” N 11° 53’ 39” E

Subaqual recordings of the changes in the range of glaciers in the Forlandsundet region (NW Spitsbergen)

Marek Grześ, Michał Król, Ireneusz Sobota*

Department of Cryology and Polar Research, Institute of Geography, N. Copernicus University, Toruń, Poland

The main goal of the studies is to try and answer the following questions: Do the cliffs of the selected glaciers in the Forlandsundet area re-advance in winter and does this result in the development of subaqual relief? Are subaqual forms concordant with the location of ice cliffs during the selected peri-

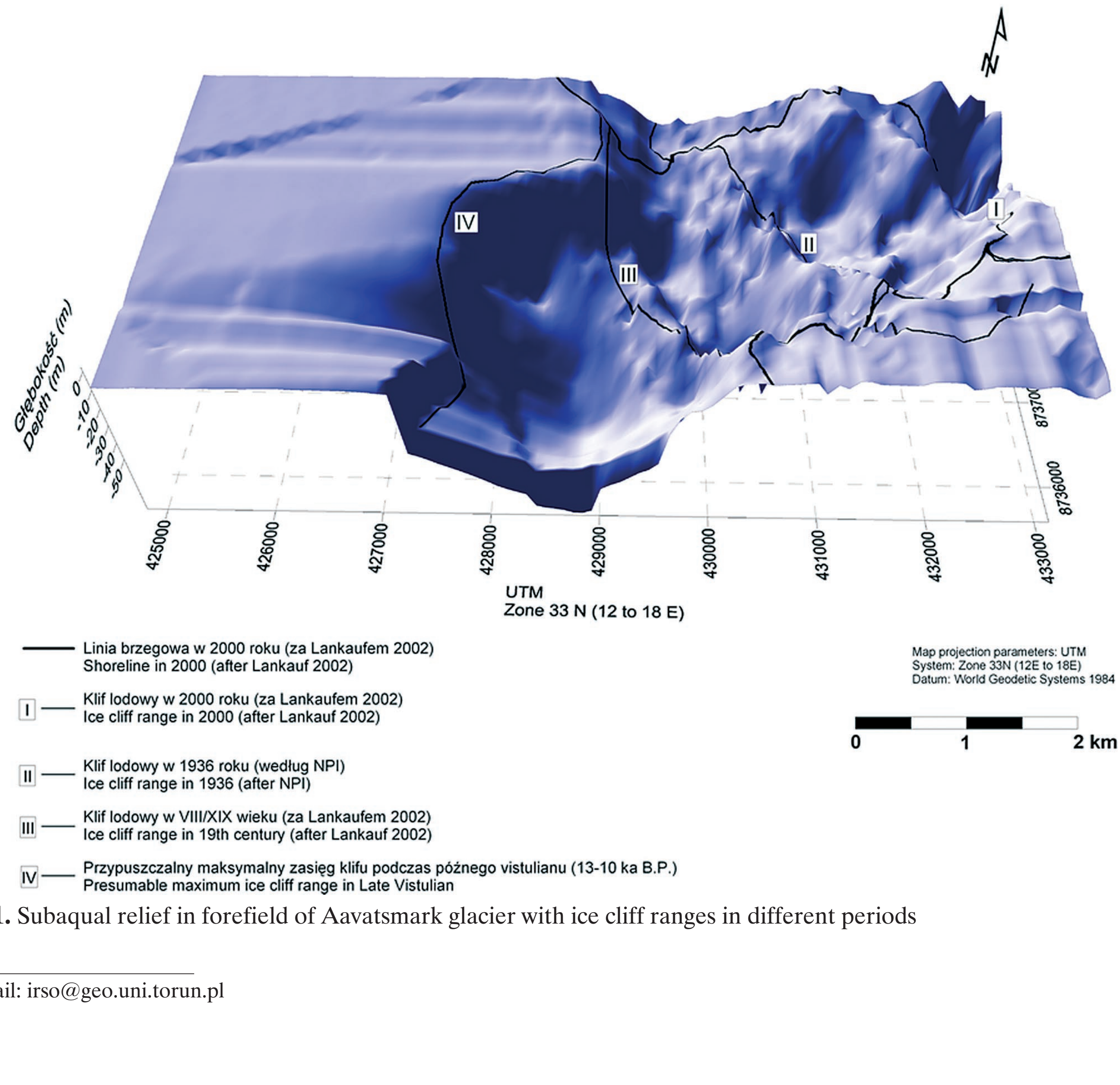


Fig. 1. Subaqual relief in forefield of Aavatsmark glacier with ice cliff ranges in different periods

* e-mail: irso@geo.uni.torun.pl

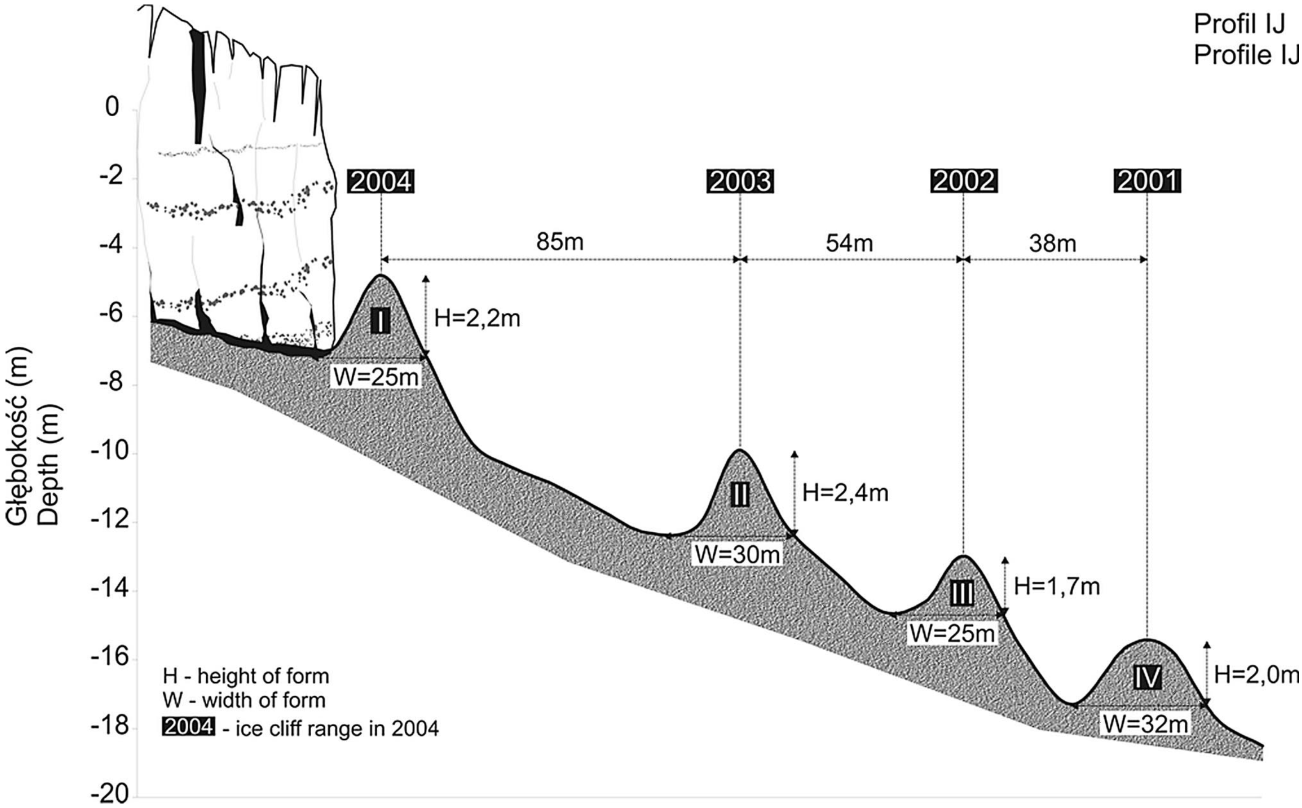


Fig. 2. Annually push moraines formed between 2001 and 2004
H – height of form, W – width of form, 2004 – ice cliff range in 2004.

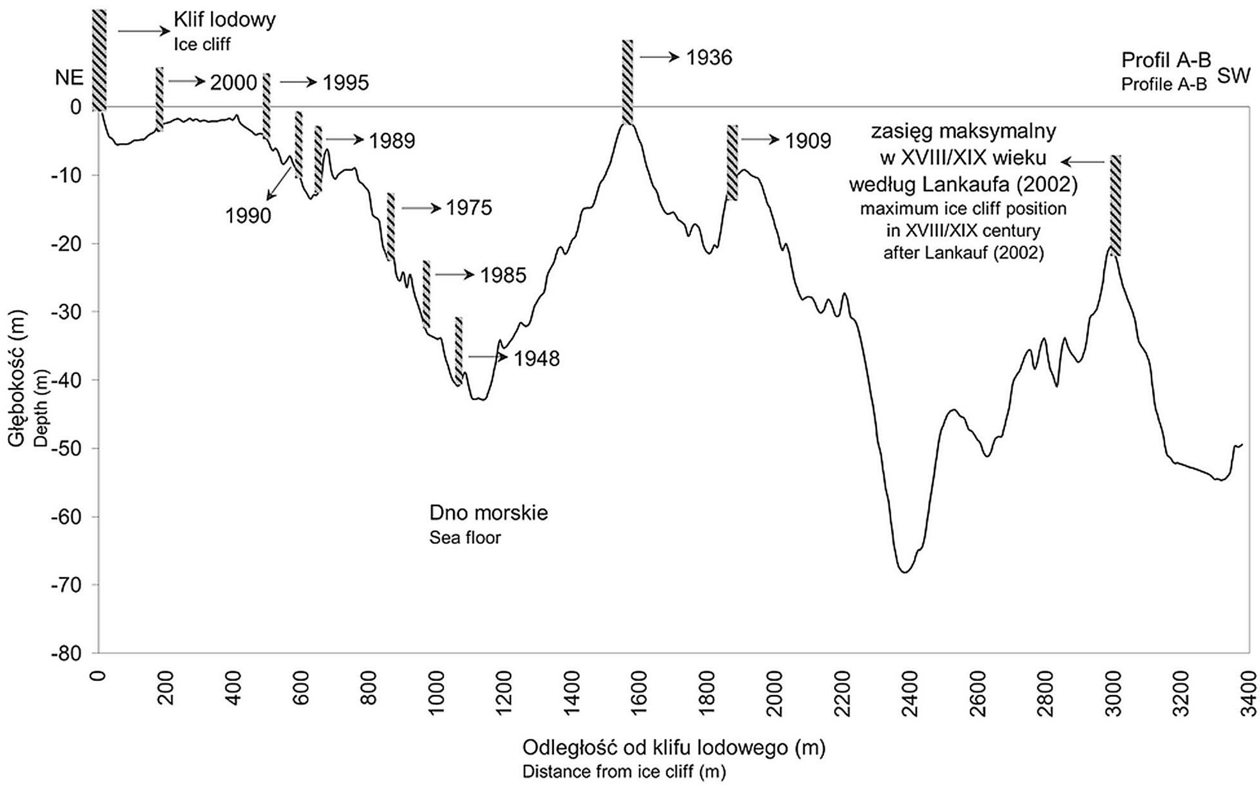


Fig. 3. Subaqual relief in forefield of Aavatsmark glacier, profile A–B

ods of advance, such as the Little Ice Age, the glacial episode (3.0–2.5 ka BP) or the late Vistulian (13–10 ka BP)? What forms are connected with the periods of significant glacier advances, such as the Little Ice Age, the glacial episode (3.0–2.5 ka BP) or the late Vistulian (13–10 ka BP)? What forms develop as a result of an annual winter re-advance of a cliff? What forms develop at a surge advance? Can bathymetry of the bays in which the glaciers end limit significantly a glacier advance? The paper presents the results of the echo sounders made in the summer seasons of both 2004 and 2005 at the selected glaciers which end in the sea in the Forlandundet region. The measurements included the following glaciers: Aavatsmarkbreen, Dahlbreen, Gaffelbreen, Konowbreen, Osbornebreen and Buchananisenbreen. According to literature and the archival cartographic materials, the changes in the range of the researched glaciers were studied. For the echo soundings at the glaciers an echo sounder correlated with the GPS Map 178C by Garmin was used. For the long profiles of the glaciers the trace method was used, while for the studies of the data the GPS Utility 4.20.4 was used.

Morphogenetic analysis of the subaqual forms of the sea bottoms where the studied glaciers end needs further research. However, their sequence as well as the fact that they correspond with the old ranges of glaciers proves their glacial genesis. Echo sounders' results of the surging Aavatsmarkbreen are also interesting. The paper contains the results of the echo sounders made at Buchananisenbreen, which in the 1930s was a piedmont glacier.

Selected problems of changes in morphometry, bathymetry and thermal conditions in the lake complex at the forefield of Aavatsmarkbreen

Ireneusz Sobota*

Department of Cryology and Polar Research, Institute of Geography, N. Copernicus University, Toruń, Poland

The lakes are located in the marginal zone of Aavatsmarkbreen. These are the following: Upper, Middle and Lower. They all have a connection with the sea and thus show untypical thermal and salinity conditions.

The vertical range of lake water temperatures shows rare thermal conditions. Heat flow in the water mass is mainly dependent on and influenced by characteristic layers of both fresh and salty water, which are the result of the water exchange between the lake and the sea. The water layer of high salinity intensifies heat accumulation, which results in a sudden temperature jump at a certain depth. The highest and most stable water temperature was found at the depth of 4 to 6 metres, irrespective to the thermal changes taking place in the layer above. It posed a barrier to heat coming in from both the layer above and from the lake bottom. The range of water temperatures was similar to the range of electrical conductivity. This means the main cause for shaping thermal phenomena in the lake was salinity. A similar layout of the heat layers in the studies lakes was also recorded by Pietrucień and Skowron (1983).

In summer 2004 (August 26) a spatial measurement of surface diversity of water temperatures was taken. According to the results, the values of temperatures were similar; the highest were recorded at the shore section of the lake as well as at the throat of the Lower Lake.

In summer 2004 GPS, a receiver with the built-in echosounder, was used to take bathymetric measurements.The results were referred to the average water level during summer. Additionally, measurements were taken in order to establish the course of the lakeshore. The bathymetric plan was used to find out that the area of the lakes is similar to the value from the year 1982 (Pietrucień, Skowron 1987) and totals 8.03 ha. Some differences stem from the natural changes in the water reservoirs as well as certain errors connected with the measurement techniques. The average depth of all the discussed water bodies was 2.6 m. The largest differences in depth were recorded in western section of the Upper Lake. Two new deeps were found in northern part of the Middle Lake and at the connection with the Upper Lake.

Morphometric changes of the analysed water bodies result in periodical disappearance of the features characteristic for meromictic lakes. This means thermal conditions of water masses of the lake have a significant influence on both physical parameters and dynamics of water, as well as the changes in its bathymetry, including parameters which describe the lake basin.

* e-mail: irso@geo.uni.torun.pl

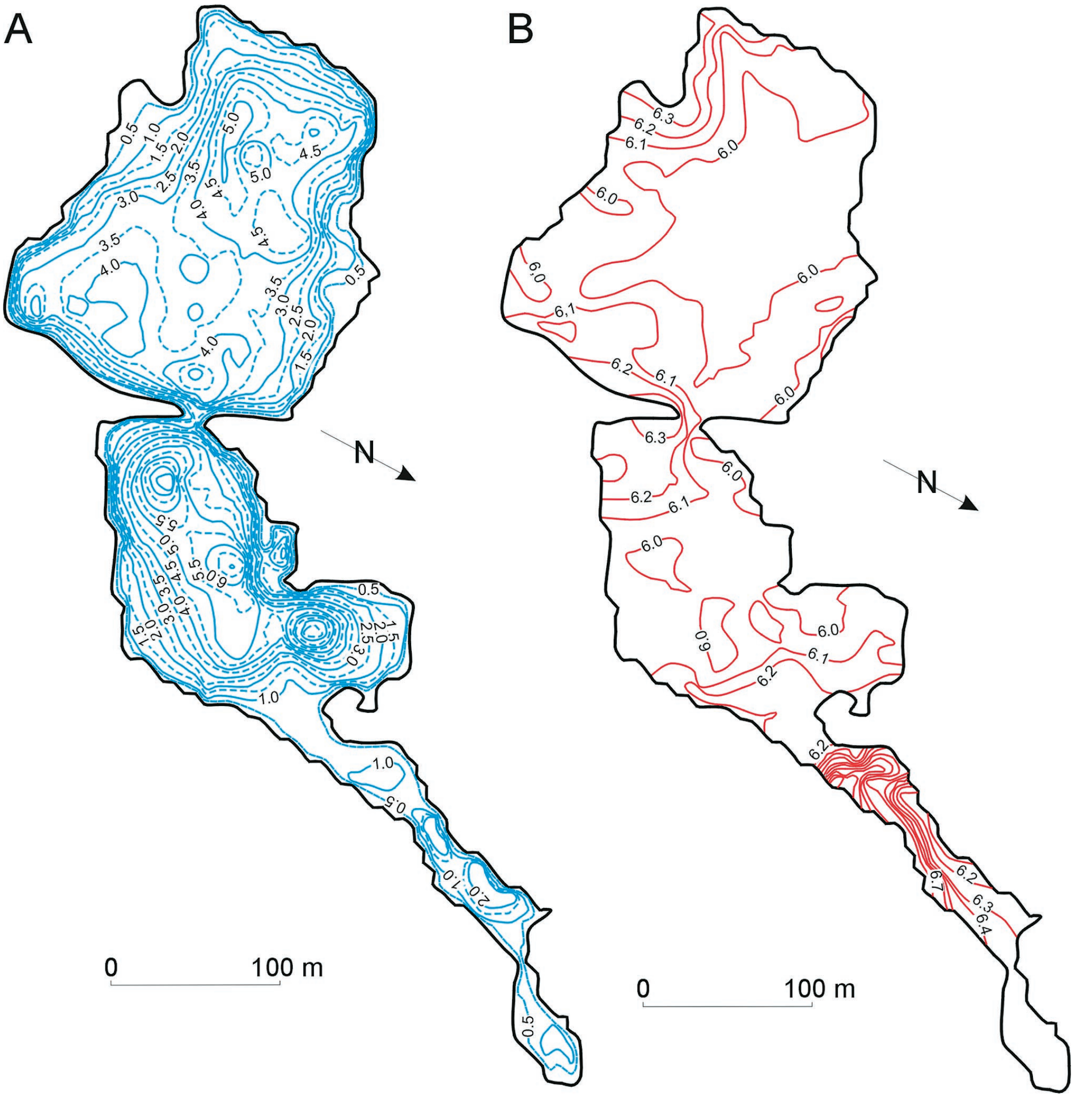


Fig. 1. A – bathymetry (m) and B – surface water temperature (°C) of moraine lakes at the southern forefield of Aavatsmarkbreen

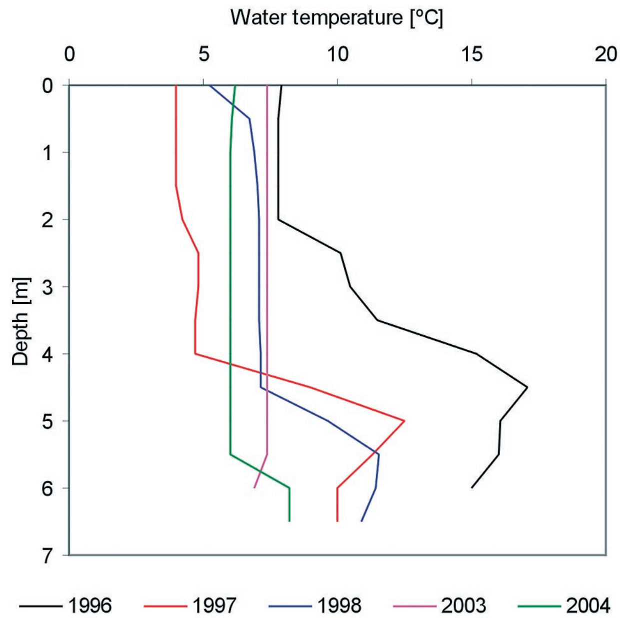


Fig. 2. Water temperature in the Upper Lake in selected years during summer season

Point 9 – Nicolaus Copernicus Polar Station *78° 40' 33" N 11° 49' 36" E*

www.stacja.arktyka.com

Marek Grześ, Ireneusz Sobota*

Department of Cryology and Polar Research, Institute of Geography, N. Copernicus University, Toruń, Poland

The first glaciological expedition to Oscar II Land was organised in 1938 on the initiative of Professor Antoni Bolesław Dobrowolski, the chairman of the Polar Club of the Exploration Expeditions Association. Ludwik Sawicki from the Geological Institute in Warsaw chose the area to be explored. Stefan Bernardzikiewicz, who took part in the 1934 expedition to Torell Land, became the person in charge of the whole expedition. Bronisław Halicki, DSc from the Stefan Batory University in Vilnus and Mieczysław Klimaszewski, DSc from the Jagiellonian University were among other participant of that expedition. They had a big motor boat sailed by a Norwegian trapper Sverre Hansen. The investigations were carried out on the glaciers and their forefields between Eidem Bay and Engels Bay (English Bay), yet predominantly in the Kaffiøyra region (the Coffee Plain). Unfortunately quite a substantial amount of the investigation results vanished during the war. The expedition to Oscar II Land remained forgotten for many years. The first investigation results, Geomorphologic Studies in the West Part of Spitsbergen between Kongs-Fijord and Eidem-Bukta, were published by Professor Mieczysław Klimaszewski only in 1960. Detailed description of glacial phenomena, post-glacial forms and deposits provided an excellent material for conducting comparative studies. In 1975 the Geography Institutes of both Nicolas Copernicus University and the Polish Academy of Sciences together with Geography Students' Scientific Society attempted to perform these investigations. Under the supervision of Professor Jan Szupryczyński a group of 12 people set off to Spitsbergen. It consisted of two groups: geomorphological and hydrological. The hydrological group of five persons brought a wooden house in separate elements. Professor Czesław Pietrucień was responsible for its design and constructions. The house was set up at the the foot of the end moraines of the Aavatsmark Glacier, in the north part of the Kaffiøyra, at latitude 78° 40'33''N and longitude 11° 49'36''E.

The Nicolas Copernicus University Polar Station is situated beyond the borders of the protected areas (parks). It allows a greater freedom for the exploration of the neigh bourning regions. During the summer there is not any blockade phenomenon of the Forland Strait due to ice pack. The straits is not covered with ice as early as at the end of June. It is very important while planning a journey. The neighbourning Ny Alesund with an internaltional exploration centre and the airfield (two flight a week) put the polar station in a favourable light. It takes 2–3 hours to cover the distance from Ny Alesund to the station by boat. In winter it takes almost the same time by scooters.

In July and August there is regular navigation traffic though the Forland Strait from Longyearbyen to Ny Alesund (once a week). Ship unloading is easy due to the sand shores near the polar station. Deep Hornbaek Bay allows even those bigger sailing units to take shelter against heavy storms.

Nicolaus Copernicus University started to take part in polar research in 1975 using its own polar station located on north-western Spitsbergen. The position of this station was chosen because of its big scientific value. People often ask us why do we do a research on polar territories, Spitsbergen in particular. The answer is that glaciers are almost ideal "climate thermometers". It refers especially to their range changes. They became matter of research being conducted by our Institute of Geography – just because they cover almost 60% of Spitsbergen. To understand post-glacial reliet of the earth's surface

* e-mail: irso@geo.uni.torun.pl

in Poland it is very useful to observe contemporary sediments and glacier forms. In such way Spitsbergen became a natural laboratory for geographers of various specialities.

Nicolaus Copernicus University Polar Station is northenmost polish scientific institution. It is situated on northern part of the Kaffiyora, close to the Aavatsmarkbreen. This station was used by 30 expeditions and 100 people so far. Effects of these expeditions are shown in 350 publications and on topographic and thematic maps.

In 1995 we started to do a sistematic study of mass balance of Waldemarbreen, and next in 2001 Irenebreen and in 2005 Elisebreen. These studies are part of the international programms and projects (WGMS, CALM).

NCU Polar Station is suitable to whole year work. It has three independent sources of energy (fuel engine, wind power station and sun battery). Means of transport are: fibre glass, rubber motor boats and snow scooters. Radio communication is ensured by FM radiostation with its call signal LH3MB.

In 30 year existence the station was visited by about 400 people: 150 Poles, 120 Norwegians and Germans, Dutch, Russians, Americans and even Australians. All other information about our station you can find on the internet: www.stacja.arktyka.com.

Fig. 1. Nicolaus Copernicus Polar Station (photo I. Sobota)

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Geomorphology of the southern side of Bellsund – Leader Piotr Zagórski

Kazimierz Pękala, Janina Repelewska-Pękalowa

Department of Geomorphology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland

Since 1986 the Bellsund Region (NW part of Wedel Jarlsberg Land), has been explored by the members of Polar Expeditions organised by M.C. Skłodowska University in Lublin. The main base was located in Calypsobyen, the western coast of the Recherche Fiord.

Within the programme of Expeditions some interdisciplinary researches of polar environment have been done. Among them there are Earth Science (geomorphology, geology, meteorology, soil science, environment protection) and Biology (botany, biochemistry) and radiochemistry. As the reflection a lot of scientists of various fields have been present in Expeditions.

The interest was the relief, cover formations and paleogeography of Pleistocene, the functioning of glacial and periglacial geoecosystems in local and global conditions of climate changes and the influence of anthropogenesis. The introduction of the latest computer technology and method of positioning allow making cartographical view of the relief.

The effect of 18th Expeditions has been numerous publications in national and international magazines, as the examples below confirm:

– Zalewski M.S. (ed.) 2000. *Bibliography of Polish Research in Spitsbergen Archipelago 1930–1996, part I*, Publications of the Institute of Geophysics Polish Academy of Sciences, Warszawa;
– Zagórski P., 1998. *Spitsbergen Bibliography: Geomorphology, Glaciology and Quaternary Geology*. IV Conference of Polish Geomorphologists II, Spitsbergen Geographical Expeditions, (ed.) J. Repelewska-Pękalowa, Wyd. UMCS, Lublin, 291–314.

The results of the studies were presented in many conferences and national sessions as well as internationally, for example International Conferences on Permafrost: Trondheim (1988), Beijing (1993), Zurich (2003), and in conferences: Frankfurt/Main (1989), Mainz (1992) and Potsdam (2005).

Introduction to guide

Piotr Zagórski*

Department of Geomorphology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland

The presented area covers western borders of Recherche Fiord from the spit in Josephbukta to Skilvika (Fig. 1). The main elements are there: extensive plain (Calypsostranda) made by the system of raised marine terraces and the forefield of Renard and Scott Glaciers. The whole makes unique and picturesque tundra landscape, extremely interesting from cognitive and scientific points of view.

The aim of first two points of the terrain session (points: 1, 2/2A) is to show the evolution of marginal zone and stages of fluctuations (advance and recession) of Renard Glacier and its influence on transformation of the shore on the base of geomorphological and archaeological studies. At the next point (3) the issues of periglacial phenomena and monitoring of dynamics of active layer of permafrost are going to be shown. A break and a short rest will be expected at Polar Station of M.C. Skłodowska University in Calypsobyen (point 4). It will also be a chance to acquaint with a history of that place, its present function and scientific programmes. The point 5 is connected with glacial issues of the Scott Glacier, which is much smaller than the Renard Glacier. At the last two points of the terrain session, it is expected to be presented the issues related to Late Weichselian and Holocene morphogenesis of Calypsostranda (point 6), with the special attention paid on conversion of shore zone at the historical time and present (point 6A).

* e-mail: pzagorsk@biotop.umcs.lublin.pl

Calypso – excursion programme

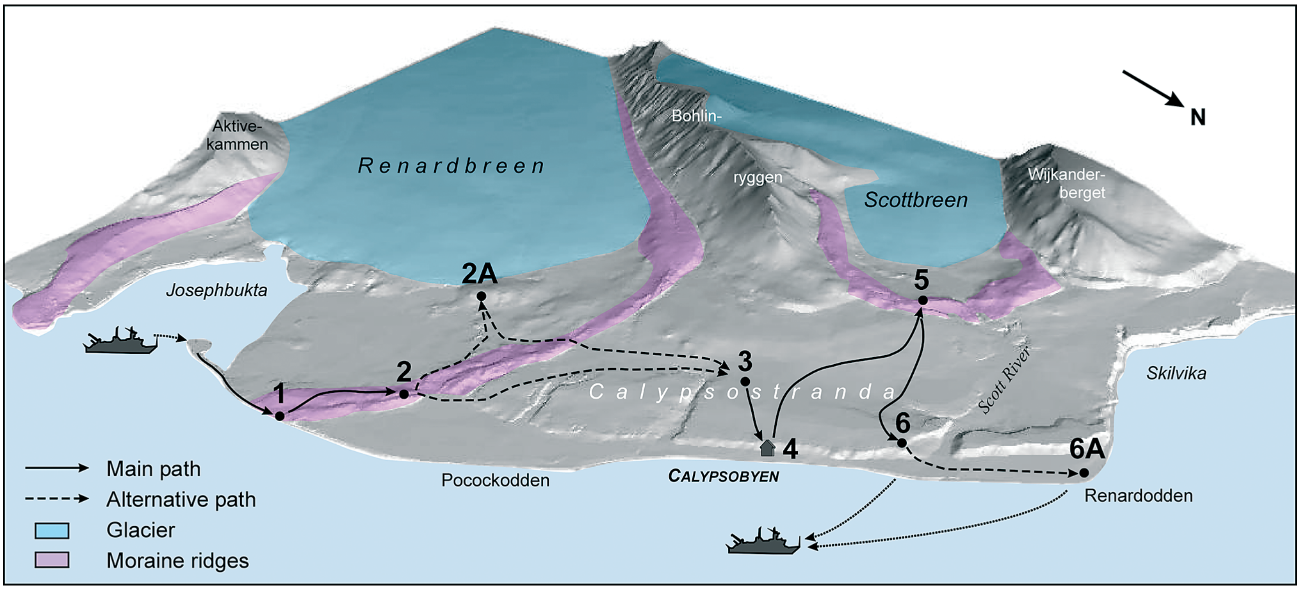


Fig. 1. The landing point, the passage path and the location of points. 3D model of the Calypsostranda Region (Zagórski 2002)

Point 1 – THE FRONTAL MORAINE OF THE RENARD GLACIER 77° 32' 22" N, 14° 34' 06" E
Piotr Zagórski, Kazimierz Pękala, Janina Repelewska-Pękalowa – The role of the Renard Glacier in forming of shore zone

Point 2/Point 2A – FOREFIELD OF THE RENARD GLACIER
2 – 77° 32' 37" N, 14° 32' 41" E; 2A – 77° 32' 23" N, 14° 29' 47" E
Jan Reder, Piotr Zagórski – Recession and development of marginal zone of the Renard Glacier

Point 3 – PERIGLACIAL POLYGON 77° 33' 20" N, 14° 29' 52" E
Kazimierz Pękala, Janina Repelewska-Pękalowa – Dynamics of active layer of permafrost

Point 4 – CALYPSOBYEN 77° 33' 31" N, 14° 31' 01" E
Kazimierz Pękala, Janina Repelewska-Pękalowa – Calypsobyen - history and the present day

Point 5 – PUSH MORAINE OF THE SCOTT GLACIER 77° 33' 36" N, 14° 26' 11" E
Jan Reder, Piotr Zagórski – Recession and development of marginal zone of the Scott Glacier

Point 6 – CALYPSOSTRANDA 77° 33' 55" N, 14° 29' 41" E
Piotr Zagórski – Relief and development of Calypsostranda

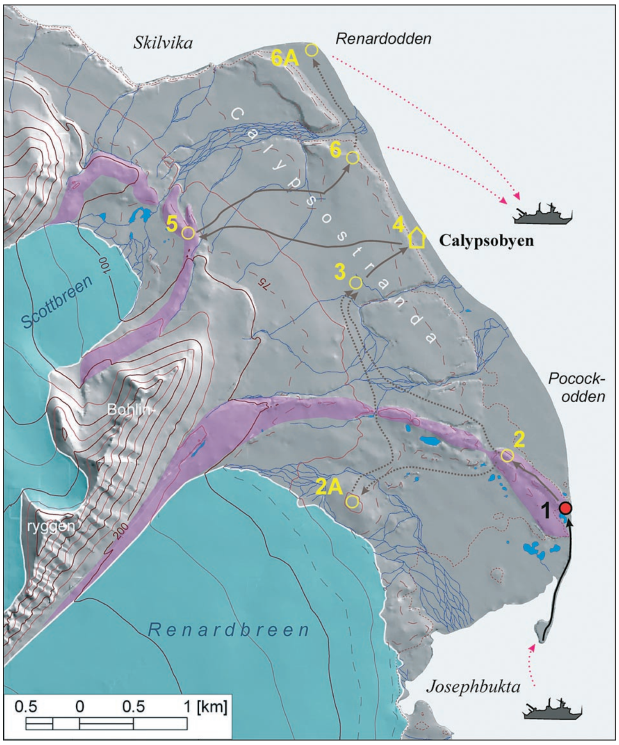
Point 6A – RENARDODDEN 77° 34' 21" N, 14° 28' 49" E
Piotr Zagórski – Present morphogenesis of the shore and the importance of archaeological sites for reconstructing the stages of development

Point 1 – The frontal moraine of the Renard Glacier *77° 32' 22" N, 14° 34' 06" E*

The role of the Renard Glacier in forming of shore zone

Piotr Zagórski*, Kazimierz Pękala, Janina Repelewska-Pękalowa

Department of Geomorphology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland



The coast of NW part of the Wedel Jarlsberg Land is exposed to various morphogenetical factors. One of the most important can be numbered among glaciers that can influence directly (destruction and transform of existing forms of relief, accumulation of moraine covers) and indirectly (with the cooperation of different factors: tectonic, fluvial, marine).

