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Spatial relationship in interaction between glacier and permafrost in different mountainous environments of high and mid latitudes, based on GPR research

Wojciech DOBIŃSKI, Mariusz GRABIEC and Bogdan GĄDEK

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Ground penetrating radar (GPR) surveys were conducted on both the glaciers and their forefields in the Tatra Mountains, Northern Scandinavia and on Spitsbergen – between the 49° and 77° latitudes. The results show that the glacial and periglacial environments interpenetrate. Permafrost is present in the glacier, and glacial ice may occur in the periglacial environment. What is common for both the environments is the perennial melting point surface, with the temperature close to 0°C. In the glacier it is the boundary of the cold-temperate transition surface and on the forefield – permafrost base.

Wojciech Dobiński, Mariusz Grabiec and Bogdan Gądek, Faculty of Earth Sciences, Department of Geomorphology, University of Silesia, Będzińska 60, PL-41-200 Sosnowiec, Poland, e-mails: dobin@wnoz.us.edu.pl; mariusz.grabiec@us.edu.pl; bogdan.gadek@us.edu.pl *(received: May 26, 2011; accepted: October 05, 2011).*

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INTRODUCTION

Glaciology and permafrost science are disciplines research areas of which are directly connected. Yet, much of the research within these disciplines is carried out separately. This is due to defining glacial and periglacial fields separately in a "classical" approach (Łoziński, 1912; Brodzikowski and van Loon, 1991; French, 2007). When we consider the glacial area as that part of the Earth's surface which is completely covered by a glacier or ice sheet, then the periglacial area comprises also the area which is to a greater or lesser extent covered with premafrost, where frost action is a dominant factor (French, 2007). Glaciology, however, is sometimes defined as the science dealing with all types of ice (Paterson, 1994), and therefore it also includes ice of the periglacial environment, which may but does not have to occur there. Closer to the synthetic description of these areas appears to be the term "cryology" (Dobrowolski, 1923). Permafrost does not need to be associated with the presence of ice when we are dealing with so-called dry permafrost, and in this case the area of permafrost research goes beyond glaciological studies. This comparison of the definitions shows, however, that the glacial and periglacial areas interpenetrate, and that a separation of glacial and permafrost research is often artificial.

Modern research in geosciences is commonly interdisciplinary, which favours a more synthetic view of the spatial relations as well as the processes and their effects taking place on land. Glaciology and permafrost science have much potential for integrative research. "The primary challenge is to overcome the historical barrier that exists between the two disciplines and to integrate rather than exclude knowledge and understanding..." (Haeberli, 2005). Such an integrated, interdisciplinary approach is a fundamental methodological assumption adopted here.

This study concerns the relations between glacial and periglacial mountainous environments of different latitudes, in conditions of long-lasting permafrost in the glacier forefield, where the glacier is subject to retreat. An attempt has been made to determine the extent to which some glacial and periglacial processes may interpenetrate, for which the course of the melting point surface (MPS) with the temperature close to 0°C in the ice and on its forefield during ablation is crucial (Fig. 1). This approach may facilitate a better understanding of the interaction of the two environments and their evolution.

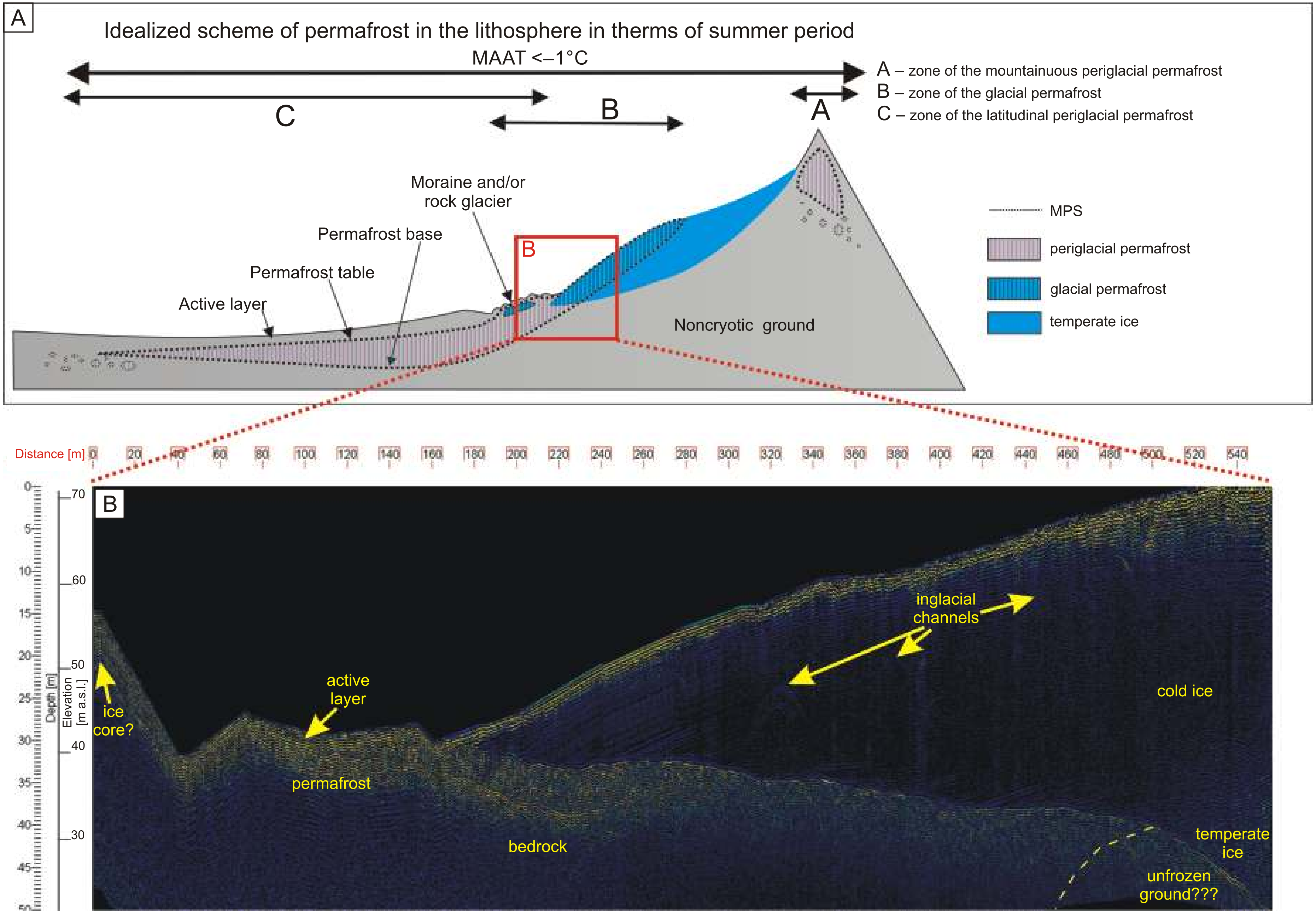

Fig. 1. Initial concept of the continuum between glacial and periglacial permafrost environments in terms of the summer period, A – general model, B – GPR result; MPS run is generalized and shown as a main hub integrating glacial and periglacial environments (compare to Dobiński, 2006)

PREVIOUS WORK AND CONCEPT OF GLACIAL PERMAFROST

Although the presence of permafrost in glaciers and ice-caps has never been completely excluded in definitions of permafrost, for practical reasons the presence of permafrost has typically been confined to the periglacial environment only (Washburn, 1973; van Everdigen, 1998). Probably the first who introduced and developed the concept of glacial permafrost in a synthetic way was Hughes (1973). He defined the concept of permafrost in physical terms, as "the physical condition" which may include in its broadest sense both a medium built exclusively from ice – at one extreme – and a medium generally devoid of ice, at the other extreme, since the principal factor is the temperature. In recent years the concept of permafrost as the physical condition seems to prevail. In a more specific approach to the concept of glacial permafrost, Hughes (1973) narrowed it down exclusively to the regolith charged basal ice layer of a glacier or ice sheet (Hughes, 1973). In 1981, Menzies indicated a significant lack of knowledge concerning the role of freezing in the relation of the glacier and postglacial sediment. He presented four hypotheses concerning the migration of frost at the bottom of the glacier, and indicated the important role of this process in how a glacier interacts with and impacts on its substrate. However, none of these publications specifically integrated research around this issue, which still remained at the interface between glaciology and permafrost science. Since then, other researchers have also used the term "permafrost", to a greater or lesser extent, when referring to the glacial environment (e.g., Björnsson *et al*., 1996; Etzelmüller *et al*., 2003; Etzelmüller and Hagen, 2005), considered glacial permafrost to be a cold layer of the polythermal glacier or a glacier with its substrate wholly frozen up to the depth of the MPS. Thus, the concept presented by Hughes has not been further developed and there is no consensus about what can or should be called glacial permafrost. The last work which attempts to define glacial permafrost is Dobiński (2006) where the author takes the thermophysical point of view.

SITE DESCRIPTION

Given the leading role of the MPS as the axis connecting both the mountain and the arctic environments, and glacial and periglacial processes, this research focusses on the fronts and forefields of the objects situated between the 49° and 77°N (Fig. 2). This choice allows for a holistic approach to the spatial context. The study examines the frontal part and the forefield of the following glaciers: (1) a temperate glacieret (Medeny) located in the Tatra Mountains at an altitude of 2000 m above sea level; (2) Storglaciären, one of the most studied polythermal glaciers in Scandinavia, the front of which ends at a height of approximately 1130 m above sea level; and (3) two polythermal glaciers – Werenskioldbreen and Hansbreen, and one cold glacier – Ariebreen, located in southern Spitsbergen, the front of which is several to dozens of metres high. These research areas are located at very similar longitudes (Fig. 2).

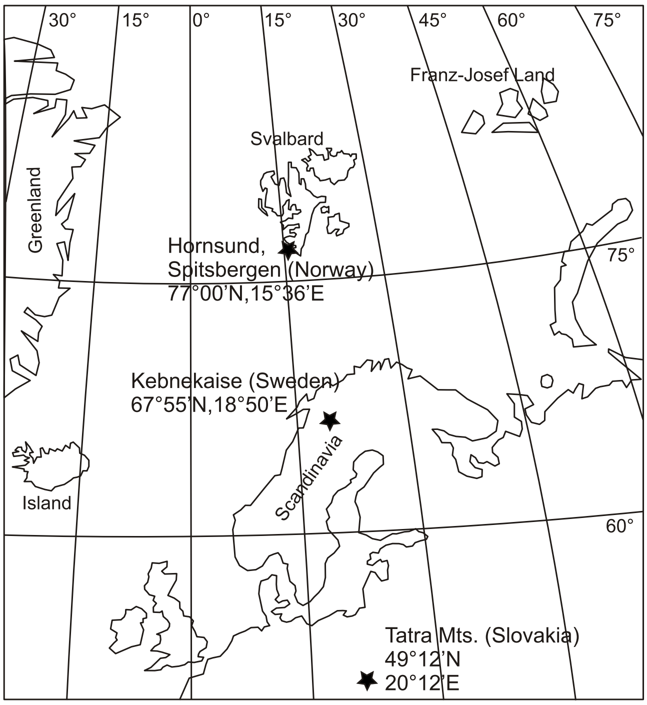

Fig. 2. Location of research areas in Central and Northern Europe

The Tatras represent the highest mountain range of the Carpathians (2655 m a.s.l.), and also the highest non-glaciated mountains between the European Alps and the Caucasus. According to Dobiński (2004) sporadic and discontinuous permafrost may occur above a height of *ca*. 1700 m a.s.l. and it is continuous probably from 2500 m.

The Medena kotlina Valley is located in the upper part of the Kežmarská Biela Voda Valley. It is a poorly developed hanging glacial cirque of northern exposure, located in a moderately cold climate at altitudes 1850–2200 m a.s.l., where MAAT = –2°C (mean annual air temperature; Hess, 1965). From the east, south and west it is shielded by rock walled peaks, whose height exceeds 2500 m above sea level. Nourished by snow avalanches, the Medeny glacieret occupies the western part of the cirque. It is the largest example of a firn-ice field in the Tatras. Its surface usually covers 2–3 ha. The location of the GPR profiles are shown in Figure 3A.