The present relief shown at point 1 was shaped fundamentally at the end of XIX and at the beginning of XX century, but the ridge of frontal moraine is built of some moraine layers of different age, that show the advance of glacier of the surge type during Holocene (Pękala, Repelewska-Pękalowa 1990, Reder 1996) (Fig. 2). The direct influence of the Renard Glacier, correlated with the advance during the Little Ice Age, caused for example redeposition of sediments and fossil flora which was dated on 660 ± 80, 1 040 ± 80 and 1 130 ± 80 BP with ^{14}C method (Dzierżek et al. 1990) (Fig. 2). Those layers are disturbed glaciotectonically and contain some fragments of woollen fabric, whalebone, animal's bones and wood – archaeological site Renardbreen 1 (Krawczyk, Reder 1989, Jasinski, Starkov 1993, Jasinski 1994). Furthermore, under the moraine, there were found some fragments of buildings from XVI century, which constitution remained intact by glacier, they were 20 cm under present sea level (Fig. 3, 4, 5). This site was studied in 1986–1993 and it is the only one in Spitsbergen where the leftovers of whale fishing buildings were covered with till. It allows us to date the activity of glaciers and changes of sea level at historical time. The terrace I was also aggradated, and the marine materials of fossil storm ridge were dated on 6.2 ± 0.9 ka BP with TL method (Pękala, Repelewska-Pękalowa 1990) (Fig. 2).

The decisive role in forming of a section of accumulative shore located on the south of abrasively cut frontal moraine of the Renard Glacier plays longshore currents (Fig. 6). At the region of the Pocockodden, there are distinguished two longshore currents; one flows northwest and the other south (Harasimiuk, Jezierski 1988, 1991). The other one is supplied with the material from conversion of sandur fans and influences the origin and remodelling of the spit developing in the shade of shore ledge – moraine ridge of the Renard Glacier (Fig. 6). Its development was also enabled in the presence of glacial sediments of marginal zone of the Renard Glacier at that part of the shore. The shape and geometry of widen, final

* e-mail: pzagorsk@biotop.umcs.lublin.pl

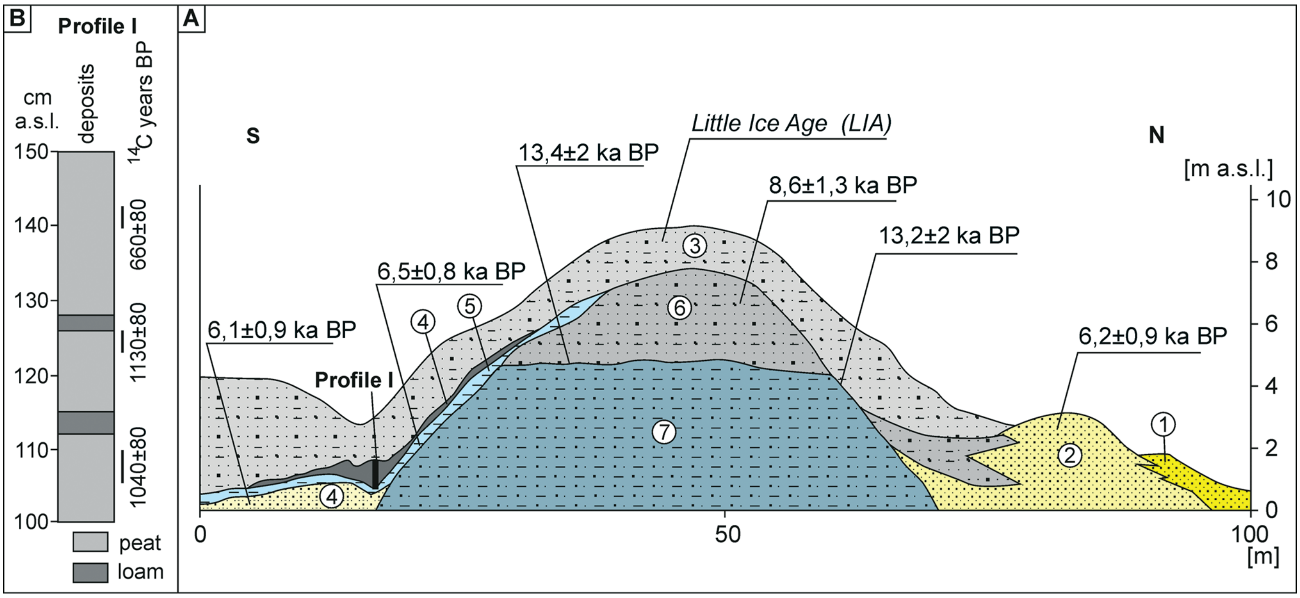


Fig. 2. The geological structure of frontal moraine of the Renard Glacier (Pękala, Repelewska-Pękalowa 1990)
A: 1 – present storm ridge 2 – fossil storm ridge, 3 – glacial till (Little Ice Age), 4 – pushed occupation level of whale settlement with fossil flora (profile 1), 5 – clay, 6 – glacial till, 7 – glacio-marine sediments; B – profile of organic sediments of the Renardbreen site (Dzierżek et al. 1990).

Fig. 3. Archaeological works at Renardbreen site 1. (photo Kazimierz Pękala, 1991)

Fig. 4. The occupation layer still visible in the northern margin of the trench (photo Kazimierz Pękala, 1991)

part of spit was and is still changing quickly. It is supported by the analysis of available cartographical materials and GPS measurements (Zagórski 2007) (Fig. 6).

Indirect role of the Renard Glacier in remodelling the shore with the help of fluvial and marine processes has been appeared fully in the section between Pocockodden and ridges of the frontal moraine of the Renard Glacier (Fig. 6). At the time of maximum range of the Renard Glacier at the Little Ice Age, the glacier waters caused the origin of gorge in the mouth where plain fluvioglacial sandur fans were developed that aggradated terrace I. Thanks to that, slightly slanting area of semi-circular outline was arisen. It is closed in the shore zone by the storm ridge. The origin of such a form shows clearly considerable advantage of fluvioglacial accumulation over the possibilities of spreading the material by waving and longshore current. Broad surfaces of fluvioglacial cones, after the recession of the Renard Glacier from the push moraine lines, became the fossil forms. Disappearance of delivery of the terrestrial material caused the increase of activity of marine processes that as an effect made gravel ridge that brought to a stop the destruction of the cone.

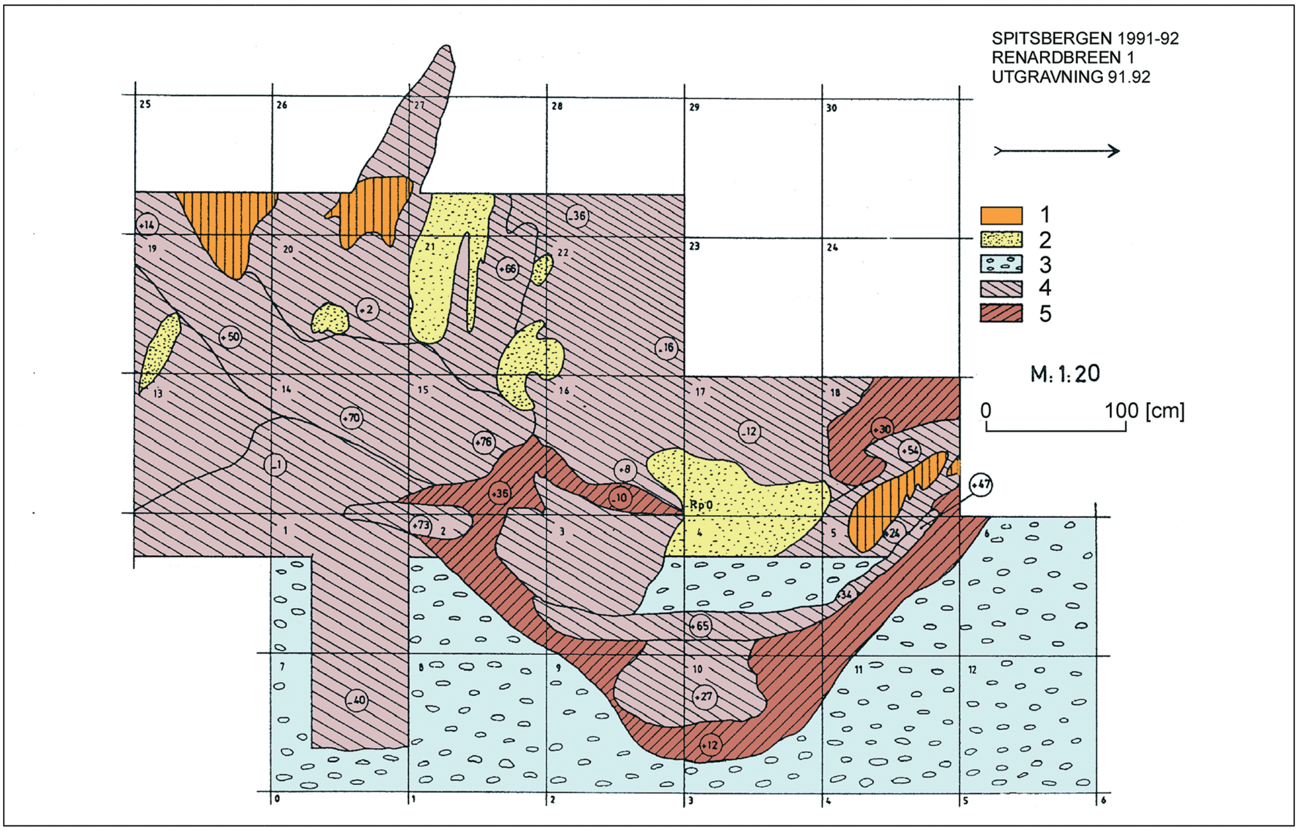


Fig. 5. The archaeological site Renardbreen 1. Excavations 1991–1992
1 – marine sediments, 2 – sand, 3 – marine gravel, 4 – brown/black occupation layer, 5 – the wall-like construction (after Jasinski, Starkov 1993)

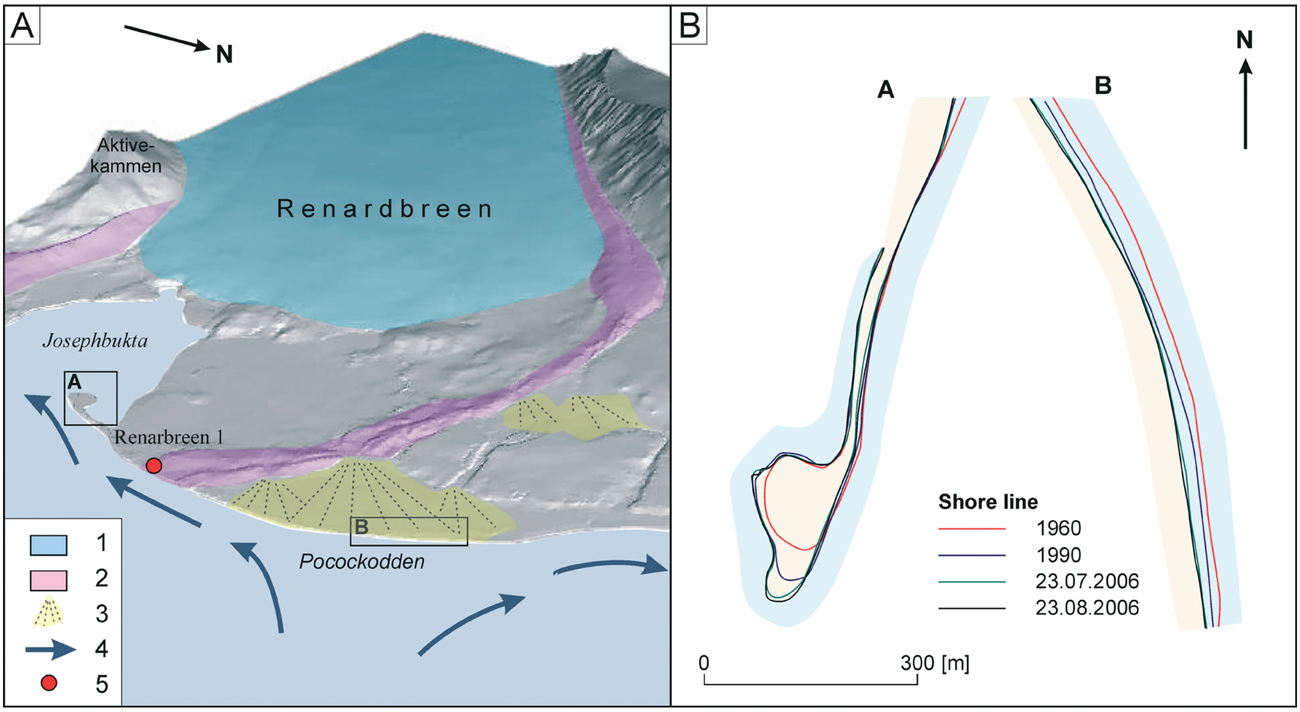
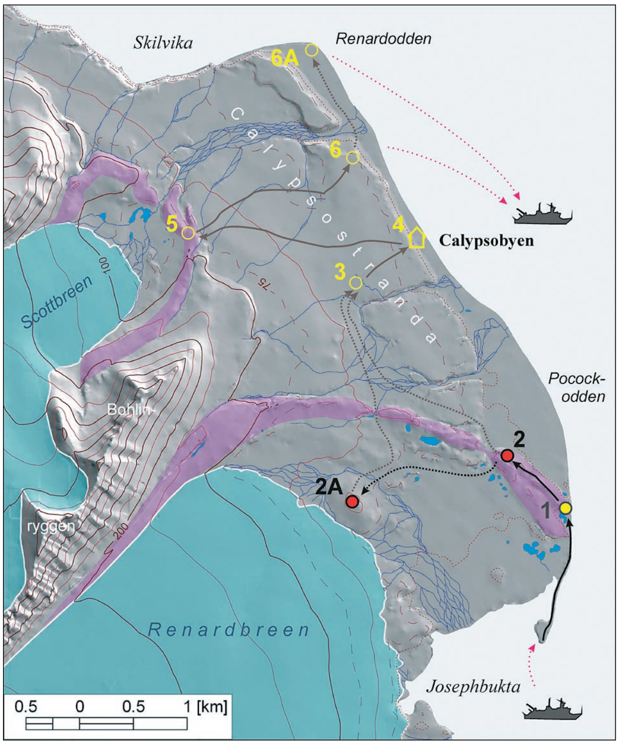


Fig. 6. A – Main factors that influence the shape of the shore in the section from Pocockodden to Josephbukta
1 – glacier surface, from 1990, 2 – frontal moraine ridge, 3 – extramarginal sandur fans, 4 – directions of the longshore currents (after Harasimiuk, Jezierski 1988, 1991), 5 – location of archaeological site Renardbreen 1. B – Changes of geometry of the shoreline made on the basis of analysis of cartographical materials and GPS measurement (Zagórski 2007).

Point 2/Point 2A – Forefield of the Renard Glacier *2 – 77° 32’ 37" N, 14° 32’ 41" E*
2A – 77° 32’ 23" N, 14° 29’ 47" E

Recession and development of marginal zone of the Renard Glacier

Jan Reder, Piotr Zagórski*
Department of Geomorphology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland



The Renard Glacier, the biggest in the NW part of Wedel Jarlsberg Land region; its area in 2006 was a little over 31 km^2. Its length in axis was about 8.3 km, the width is various, from 2.5 km at the lower part, 7–8 km at the central firn field and its side arms. Tongue of the Renard Glacier covers the valley limited from NW by the Bohlinryggen and Activekammen from SE (Fig. 7, 8).

The largest size of the Renard Glacier was during its maximum spread at the end of XIX century when the glacier front was staying on the line of frontal moraine range and finally formed during the Little Ice Age (Fig. 9). Then the glacier filled the whole area of Josephbukta and its area was 38 km^2. Till 1936 on that area, there was no major change. The glacier was still filling the whole area to the inner side of moraine range. The part escaping right into the fiord underwent the significant recession of nearly 1000 m and exposed a considerable part of the Josephbukta (Fig. 9). In the following period of 1936–1960 the much quicker recession began especially in the land part without direct contact with fiord water (Reder 1996). That recession occurred by frontal receding of the glacier front of 780 m (33 m a^{-1}), and on the southern side of the bay – 1200 m (50 m a^{-1}). Also the receding of 560 m (23 m a^{-1}) was present in the Josephbukta revealing almost all of it. Between 1936 and 1960 the direction of the proglacial water outflow changed. Till that time active outside wide sandur fans became dead and the outflow made directly for Josephbukta (Harasimiuk 1987, Reder 1996, Zagórski 2004).

In the following years, till 1990 the quicker recession of the glacier front underwent mainly land part of maximally 720 m (24 m a^{-1}), while much slower was the recession of the part connected with the bay mouth – maximum up to 450 m (15 m a^{-1}). The Renard Glacier had a mouth to the fiord in the Josephbukta and its front made some metres high ice cliff (Fig. 7, 9). Now the deglaciation of the Renard Glacier has generally a frontal character. Based on observations and GPS measurements from 1990–2006 the glacier front receded of maximum almost 340 m (21 m a^{-1}). Starting from the end of XIX century till 2006, area of 7 km^2 was exposed

* e-mail: pzagorsk@biotop.umcs.lublin.pl

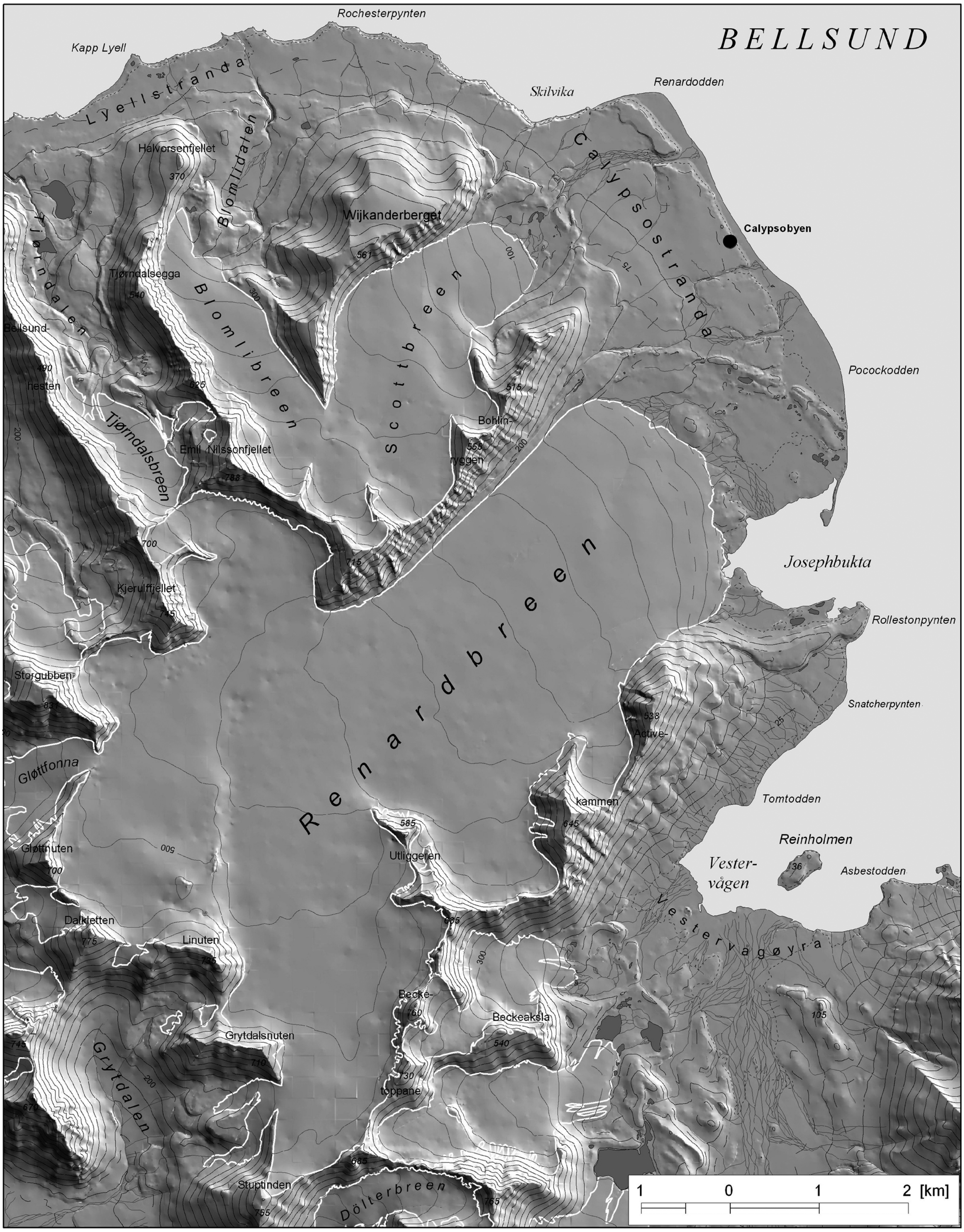


Fig. 7. The Renard Glacier and Calypsostranda Region. The shade map made using the Digital Terrain Model (DTM) obtained from the aerial photos from 1990 (Zagórski 2002)

where 1.5 km^2 was Josephbukta. It has its consequences in the origin and formation of the surface of the forefield of the Renard Glacier limited by frontal moraine ridges (Fig. 9).

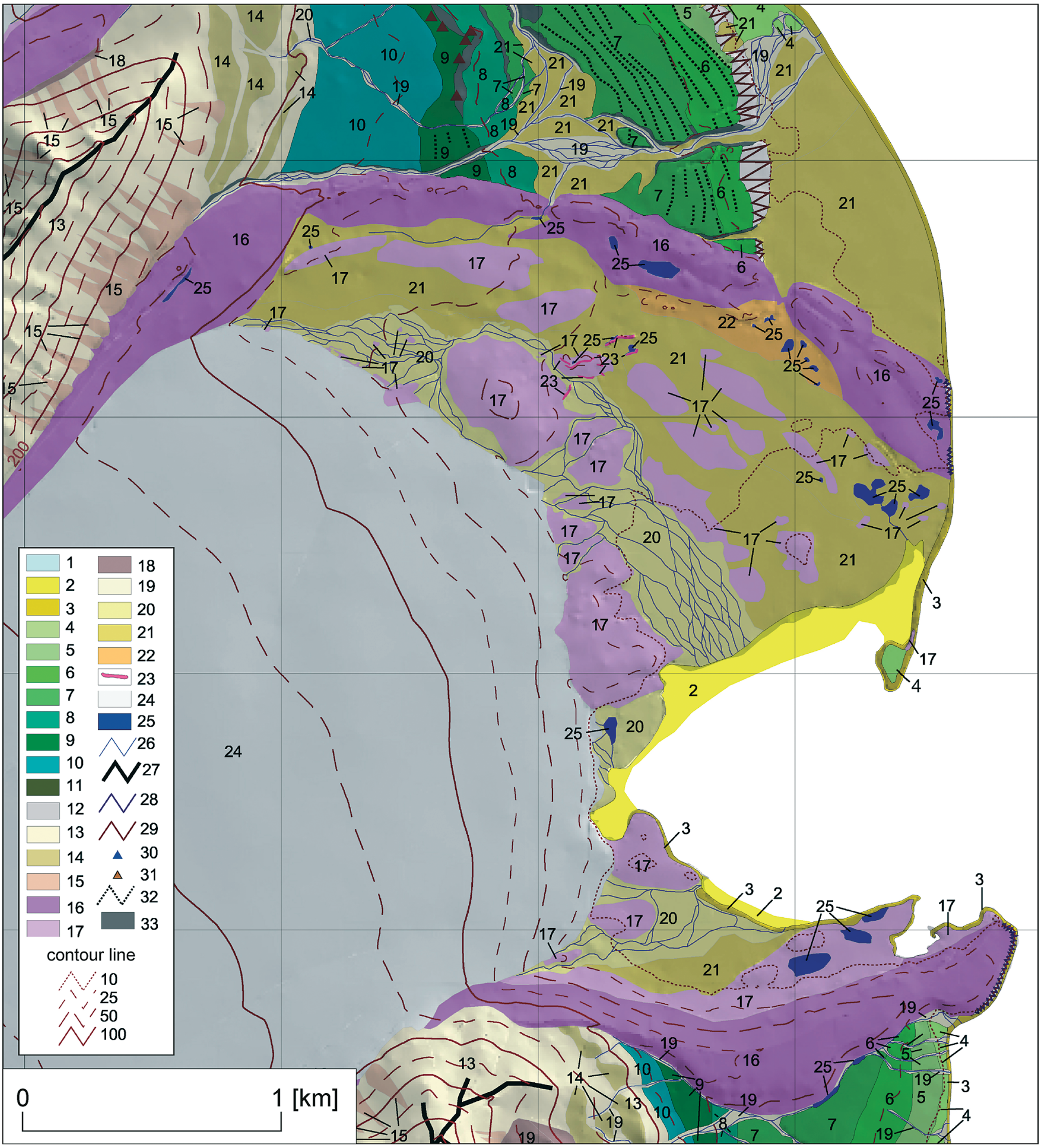


Fig. 8. Geomorphological map of the forefield of the Renard Glacier (Zagórski 2002)
1 – contemporary abrasion platform, 2 – tidal flat, cone of delta, 3 – contemporary storm ridge, 4 – terrace I (2–8 m), 5 – terrace II (10–20 m), 6 – terrace III (25–30 m), 7 – terrace IV (30–40 m), 8-terrace V (40–50 m), 9 – terrace VI (50–65 m), 10 – terrace VII (70–85 m), 11 – terrace VIII (105–120 m), 12 – superficial flattening, 13 – slopes, 14 – denudation-structure level, 15 – talus cones, 16 – ice-cored moraine ridges, push and lateral moraines, 17 – ground and ablation moraines, 18 – rock glaciers (nival), 19 – floors of pronival valleys, 20 – contemporary sandur plains and fans, alluvial cones, 22 – kame, 23 – esker, 24 – glaciers, 25 – lakes, 26 – rivers, 27 – ridges, 28 – active marine cliffs, 29 – dead marine cliffs, 30 – skerries, 31 – paleoskerries, 32 – old storm ridges, 33 – edges.

The frontal moraine of Renard Glacier consists of two genetically and age-old distinct parts: inside of push moraine character and inside neighbouring ice-cored moraine ridges (Fig. 2, 8). The push moraine at N and NW part of the forefield has the surface of mild character, slopes are mild and the tops are not marked sharply. The more varied is southern area with that is only fragmentally preserved part of the moraine. Its surface is characterised by very intensive line of relief in the shape of longitudinal parallel swellings and lowerings. Similar morphological features show frontal moraines of glaciers accumu-

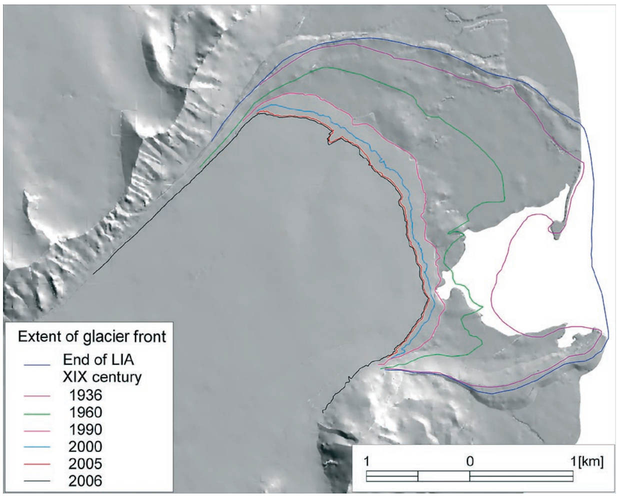


Fig. 9. The extent of the Renard Glacier fronts combined on the basis of archival data (Reder 1996, Szczęsny et al. 1989, Zagórski 2005) and GPS measurement

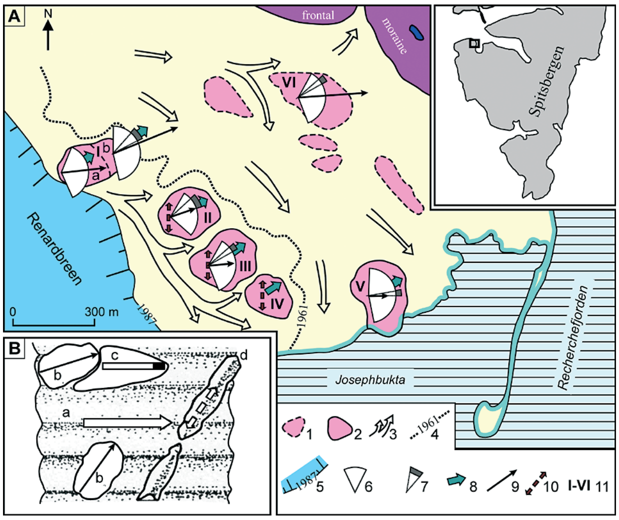


Fig. 10. The sketch of the forefield of the Renard Glacier (Merta 1988)
1 – patches of erosive moraine of compact texture of "fluted" type, 2 – patches of fresh relief of "fluted type", 3 – directions of outflow of proglacial water, 4 – location of the glacier front in 1961, 5 – location of the glacier front in the study season, 6 – the range of orientation of the longer axis of free stones (type b2), 7 – the range of orientation of extension of moraine accumulates of type c, 8 – directions of setting of moraine ridges and grooves of the fluted type, 9 – resultant factor of orientation of the longer axis of stones (type b1), 10 – location of uncompleted ridges, 11 – measurement domains I–VI, B: scheme of location of respective types of directional elements, their symbolic and the way of measurement: a – ridges and grooves, b1 – stones with the sediment at their hinterland, b2 – free stones, c – moraine deposits in the shade of stones b1, d – uncompleted ridges.

lated in the conditions of strong compression, so at the surge stage. At the area of ice-cored moraine ridges, even huge denivelations can be seen. Sharp, pyramidal tops and considerable number of cracks and lowerings of thermokarst character (often filled with water) prove the existence occurrence of relict ice inside (Reder 1996).