Storglaciären is a small valley glacier located in the Kebnekaise massif (67°55' N, 18°50' E), in the northern part of the Scandinavian Mountains. Lying at an altitude between 1130–1700 m above sea level, it covers an area of 3.1 km^2. It is classified as a polythermal glacier, and has a cold ice layer in its ablation zone (Jansson, 1996; Holmlund and Eriksson, 1989; Pettersson *et al.*, 2003). This glacier is located in the area of mountain permafrost (King, 1986; Kneisel, 1999). The location of the GPR profiles are shown in Figure 3B.

Our study in southern Svalbard (Wedel Jarlsberg Land) was carried out on three sites located on selected glaciers (Ariebreen, Werenskioldbreen and Hansbreen) and their forefields (Fig. 3).

Ariebreen is a small (0.36 km^2), entirely cold valley glacier (Fig. 3C). There is virtually no firn layer or it is very thin

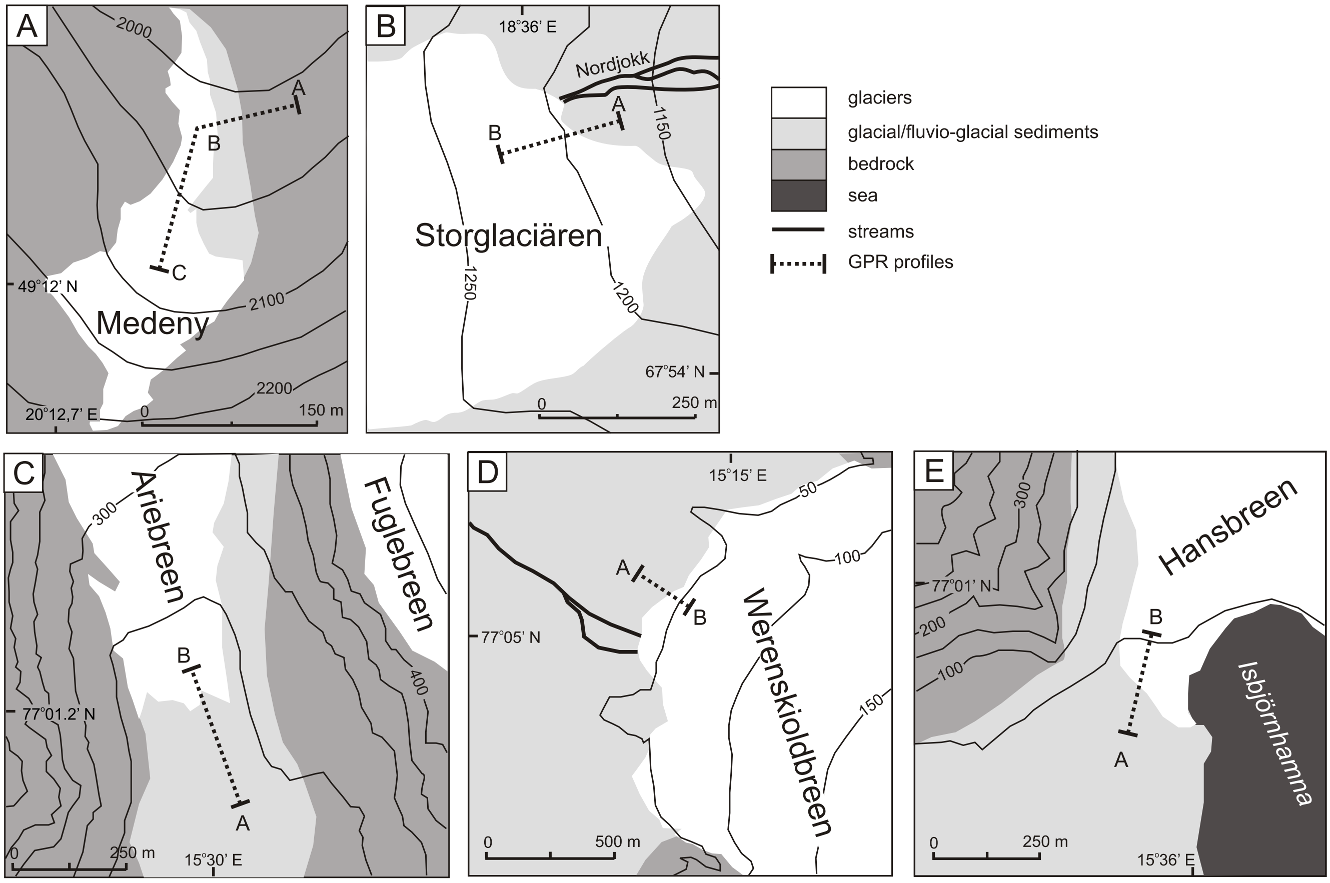

Fig. 3. Location of the GPR profiles on glaciers and their forefields

A – Medeny glacieret; **B** – Storglaciären; glaciers of the Hornsund area, Spitsbergen: **C** – Ariebreen, **D** – Werenskioldbreen, **E** – Hansbreen

(2–3 m; Navarro *et al*., 2008). Ice velocities are low and do not exceed 1.7 m/year. Ariebreen has significantly retreated during recent years. In 1990–2007 its surface area shrank by 28% whereas its volume decreased by 43% in the same period (Pętlicki *et al*., 2008). The forefield of this glacier contains an ice-cored moraine dated back to the Little Ice Age and significantly older lateral as well as basal moraines (Szponar, 1975).

Werenskioldbreen (Fig. 3D), is a valley-type glacier that has a well-defined basin boundary. Its accumulation field consists of three sections: the northern section producing the Skilryggbreen tongue, the central stream of Werenskioldbreen and the smallest southern section forming the Angellisen tongue. The tongues are separated by medial moraines, the largest of which separates the central flow from Skilryggbreen. The glacier runs longitudinally with its snout veering to the north. In 1990, the glacier surface measured *ca*. 28 km^2 (Jania *et al*., 2002). The glacier is moving at a slow pace of a few centimetres a day (Kosiba, 1960; Baranowski, 1977). A distortion of the medial moraine may suggest a glacial surge. The glacier represents a polythermal type with a cold ice layer on the top (Pälli *et al*., 2003).