At the first stage of the recession of the glacier the outflow from moraine ridges was blocked and at its internal side some marginal troughs begin their kelter. Ablation water was taking them to Josephbukta direction (it was parallel to the glacier front). With the growing distance from the glacier tongue in the SW direction, in the lowering between its edge and ice-cored moraine ridges some intensive accumulation processes of the material carried by ablation water began to happen. Then, the kame terrace was formed made of sandy deposits with some gravel infillings and ablation till (Fig. 8). The total obstruction of the outflow in northern direction through the frontal moraine and kame terraces, caused creation of the marginal river, flowing along glacier front in direction of Josephbukta. On the hinterland of moraine series and kame terrace the ablation waters have cut down the deep valley, which present dry floor is covered with sandy-gravel sediments. As the result of progressive recession the set of ground moraine of fluted type was made. That set is on the outcrop of bedrocks of roche moutonnée type (Merta 1988, Reder 1996) (Fig. 10, 11).

The inside set of marginal sandurs consists of three layers correspond with the stages of recession of the glacier. Two upper layers, not active now, compose the forms of the shelf type or terraces connected to the inner slopes of ice-cored moraine ridges. Single packs of sediments that belong to the upper system of cones are universally met at lowerings of the fluted moraine. The third, contemporary sandur layer is made of the series of cones that are in the lowerings between roche moutonnée on the hinterland of the glacier tongue edge. The surface of that sandur is formed by proglacial water of marginal rivers (Fig. 8). At the direct neighbourhood of tongue they have concentrated confluence, huge fall and considerable erosive abilities. Due to a progressive recession of the glacier causes the marginal rivers to move towards the glacier front that receding every year. The traces of older flows recorded as dead, hung riverbeds which location can reconstruct the advance of the glacier front with high probability (Fig. 8).

During the last thirty years the large island mountains of roche moutonnée character were unveiled from the ice, as well as moraine cover of fluted type, which was on. The glacier gradually recedes towards West lost the contact with the water of Josephbukta (Fig. 8, 9, 10). At the direct forefield of the glaciers, between the taking back tongues and frontal moraines (ice-cored moraine ridges) that mark the maximum extend of the last transgression, there were created the zones of ground and ablation moraines, similar to drumlins forms, inner sandurs and sometimes concurrent crevasse forms. Ground moraines, often fully developed as the moraine of the fluted type, stay mainly on the roche moutonnée (Merta

Fig. 11. The fluted moraine covers the proximal slope of roche moutonnée (photo Piotr Zagórski 2006)

Fig. 12. Covers of naledi and the lump of dead ice buried in the sandur sediments (photo Piotr Zagórski 2006)

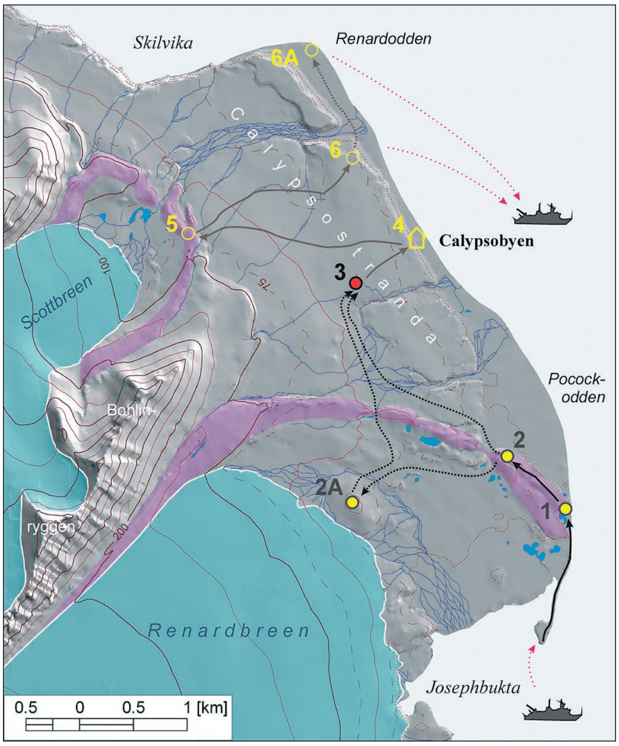
1988, Reder 1996) (Fig. 11). Proglacial rivers that aggrade and cut sandur surface use the lowerings between them. On the distal side of roche moutonnée, from time to time the ridges of eskers are preserved that were formed in the middle of XX century and their orientation correspond to the direction of crevasses on the glacier and the directions of grooves of the ground moraines on its forefield (Fig. 8). Hypsometric domination of moraine ridges was softened by neighbouring from inside kame terraces. In that zone big morphological importance has universally appeared vast covers of naledi and rarely present clods of dead ice buried in sandur sediments (Fig 12).

Point 3 – Periglacial polygon *Localization: 77° 33’ 20" N, 14° 29’ 52" E*

Dynamics of active layer of permafrost

Kazimierz Pękala, Janina Repelewska-Pękalowa* – presented by Piotr Zagórski

Department of Geomorphology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland



The fragment of Calypsostranda that was formed by cryogenic processes connected with frost segregation in different moisture conditions is characterised by the presence of structure soil (garland terraces, stone circle) of different size, shape and present active processes. That area is the polygon of periglacial study and monitoring of active layer of permafrost (Fig. 13, Table 1).

Within the confines of scientific programme of polar expeditions of M.C. Skłodowska University, during almost twenty seasons (1986–2005) the measurements of thickness of active layer were conducted. The main study polygon was Calypsostranda, the moraine plain located in the neighbour of Renard and Scott Glaciers (Fig. 13). The thickness of active layer of permafrost was stated with the use of the method of sounding with the metal rod and some Danilin’s frostmeters were used, too. The measurement point’s representative for tundra survey were located in various places of different degree of water mobility in covers, flora cover, inclination and exposition. They were on the surface of raised marine terrace of the height to 20–40 m a.s.l., and on the slopes of valleys cutting that terrace and on inclinated surfaces of dead cliff transformed by periglacial processes (Fig. 13). The maximum of Summer ground thaw were diverse (Table 1).

The maximum sizes of thawing were noticed at the point with movable water in covers (225 cm) while minimum – at the peat island (45 cm). For the inclined surface it was stated that except for obvious thermal privilege of the south exposed slope, also warming up was influenced by winds of foehn type which effect touched the slope III (S exposition). The speed of thaw was diverse, at the range from 0.25 to 6.0 cm per 24h. The biggest – at the first stage.

The studies on Calypsostranda show that diverse amounts of Summer thaw of the ground have also some local factors, like foehn phenomenon, mobility of non-permafrost water, flora, exposition and snow cover (Repelewska-Pękalowa et al. 1988).

The data from Calypsostranda area are included into International Monitoring System of permafrost active layer: CALM (*Circumpolar Active Layer Monitoring – Site P1 Calypsostranda*) and can be found in the database of *National Snow and Ice Data Center,*

* e-mail: geomorf@biotop.umcs.lublin.pl

Table 1. Maximum thickness of active layer in Calypsostranda in chosen points (in cm)

Point / Year	1	2	3	4	5	I	II	III	IV
1986	90	125	120	–	60	130	–	145	122
1987	111	175	175	175	68	124	150	165	130
1988	108	163	168	193	70	121	180	177	135
1989	145	165	157	180	83	135	160	186	139
1990	130	165	165	165	56	118	135	170	122
1991	127	148	163	170	75	141	150	165	121
1992	140	170	165	180	70	140	180	155	125
1993	112	180	180	196	70	130	180	180	140
1995	125	176	180	174	68	135	170	160	160
1996	125	154	178	168	65	132	160	151	128
1998	130	124	121	170	75	–	–	160	–
2000	108	175	155	130	45	126	135	160	150
2001	116	131	180	165	73	150	170	132	155
2002	130	155	170	154	81	139	160	150	143
2005	150	225	220	210	115	157	195	200	145

Points 1–5 along the NS and WE transects WE: 1 – flat marine terrace (sands and gravels, dry tundra), 2 – structure soils with movable water, sandy-gravel cover, moss on the peat surface, 3 – and 4 – patterned ground with movable water in covers, sands and gavels, without flora, 5 – peat island on little water basin.
Slopes: I – N exposition, I – S exposition, III – E exposition, IV – W exposition.

Boulder, Colorado (Repelewska-Pękalowa 2002, Repelewska-Pękalowa, Pękala 2003, Christiansen et al. 2003) (Fig. 14). The aim of CALM programme is to collect and share data which document the process of Summer thaw of the ground in zones of occurrence of permafrost on both hemispheres. The measurements are done in 117 areas and 15 countries are involved. Only two areas, not long ago did represent Spitsbergen: Kapp Linnee (S1) and Calypsostranda (P1). In 2000 the measurements were begun in Longyearbyen and Ny Ålesund, and very recently site P2 (Kaffiøyra). The CALM programme is designed for observation the reaction of active layer of permafrost to climate changes and by the decision of IPA it will be executed within the confines of projects of International Polar Year 2007–2008.

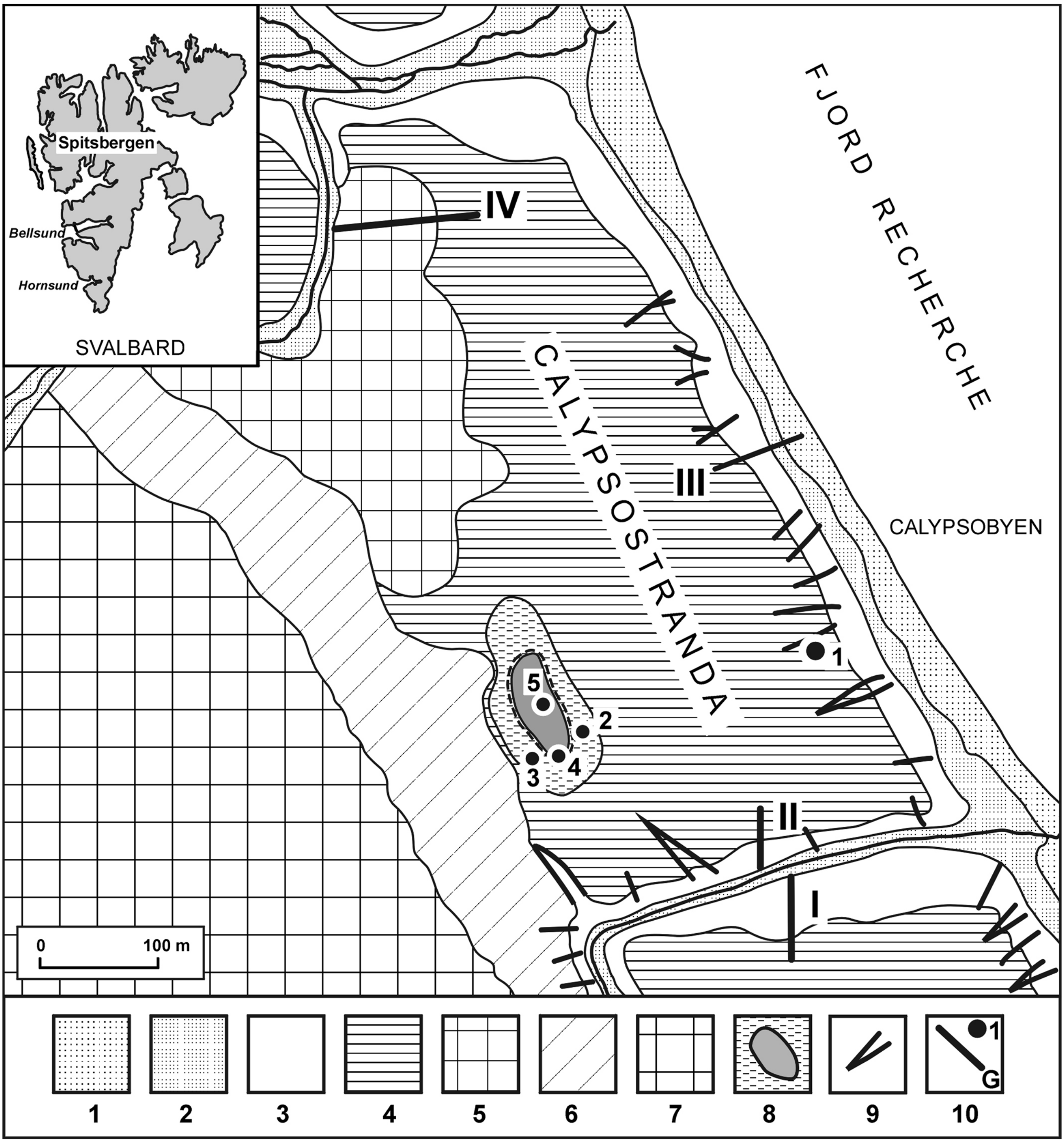


Fig. 13. Main sets of forms and localization of measurement points of active layer of permafrost (Repelewska-Pękalowa, Pękala, 2003)
1 – beach, 2 – floors of valleys and zones of alluvial cones at the cliff base, 3 – cliff and erosive edges of valleys, 4 – dry surfaces of marine terraces, 5 – zones of active solifluction, 6 – periodically wet terraces aggradated with alluvial cones, 7 – slopes and high marine terraces converted by weathering, cryoplanation and erosive processes, 8 – seasonal lake, 9 – erosive dissection, 10 – measurement points.

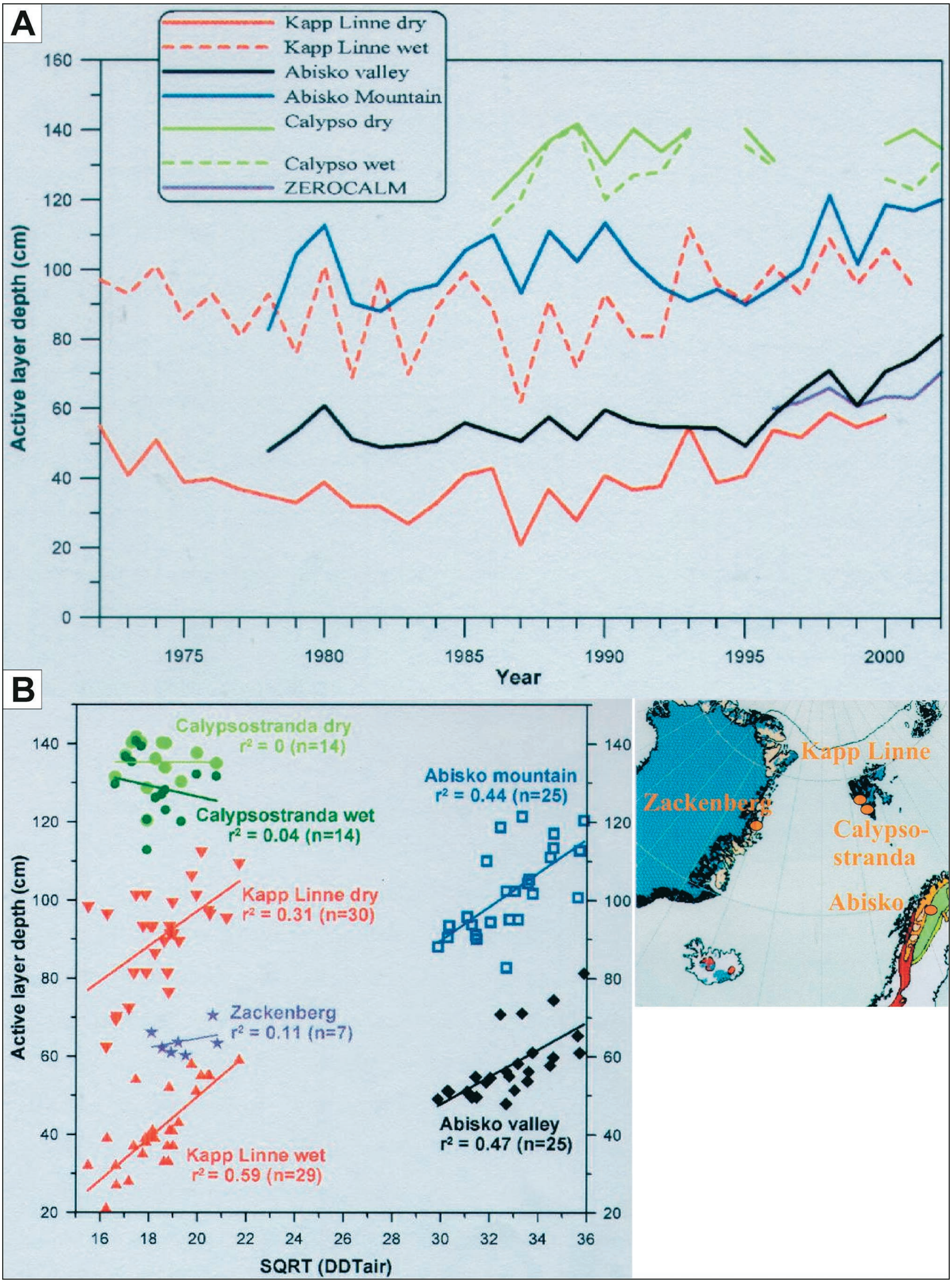


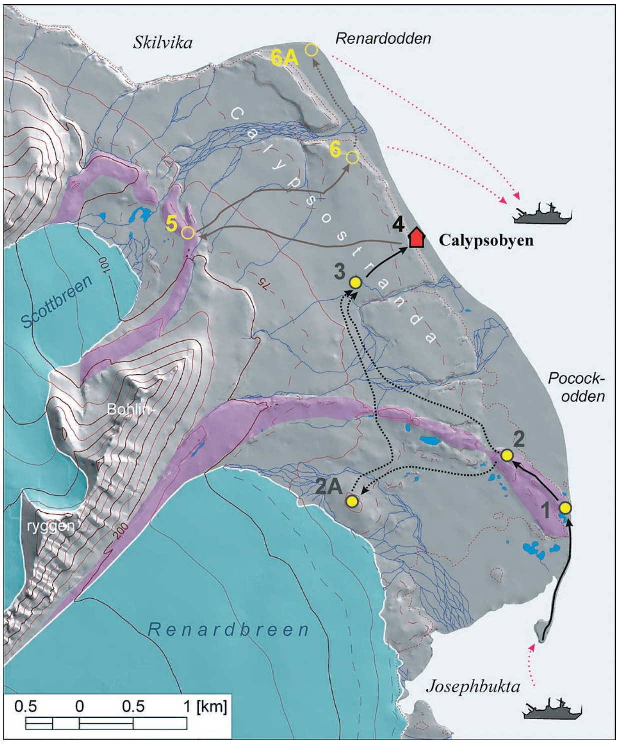
Fig. 14. Thickness of active layer of permafrost in dry and wet conditions, B: Correlation between thickness and air temperature (DDT – Daily Degree Thaw) (Christiansen et al. 2003)

Point 4 – Calypsobyen 77° 33’ 31" N, 14° 31’ 01" E

Calypsobyen – history and the present day

Kazimierz Pękala, Janina Repelewska-Pękalowa* – presented by Piotr Zagórski

Department of Geomorphology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland



A mining settlement Calypsobyen is situated opposite the mouth of Van Keulen Fiord to Bellsund (Krawczyk, Reder 1989, Roll 1993). It consists of wooden buildings preserved in different conditions (Fig. 1, 7, 15). The oldest buildings reach date back to first years of XX century. They are not big but covered with a ridge roof. That generation of buildings is represented by house on the slope near the mouth of Wydrzyca Stream (E). It was once covered with birch bark and some buildings in the “centre” of the village. Only one of those with two rooms (C) is suitable to live in. The rest (D) were used as the farm buildings.

A bit latter, after 1911 the London company: “*The Northern Exploration Company*” began the economic activity. It planned to exploit out coal and marble in the Bellsund region. Up till now it is possible to find the signs with ‘NEC’ on. They were used do mark the area that belonged to the company. At the end of First World War, some big buildings were built for mine needs. Quite quickly the mine activity was stopped and trappers used existing buildings. Their presence is still noticeable by equipment and traces they had left behind.

Till present only one building on the beach has been preserved. The longer axis is perpendicular to the shore and now it is usable to live in (A) and two-part building that is a bit higher on the slope (B), which has been turned into a store (Fig. 15). At the near surroundings of the buildings there are still some traces to the entrance to the mining shaft, track, coal truck and some mine tools. The relict from that epoch is a big wooden transport boat called “*Maria Teresa”*. There is also a partly ruined building on the raised marine terrace (F). There is a very good view over the fiords, so during the Second World War Germans built a broadcasting station. Its fallen aerial mast has been here near the entrance (Fig. 16).

The buildings in Calypsobyen have been left untouched because according to law all traces of human activity, from before 1946 year, are under legal protection (Roll 1993). They are the heritage park of industrial buildings from the beginning of XX century. The Calypsobyen and the whole NW part of Wedel Jarlsberg Land are within the border of the National

* e-mail: geomorf@biotop.umcs.lublin.pl

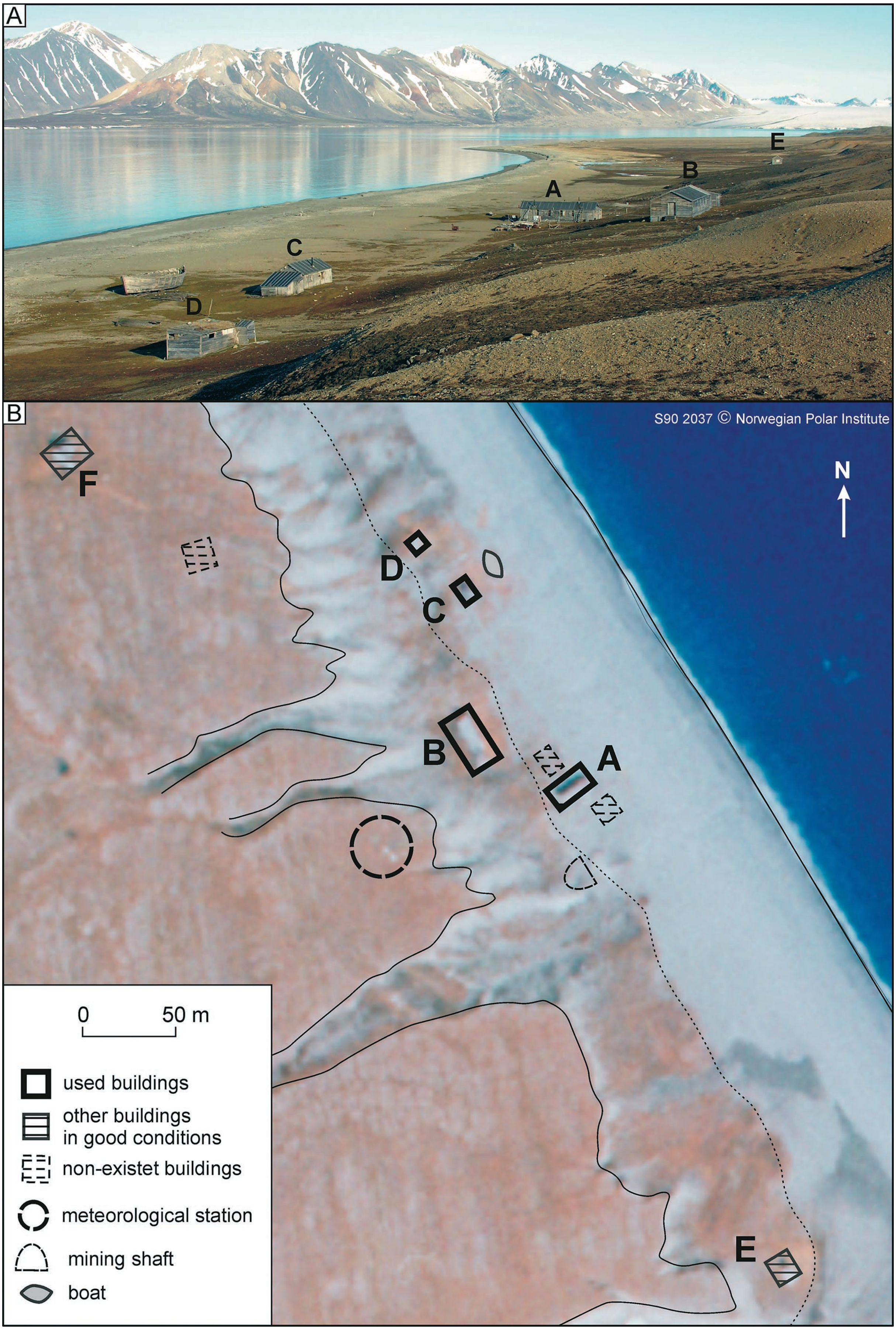


Fig. 15. Calypsobyen. A – general view (photo Piotr Zagórski 2005), B – localisation of buildings (Orthophotomap, Zagórski 2005)

Fig. 16. The building of broadcast station (F) from the Second World War (photo Piotr Zagórski 2006)

Fig. 17. The repair works on the building C (photo Janina Repelewska-Pękalowa 1986)

Park formed in 1973. Because of it there are some important limits for staying and working there.

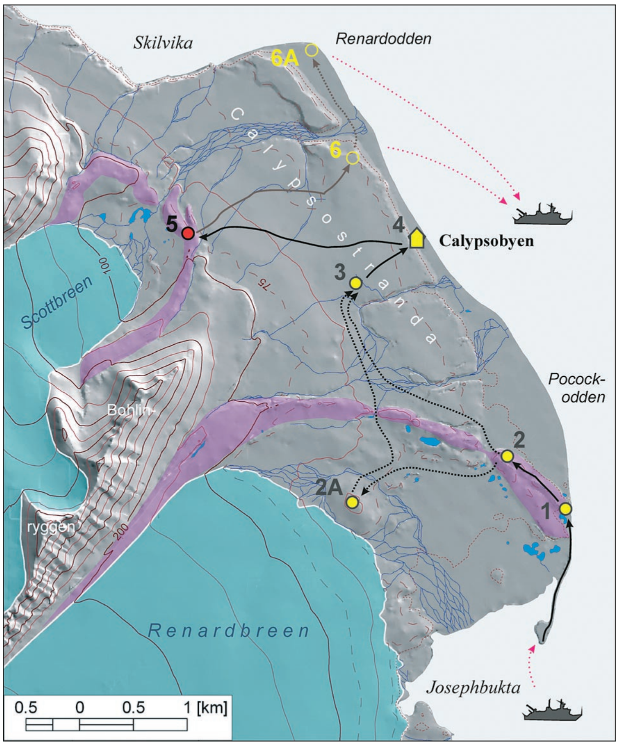
Under the permit of Governor of Svalbard, since 1986 the buildings in Calypsobyen have been the main bases for Polar Expeditions of M. C. Skłodowska University. The participants of 16 expeditions who have worked here did a lot of necessary repair work to live and work here (Fig. 17). All work was done with a great care to preserve the original look. For a few years the Norwegian administration is responsible for all renovation.

Point 5 – Push moraine of the Scott Glacier *77° 33' 36" N, 14° 26' 11" E*

Recession and development of marginal zone of the Scott Glacier

Jan Reder, Piotr Zagórski*

Department of Geomorphology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland



The Scott Glacier filled the valley of NW-SE direction at the lower part and higher – meridional one. From East the Scott Glacier is limited by Bohlinryggen range while from West – Wijkanderberget (Fig. 18). From Southwest, in the zone of low passes it connects with the Blomli Glacier, which fills the upper part of the Blomli Valley. The mouth of the valley of the Scott Glacier closes a few meters' push moraine ridge (ice-cored moraine ridges) cut by the gorge made by outflow of proglacial water. In axial parts its length was about 3.5 km and the wideness at the lower part exceeded 1 km when in the zone of firnfield reached about 1.5 km (Fig. 7).

The Scott Glacier in 2006 included the area of 4.7 km^2, but the largest area was at the end of XIX century (decline of Little Ice Age), and the episode of the surge is dated on 1880 (Liestøl 1993). Within the reach of it was probably all present inside part of the forefield to push moraines, and its area could have been over 6 km^2 (Zagórski, Bartoszewski 2004) (Fig. 19, 20). Since the Little Ice Age till 1936 the average for the whole length of the glacier front the distance of recession was 57 m – maximum 148 m. For the period of 1936–1960 the speed of recession was 1.8 m a^{-1}, as the mean recession – 44 m (maximum 120m – 5 m a^{-1}).

For the period 1960–1987 the mean recession of the glacier front on its whole distance was 162 m, (6–7 m a^{-1}), maximum 400 m (15 m a^{-1}). Those data can be incompleted because according to some archival data (the photo taken in 1963 and published in the book by J.Landvik et el. 1992, page 337), the Scott Glacier was just after the stage of advance (surge type). So in fact since 1960s we can talk about the beginning of the fast recession of the Scott Glacier and revealing inside part of the forefield. The following period 1987–1990 was characterised by the acceleration of the recession for the whole length up to 28 m, what corresponds to 9.3 m a^{-1} (maximum 68 m, 23 m a^{-1}). Since the end of XIX century till 1990 the surface of the Scott Glacier was reduced by 13% of the primary area (Zagórski, Bartoszewski, 2004).

Systematic studies and measurements of the Scott Glacier are conducted since 2000 and show that its front during the period of 1990–2006 moved back on

* e-mail: pzagorsk@biotop.umcs.lublin.pl

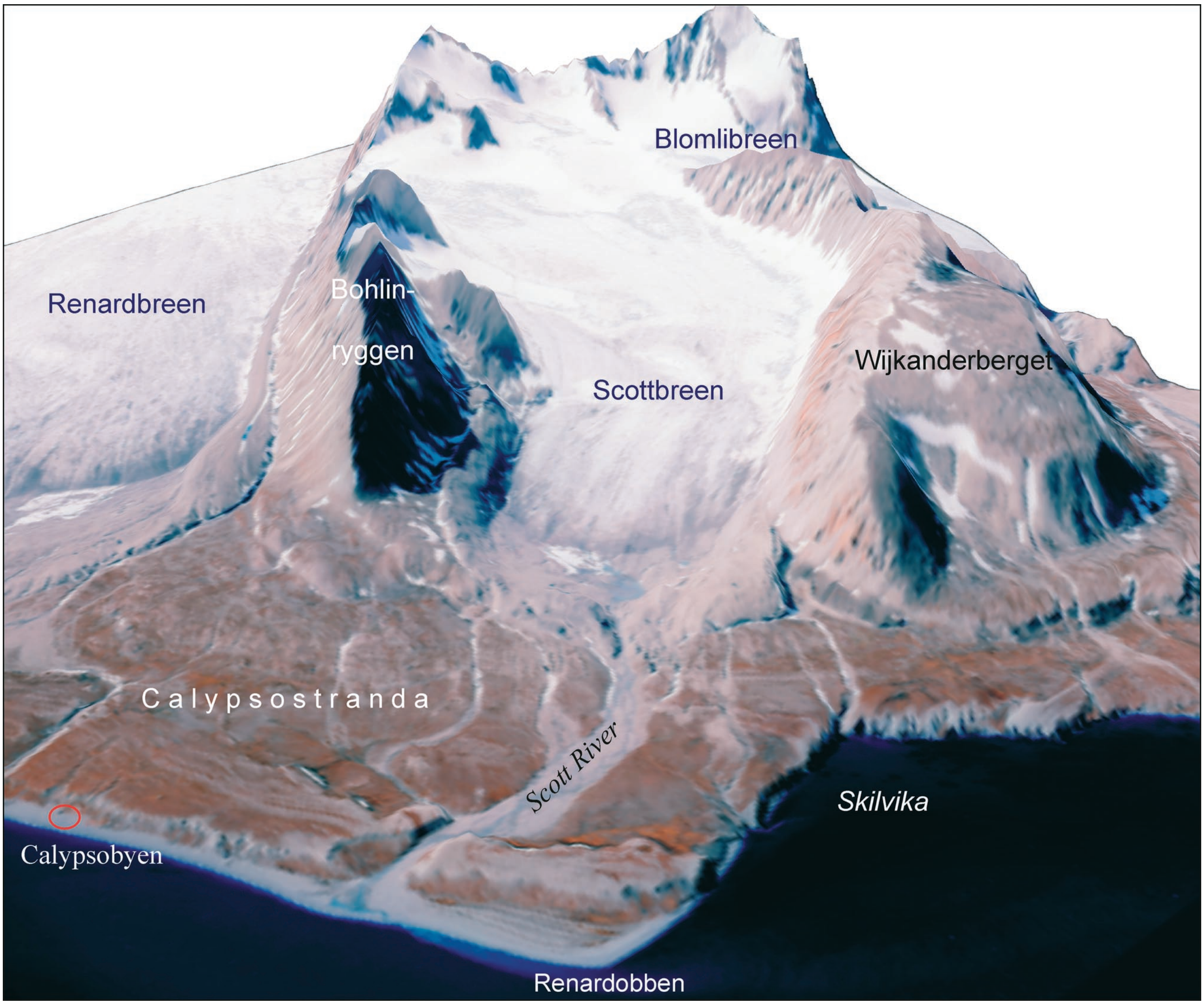


Fig. 18. Spatial model of the Scott Glacier and Calypsostranda made on the basis of combined the digital model of the terrain with aerial photo from 1990' (Zagórski 2002)

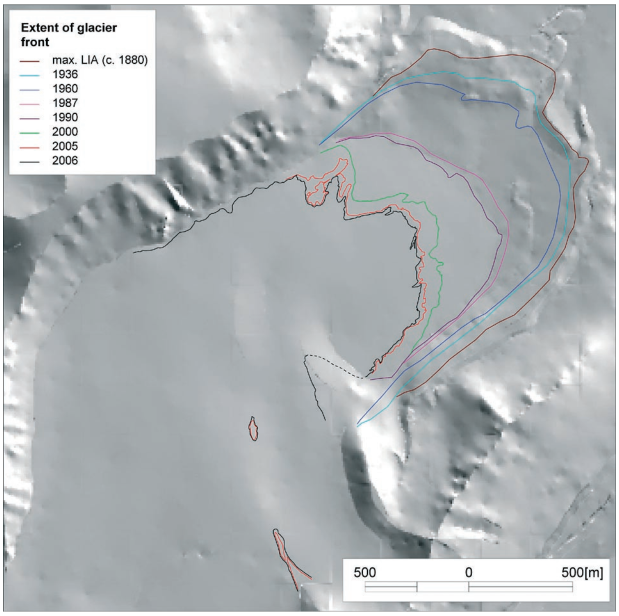


Fig. 19. The extent of the Scott Glacier fronts combined on the basis of archival data and GPS measurements (Szczęsny et al. 1989, Merta et al. 1990, Zagórski, Bartoszewski 2004, Zagórski 2005)

average 230 m on its whole distance, while the maximum was 440 m (28 m a^{-1}). The last measure period is marked especially clearly when the mean speed of recession of the glacier front reached 21 m, and maximum even 140 m. The main reason for quick decrease of the thickness of the glacier ice is the relief of the bedrock with the zone of rocky steps at the bottom of the Wijkanderberget (Fig. 19, 20).

The relief of the forefield of the Scott Glacier is much less diverse than the forefield of the Renard Glacier. The dominant element is frontal moraine ridge (Fig. 18, 21). The material of the moraine is on the bedrock of former denudation-structure layers of roche moutonnée character. The ridge of the lateral moraine accompanying the glacier tongue from southeast rises 60 m above the surface of the glacier.

The frontal moraine, but especially the lateral one deposited along the slopes of Bohlinryggen are compound forms came into being in two different phases

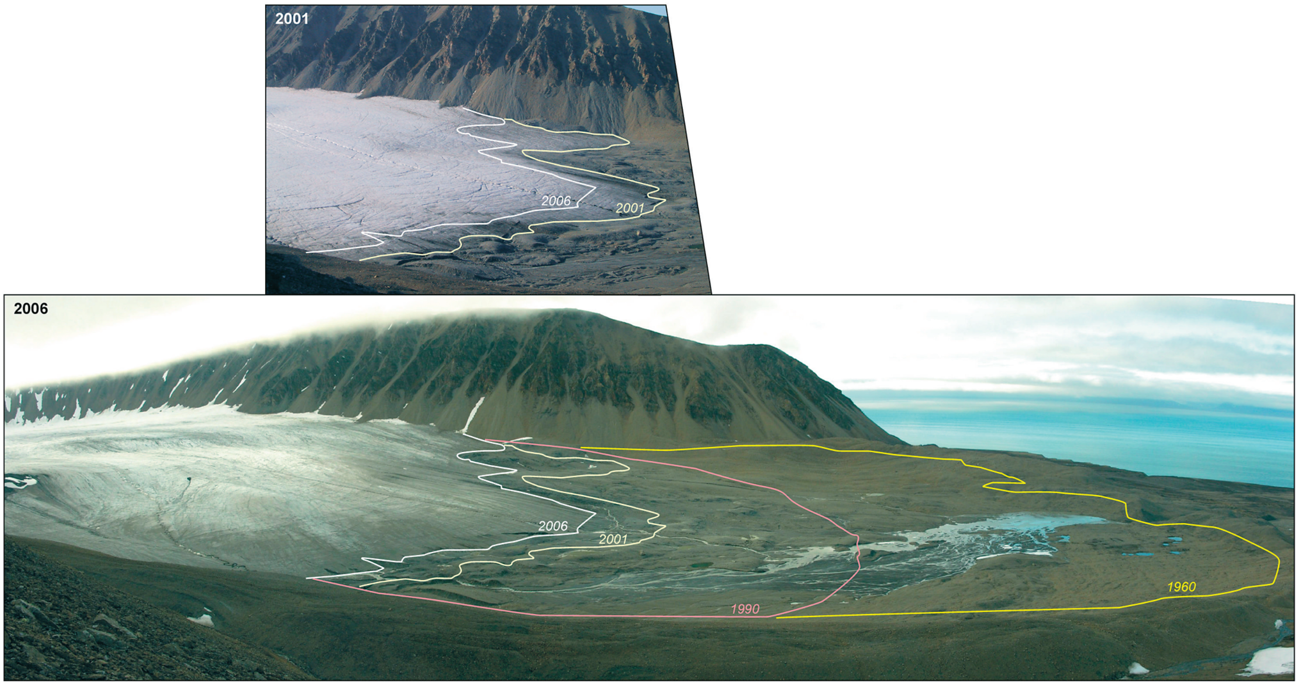


Fig. 20. The forefield of the Scott Glacier. The view from Bohlinryggen (photo Stefan Bartoszewski 2001, Piotr Zagórski 2006)

of the glacier transgression. The young ice-moraine sediments, partially pushed, were accumulated during the fast advance of the glacier onto its forefield in XIX century. They cover here a bit older moraine series, which probably arose during the advance of the Scott Glacier in the earlier phase of the Little Ice Age (Fig. 20). The frontal moraine from inside gradually and softly came onto the ground moraine with the traces of flow in the direction of the only active gorge. The more distinct element of the relief here is only the course of little hills that marks one of stages of the glacier recession (Fig. 21).

The internal marginal zone looks like a hollow: south-east part of the area is lower and a lot of it is flooded with water, north-east part is some meters higher from about the line of the gorge (Fig. 21, 22). This zone is made of mainly ground moraine, locally distinctly fluted. Some active riverbeds of the 2–4 m

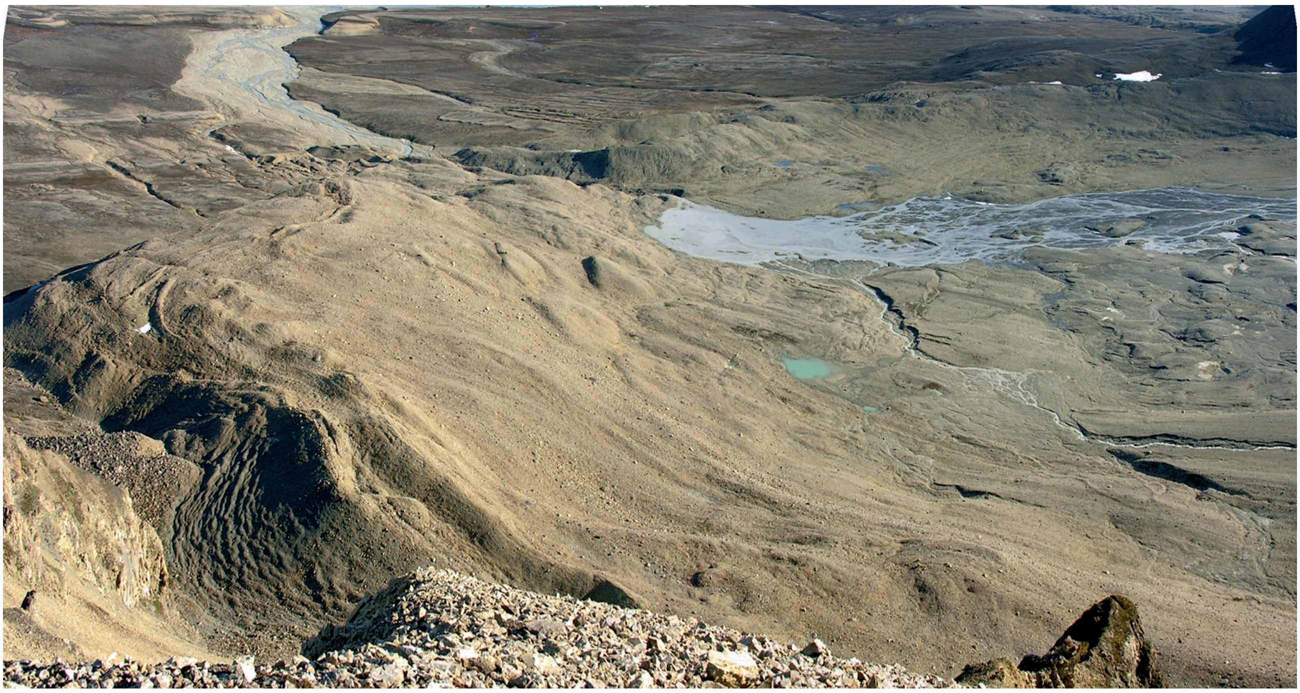


Fig. 21. The forefield of the Scott Glacier – the view from Wijkanderberget. Varied zone of the push moraine and the area of contemporary forming inner sandur (photo Piotr Zagórski 2006)

Fig. 22. The marginal zone of the Scott Glacier (photo Piotr Zagórski 2006)

depth cut it. The grooves of fluted moraine in this region are of different character than those observed on the forefield of the Renard Glacier. They are much bigger, and the height of single ridges reaches 50–60 cm. The orientation of the grooves follows the axis of the valley and the main direction of the glacier recession. The orientation, material layout and size can show that in that forefield Renard Glacier region they came into being as the result of filling former supraglacial troughs with material from ablation moraine. The depressions on the surface of the moraine are filled with stagnating water, which is the result of melt out phase. Sometimes the thin layer of silt is accumulated. The other fragments of moraine do not have signs of washout.

In the central part of the forefield, in the axial part of the valley the floodwaters exist where fine-grained material is sedimented. In the zone between frontal moraine and the glacier the typical sandur has not been developed yet (Fig. 20, 21). This is rather the zone of cut and washout of the ground moraine and sand little cons.

At the final part of the glacier tongue some ridges were observed. They are large and accumulative, similar to kame, made of fluvioglacial material and deposited on ice that melts out slowly. They are transverse to the axis and movement of the glacier. Their height is from some centimetres to about 2 m. They are built with fine-grained, irregular stratified material originated from the washout of the ablation moraine. They have significant asymmetric structure: proximal slope is a very gentle continuation of the slope, the distal slope is steep and falls at the angle of 30°–45° into the direction of the inside sandur. The similar set of forms of similar topographic layout due to glacier front is observed on the distance of some tens of metres on the northwest from the present edge of ice. Their origin should be connected with some phenomena that are noticed only sporadically. It is probable they are of extreme character as the result of unusually dynamic and efficient morphologically water flow on the surface of the glacier. It accompanies the beginning stage of ablation during Spring and early Summer. The steep distal slope could arise as the result of damming the outside part of the ridge against the thick cover of naledi or thick cover of snow on the glacier forefield.

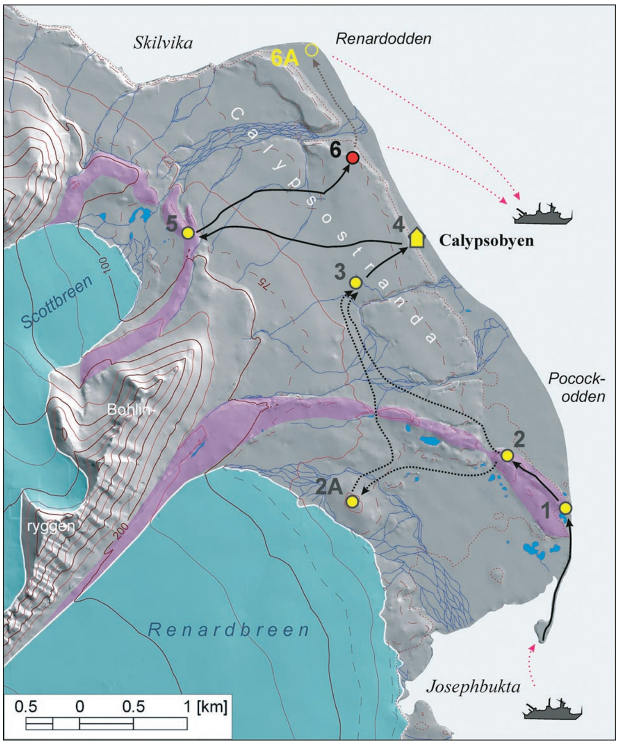
As it was marked earlier the main outflow from the glacier takes place from the SE side. Between the glacier and the lateral moraine the kame terrace was made. In the middle part of the forefield, there is a large sandur fan located aslant to the glacier front and only periodically active. The flowing river in the edge zone of the cone cut into the moraine sediments and now the cone rises 40–50 cm higher than the level of the ground moraine.

Point 6 – Calypsostranda *77° 33’ 55" N, 14° 29’ 41" E*

Relief and development of Calypsostranda

Piotr Zagórski*

Department of Geomorphology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland



The direct effect of sea level changes connected with glacial-interglacial cycles and glacioisostasy are the raised marine terraces. Very often they develop systems of steps within which there are characteristic storm ridges marking former shoreline, dead cliffs and paleoskerries related to marine abrasion. On Calypsostranda area seven terraces can be distinguished. The range of their height is between 2 and 85 m a.s.l. (Zagórski 2002, Zagórski et al. 2006) (Fig. 23).

The highest is the terrace VII (70–85 m) developed as slightly slanting abrasion platform. In the region of the Bohlinryggen, it neighbours to denudation level (80–90 m and 125–140 m) and shows clear traces of glacial remodelling. On the forefield of the Scott Glacier it was aggradated partly with ice-cored moraine ridges (Fig. 24). The abrasive character belongs to the terrace VI (height 50–60 cm). The age of those terraces is difficult to state due to a lack or vestigial occurrence of accumulative sediments. Their surfaces show traces of distinct glacial remodelling so the guess of pre-Weichselian age.

The marine terraces (V–I), which are located lower, are of accumulative character. They are made of various sediments as regards genesis and stratigraphy. It indicates multistage of development of the surfaces in Late Pleistocene when the periods of marine inundation interlaced with the advances of glaciers.

The marine terrace V (40–50 cm) probably marks the maximum limit of sea inundation from about 12 ka BP, what means after the last maximum Weichselian deglaciation (Fig. 25). That is the slightly slanting plain in the lower part accumulative changing into abrasive-accumulative one. In its morphology it can clearly distinguished the storm ridge of maximum width 60 m. Its length is at the foot of a denudation level 110–130 m (Wijkanderberget region) to Skilvika where it was cut abrasively.

The dominant terrace IV (30–40 m) is accumulative, nearly flat with fossil storm ridges and covered by fluvioglacial and marine sediments (Pleistocene and Holocene), which are lying on Palaeogene and Precambrian bedrock. Glacial sediments (medial moraine) are connected with the conjunction of glacier tongues from the region: Recherche Fiord and Van Keulen Fiord (Fig. 25). Near the Renardodden the terrace IV is limited by dead marine cliff modelling by solifluction. From the Skilvika terrace IV is

* e-mail: pzagorsk@biotop.umcs.lublin.pl

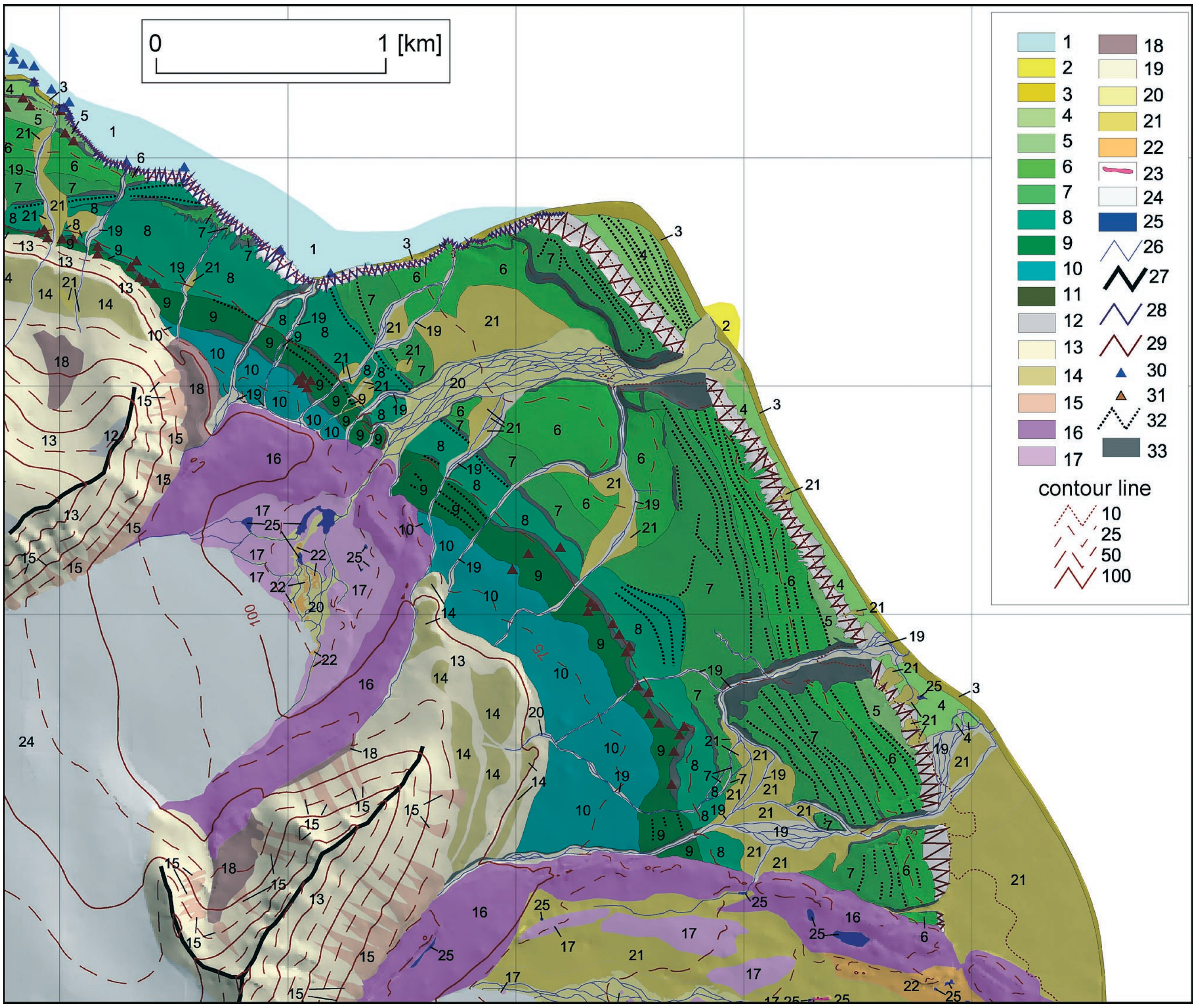


Fig. 23. Geomorphological map of the forefield of the Renard Glacier (Zagórski 2002)
1 – contemporary abrasion platform, 2 – tidal flat, delta cons, 3 – contemporary storm ridge, 4-terrace I (2–8 m), 5 – terrace II (10–20 m), 6 – terrace III (25–30 m), 7 – terrace IV (30–40 m), 8 – terrace V (40–50 m), 9 – terrace VI (50–65 m), 10 – terrace VII (70–85 m), 11 – terrace VIII (105–120 m), 12 – superficial flattening, 13 – slopes, 14 – denudation-structure level, 15 – talus cones, 16 – ice-cored moraine ridges, push and lateral moraines, 17 – ground and ablation moraines, 18 – rock glaciers (nival), 19 – floors of pronival valleys, 20 – contemporary sandur plains and fans, alluvial cones, 22 – kame, 23 – esker, 24 – glaciers, 25 – lakes, 26 – rivers, 27 – ridges, 28 – active marine cliffs, 29 – dead marine cliffs, 30 – skerries, 31 – paleoskerries, 32 – old storm ridges, 33 – edges.

Fig. 24. The view of Calypsostranda from Wijkanderberget (Photo Piotr Zagórski 2006)

destroyed intensely by abrasion. In that part of Calypsostranda it surrounds circularly distinct plain depression within which the lower terrace III of 25–30 m height was distinguished. Its fragments occur also between the valley of the Scott River and moraine ridges of the Renard Glacier and have character of slightly inclined accumulative surface with fossil storm ridges (Fig. 23, 25). On the distance from the Calypsobyen to extramarginal sandur fans of the Renard Glacier the terrace III merge into the lower terrace II (10–20 m). The dead cliff from east of Calypsostranda proves the intensity of abrasion of both terraces III and II in early Holocene.

The lowest terrace I (2–8 m) is a beach around the whole shore between Josephbukta and the Renardodden (Fig. 23). On the distance from the vast extramarginal sandur fans of the Renard Gla-

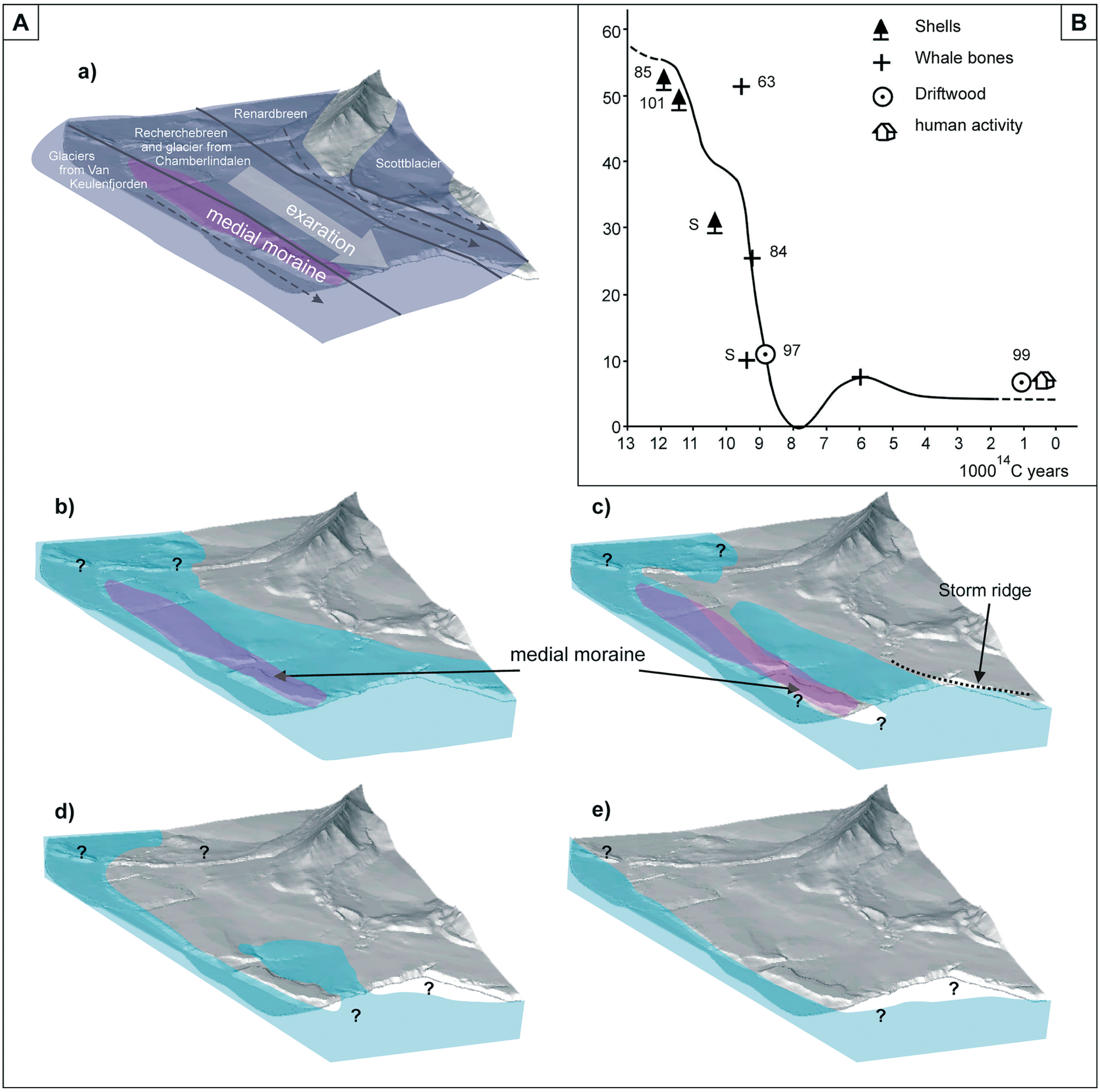


Fig. 25. A. Phases of development of Calypsostranda at the decline of Weichselian and in Holocene (Zagórski 2002)
a – the zone of influences of the glaciers during the glacial maximum of the Late Weichselian (about 20 ka BP), b – shoreline at 12 ka BP (development of terrace V), c – shoreline at 11–10 ka BP (development of terrace IV), d – shoreline at 10–9 ka BP (development of terrace III), e – shoreline at 8 ka BP (development of terrace II); B. Shoreline displacement curve for north-western Wedel Jarlsberg Land (Lognedallen) (after: Salvigsen et al. 1991)

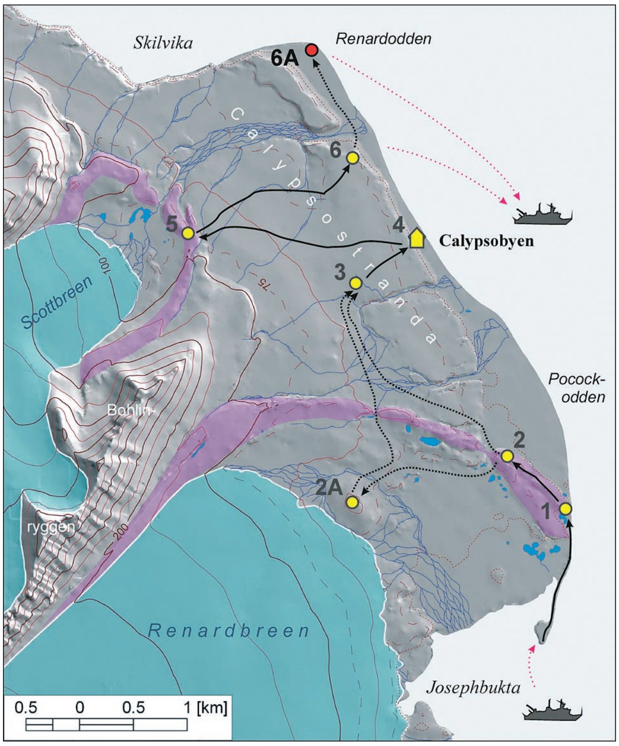
cier in the Pocockodden region to the mouth of the Scott River, the terrace I is build of two old storm ridges divided by two depressions developed as lagoons. In the neighbourhood of the Renardodden, as the result of intensive accumulation some now fossil ridges were made. On their surface there are numerous settlements sites from XVII and XIX century (Krawczyk, Reder 1989).

Point 6A – Renardodden *77° 34’ 21" N, 14° 28’ 49" E*

Present morphogenesis of the shore and the importance of archaeological sites for reconstructing the stages of development

Piotr Zagórski*

Department of Geomorphology, Institute of Earth Sciences, Maria Curie-Skłodowska University, Lublin, Poland



The Renardodden Region is unique example of the influence of marine factors on development and conversion of shore zone of accumulative character. The development of terrace I (2–8 m) was connected with the rise of the delivery of the material by the proglacial river of the Scott Glacier during the Little Ice Age. The additional reason of so big deposition could have been the change of the angle of the pass of the waves to the shore, which force accumulation. Here, longshore currents play the important role. Their zone of convergence exists in the section of the highest bend of the shore (Harasimiuk 1987, Harasimiuk, Jezierski 1991, Harasimiuk, Król 1992, Jezierski 1992, Zagórski 2004) (Fig. 26). The old storm ridges are well developed in that part of the shore are cut abrasively from the north and are aggradated with present storm ridge.