Hansbreen (Fig. 3E) represents a valley-type glacier with a complex basin (Jania, 1988), which ends in a cliff in Hornsund. The surface of the glacier covers *ca*. 56 km^2 and the average inclination angle is 2° (Jania *et al*., 1996). Mainly due to highly negative values of summer balance, an average of –1.3 m of water equivalent, the net balance is generally negative (–0.38 m; Szafraniec, 2002). Hansbreen represents the typical two-layered thermal structure of a polythermal glacier (Macheret *et al*., 1993; Jania *et al*., 1996; Moore *et al*., 1999).

METHODS

The study of the area of glacial and periglacial environment merging into the fronts and forefields of the glaciers has been carried out using the method of radio-echo sounding of deep structures. Ground penetrating radar (GPR) is an effective tool for establishing the boundaries between materials such as ice and moraine material as well as distinguishing between a dry medium and saturated one. Therefore, this method is used in the research of both glacial and periglacial environments. Thanks to its properties, radio-echo sounding allows deep penetration of ice and detection of internal reflecting horizons between dry cold ice, and ice filled with water in a liquid state (temperate ice). The GPR method allow to identification of discontinuities in the surface layer of the lithosphere through generation and propagation of electromagnetic impulses, followed by registration of the reflected impulses. The radar image is generated according to the ratios between the power of the

Table 1

Dielectric properties and radio-wave velocity in selected materials

Material	Relative dielectric permitivitty ε_r	Radio-wave velocity V [mns^{-1}]
Air*	1	0.3
Freshwater*	80	0.03
Bedrock*	4–6	0.12–0.13
Dry clay*	4	0.15
Saturated clay*	25	0.06
Frozen sediment*	6	0.12
Ice*	3.2	0.17
Snow (ρ = 500 kg/m^3, w = 0%)**	2	0.21

* – after Neal (2004) and Moorman *et al*. (2003); ** – Grabiec *et al*. (2011)

transmitted and received signal, which are the result of changing dielectric properties of the material probed (Table 1). Dielectric properties of crustal surface layers may be the result of lithology, facial structure, density, physical state of the material, water content or sedimentological variation, *etc*. (Neal, 2004). The radar method provides the best results when sounding structures of significantly different dielectric properties.

The measurements were performed by an impulse GPR, consisting of a control unit and unshielded antenna of a centre frequency of 200 MHz. In GPR study a common offset mode has been used. The GPR set was moved along specific profiles. The location of the profiles and their length were provided by a signal from the GPS receiver cooperating with the measurement unit. The course of profile recording was then verified based on identification of the beginning and end of the measurement on maps. In total, nine profiles of a total length of 808 m were performed. The traces along the distance interval were recorded at intervals of 0.2 s, or every 10 cm. The separation of the 200 MHz antenna was maintained constant at 0.6 m. Each trace was created on the basis of 512 samples with a time window of 501 ns or 286 ns.

The radar images obtained in the fieldwork were then processed by using the following procedures: DC removal, time-zero adjustment, background removal, amplitude correction, trace edit, radio-wave velocity models.

For the calculation of the depth scale of the GPR images recorded, measurements of the radio-wave velocity (RWV) in the medium by the CMP method (common-midpoint were made). The measurement consisted of recording an electromagnetic pulse, while systematically increasing the distance between the antennae by 0.6 m, to a maximum of 20 m. The measurements of this type were performed in characteristic points in the area of the measurement, i.e. on the forefield of the glacier with an ice core, on the forefield without an ice core as well as on the glacier in the zone of temperate ice and cold ice.

Where there was no measurement of CMP, the RWV has been estimated on the basis of matching the shape of hyperbolas resulting from wave diffraction by the objects located at a certain depth, to the theoretical hyperbolas.

The average RWV from the surface to the level of the reflector (v) is calculated as follows:

$$v = \sqrt{\frac{x_2^2 - x_1^2}{t_{x2}^2 - t_{x1}^2}}$$

where: t_{x1}, t_{x2} – two-way travel time of reflected wave where the distance between the antennae (in the CMP method) or the horizontal distance from the reflector (in the method of matching hyperbolas), respectively x_1, x_2 (Robinson and Coruh, 1988; Moorman *et al*., 2003). In the absence of applicability of the above methods, there were adopted velocities obtained in CMP measurements or hyperbola matching performed in similar terrain conditions.

RESULTS

The results obtained allow presentation of the hydrothermal characteristics of selected glaciers and the structure of their forefields in mountainous environments at different latitudes.

GLACIERET IN MEDENA KOTLINA VALLEY AND ITS FOREFIELD

In thermal terms the glacieret in the Medena kotlina Valley is made of temperate ice. The profile analysed runs along the axis of the glacieret, then at the bottom it changes its direction to transverse, passing through the frontal – lateral moraine (Fig. 4). The internal structure of the glacieret is complex. At a depth of about 4 metres a clear reflection horizon generated on a layer of coarse-grained material tens of centimetres thick was recorded. In 2003, this material formed the surface moraine of the glacieret. During the study period, above this layer the glacieret consisted of firn with distinct annual layers. However, in a deeper layer it was made of ice of density of 800 kgm^{-3} (Gądek and Kotyrba, 2003). In the section from 0 to 60 m of the measurement profile clear hyperbolic structures, interpreted as diffractions caused by an englacial channel ceiling, were recorded. The substrate measurement profile of the glacieret was clear only between 0 to 40 m of the measurement. The hyperbolic structures visible on the radar image in the section between 40 to 60 m can be interpreted rather as diffractions generated by the englacial channel ceiling or single boulders. A reflection horizon generated on the surface of massive buried ice was registered in the glacieret forefield (Gądek and Grabiec, 2008). It is covered by 2.7–0.8 m sediments that probably form the unfrozen part of the active layer.