To estimate the role of marine processes in the Renardodden region it was crucial to recognise numerous archaeological sites here (Krawczyk, Reder 1989, Jasinski et al. 1993). Archaeological data show intensive exploration of this area since XVII century. The nearest to the shoreline zone is located the site Renardodden 1 (Fig. 27). It is remain of the Russian station of walrus hunters dated on the first half of XIX century. Probably, the building was out of reach of storm waving, but after the latter rise of activity of the abrasive processes caused the most probably by the changes of the sea level, the old storm ridge was destroyed and storm waves dragged pieces of bricks and organic remnants over the tidal flat zone (Jasinski, Zagórski 1996). The sediments of the following storm ridge, now intensively transformed, cover traces of the dragged occupation layer. Such conditions was kept till the beginning of 60s, so since the moment of start of quick recession of the Scott Glacier (Reder 1996, Zagórski, Bartoszewski 2004). Till 1990 the intensification of the delivery of the material caused aggradation of the cape of over 20 cm (Fig. 26). Yet the last years show that the delivery of

* e-mail: pzagorsk@biotop.umcs.lublin.pl

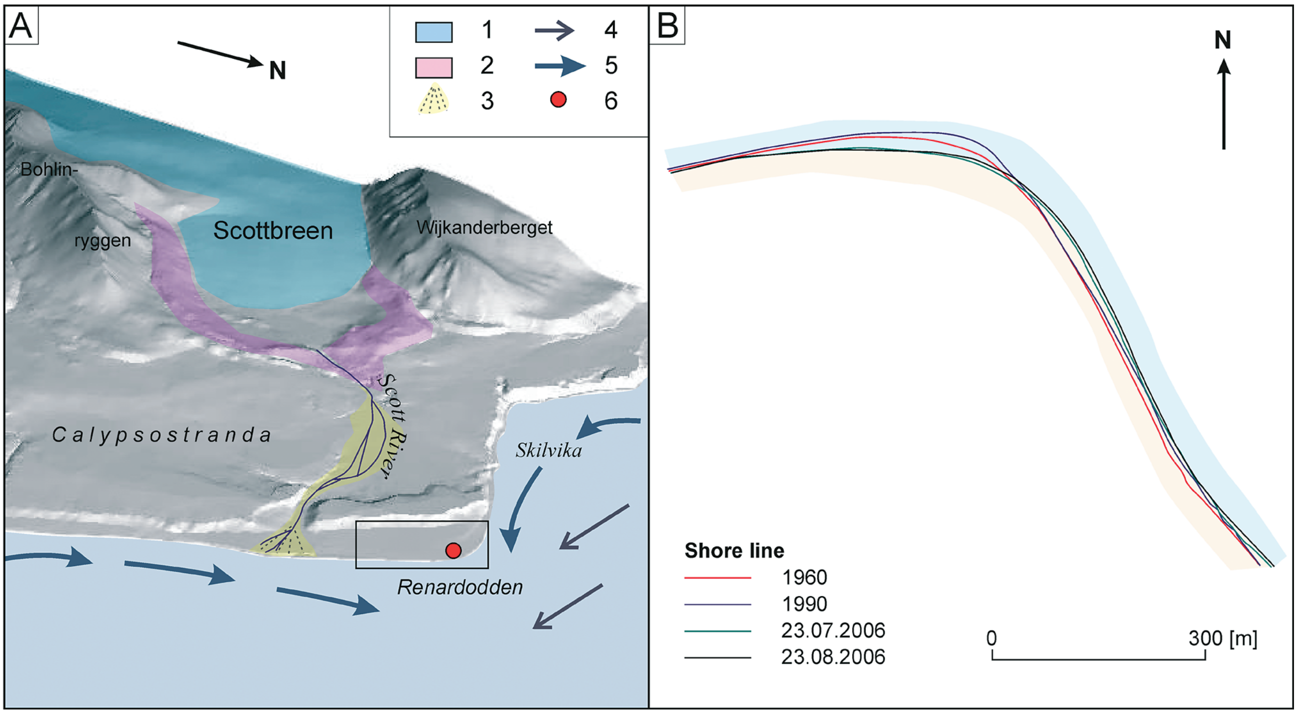


Fig. 26. A. Main factors that influence on formation of the shore in the Renardodden region
1 – glacier surface in 1990, 2 – frontal moraine ridges, 3 – sandurs, 4 – drift of western winds, 5 – directions of displacement of the longshore currents (after: Harasimiuk, Jezierski 1998, 1991), 6 – localisation of the archaeological site Renardodden 1.
B. Changes of geometry of the shoreline combined on the basis of archival data and GPS measurement (Zagórski 2007)

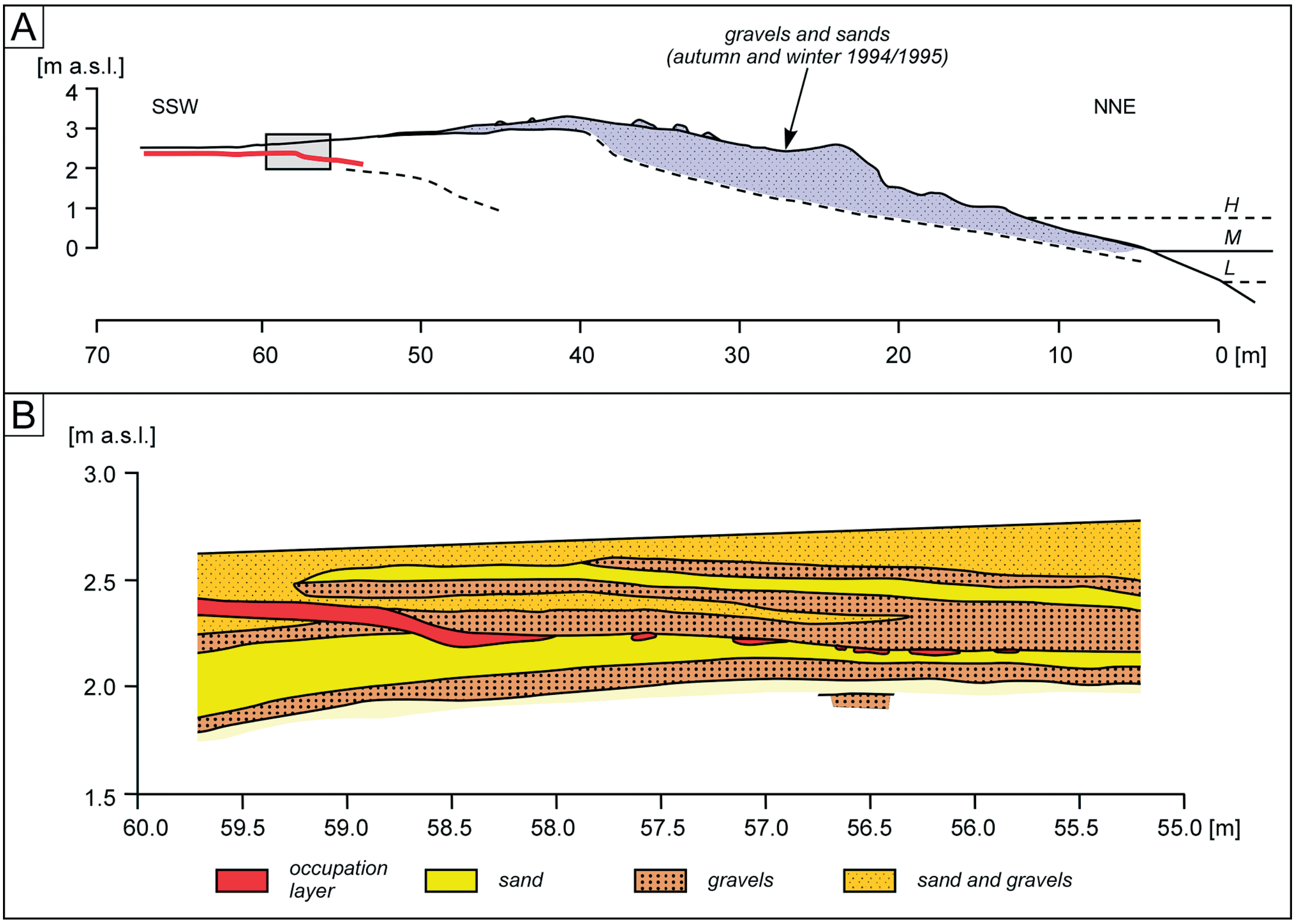


Fig. 27. Archaeological site Renardodden 1
A – Geological profile across the storm ridge, B – Geological profile across the fragment of storm ridge with dragged occupation layer (after: Jasinski, Zagórski 1996).

the material from the marginal zone of the Scott Glacier falls but the importance of marine processes rises (waving, longshore currents). The archival data (maps, air photos) and GPS measurement show the changes of geometry of the Renardodden. Strong cut of the part from the Skilvika is noticed but the section in the direction of the mouth of the Scott River is aggradated (Zagórski 2007).

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Recent and present-day glaciological and geomorphological processes at Hornsund – Leader Piotr Głowacki

Piotr Głowacki*

Polar and Marine Research Department, Institute of Geophysics, Polish Academy of Sciences, Warszawa, Poland

Zbigniew Zwoliński

Institute of Paleogeography and Geoecology, Adam Mickiewicz University, Poznań, Poland

The Station

The Station was founded in July 1957 by members of the Expedition of the Polish Academy of Sciences working under the auspices of the International Geophysical Year 1957–1958. The expedition was led by Stanisław Siedlecki, a well – known Polish polar scientist and explorer who had earlier taken part in a 1932 – 1933 winter expedition to Bjornaya organized as part of the Second International Polar Year.

The Polish Polar Station is situated on the Isbjornhamna in the Hornsund Fiord on Wedel Jarlsberg Land (Fig. 1) which comprises the southern part of West Spitsbergen. The geographical coordinates at the main building are 77° 00' 04"N and 15° 33' 37"E. The Station stands on a terrace rising 9 m a.s.l. The Polish Polar Station Hornsund is part of the Institute of Geophysics, Polish Academy of Sciences. Direct scientific and logistic management is made by the Polar and Marine Department of the Institute. The station is supported by the Ministry of Scientific Research and High School Education.

The vast expanse surrounding the Station – Torell Land and Wedel Jarlsberg Land – has been studied by Polish expeditions since 1934. The Poles were the first to prepare topographical maps of this region. Polish names of mountains and glaciers are officially recognized in Spitsbergen.

Fig. 1. Polish Polar Station at Hornsund during summer time (photo Archive IGF)

In the early period, research teams residing at the Station conducted the following types of observations: meteorological (the data obtained were used in climatology and in weather forecasts and was transmitted several times daily to the world network); all sky – camera (which included the observation of the ionosphere and aurora polaris); astronomical; actinometrical; glaciological and the study of permafrost; the scanning of radioactive fall – out and of CO_2 content in free atmosphere; and in the summer time geological and geomorphological.

* e-mail: glowacki@igf.edu.pl

Fig. 2. Polish Polar Station at Hornsund, April 25, 2007 (photo A. Nawrot)

These studies were part of a global cooperative effort initiated during the International Geophysical Year 1957 – 1958. It was then that the first yearlong research program was carried out, with 10 scientists and technicians, led by Siedlecki, wintering at the Station. In the next twenty years, the Station was used by a summer expeditions sponsored by Polish Universities and the Polish Academy of Sciences. In 1978, the Presidium Committee of the PAS decided to have the Station modernized and enlarged to make it suitable for use in winter again.

The Polish Polar Station Hornsund (Fig. 2) is situated in the centre of Spitsbergen Archipelago, at the junction of Asiatic and American Arctic. This location is very suitable for geophysical investigations. It is ideal for studying lithospheric structure as well as the physical processes in the atmosphere and extraterrestrial space. The glaciers placed nearby the Station are an excellent research polygon for determining the intraglacial physical processes. Seismological recording is the basis for the study of Arctic seismicity and glacial seismic events. The location of Spitsbergen is optimal for studying physical processes in the region of aurora and polar cusp. Measurements of geomagnetic field components together with those of atmospheric electricity elements, ionospheric absorption, and observations of aurora provide information on the processes occurring in the magnetosphere and ionosphere under the influence of solar wind. Another important problem studied there is the determination of factors affecting the solar radiation inflow to the Earth's surface in Arctic. Many programs performed at the Station concern the physical parameters investigated in the framework of the International Global Change Program.

The research is done in the following main directions:

- Geophysical investigations, including:
  - geomagnetic field;
  - seismicity of the Arctic Sea basin;
  - atmospheric electricity;
  - optics of the atmosphere;
  - ionospheric events in the "Polar cusp" region.

  Environmental research:
  - mass balance of glaciers (Fig. 3);
  - long-range transport of pollutants from Europe to the Arctic;
  - local and regional evolution of the environment;
  - geomorphological investigations;
  - biological research.
- Transmission of data to international data centres:
  - IMAGE – International Monitoring for Aurora Geomagnetic Effects;
  - INTERMAGNET – International Real-time Magnetic Observatory Network;
  - WMGS – World Glacial Monitoring Service;
  - WMO – World Meteorological Organization.

Since 1978, there have been some small changes in the scientific programs conducted at the Station and the research has always been carried out in strict compliance with the international standards. The main areas of research interest are as follows:

- Meteorology – gathering data for synoptic purposes and to detect climatic changes (Fig. 4);

Fig. 3. Change of frontal zone of Hansbreen between 1957 and 2003 (photo Archive IGF)

- Seismology – monitoring of world earthquakes, measuring the seismicity of the Spitsbergen Archipelago region, and registering tremors linked to the dynamics of the Hans Glacier;
- Magnetism – registering changes in the XYZ components of the earth's magnetic field;
- Ionosphere sounding to determine the structure of the ionosphere;
- Glaciology – photogrammetric measurements of the head of the Hans Glacier and its dynamic (Fig. 5);
- Permafrost – studding the dynamics of attendant processes;
- Atmospheric electricity – determining the magnitude of the electric field and registering its vertical component;
- Regular environmental monitoring – automatic registering of selected climatic features and conducting analyses of chemical buildup air and water pollution and the isotopic content of the snowcap.
- Geomorphology – investigations are conducted by numerous scientific teams from different universities and countries:
- morphogenesis of marginal zones for retreating both land based and tidewater glaciers (Fig. 6);
- processes of mechanical and chemical denudation;
- contemporary morphogenetic processes on paraglacial areas.

Some of the information's obtained are then regularly forwarded to international Data Centers. The other results are returned to Poland where they are processed and published in national and international journals or in the form of annuals or special issues. In the summer, the Station serves as the base for special programs in earth sciences, biology, ecology and oceanography. The studies are carried out by a number of university research centers as well as by the Polish Academy of Sciences.

In its present state, the Polish Polar Station at Hornsund is a modern, conveniently equipped complex. The Station consists of the main building, the building of the power station, three separate scientific cabins, the store-house for marine equipment for local boat transport.

Scientists will have the access to:

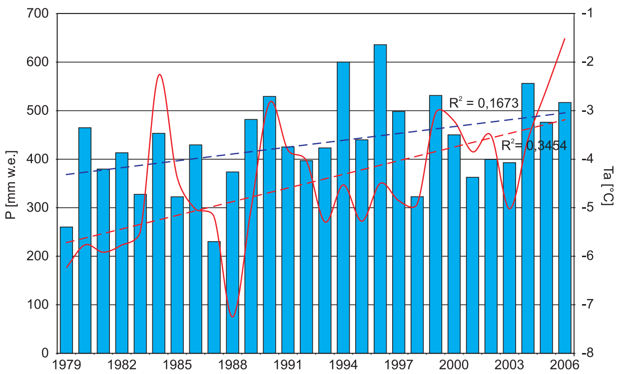


Fig. 4. Course of annual precipitation (P) and air temperature (Ta) in Polish Polar Station at Hornsund

Fig. 5. Frontal zone of Hansbreen, April 3, 2007 (photo A. Nawrot)

Fig. 6. Front of Hansbreen, May 3, 2007 (photo A. Nawrot)

– 13 single bedrooms (for round-year use),
– 7 larger bedrooms (for spring-summer season) – 30 places,
– 8 laboratories,
– a kitchen, a dining-room, bathrooms facilities with running hot and cold water, toilets, a medical room and a storage rooms with freezers, workshop,
– two tractors (one suitable for excavating and bulldozing), a mobile crane, trailers, six snow scooters and two large rubber boats Zodiacs (Fig. 7) with Yamaha 30 engines, two caterpillar amphibians (capable of carrying up to 4 tons each) internet connection.

The Station is provided with electricity 230/400 V, three satellite-communication systems (Iridium, Inmarsat-B). In 1978 the Station acquired a new sewage-treatment plant which can process up to 5 cubic meters of sewage per day. In the summer, the nearby lakes serve as the source of drinking water; in winter the water is obtained by means of melting snow. The Station is serviced by ships specially chartered out during summer (Fig. 8). The Station has the necessary equipment for land-based rescue work. In January 1990, by decision of the Royal Norwegian Mail the Polish Polar Station became a „postal point", and ever since has become an attraction for philatelists from all over the world.

More important instruments working at the Station:
– Automatic weather station Vaisala QLC50
– Torsion photoelectric magnetometer PSM–8911-08, LEMI-004P/96 fluxgate magnetometer, digital logger DR-02, analog recording system PSM/R–8111, proton magnetometer PMP–5–115, fluxgate magnetometer declinometer/inclinometer
– Ionic Chromatograph Compact IC–716, Methrom
– Laser total station TCR–1105, Leica
– Seismological station MK–6
– Ionosonde
– Multifunction Computer Meters CX–742, Elmetron
– Sunphotometer Cimel CE 318
– GPS permanent station Leica GRX 1200 and Leica GX 1230

The Station cooperates with 25 scientific institutions in Poland and 35 institutions from other countries. From 2002 it is one of six flagship sites for biodiversity of Europe.

The Station is open for students from various universities, to collect materials for master's and PhD dissertations and professional training. Courses in connection with the EUROPOLAR ERA-NET project and field workshops for glaciology, geomorphology, geology and biology are also planned.

Our research activity in Spitsbergen is made possible by the Treaty of Paris ratified by the various countries and major world powers in 1920. The Treaty proclaims Norwegian sovereignty in Spitsbergen and confirms the right of the signatories to conduct scientific research in that area. Poland joined the Treaty of Paris in 1931. Today however, Poland is the only non-arctic country to maintain a stationary research station in the Arctic region.

Fig. 8. Research ship SSB Horyzont (Maritime Academy in Gdynia, Poland) and front of Hansbreen in the background (photo Archive IGF)

Fig. 7. Transportation of research groups on boats (photo Archive IGF)

The active participation in studying the region effected in the invitation of Poland by the eight Arctic countries to take a part in the works of the International Arctic Sciences Committee – IASC. Since 1991 Poland has been a full member of this organization. IASC prepares and coordinates all major research programs in the Arctic.

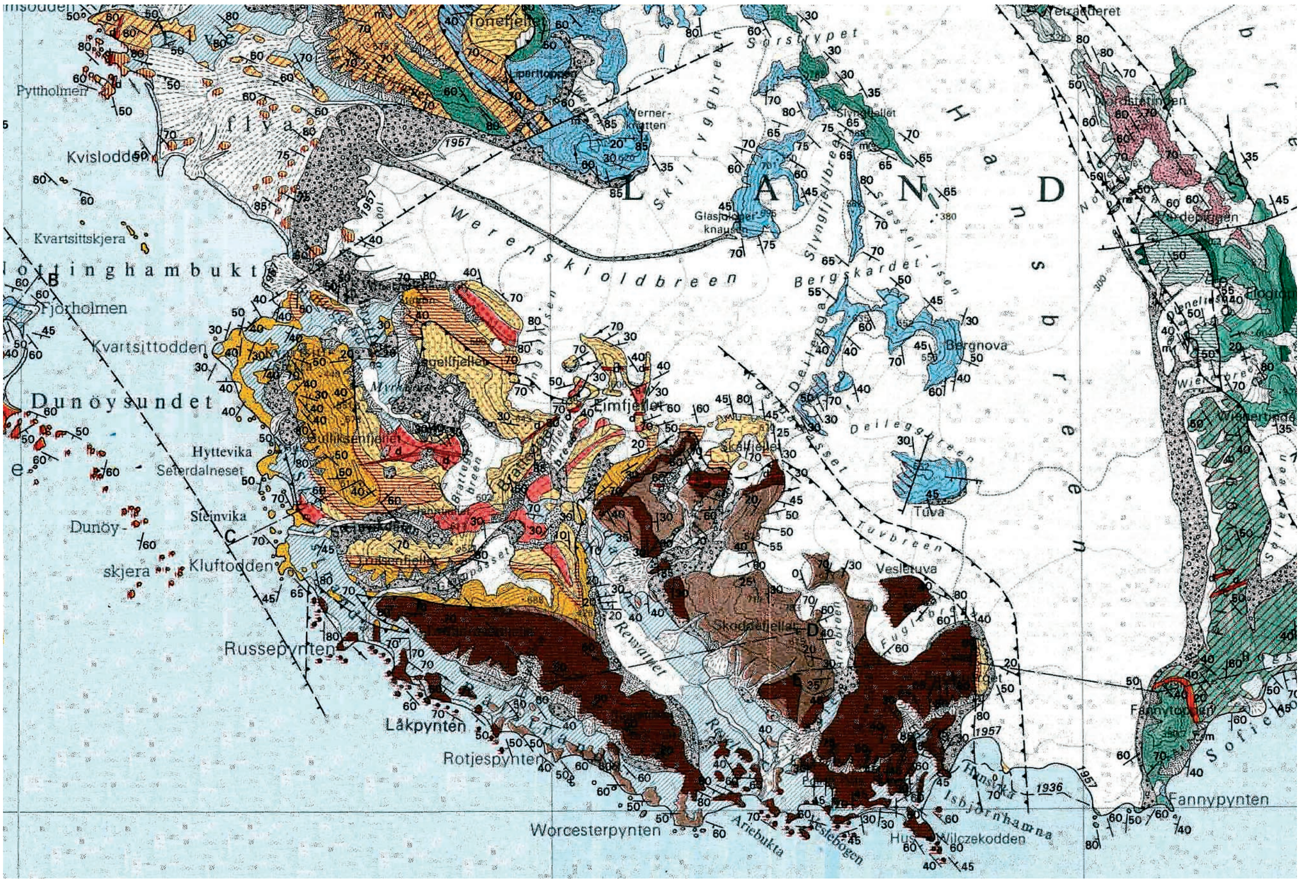
More information: http://hornsund.igf.edu.pl.

The excursion

Fluctuations of glaciers in Southern Spitsbergen are well documented in the past and currently studied. Glaciers were mapped first time by the Russian-Swedish expedition measuring arc of meridian in 1899–1901. Later maps were prepared by terrestrial and aerial photogrammetric survey (the Polish Expedition of 1934) and oblique aerial photos of 1936. There are also maps from German expedition 1938 and number of Polish expedition since the 3rd International Geophysical Year 1957/1958.

Significant recession of majority of glaciers in the vicinity of Hornsund is well visible. Mean annual retreat is in order of 25–50 m. There are also surging glaciers. Hansbreen are systematically observed thanks to permanent operation of the Polish Polar Station in Hornsund (since 1978).

This excursion will demonstrate marginal zones of retreating tidewater Hansbreen. Mass balance, velocity and calving intensity are regularly measured on glacier. Hydrothermal structure, including drainage system is studied by means of radar sounding and direct speleological exploration. During the excursion will be demonstrated glaciological (dynamics of calving glacier, mass balance) and environmental programmes (marginal zone) conducted by the Polish Polar Station, Hornsund – visit to the Station and research sites near and on Hansbreen.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Kluftodden
Russepynten
Låkpynten
Rotjespynten
Worcesterpynten
Ariebukta
Hus
Wilczekodden
Isbjørnhamna
Fannypynten
Werenskioldbreen
Angellfjellet
Eimfjellet
Skålfjellet
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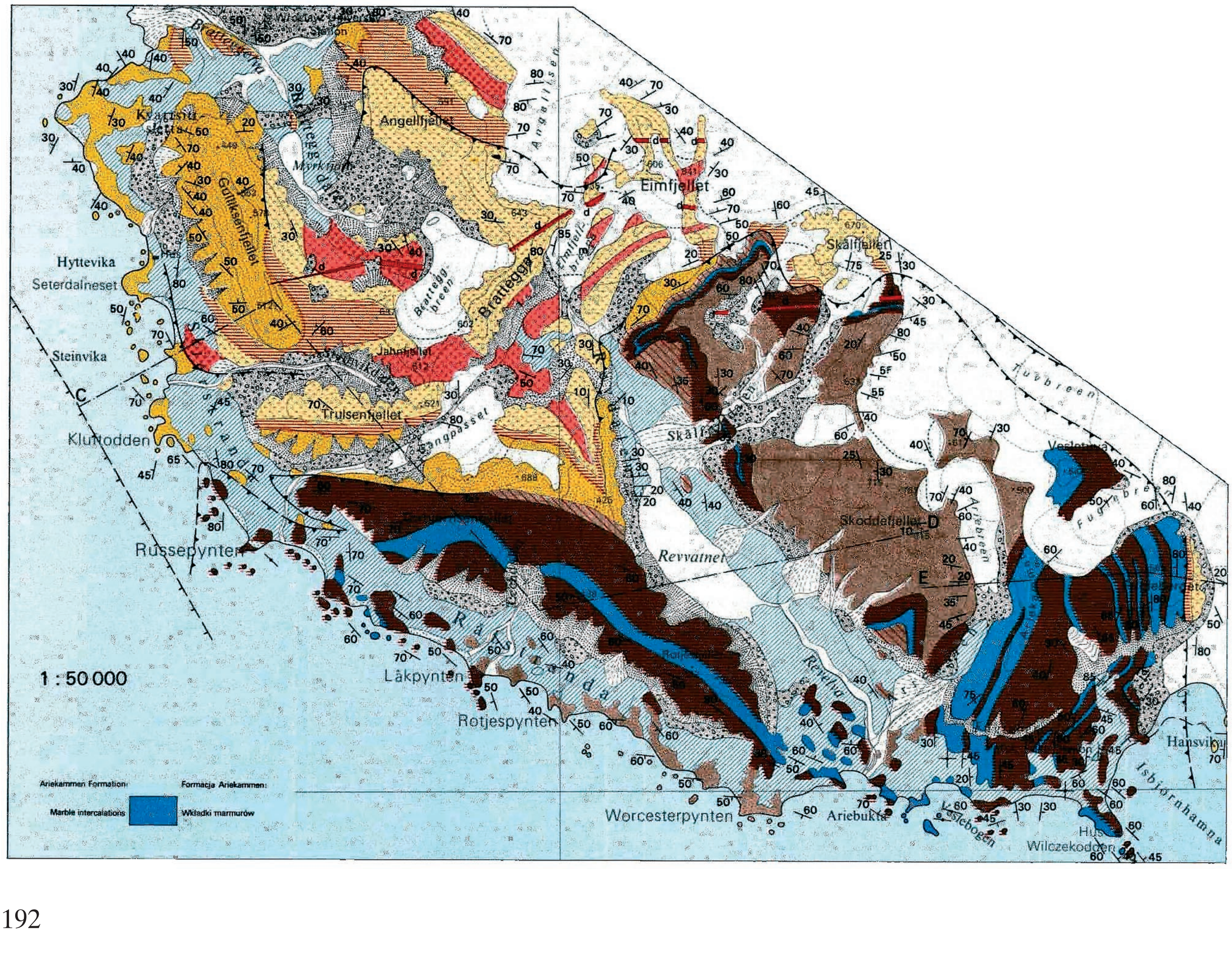
Tuva
Tuvbreen
Bergnova
L A N D
Hansbreen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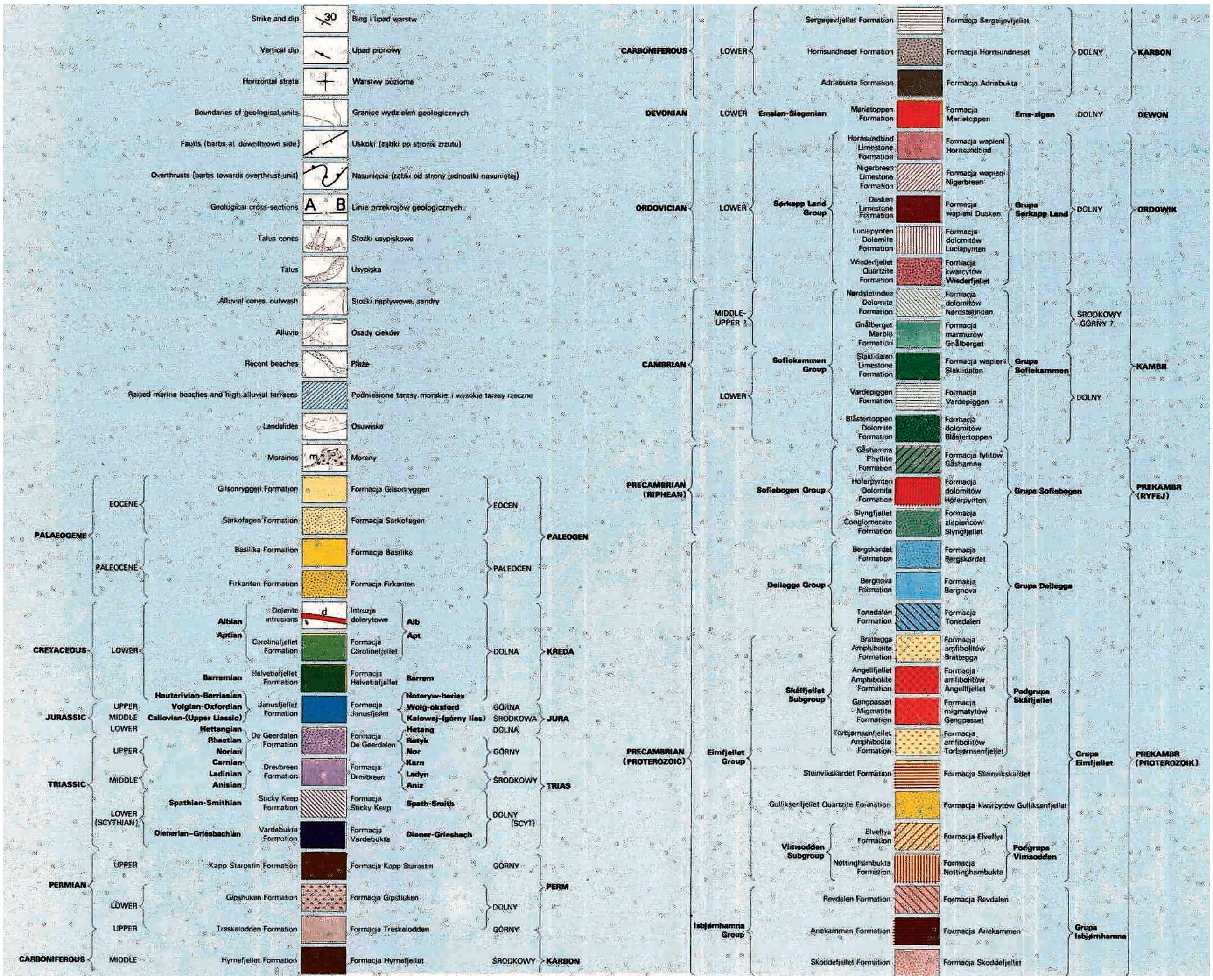
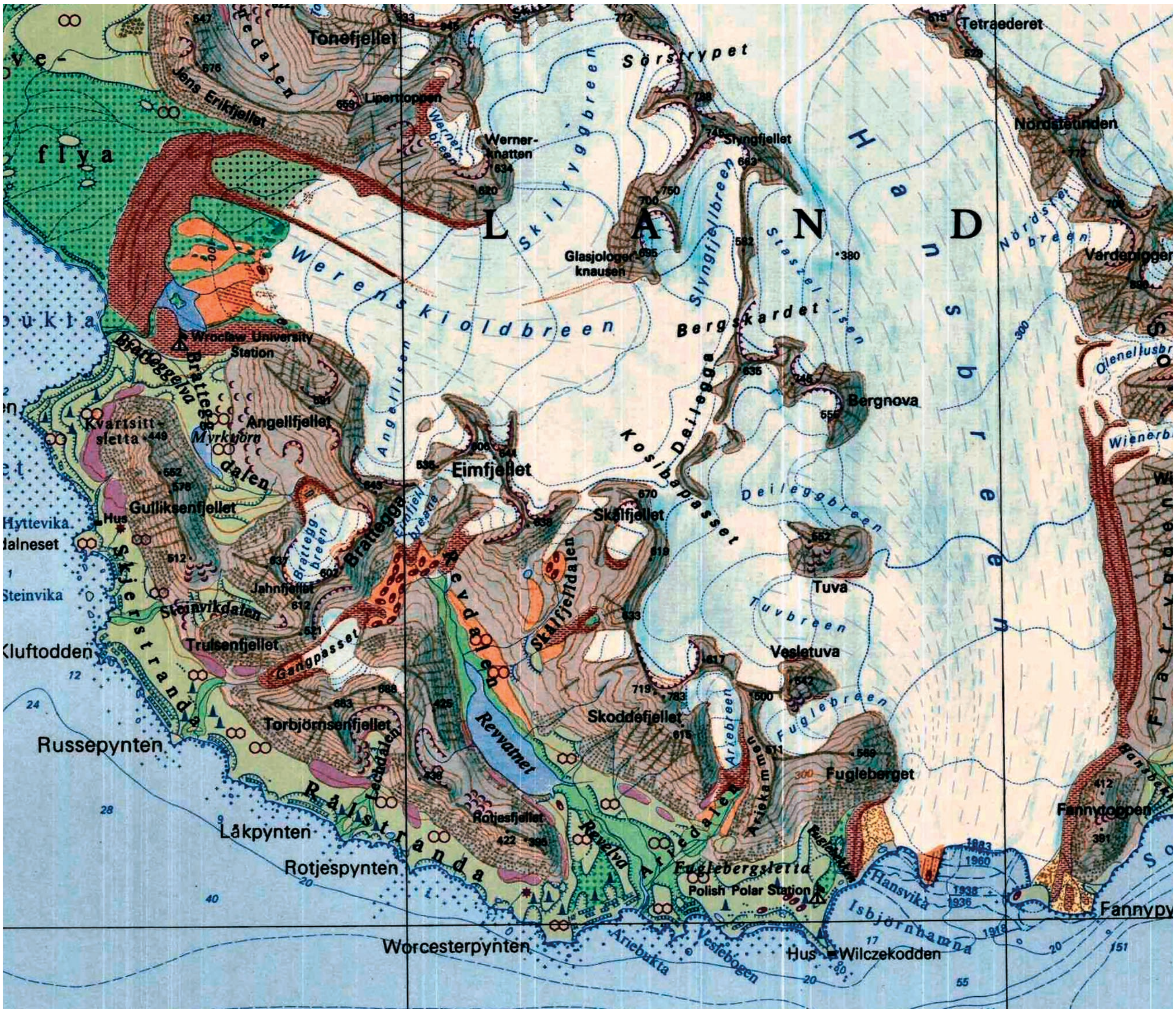