STORGLACIÄREN FRONT AND ITS FOREFIELD

The profile in the centre of the front and forefield of Storglaciären measures 539 m, 120 m of which are located on the forefield (Fig. 5 shows first 190 m of the GPR profile). A distinct two-layer hydrothermal structure of the glacier can be discerned. The length of the freezing zone is about 70 m, at this

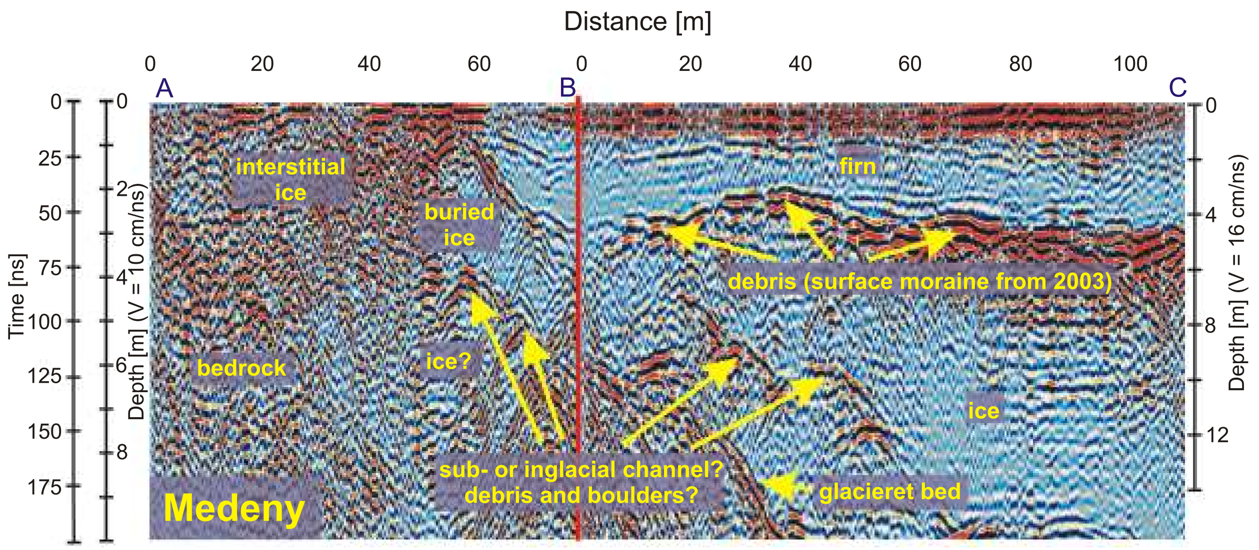

Fig. 4. Result of the GPR profiling performed on the Medeny glacieret and its forefield, Medena kotlina, Tatra Mts., Slovakia

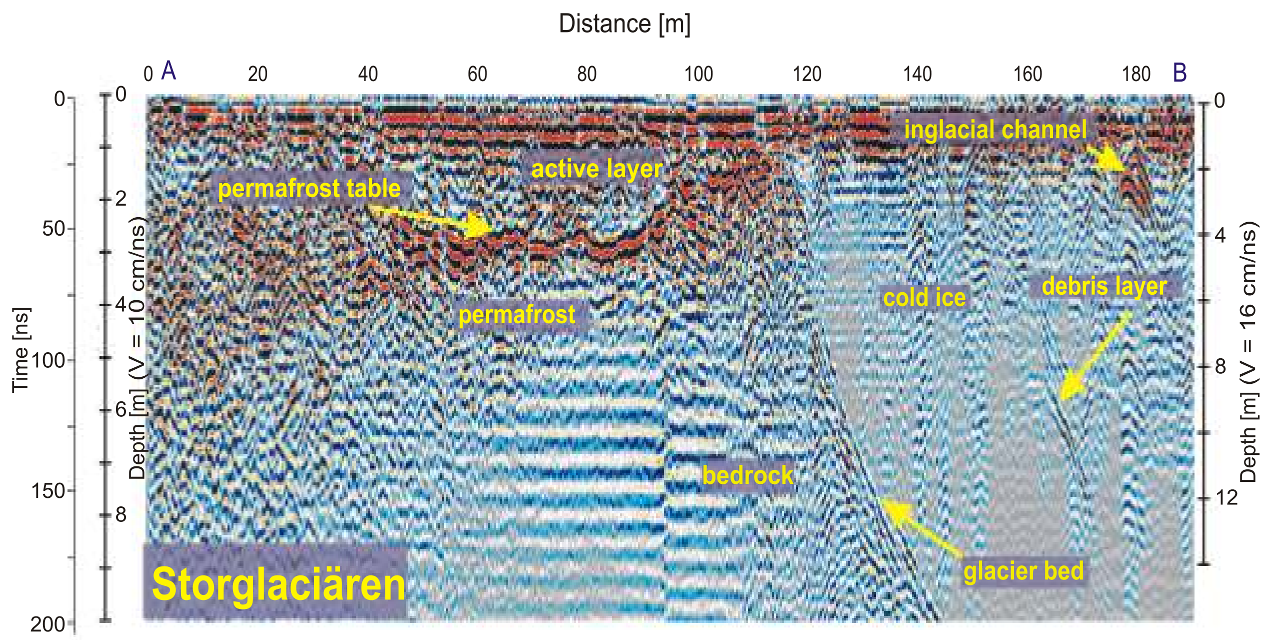

Fig. 5. Result of the GPR profiling performed on Storglaciären and its forefield, Kebnekaise Massif, Northern Sweden

distance the movement of the glacier is compressive and consists mainly of plastic deformation of the glacier. Shearing tensions and visible ice slip planes also occur in this section. The thickness of the cold ice layer decreases upwards, which is consistent with the results obtained by other researchers (Pettersson *et al*., 2007). Belts visible in the frontal part of the glacier are associated with the slip lines. Beneath, there is a layer of temperate ice with a high content of water. A ten-metre-long section of the glacier front is covered with surface moraine from melting. In the forefield on the extension of the glacier front there is a distinctive horizon identified as the active permafrost table. This layer increases in thickness from *ca*. 0 m at the contact point with the glacier to about 3 m at a distance of about 30 m from the glacier front. In the further part of the profile the horizon remains at a similar level. At a distance of about 70 m from the glacier front, in the active layer there occur the reflections from material of the rubble fraction (unsorted moraine material), which blurs the boundary between the active

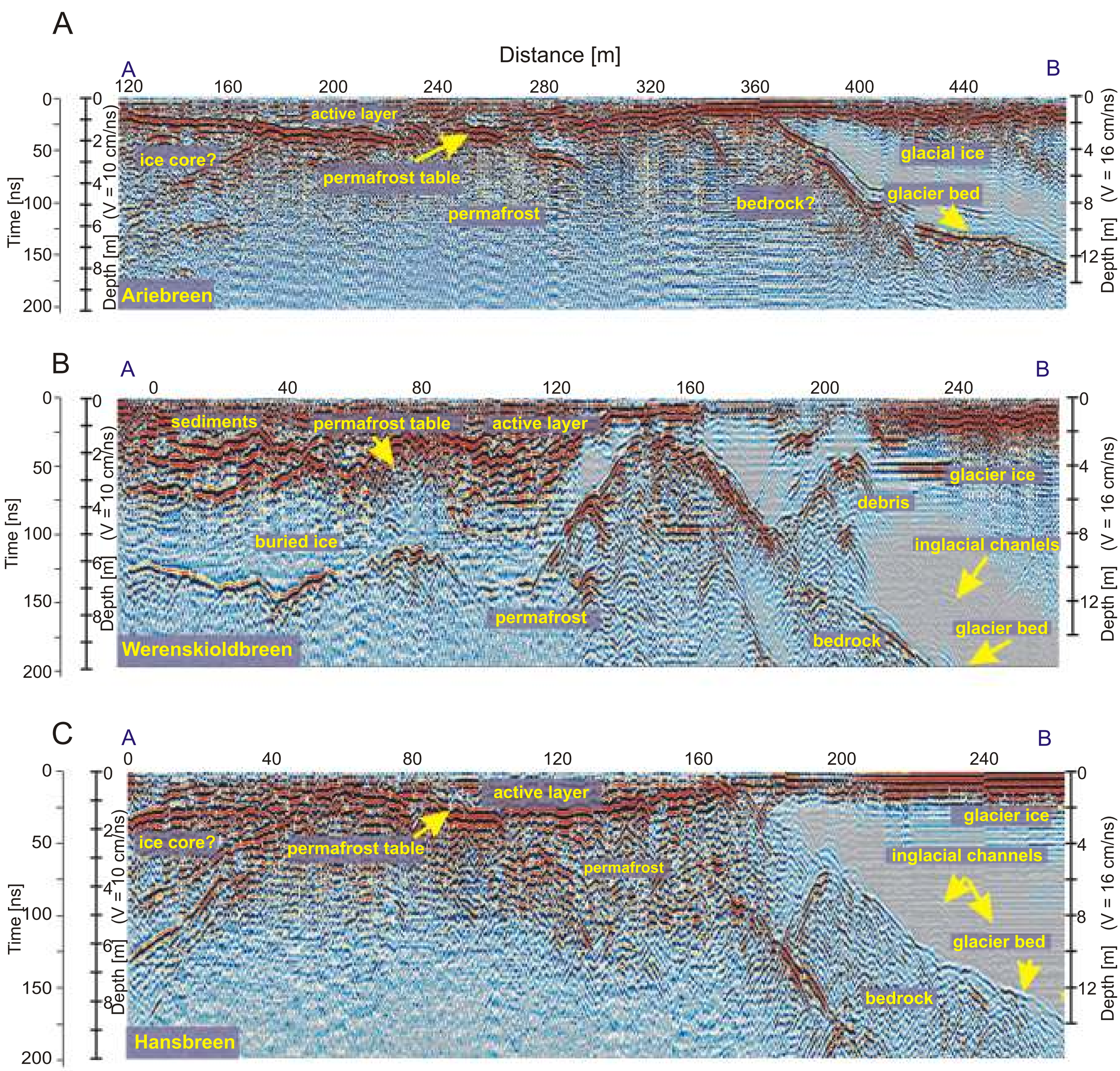

Fig. 6. Results of the GPR profiling performed on Ariebreen (A), Werenskioldbreen (B) and Hansbreen (C), and its forefields, Hornsund area, Spitsbergen

layer and permafrost, and makes interpretation rather difficult. The lower boundary of the active layer in this part of the profile lies at a greater depth (about 3.5 m).

ARIEBREEN FRONT AND ITS FOREFIELD

The radar image shows the glacier front in the section that is approximately 110 m long and of a maximum thickness of about 12 m (Fig. 6A). Its internal structure is typical for cold ice with a few diffractions. Hyperbolas appearing in the top layer indicate the transport of englacial material within the slip planes. The boundary between the glacier and the ground is clear, underlain by numerous diffractions generated by the coarse fraction subglacial material (boulders of diameter not less than 0.4 m). There are no reflections from subglacial channels. Below the glacier floor the structure of the image is distorted by overlapping hyperbola shoulders, and in parts where they are not present the structure is formed by multiple diffractions caused by material of the finer fraction. Such a structure stretches over the forefield of the glacier and is typical of glacial sediments (moraine). The distinctive horizon on the glacier forefield is connected with the bottom of the permafrost active layer. It reaches a maximum of 1.6 m (220 m of the profile) and becoming thinner towards the front of the glacier. The thickness of the permafrost active layer also decreases towards the beginning of the profile on the slope of the terminal moraine, reaching a minimum of about 0.5 m at the peak section. In the permafrost, below the active layer, occur unidentified structures (270–290 m, 310–320 m, 340–350 m) which disappear towards the glacier margin. They may be due to sedimentation in successive phases of glacier recession. In the frontal moraine the thickness of the active layer is limited by the occurrence of the ice core underneath. The buried ice is visible on the radar image between 120 and 210 m of the profile. Interpreting this structure as the ice core is justified by much weaker wave attenuation, and consequently a greater range of penetration as well as the structure with a few diffractions formed by point objects. The ice buried in the zone between 120 and 170 m is layered with 3 sediment layers occurring at approximately 80, 125, 160 ns. Assuming that the dielectric properties of buried ice are similar to glacier ice, and assuming the same RWV, the thickness of the ice core in Ariebreen terminal moraine should be estimated at not less than 10 m. In the lower part of the terminal moraine (170–210 m) the buried ice is visible only to the depth of the shallowest layer of sediment. At the peak part of the moraine (140 m) below the active layer there is a visible hyperbola generated by an object e.g., a boulder buried in the moraine.