Fig. 9. Geological map of the northern part of Hornsund (Birkenmajer 1990)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Jahnfjellet
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Revdalen
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Skoddefjellet
Fuglebreen
Torbjörnsenfjellet
Russepynten
Revvatnet
Fugleberget
Ariebreen
Fannytoppen
Låkpynten
Rotjesfjellet
Rotjespynten
Revelva
Fuglebergsletta
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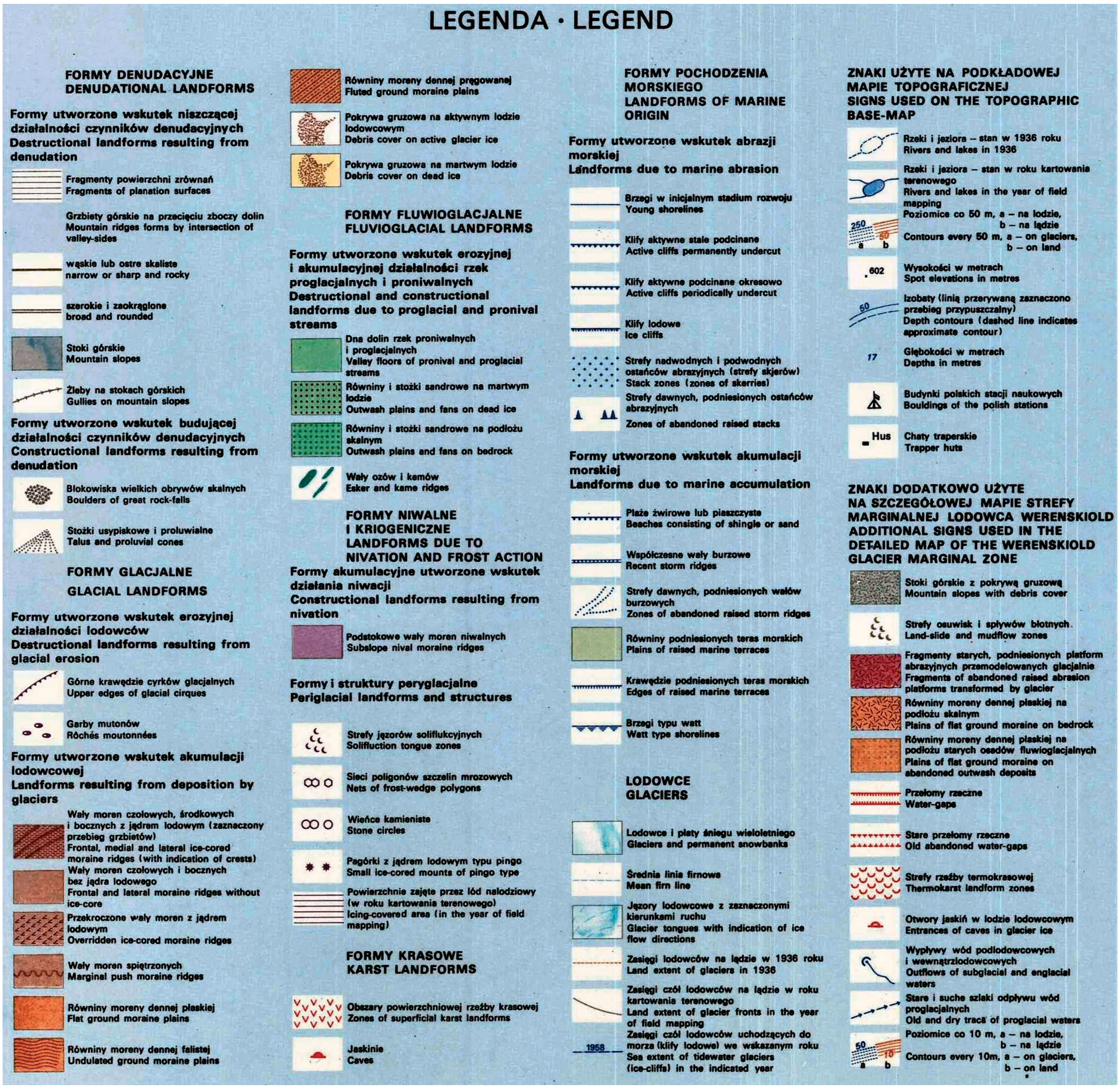


Fig. 10. Geomorphological map of the northern part of Hornsund (Karczewski 1984)

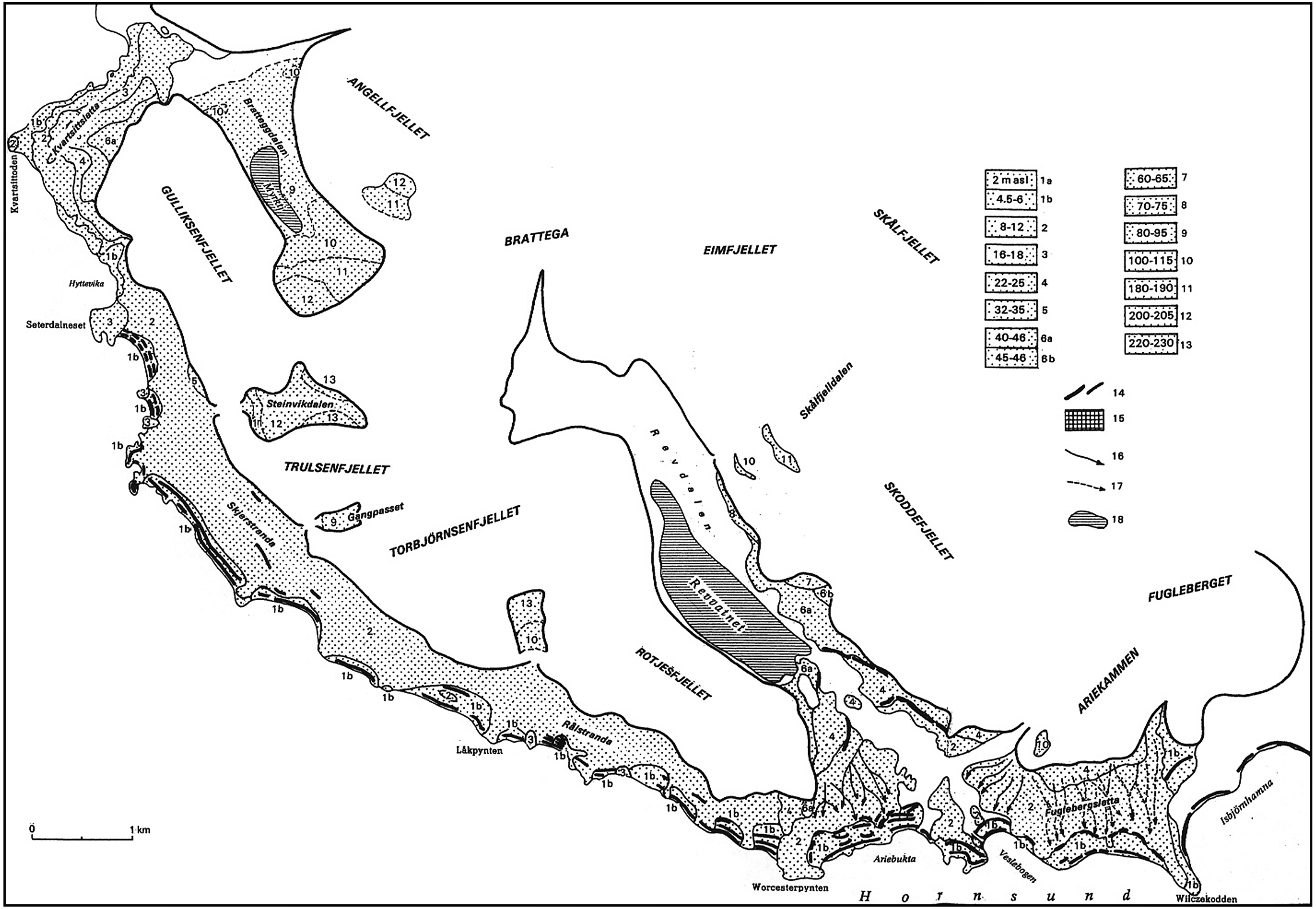


Fig. 11. Raised marine terraces of northern part of Hornsund (Karczewski et al. 1979)
1–13 – levels of raised marine terraces, 14 – front ice-moraine ridges, 15 – storm bank, 16 – channels of melt-out waters, 17 – channels of pronival waters, 18 – lakes

Fig. 12. Hydro-morphological scheme of Sofiekammen Ridge (Pulina 1977)
1 – mountain ridges and hypsometric spots (m a.s.l.), 2 – glaciers, 3 – moraine deposits, 4 – gullies, 5 – talus cones, 6 – debris deposits, 7 – surface with active solifluction, 8 – edges of valleys, 9 – rock gorge or river gap, 10 – edges of marine terraces: I – low terrace with marine gravel, a – narrow beach, b – limestone islands, II – middle terrace with numerous karstic micro- and macroforms, c – limestone cliff, III – high terrace covered by slope deposits, 11 – caves, 12 – surface streams, 13 – subsurface flows within ice and snow cover, a – ponor, b – outflow, 14 – supraglacial streams, 15 – lakes, 16 – karst spring "Orvin", 17 – other springs.

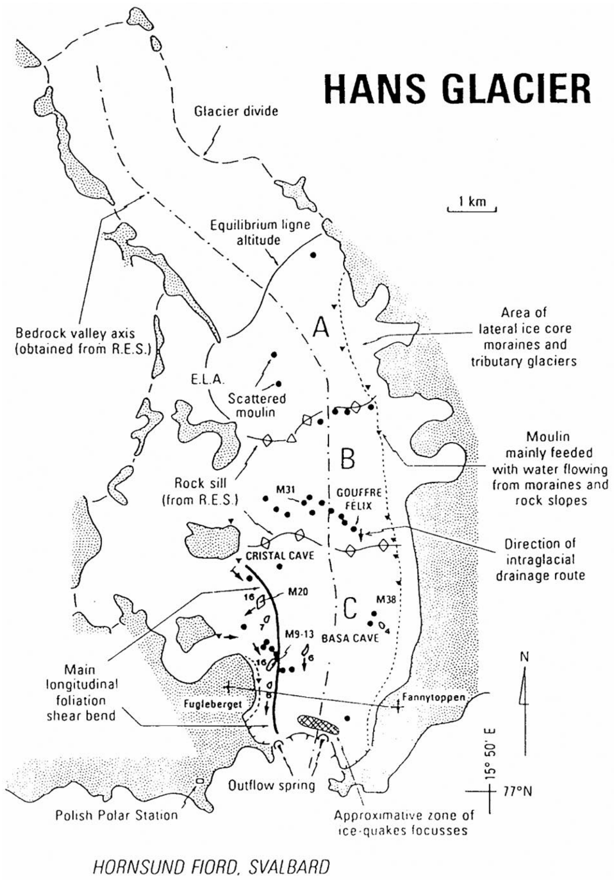


Fig. 13. Location of ice caves in Hansbreen (Schroeder 1995)

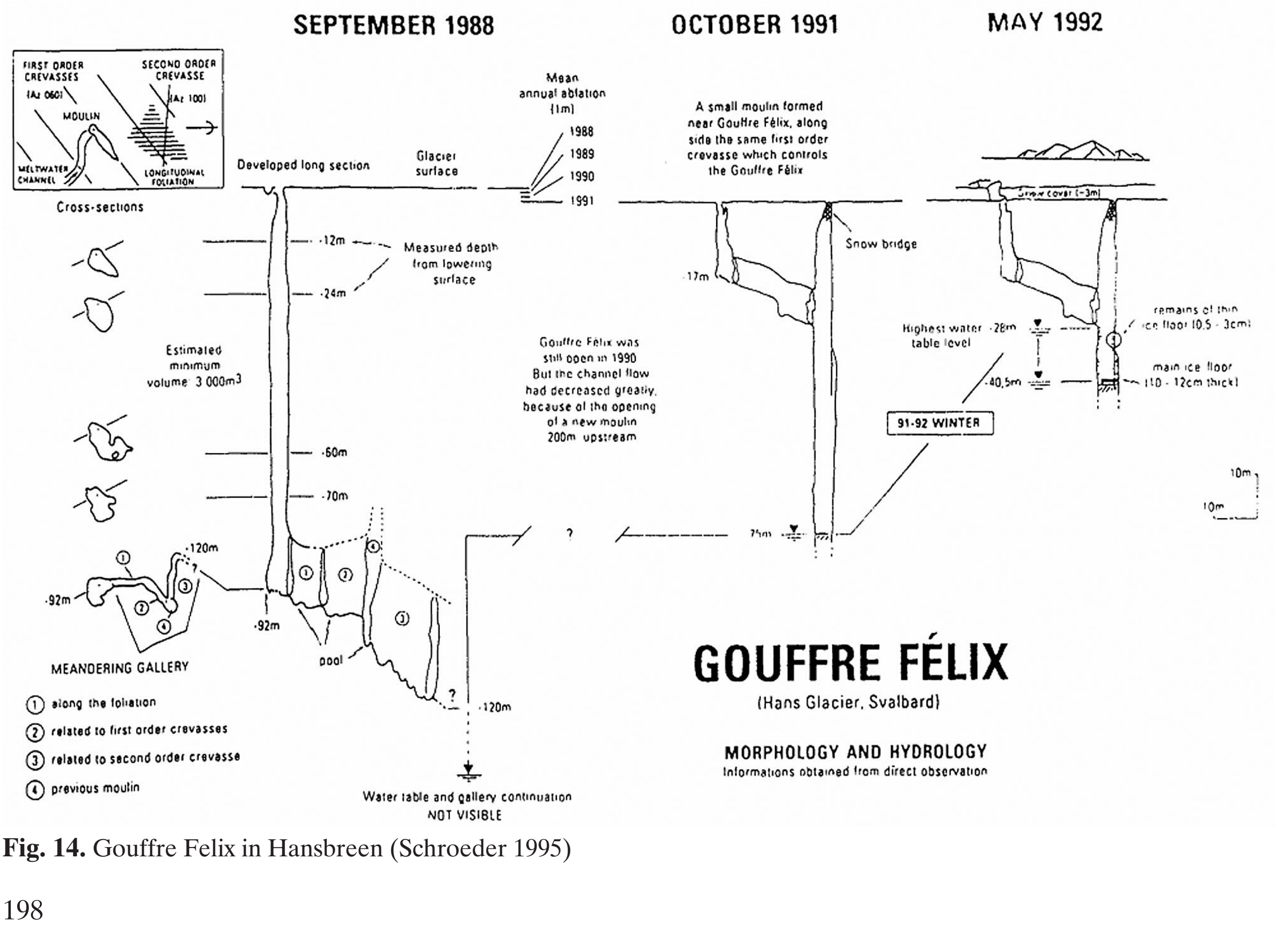


Fig. 14. Gouffre Felix in Hansbreen (Schroeder 1995)

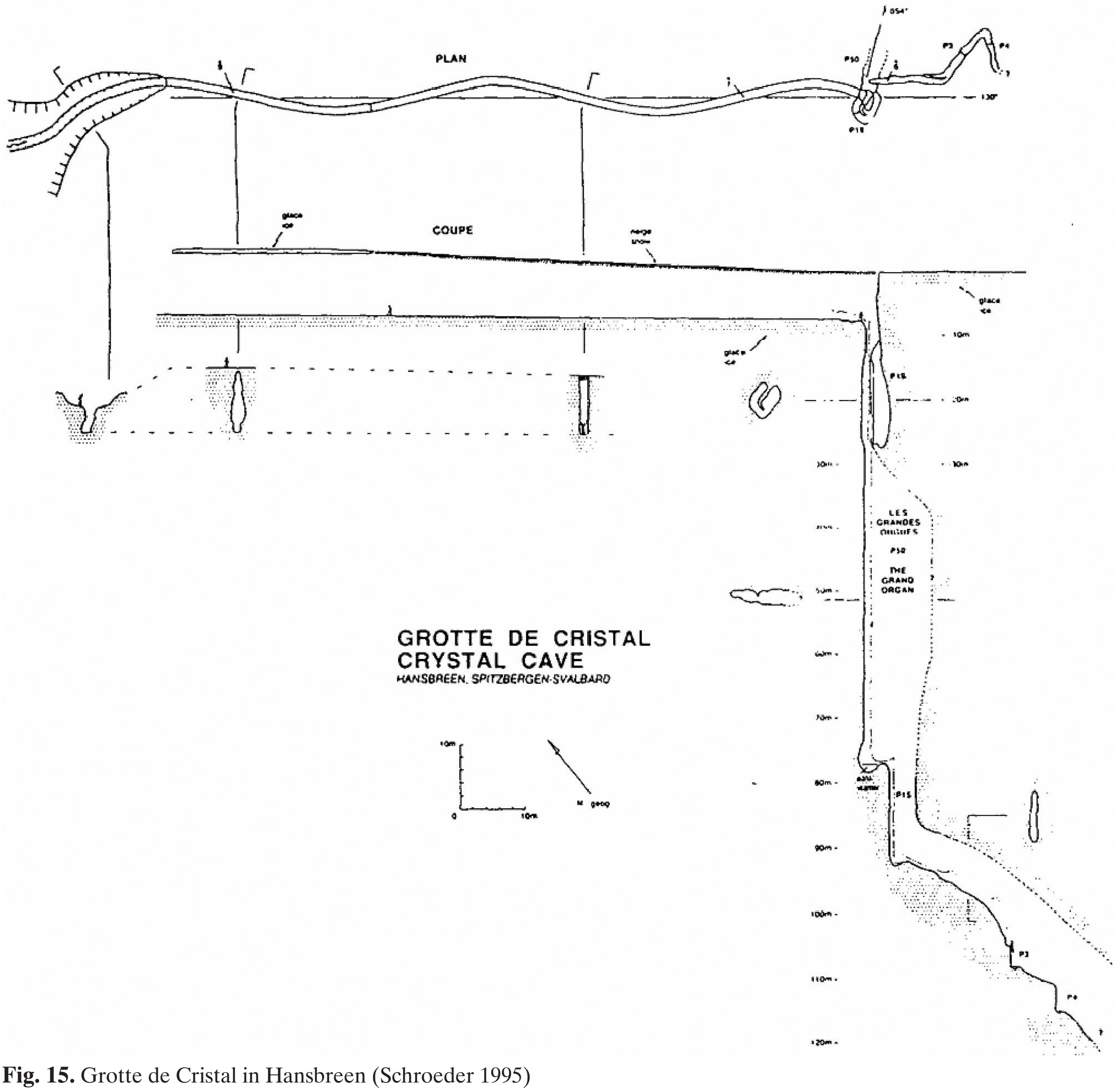


Fig. 15. Grotte de Cristal in Hansbreen (Schroeder 1995)

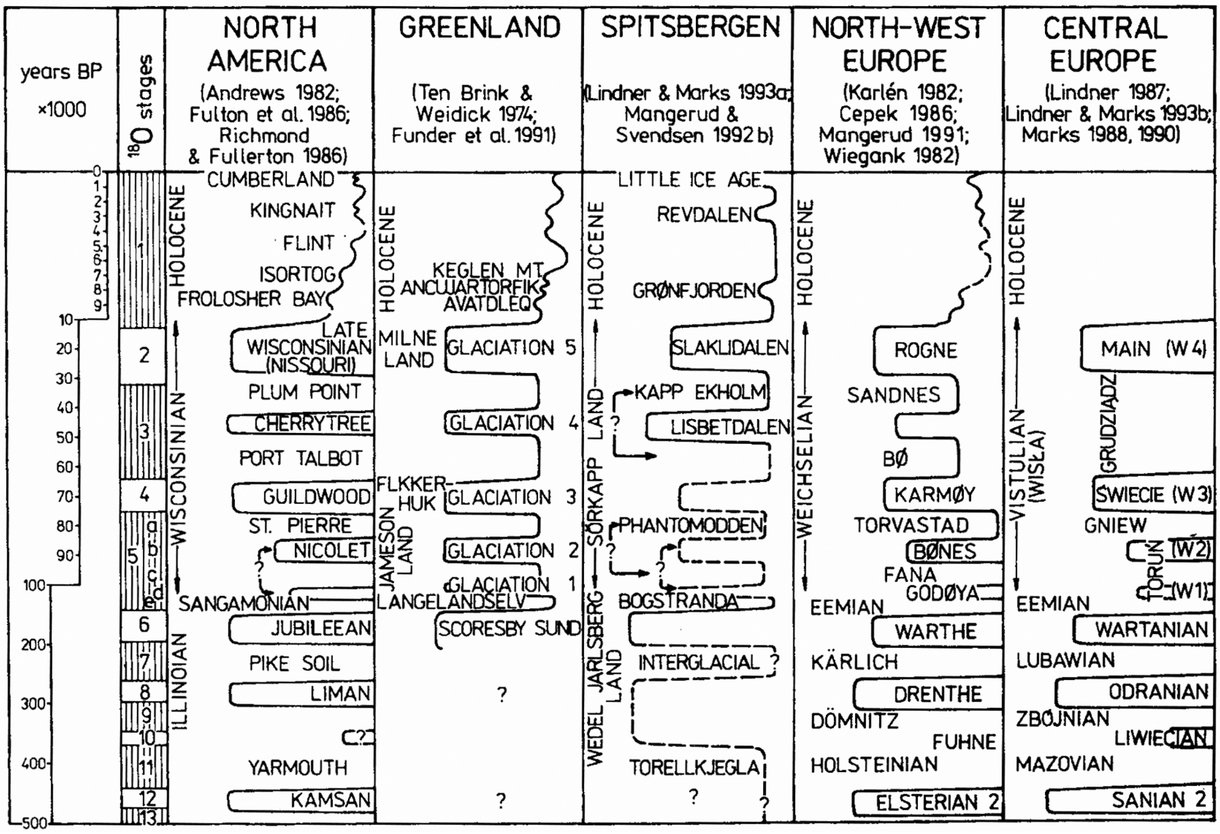


Fig. 16. Correlation of Quaternary stratigraphic schemes for Northern America, Greenland, Spitsbergen and Europe (Lindenr, Marks 1993)

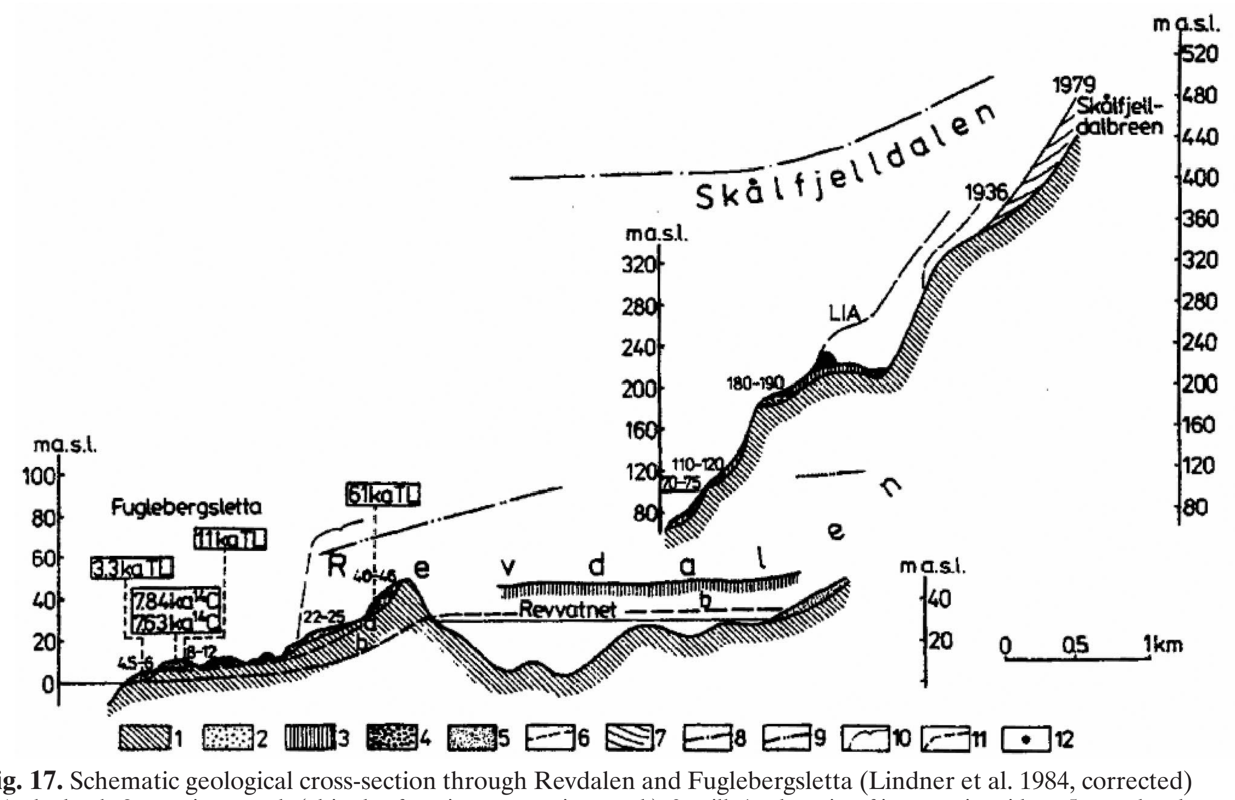


Fig. 17. Schematic geological cross-section through Revdalen and Fuglebergsletta (Lindner et al. 1984, corrected)
1 – bedrock, 2 – marine gravels (altitude of marine terraces in m a.s.l.), 3 – till, 4 – deposits of ice-moraine ridges, 5 – sand and gravel of outwash-plains, 6 – older (a) and younger (b) outwash-plain, 7 – glacier ice, 8 – old glacier fronts, 9 – young glacier fronts, 10 – the biggest range of glaciers in Holocene, 11 – range of glaciers during LIA (Little Ice Age), 12 – sampling sites for dating

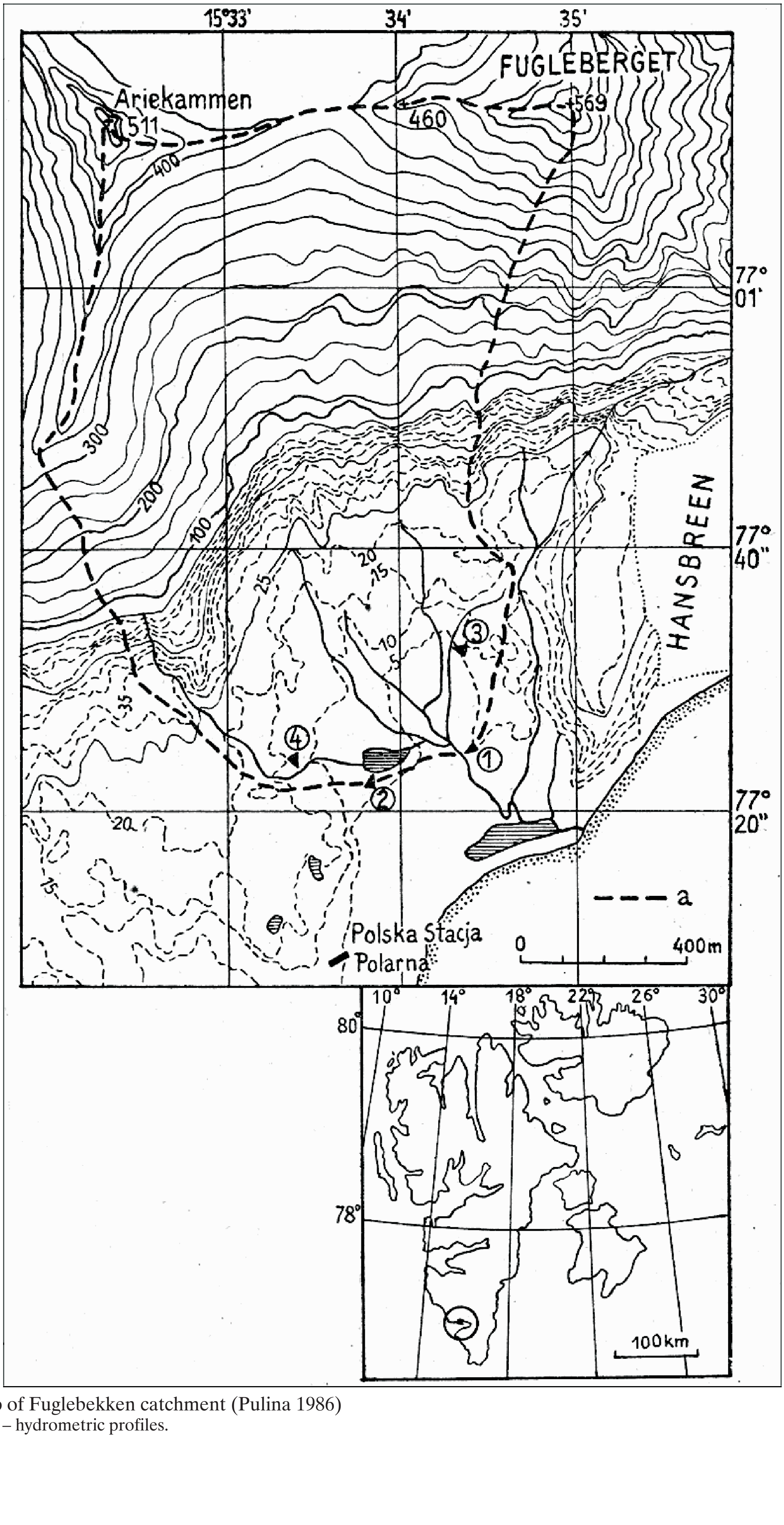


Fig. 18. Map of Fuglebekken catchment (Pulina 1986)
1, 2, 3 and 4 – hydrometric profiles.

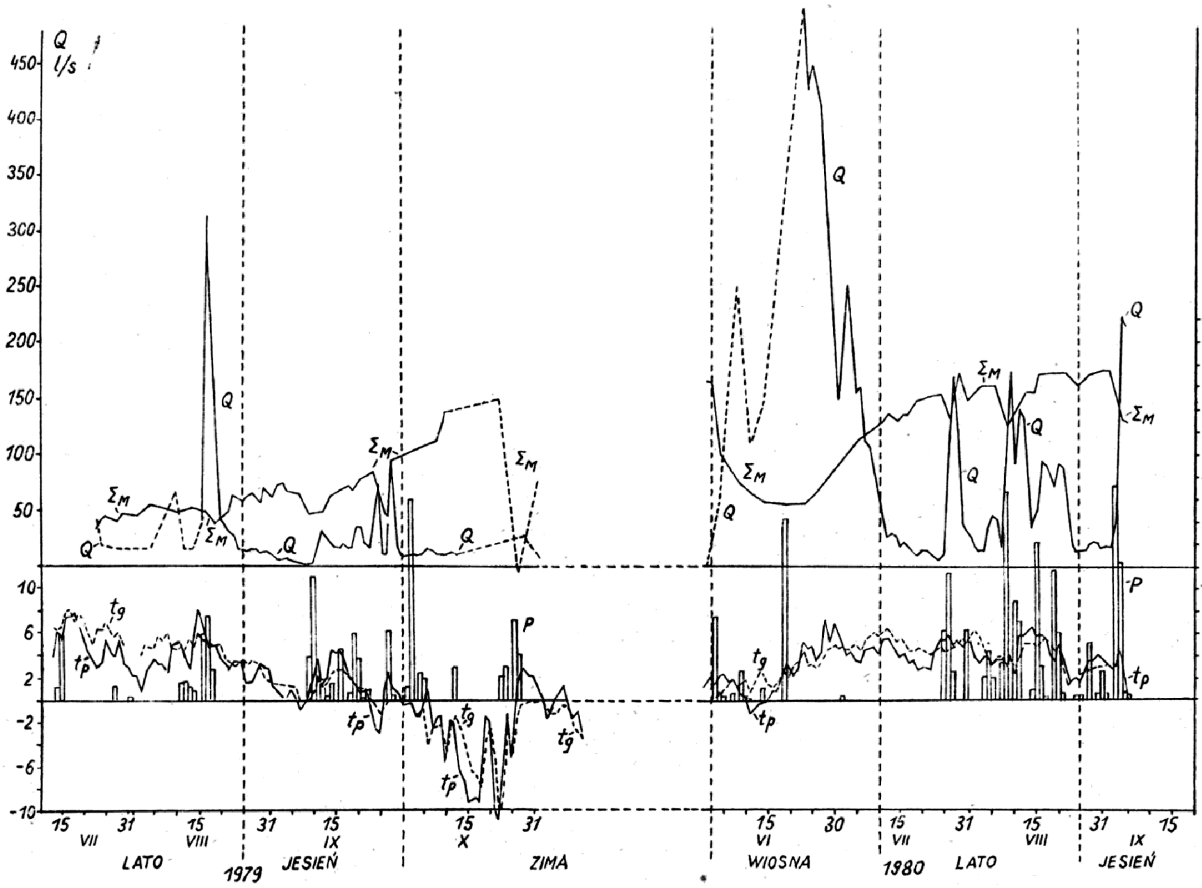


Fig. 19. Course of discharge (Q) in the background of physicochemical properties of stream water and meteorological elements within the Fuglebekken catchment during hydrological year 1979/80 (Pulina 1986)
tg – ground temperature at depth 5 cm, ÓM – total mineralization of stream water, P – precipitation, tp – air temperature.

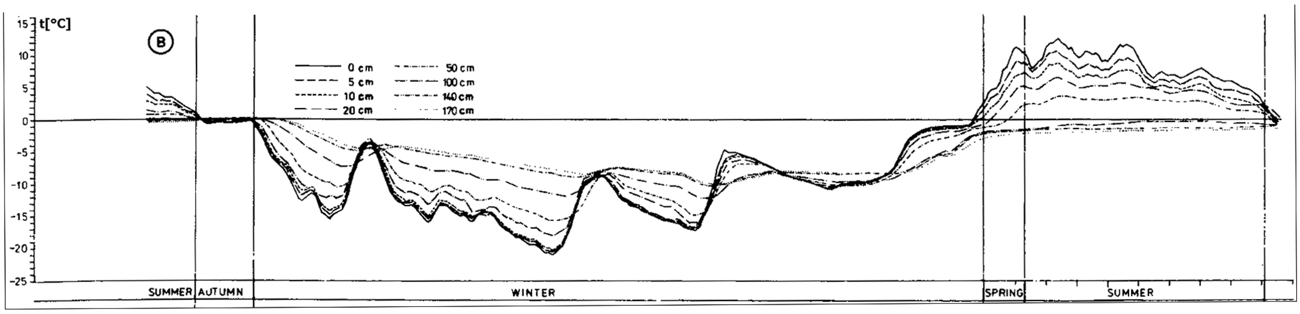


Fig. 20. Annual (1985/86) course of ground temperature at different depth in Fuglebekken catchment (Głowacki et al. 1990)

Fig. 21. Sedimentary covers of the south-west foreground of Hansbreen 1998 (Karczewski, Rachlewicz 2004)
1 – moraine cover of ice-moraine ridges, 2 – supraglacial moraine deposits on the dead ice, 3 – lodgement till on the bedrock, 4 – hydrated suprglacial covers on passive ice, 5 – fluvioglacial sand and gravel, 6 – marginal lake deposits of glacier waters, 7a – bedrock, 7b – fluted moraine, 8 – glacier.

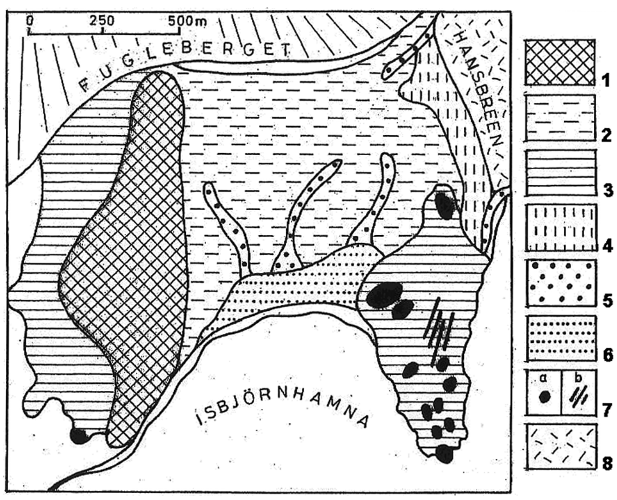


Fig. 22. Glaciers of ice basin Hornsund (J. Jania 2004)
Surge glaciers are marked by asterisk.

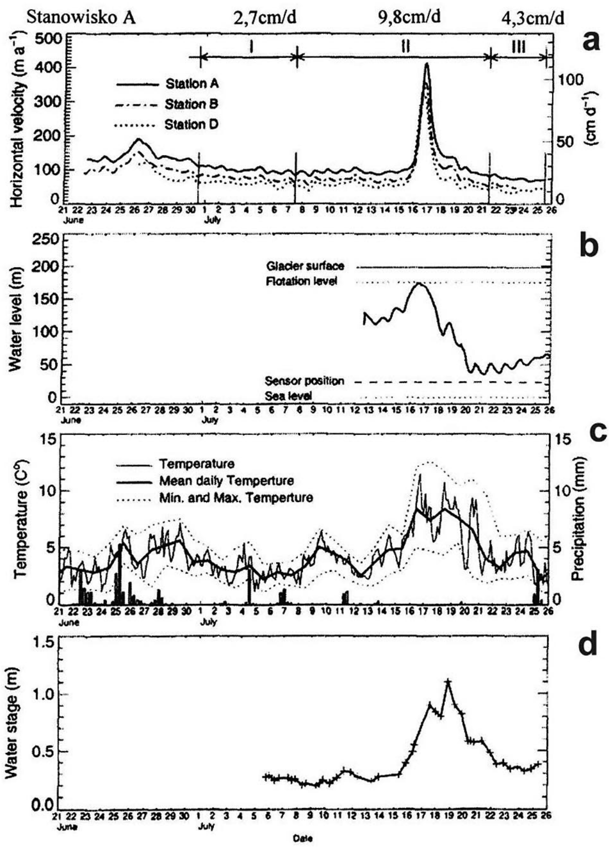


Fig. 23. Dynamics of Hansbreen movements during summer 1999 (Vieli et al. 2003)

a – horizontal velocity, I-III – periods of ablation, b – pressure in ice aven as a water level, c – air temperature and precipitation at Polish Polar Station, d – water level in glacial river of Werenskioldbreen

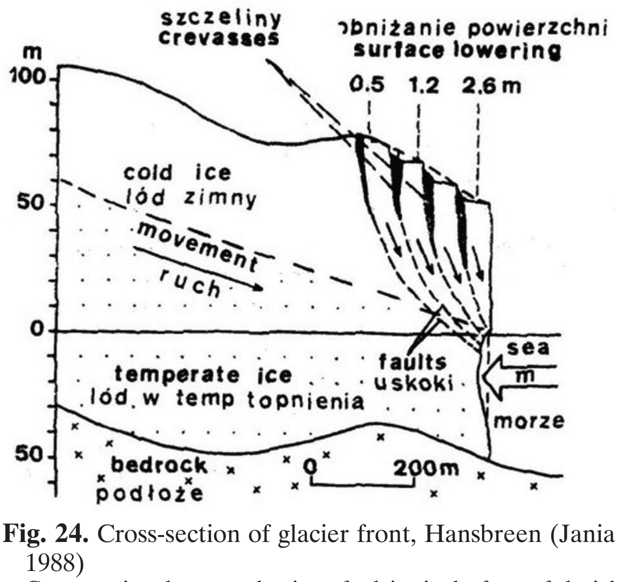


Fig. 24. Cross-section of glacier front, Hansbreen (Jania 1988)

Cross-section shows mechanism of calving in the form of glacial slips; m – subwater melting.

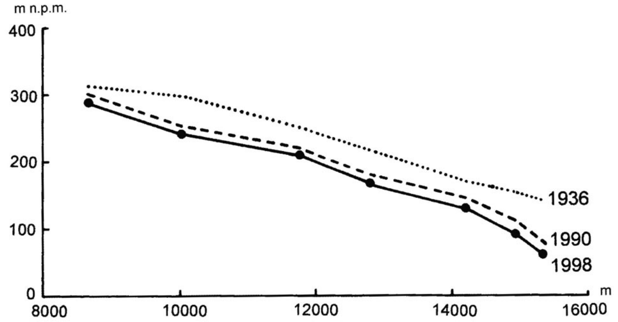


Fig. 25. Changes in thickness of Hansbreen along the ablation zone (Jania 2004)

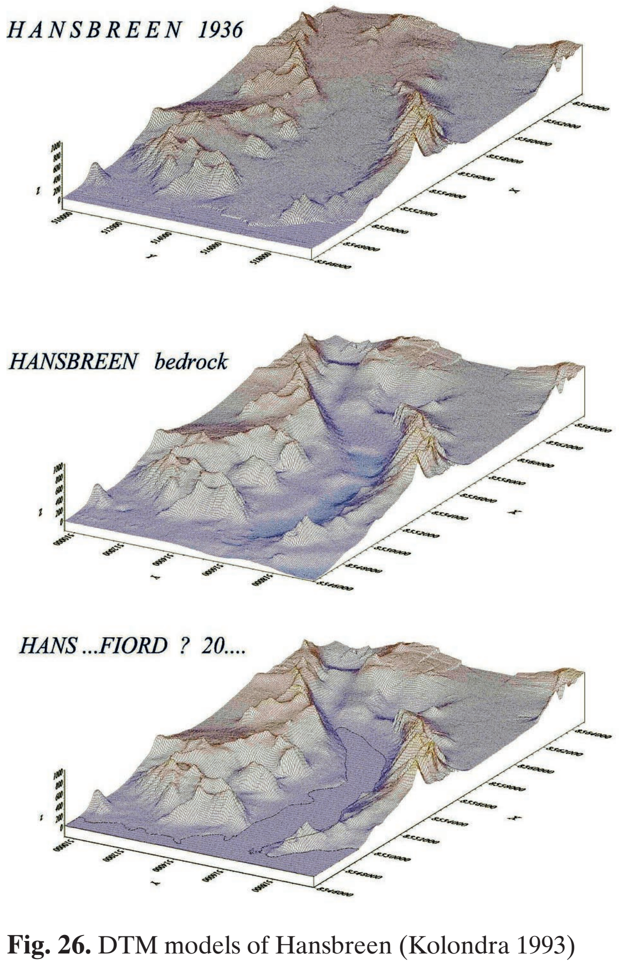


Fig. 26. DTM models of Hansbreen (Kolondra 1993)

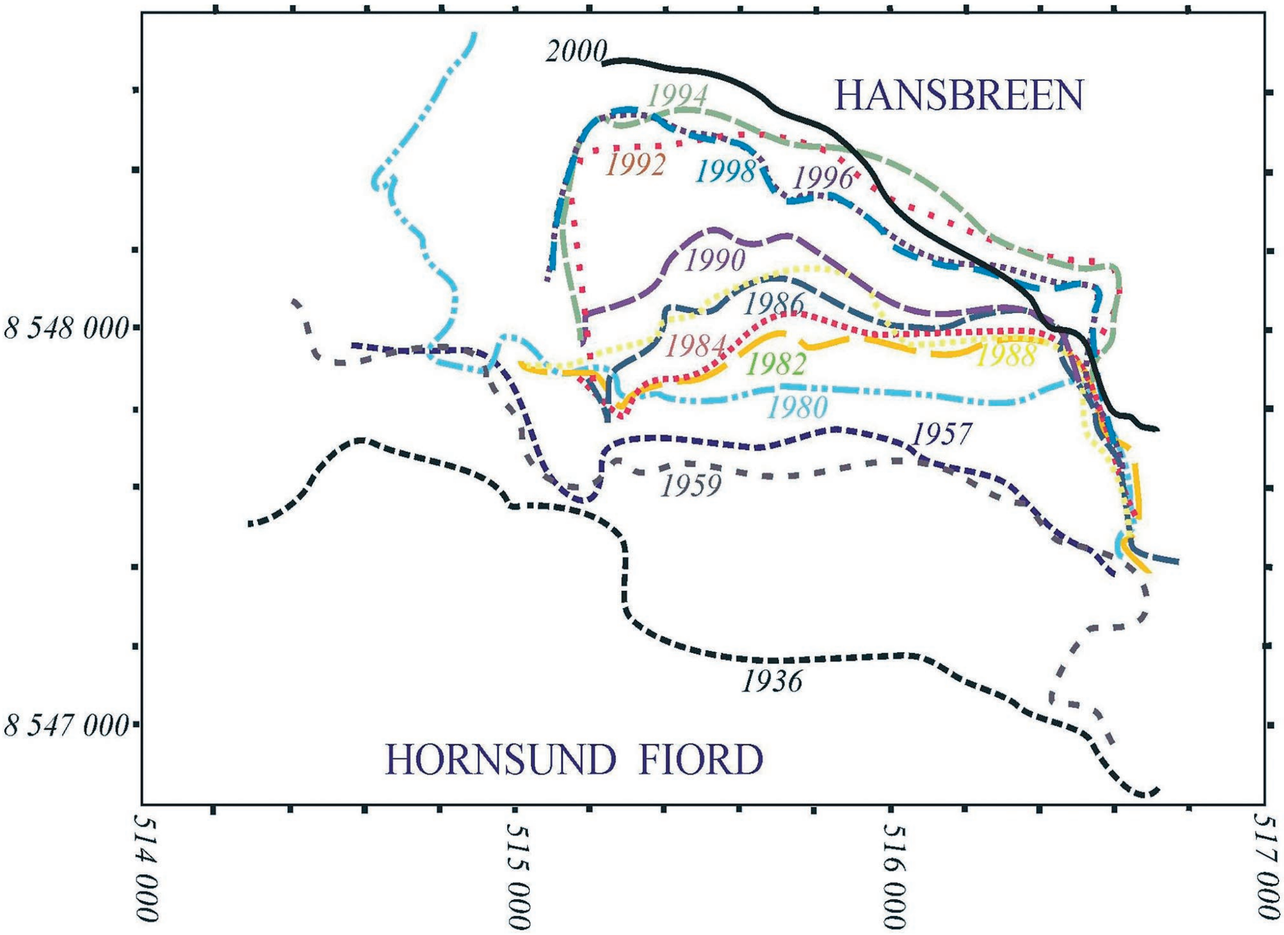


Fig. 27. Changes in front location for Hansbreen (Kolondra 1993)

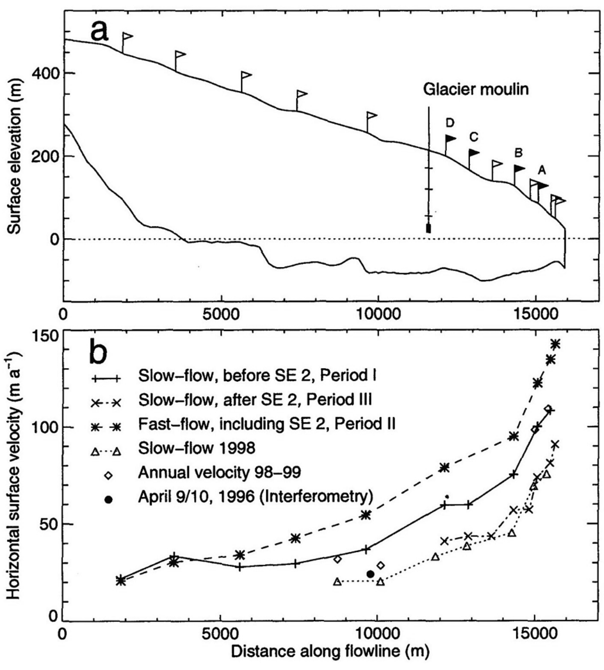


Fig. 28. Velocity of ice surface for Hansbreen (Vieli et al. 2003)

a – sampling sites in glacial avens, b – surface velocity of ice.

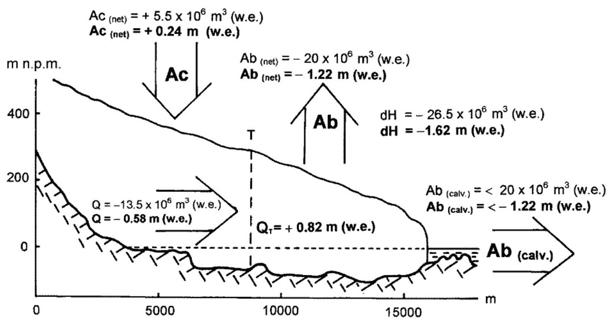


Fig. 29. Mass balance for Hansbreen along longitudinal profile of glacier (Jania 2004) – average values of water equivalent (w.e.) for 1990–1998

Ac – accumulation netto, Ab – ablation netto, Ab (calv.) – ablation by calving, Q_T – flow of ice below ELA, dH – decrease in glacier thickness.

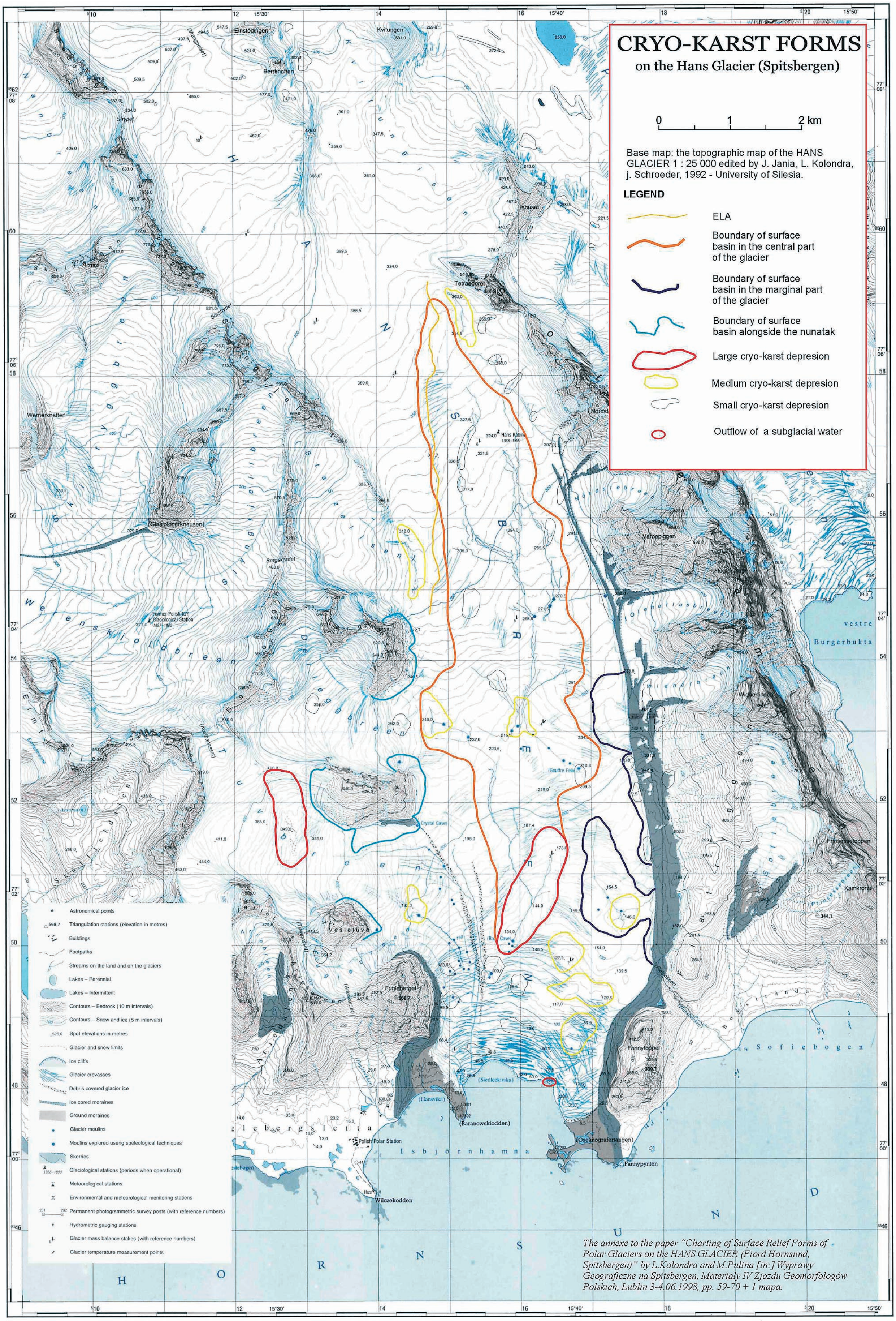


Fig. 30. Distribution of cryo-karst forms on the Hansbreen (Kolondra, Pulina 1998)

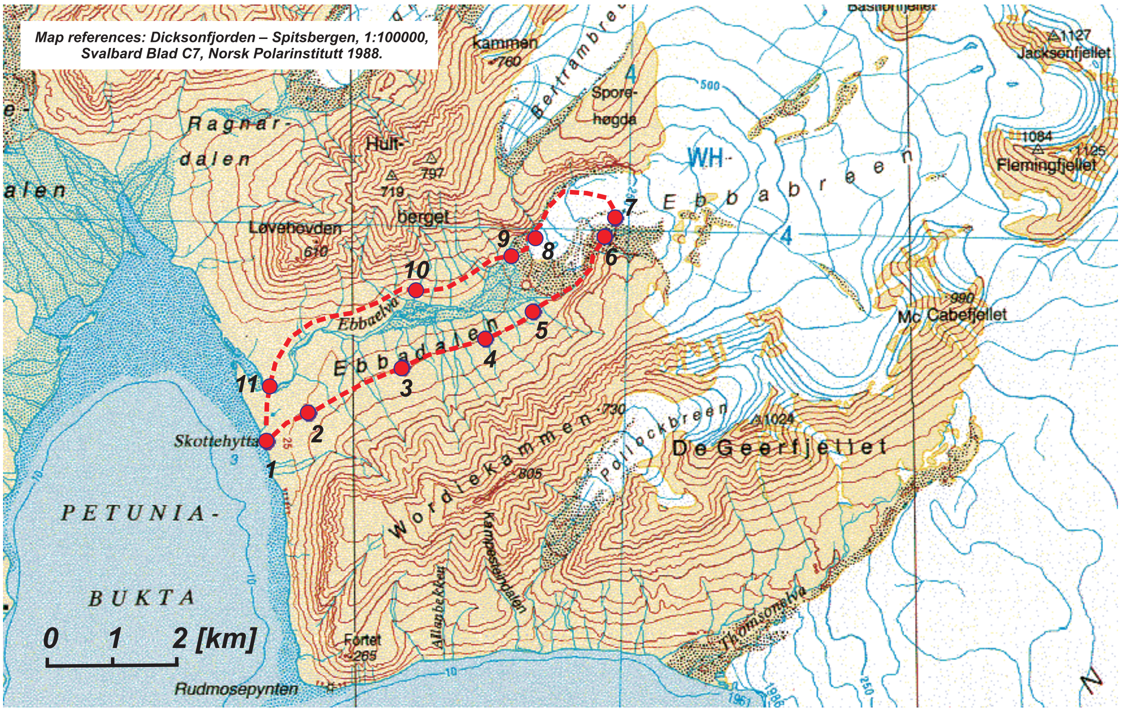
Fig. 31. Stone rings (polygonal soils) in Środoń' garden (photo J. Borysiak)

Fig. 32. Tundra vegetation on the northern bank of Hornsund (photo J. Borysiak)

Petuniabukta: from glacial to paraglacial processes in Ebbadalen – Leader Grzegorz Rachlewicz

Grzegorz Rachlewicz*

Institute of Paleogeography and Geoecology, Adam Mickiewicz University, Poznań, Poland



Petuniabukta – excursion programme

Point 1 – Skottehytta 78°41.944'N 16°36.932'E
Introduction, environmental background
Point 2 – Terrace 78°42.143'N 16°38.428'E
Raised marine terraces, non-glacial catchments
Point 3 – Ebbadalen 78°42.458'N 16°41.790'E
Slope processes, periglacial phenomena
Point 4 – Wordiekammen W 78°42.654'N 16°44.777'E
Sedimentary rocks – Gipsdalen Group
Point 5 – Wordiekammen E 78°42.843'N 16°46.509'E
Crystalline bedrock – Hekla Hoeck Formation
Point 6 – Ebbabreen marginal zone 78°43.368'N 16°49.098'E
Glacial geomorphology and deglaciation since Little Ice Age
Point 7 – Ebbabreen 78°3.495'N 16°49.479'E
Glacial phenomena
Point 8 – Ebbabreen marginal zone, Ebbaelva waterfall 78°43.391'N 16°46.583'E
Marginal glacifluvial outflow
Point 9 – Bertilelva 78°43.219'N 16°45.736'E
Bertilbreen characteristic and glacifluvial processes
Point 10 – Hultberget 78°42.962'N 16°40.928'E
Complex view on Arctic valley system
Point 11 – Ebbaelva mouth 78°42.319'N 16°37.039'E
Catchment closing point and interferences with coastal processes

* e-mail: grzera@amu.edu.pl

Introduction to the excursion

The surroundings of Petuniabukta is an area easy-accessible from Svalbard West coast, in the inner-fiord region of Central Spitsbergen. It is the most North-Eastern tip of the Isfjorden system, what through its location determines geology, morphology and climate features.

The exploration following the search for natural resources and investigations of natural environment in this region started at the turn of 19th and 20th century with expeditions of Baron N.A.E. Nordenskjöld, who gave most of place-names in this area, and activities of Scottish Spitsbergen Syndicate. The share of the control over this territory was crowned by the establishment of settlements and coal-mines Pyramiden by Swedes in 1910 (in 1927 sold to the Soviet Union, operating until 1999) and Brucebyen by Scotts in 1919. The prospection and exploitation of mineral resources, inspite of coal, included also gypsum, uranium and petroleum. About the same age (from 1917) is the cabin Skottehytta, on the eastern coast of Petuniabukta. Later, in the 50. and 60., it was used by Cambridge Geological Expeditions and

Fig. 2. Poznań research station at Petuniabukta

since the 80., "discovered" for scientific use by P. Kłysz in 1979, became a base of Polish expeditions from Adam Mickiewicz University in Poznań (Fig. 2); consecutive leaders: 1984 – W. Stankowski, 1985 – A. Kostrzewski, 1986 – A. Karczewski, 1987 – W. Stankowski, 1989 – A. Karczewski, 2000–2003 – G. Rachlewicz, 2005 – L. Kasprzak, G. Rachlewicz, 2006 – Zb. Zwoliński, 2007 – L. Kasprzak, K. Dragon.

First period of Polish investigations in Petuniabukta (1984–1989) realized geomorphological mapping around Petuniabukta and a general subject "Quaternary palaeogeography and present-day processes in an area between Billefjorden and Austfjorden, central Spitsbergen". Among other publications a map and a volume of papers cited below were published. In the year 2000 a new project "Matter circulation in the Arctic terrestrial-marine geoecosystem on the example of Billefjorden" has been started. It is continued until present days preparing and realizing a part of Polish National Project for the International Polar Year 2007–2008 "Structure, evolution and dynamics of lithosphere, cryosphere and biosphere in the European sector of the Arctic and in the Antarctic".

Main topics of the current project in Petuniabukta are covering:

– geology, geomorphology and Quaternary paleogeography;
– meteorology and environment reactions to climate changes with special attention paid to glaciers and permafrost;
– morphology and functioning of glaciers marginal zones;
– spatial and temporal mass fluxes in terrestrial and marine environments.

Petuniabukta, within an area of average glaciers coverage and specific alteration of quasi-continental climate, not observed on the Western coast of the Island, offers a variety of interesting examples to study past and present features of the natural environment, like: unique and diversified geology, activity of various morphogenetic processes, unequivocal linkage between terrestrial and marine systems, limited human impact, and finally easy access to investigation sites (Fig. 3). Further investigations in the Billefjorden region, with possibilities of their expansion on neighboring areas, will continue with more detailed approach and advanced instrumentation to obtain long-term observation databases.

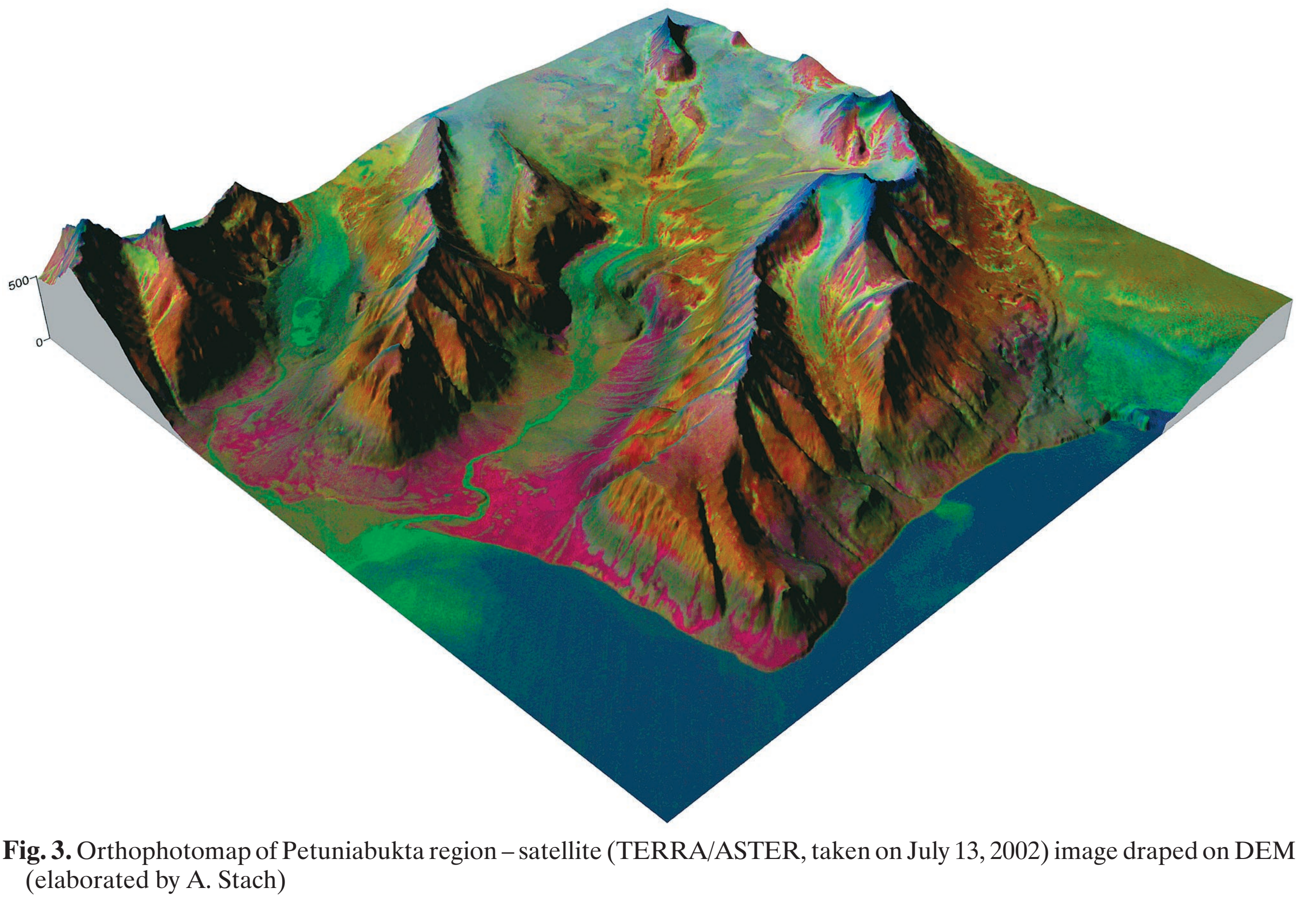


Fig. 3. Orthophotomap of Petuniabukta region – satellite (TERRA/ASTER, taken on July 13, 2002) image draped on DEM (elaborated by A. Stach)

Geological setting of the Petuniabukta Region

Witold Szczuciński
Institute of Geology, Adam Mickiewicz University, Poznań, Poland

Grzegorz Rachlewicz*
Institute of Paleogeography and Geoecology, Adam Mickiewicz University, Poznań, Poland

Surrounding of the Petuniabukta (Petunia Bay) offers a spectacular insight into geological history of Svalbard (Fig. 4) and modern geological processes. Due to being cut by one of the most important fault zones in the region – N-S trending Billefjorden Fault Zone (Fig. 5) a complex set of rocks is visible now on the earth surface. One can see old crystalline rocks next to clastic sandstones, coal measures and created in hot and dry conditions sequences of carbonates, anhydrites and gypsum. Valley floors and fjord bottom are covered by products of the youngest processes dominated in contrast to older rocks by facies typical for polar climates. These processes are still active and intensive formation of new sediments and alteration of older rocks are well visible. This region was in focus of geological investigations from the beginning of the 20th century – partly due to long lasting exploitation of coal (mainly in Pyramiden). That interest is documented by production of geological maps, books and many scientific papers related to that region (selected further reading is attached at the end).