WERENSKIOLDBREEN FRONT AND ITS FOREFIELD

Werenskioldbreen represents a type of polythermal glacier with a relatively wide zone of the glacier front frozen to the ground (800 m). For this reason, the radar profile includes a transition zone of the forefield and the glacier in the zone of cold ice. In the frontal part of the glacier, numerous point diffractions (Fig. 6B) are related to the englacial channels, or more frequently to the material inside the glacier. Moraine material deposited on the surface of the glacier refers to the zones of englacial sediments (e.g., 200–220 m in the profile). The glacier substrate is clearly separated, with numerous diffraction hyperbolas caused by coarse-grained subglacial material. In the forefield, between 60 and 140 ns, there is a visible structure, limited by two distinct horizons, which has been identified as buried ice. The thickness of this layer is estimated at about 6 m. Beneath, there is probably coarse-grained moraine material (such as in the substrate of the glacier margin), as inferred from hyperbolas located at the lower boundary of this layer. This suggests a lack of ice layers below. Between 120 and 160 m of the profile the layer thickness gradually decreases to 1 m. The ice buried on the forefield is in contact with glacial ice at the front. The presence of the outflow of water under pressure on the glacier forefield in the area analysed also suggests hydraulic contact between the two structures. Ice buried on the forefield is the result of sudden overlaying of the glacier front with fluvio-glacial material, and then a gradual process of separating the glacier front from the ice under the sandur. As a result of further recession of the glacier front the connection will be broken in the near future. A parallel structure of the image above the layer of buried ice on the glacier front is the result of the presence of sorted fine sediment layers of fluvio-glacial origin (sandur). Between the separate layers associated with episodes of fluvio-glacial sedimentation, especially in the lower part of this structure, there is the possibility of thin, gradually melting ice layers. This is seen in the upward decreasing of the distance between the horizons (visible in the 0–60 m segment of the profile). In the 70–100 m segment the laminar structure of the upper part of the profile is replaced by a multi-reflective one, which suggests the presence of non-sorted moraine material. The thickness of sandur deposit imaged by GPR is estimated to be about 2–3 m. As several horizons may be distinguished it was assumed that the bottom of the permafrost active layer does not occur deeper than the ceiling of buried ice, which means that the ice is likely to be subject to seasonal degradation.

HANSBREEN FRONT AND ITS FOREFIELD

This glacier of polythermal type has its front terminating in the sea. The profile, however, runs across its inactive western part which ends on land. The polythermal structure of the glacier is shown in Figure 1B. In the cold ice on the glacier margin a few diffraction hyperbolas generated by the englacial channels may be seen, while the substrate is composed of diffractions (Fig. 6C) caused by coarse-grained subglacial material. Hyperbolas occuring below are likely to be the echo of the diffractions by the material at the bottom of the glacier. The glacier forefield represents a multi-reflective structure, typical of moraine material. The layer separated by a clear horizon in the profile's upper part is interpreted as the active layer of permafrost. The thickness of the active layer decreases towards the glacier front reaching a maximum thickness of about 1.4 m at the beginning of the profile. From 100 to 130 m of the profile there is a visible unrecognized structure going down from the bottom of the active layer towards the glacier margin to the level of approximately 80 ns. Then, this structure runs parallel to the ground surface on the glacier forefield. The profile starts at the top of the terminal moraine dating from the last phase glacier stagnation. In this zone (0–40 m of the profile) beneath the active layer of permafrost there are structures typical of ice core with layering of moraine material (three distinctive horizons). The thickness of the ice cores measured from the bottom of the active layer to the visible lowest horizon is about 12 m, on the assumption that a typical velocity of radar wave propagation is 0.16 m ns^{-1}.

DISCUSSION

The GPR studies conducted allow a distinction between the processes taking place in the top and bottom part of the glacier, with particular emphasis on the freezing process occuring both in the glacier and its forefield in the areas surveyed. In the upper part it is primarily a characteristic of the relationship between cold ice and temperate ice. In the bottom part it is interaction of the glacier ice with the englacial and moraine material including permafrost occurring on the glacier forefield.

Repeated over several years GPR research results show the dynamics of the Medeny glacieret in the Tatras and its rotational motion (Gądek and Kotyrba, 2007) on both the bedrock and the dead ice underneath the glacieret margin. The dead ice can be called glacial permafrost according to the definition proposed by Hughes (1973). Its long-term persistence is possible thanks to the climate here with an average annual temperature of –2°. The presence of permafrost in the glacieret margin and in other parts of the valley in which the glacieret is located has also been found with other geophysical methods (Gądek *et al.*, 2009). In the climate of the Tatras, where there is not sufficiently low average annual air temperature, and where freezing is hindered due to the thickness of the winter snow cover, the recession of the glacieret will not be accompanied by the transformation of its thermal structure. Therefore, glacial permafrost, understood as a layer of glacial cold ice occurring in the upper part of the glacier does not exist within the Medeny glacieret and it should not be expected to be completely frozen (Fig. 7A).

The results of GPR research carried out on Storglaciären confirm shrinking of the cold ice layer in the bottom part of the glacier. At the same time, according to published data, the thickness of cold ice generally decreased an average of 8.3 ±1.3 m, from about 22% to a maximum of 57% on the entire glacier surface (Pettersson *et al.*, 2003). Such an effect results from a locally faster inwards migration in the glacier of the cold-temperate transition surface (CTS), due to a larger temperature gradient in a thinner layer of cold ice. This means that the climate controls when the thickness of the cold ice layer reaches its minimum, and even if a recession caused by loss of glacier mass follows, the cold ice layer will not continue to shrink, which in turn may lead to a progressive freezing of the entire glacier and its substrate, beginning from its frozen front. The sufficiently cold climate that prevails in this region, with an average annual air temperature below –4°C, allows deep penetration of sub-zero temperatures. It is an essential prerequisite for the thermal transformation of the glacier (Fig. 7B).