The geological setting is dominated by N-S trending Billefjorden Fault Zone – (BFZ) and related Billefjorden Trough. The complex history of the fault zone activity caused that western and eastern coasts of Petuniabukta are dominated by different lithological units – Devonian clastic rocks on the west and carbonate Carboniferous sequences on the east. The faults of the BFZ are well visible in many places (Figs. 5–7) and sedimentary rocks adjacent to them are often also deformed due to movements along the faults. The BFZ have been active with various intensity since Precambrian times, but the most intense movements along it occurred during early Paleozoic when horizontal dislocation of rocks for several thousands of kilometers and vertical dislocation up to 20 km took place. In the recent times small and rare earthquakes and hydrothermal springs are observed along this zone. The complex tectonic history of the region resulted in four structural units, which are separated by unconformities:

- The oldest structural units is composed of Precambrian rocks often called the Hecla Hoek Succession (or Pre-Old Red rocks), which were engaged in the Caledonian Orogeny. They are represented mainly by various crystalline (igneous and metamorphic) rocks;
- Next unit is of Devonian age and comprise of sedimentary rocks – mainly sandstones and mudstones. It is preserved only on the western side of BFZ;
- Carboniferous-Permian rocks are represented by various sedimentary rocks: conglomerates, sandstones, mudstones, limestones, coal (exploited till 1998), gypsum, anhydrites and dolomites. The lateral and vertical variety of rock types in the unit is due to BFZ activity during that time;
- The youngest structural unit is of Quaternary age. It consists of various sediments: glacimarine muds, beach sands and gravels, intertidal deposits, glacial and glacifluvial terrestrial deposits in marginal zones of glaciers and sedimentary covers on slopes.

* e-mail: grzera@amu.edu.pl

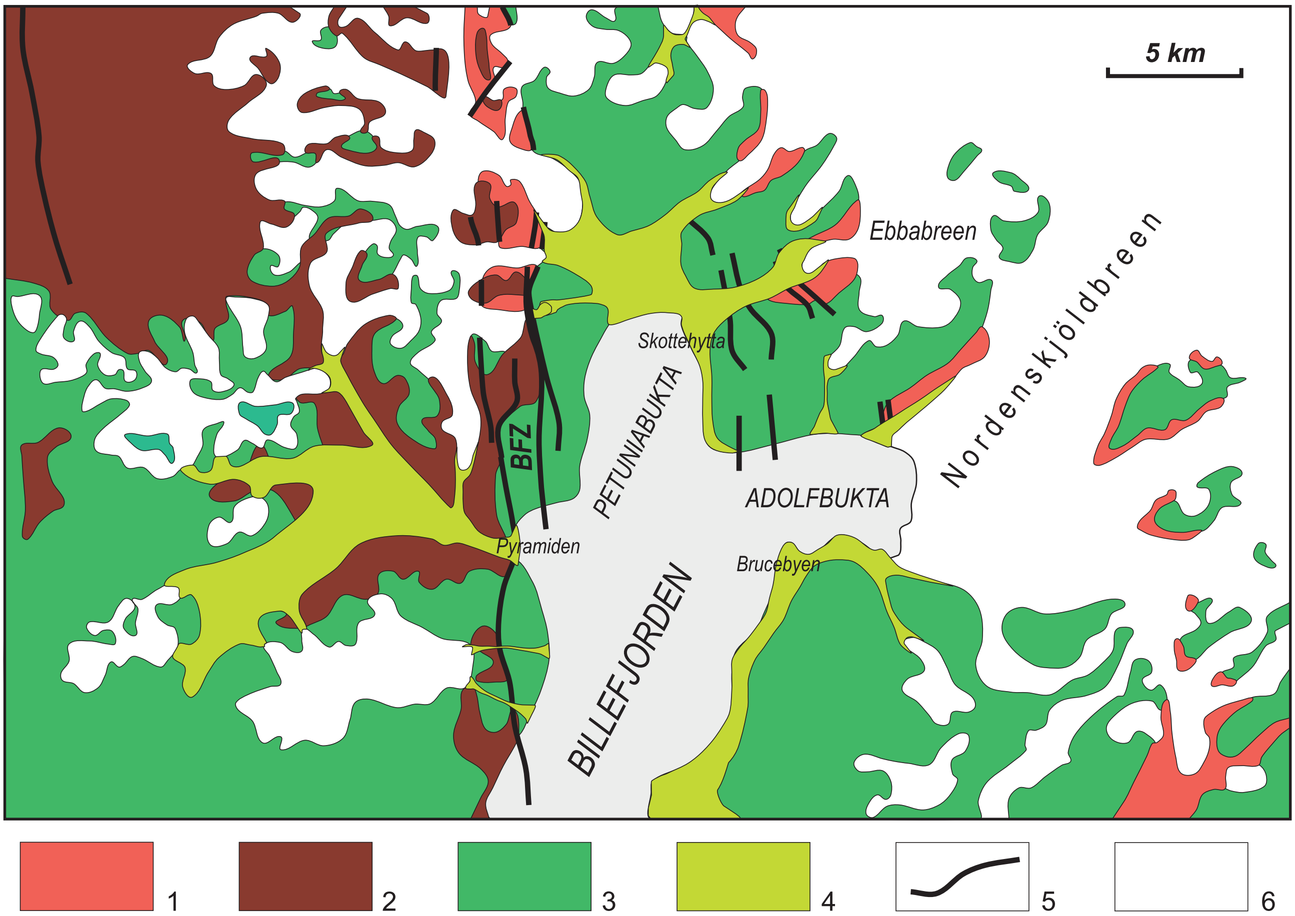


Fig. 4. Simplified geological sketch of Billefjorden, after Dallmann et al. (1999)
1 – Precambrian crystalline basement; 2 – Devonian sedimentary rocks; 3 – Carboniferous-Permian sedimentary rocks; 4 – Quaternary covers; 5 – main faults; 6 – glaciers; BFZ – Billefjorden Fault Zone.

Pre-Old Red rocks are visible as isolated outcrops of usually dark and resistent rocks. The most common rock types are: gneiss, schist, phyllite, amphibolite and syenite, but granite, quartzite and marble are also documented. The pre-Devonian formations were subjected to block tectonics and multiple folding. The whole complex formed for about a billion of years, reaching thickness up to 18 km . The complex starts with transformed into amphibolites former volcanic rocks (the lowest 12 km) covered with clastic rocks (gneisses, schists, phyllites, quartzites), carbonate rocks (marbles), tillites and finally carbonate-dominated layer. At the time of their formation and after that several igneous intrusions took place (granites, syenites). Isolated outcrops of the Pre-Old Red rocks offered an opportunity to use them as indicators of the direction of glacial transport. Erratic boulders derived from them were found on mountain fields (built of younger sedimentary rocks) in valleys and in fjord sediments (drop stones).

The next structural unit is composed of Devonian clastic rocks with admixture of carbonates and coals. Most of them belong to Wood Bay Formation, which consists of typical 'old red facies' – red shales intercalated with sandstones and conglomerates. They are famous because of common fish fossils. In the late Devonian rocks appear coal seems, sandstones with common fossil plants and rare limestones.

Most of the rocks visible around Petuniabukta is of late Devonian – Carboniferous – early Pemian age and belong to Billefjorden and Gipshuken (or Gipsdalen) Group. They consist of several formations, which are represented by clastic rocks (conglomerates, sandstones and mudstones) with coal seams (mined in Pyramiden) and the most common carbonate rocks (limestones, dolomites) with anhydrite and gypsum strata. Most of these rock types are represented in Ebbadalen Formation, which has its stratotype on the northern slopes of Wordiekammen in Ebbadalen (Ebba valley). The age of this formation was established on the base of brachiopoda and foraminifera fossils to the mid-Carboniferous (Bashkirian or even slightly earlier). The thickness is from 0 to 550 m, of which 280 m is present in the stratotype. The formation lies in an asymmetric basin, about 18 km wide, elongated parallel to the BFZ. Its largest thickness is observed in the near-fault area and is thinning eastward. Facies in the Ebbadalen Formation are variable and their

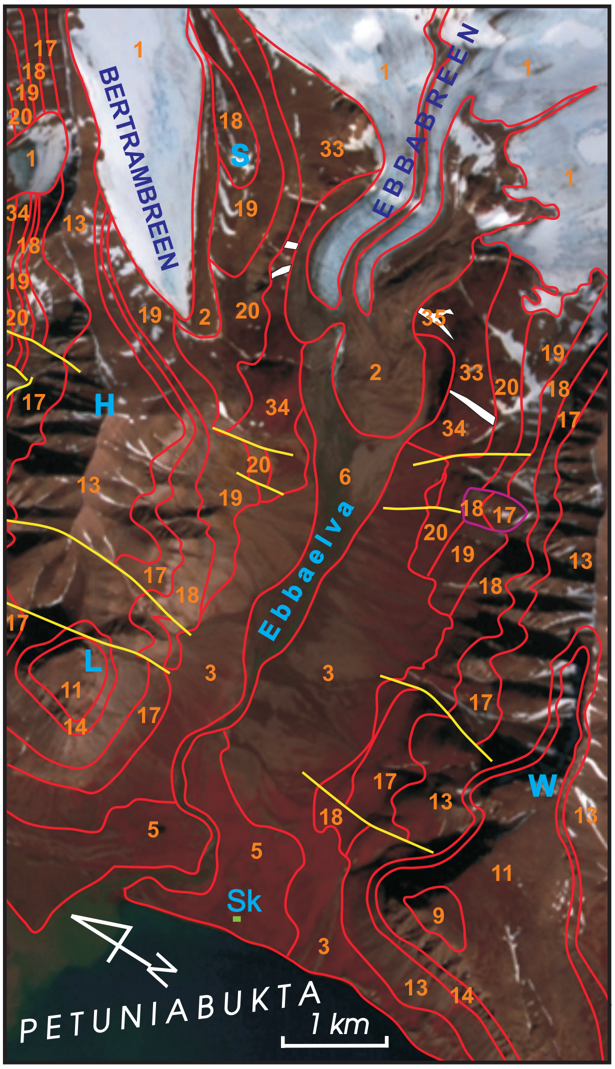


Fig. 5. Geology of Ebbadalen (after Dallmann et al. 2004) on the background of a part of TERRA/ASTER satellite image from 2002-07–13

Red lines – rock boundaries; yellow lines – faults; magenta lines – boundary of landslides. QUATERNARY: 1 – glaciers; 2 – moraines (Holocene); 3 – slope deposits (talus and undifferentiated material, Holocene); 5 – marine shore deposits (Holocene); 6 – Fluvial and glacifluvial deposits (Holocene); Carboniferous: Gipsdalen Group – Dickson Land Subgroup: 9 – Gipshuken Fm. (gypsum/anhydrite, dolomite breccia, dolomite and limestone); 11 – Wordiekammen Fm. (limestone and dolomite, sandstone, mudstone); Campbellryggen Subgroup: 13 – Minkinfjellet Fm. (sandstone, dolomite, gypsum/anhydrite); 14 – Fortet Mb. (dolomite solution breccia); Ebbadalen Fm.: 17 – Trikolorfjellet Mb. (gypsum/anhydrite, dark limestone); 18 – Ebbaelva Mb. (multicolored sandstone, shale, limestone, dolomite, gypsum/anhydrite); 19 – Hultberget Fm. (red sandstone, shale and conglomerate); Billefjorden Group: 20 – Hörbyebreen and Mumien Fms. (sandstone, conglomerate, shale and coal); Paleo- and Mesoproterozoic: 33 – Smutsbreen Unit (garnet-mica schist, calc-peltic schist and marble); 34 – Eskolabreen Unit (biotite (amphibole) gneiss, amphibolite, granitic gneiss); 35 – distinct marble layers within other basement units. S – Sporehøgda; H – Hultberget; L – Løvehovden; W – Wordiekammen; Sk – Skottehytta.

Fig. 6. Billefjorden Fault Zone – a pronounce fault separating Proterozoic rocks (on the left on slopes of Faraonfjellet) from light Carboniferous carbonate rocks (Cheopsfjellet) In the very back, a snow covered Devonian sedimentary rocks (Karnakfjellet)

probable sedimentary environments were lakes, alluvial fans, braided rivers, estuaries, deltas, sebkhas, lagoons and beaches. Its lower part is composed of gray and yellow sandstones interbedded with grayish-green schists, anhydrites, conglomerates and red sandstones. The upper part is mainly consists of carbonate and sulphate rocks formed probably in sebkha environment. Due to its relatively higher resistance they are very well visible on mountain slopes forming cliffs. Within the carbonate and sulphate rocks karst forms have developed (Fig. 8).

The youngest unit is represented by Quaternary sediments. They are mainly from the Holocene period because the fjord was deglaciated about 10,000 years ago. Only in few places older sediments are preserved in raised marine terraces (in Hörbyedalen, Ebbadalen and in well known Kapp Ekholm section in the middle part of Billefjorden). During the early Holocene the whole region was glacioisostatically uplifted and associated relative sea level fall was more than 90 m. Due to that along the coast of Petuniabukta are well preserved raised marine terraces composed mostly of sand and gravels, but locally also of finer deposits. Their thickness is usually within 1 to 2 m, but in some cases even about 20 m high paleo-spits are preserved. During the Holocene extensive slope covers and alluvial fans have developed. Several erosional cuts show that their thickness is up to 10 m. They are composed of poorly sorted debris, which is locally intercalated by well sorted finer material. Paleosoil found within them suggest complex evolution with periods of slope stabilization. It is believed that during most of the Holocene glaciers were much smaller than now. Their advance started probably around 3000 years ago and maximum extent was reached during the Little Ice Age, which terminated at the end of 19th century. Since that time the glaciers are continuously retreating with rates from few to about 50 m per year. Their recession is associated with deposition of glacial sediments forming ice-cored terminal moraines and lat-

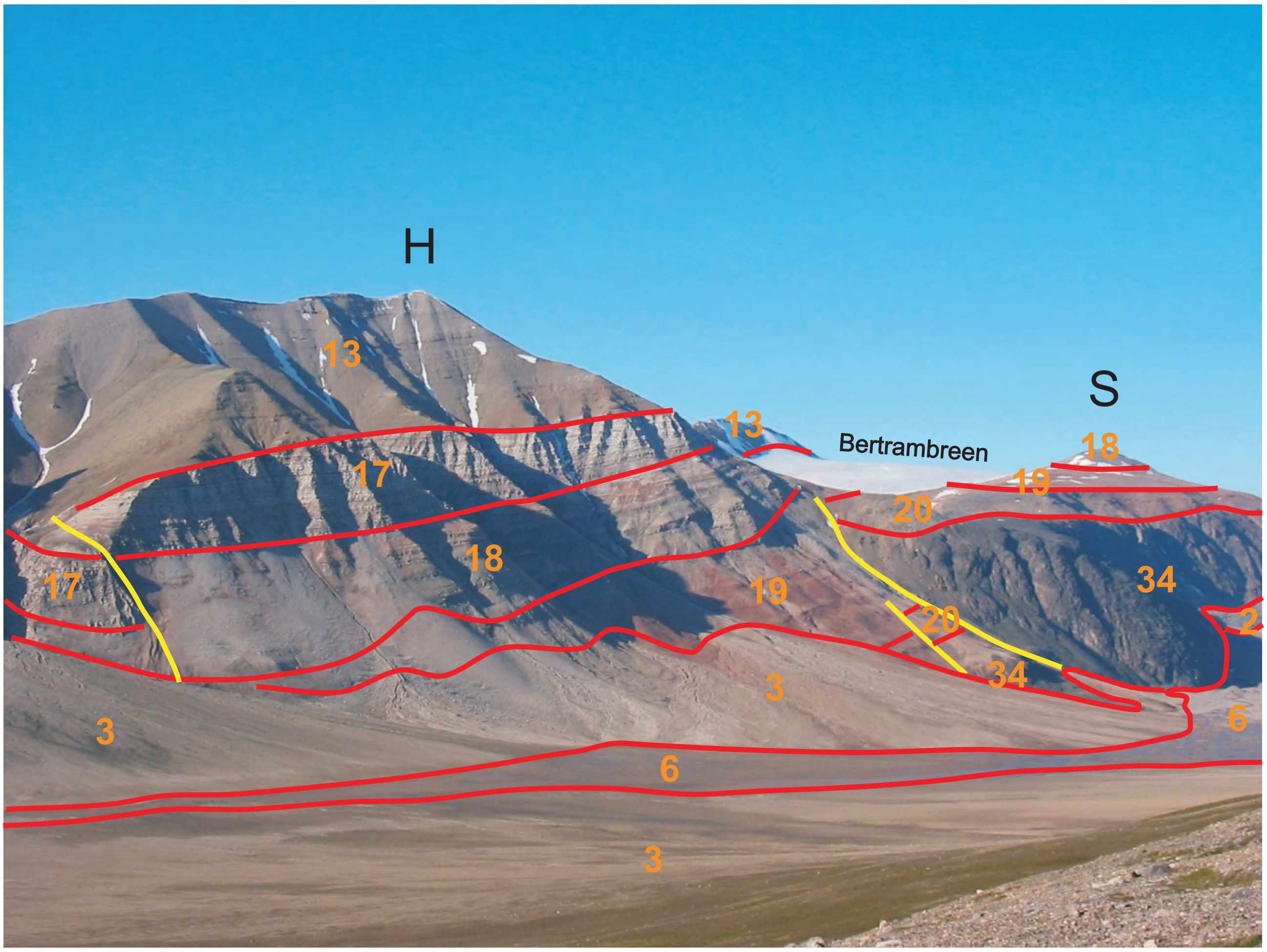


Fig. 7. Northern slopes of Ebbadalen: Hultberget and Sporehøgda massifs. Explanations on fig. 5

eral moraines covered by about 2.5 m thick debris mantle. There are also deposits related to basal deposition (lodgement till) and push moraines. Due to meltwater circulation sediments were left in form of eskers, kames and extensive outwash plains. The latter contain up to 20 m of sediments, so they play important role in sediment storage in a glacial system. Glaciers around Petuniabukta are land terminating and glacial rivers enter the fjord through up to 2 km wide tidal flat. It serves as transfer and storage zone for sediments and is shaped by tidal action (tidal amplitude up to 1.5 m), waves, shore ice, and glacial rivers. The intertidal zone is built mostly of silts and sands and most of sedimentation occurs at its margin (with particulate matter flux up to 90 $gm^{-2}hour^{-1}$) causing its successive progradation. Further in the bay the accumulation rate is much lower and the annual average sediment accumulation is in order of mm per year in the main fjord basin. The fjord floor is covered with glacimarine muds and their maximum thickness is in Adolfbukta (next to tidewater glacier – Nordenskiöldbreen) and reach up to 25 m.

Fig. 8. Weathered surface of Carboniferous anhydrite from Ebbadalen Formation (Ebbadalen)

Geomorphology outline of the vicinity of Petuniabukta

Grzegorz Rachlewicz*

Institute of Paleogeography and Geoecology, Adam Mickiewicz University, Poznań, Poland

The relief of coasts, valleys and mountain massifs around Petuniabukta reveals a variety of interesting and unique features. Landscape associations are diversified according to their genesis, intensity of geomorphic processes and age. Main stream of this branch of research arose here as the aftermath from Poznań University expeditions in the last three decades.

Main agents in shaping primary features of landscape were associated with extensive Quaternary glaciations finished ultimately 10 ka BP. The traces of at least four major advances of Spitsbergen – Barents Sea ice-sheet were detected in the not far Kapp Ekholm section. The most spectacular effects of their activity are large valleys and fiords. The last, widespread episode of glaciers advances during the Little Ice Age (LIA – 600–100 BP) was responsible only for the architecture of valley marginal zones.

In Billefjorden, which is glaciated in about 44%, among 23 existing glaciers only one (Skansdalsbreen) was reported to surge after the LIA. Some premises within the wide area of sharp, non ice-cored ramparts, suggest also a possible surge in the case of Hörbyebreen. Non-surging glaciers commonly leave marginal zones in form of a set of ice-cored morainic ridges. Their setting is closely connected with the layout of the hard-rock basement. In Petuniabukta it can be observed on examples of Svenbreen and Ebbabreen, terminating next to hardly resistant crystalline thresholds. The Ebbabreen LIA marginal zone is located beneath a 50 m high gneiss step, transverse to the valley axis. Valley slopes are dominated by egzaration relief with polished surfaces, striae and glacial undercuts at the height of 50 m above the valley floor. In the upper part this level is marked by belts of lateral moraines. The marginal zone is shaped in form of an asymmetric oval. Maximum heights of frontal moranic rampart, elevated 20–25 m above the valley floor, lined with outwash sediments, are located in the southern wing. Mass movements on slopes of ice-cored moraines are the most intensive here, filling up englacial voids and crevasses with debris-slides and melt-water derived material. Central part of marginal zone is occupied by a depression with small lakes, drained through a system of ice-cracks, to the springs on the edge of the marginal zone. The central part of terminal moraine continues up the glacier in the form of supraglacial belt, connecting the Bastion nunatak in the central part of accumulation area with the edge of ice. The northern part of marginal zone reveals confined amount of morainic material, as a discontinuous cover on roches moutonné of the crystalline threshold. A spectacular waterfall of the main subglacial outflow from the glacier margin is located beneath it. This outflow generates in majority outwash series at the bottom of the valley. Some smaller hillocks in this part may suggest earlier abrupt slide of ice in the steeper part of the basement rocks. Another type of marginal zone can be observed in the case of of Ragnarbreen, showing erosional features in the form of vast depression, taken by the proglacial lake framed by a garland of ice-cored moraines.

Rock walls dominating over valleys, glacieted mostly in upper parts, undergo intensive weathering processes. Beneath flat field surfaces develop structural features of hardness dependent rock outcrops

* e-mail: grzera@amu.edu.pl

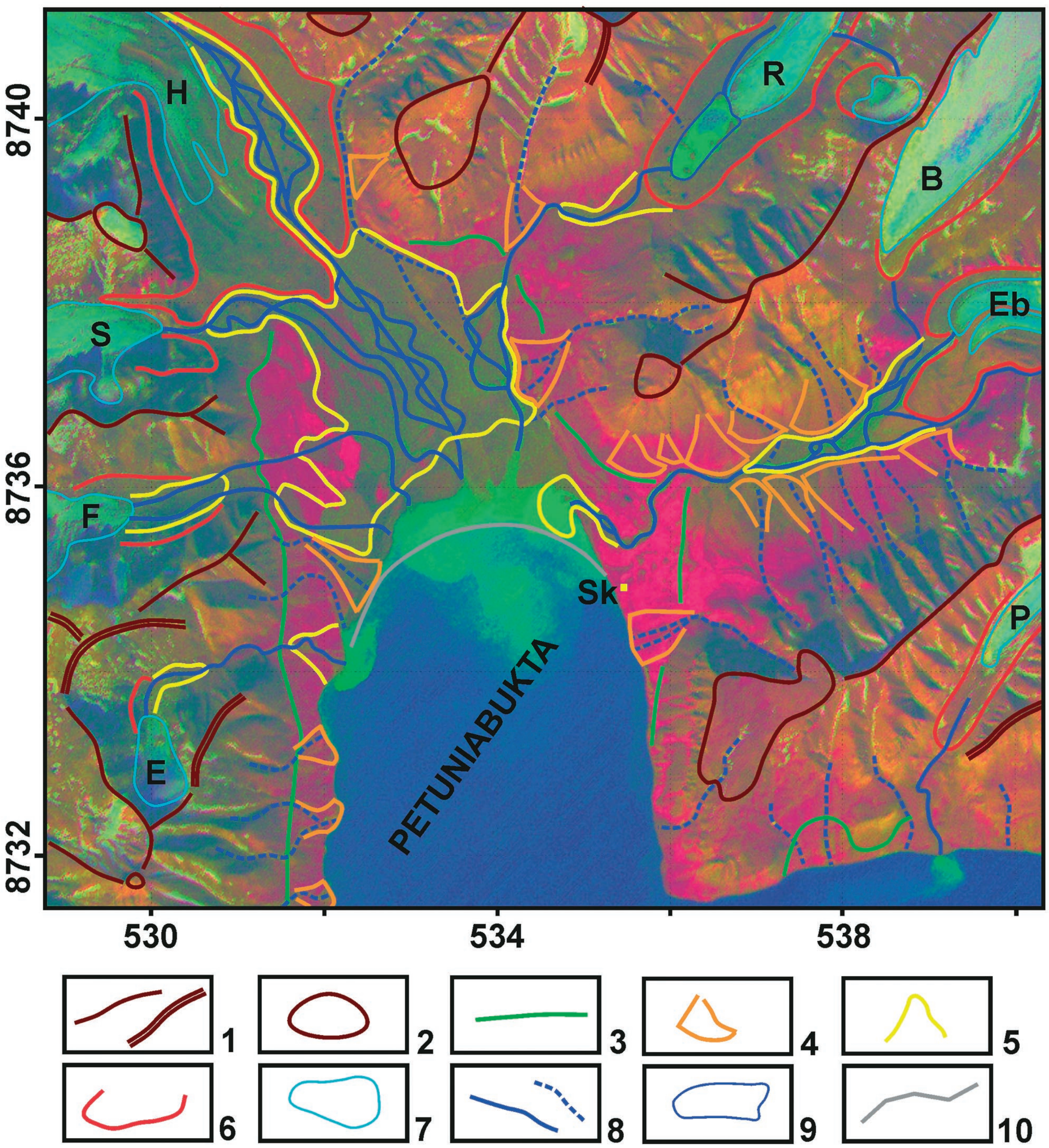


Fig. 9. Main features of geomorphology of Petuniabukta (updated after Karczewski et al. 1990)
1 – Narrow and rounded mountain crests; 2 – Flat top structural surfaces; 3 – Extent of raised marine forms; 4 – Alluvial fans; 5 – Outwash planes; 6 – Marginal zones of glaciers; 7 – Glaciers; 8 – Periodic (proglacial) and episodic streams; 9 – Lakes; 10 – tidal flat. E – Elsabreen; F – Ferdinandbreen; S – Svenbreen; H – Hörbyebreen; R – Ragnarbreen; B – Bertrambreen; Eb – Ebbabreen; P – Pollockbreen; Sk – Skottehytta.
The background satellite (TERRA/ASTER, taken on July 13, 2002) ortophotomap prepared by A. Stach.

underlined by talus cones and solifluction slopes. Their boulder and debris cover is transformed by mass movements associated with snow and rock avalanches and locally by episodic streams.

Lower parts of valleys, especially on the eastern coast of Petuniabukta are developed in form of raised marine terraces to the level of about 80 m a.s.l. The highest terraces in Ebbadalen, where associations with Pleistocene glaciations are visible, were 14C dated for 37860±1000 yBP. Younger terraces sequence descending from 45 m a.s.l. to the actual coast-line is associated with sea level changes since mid-Holocene. During the younger Holocene, with a progressing warming, central part of the valley was flooded by a sea transgression, recorded in form of a lagoon in the Ebba river mouth. Outflowing glacial

Fig. 10. Mountain walls built of carbonate rocks, talus slopes and raised marine forms (terrace, spit) on the Eastern coast of Petuniabukta (Wordiekammen massive)

rivers in the tide zone accumulate part of bedload and suspended material forming broad tidal flat coupling with outwash cones and planes revealing the greatest intensity of eolian processes.

In Petuniabukta there is a small, although visible range of human-induced landscape changes. Most of them are effects of mining and explorative activity around the settlement Pyramiden as roads and mine waste dumps.

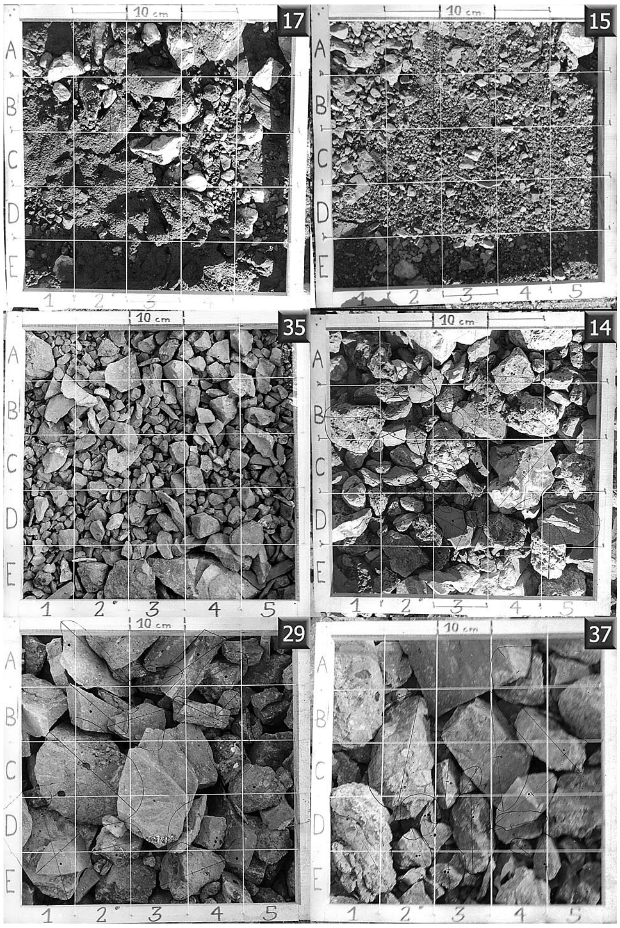


Fig. 11. Facies of slope deposits on western side of Wordiekammen massive (photo Zb. Zwoliński)
sample 17 – incorporation of debris facies into mud facies, sample 15 – fine debris facies, samples 35 and 14 – medium debris facies, samples 29 and 37 – coarse debris facies.

Fig. 12. Upper part of Ebbadalen with the marginal zone of Ebbabreen, higher located Bertrambreen and Mittag-Lefflerbreen in the back

Fig. 13. The front of Ragnarbreen with a marginal lake, seen from the morainic ridge of Little Ice Age

Fig. 14. Tidal flat and outwash plain of the inner part of Petuniabukta. Ebbadalen visible in the front and valley glacier Hörbye in the back

Fig. 15. Fifth level of raised marine terrace near Skottehytta, 20–25 m a.s.l. (photo Zb. Zwoliński)

Fig. 16. Supraglacial stream on the Ebbabreen (photo Zb. Zwoliński)

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