The most advanced freezing process of such a relatively small glacier occurs on Spitsbergen, affecting Ariebreen glacier (Fig. 7E). The loss of mass, the significant inclination of the surface, which favours supraglacial runoff and hinders percolation of ablation water, as well as low the MAAT of approx. –5°C in this region have led to the total freezing-up of the glacier. An attempt to classify this type of mountain glaciers' thermal structures, which refers to a different course of CTS/permafrost base (PB), is given in the work of Etzelmüller and Hagen (2005).

Larger glaciers, which include Werenskioldbreen (Fig. 7D) and Hansbreen (Fig. 7C) remain polythermal glaciers. Their recession in a cold climate and a much lower dynamic of geomorphological processes causes the formation of a broad dead ice deposit zone on the forefield, which is not present on such a scale in the mountainous environment of lower latitudes – in the Tatra Mountains or in Scandinavia. The ice of glacial genesis occurring in this area is part of the periglacial environment. As in the case of Medeny glacieret in the Tatras, this kind of ice accumulation in the soil may be called, after Hughes

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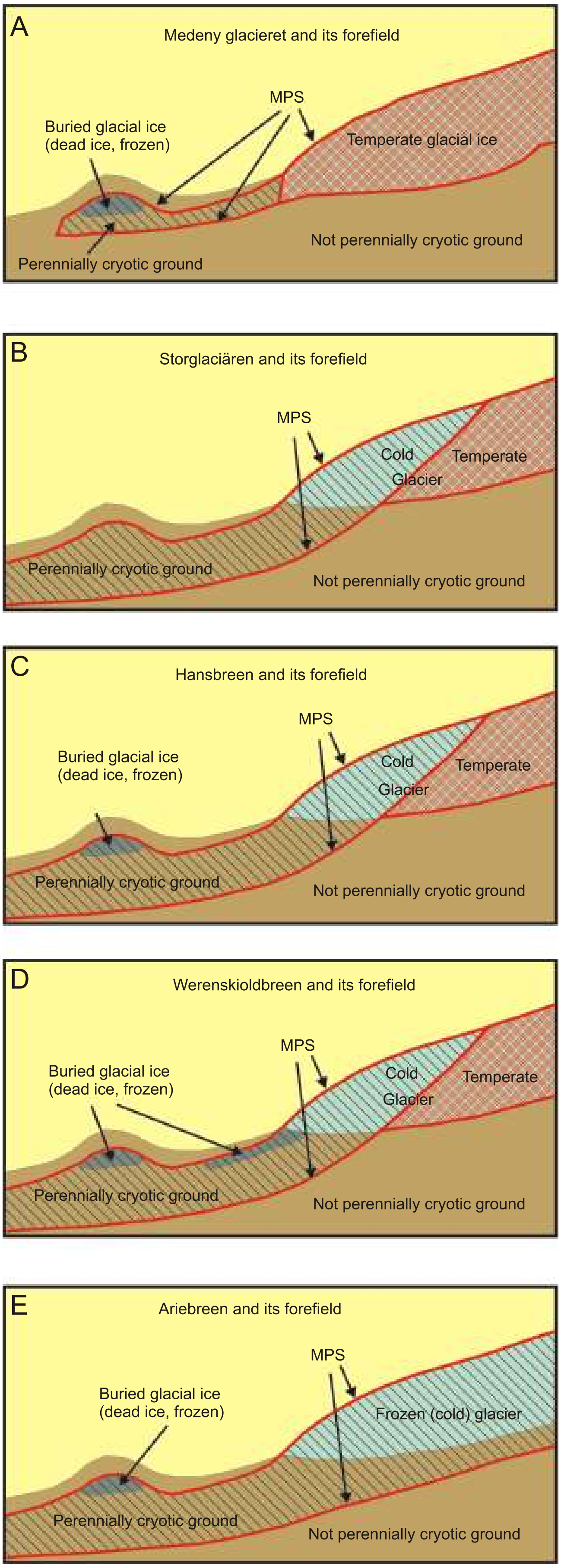

Fig. 7. Schematic diagrams illustrating the structural models of glacier – permafrost relationship based on the interpretation of the field data shown in Figures 4–6

A – Medeny glacieret and its forefield, **B** – Storglaciären and its forefield, **C** – Hansbreen and its forefield, **D** – Werenskioldbreen and its forefield, **E** – Ariebreen and its forefield

(1973), glacial permafrost. The specific pattern of deglaciation of Werenskioldbreen can be associated with a slight inclination of its front, and a relatively flat terrain on the glacier forefield as well as stabilizing by agradating permafrost on the forefield till melting on the margin (e.g., Murton *et al*., 2005).

Based on the research, three types of relations between the glacier and its forefield can be indicated:

1. A temperate glacier with permafrost on its forefield (Fig. 7A). The glacier in its whole volume is temperate and frozen ground does not occur underneath. Permafrost occurrence is possible in the forefield; it may also encompass massive ice of glacial genesis. This type may occur in the high mountains of mid-latitudes, in a humid climate with an average annual air temperature (MAAT) slightly below 0°C (about –1, –2°C). The Medeny glacieret may be an example of this type.

2. A polythermal glacier with a frozen glacier margin penetrating its substrate in the frozen section and in the immediate forefield (Fig. 7B–D). Buried ice of glacial origin (Fig. 7C) and inactive ice, covered with fluvio-glacial sediment and connected directly with the glacier tongue (Fig. 7D) may be present in the permafrost on the forefield. The evolution of the frontal zone of the glacier in the current climate (recession) results in separation of this inactive ice. This type of relation: glacier – permafrost may occur in an arctic mountainous environment (Storglaciären) and in the subpolar climate of southern Spitsbergen with a MAAT of about –4, –5°C (Hansbreen, Werenskioldbreen).

3. A cold glacier with permafrost straddling both the forefield and the glacier with its substrate (Fig. 7E). This type of relation exists in the mountainous environment, and includes glaciers of small thickness and of significant surface inclination in cold climates (MAAT: –4, –5°C). The presence of glacial ice detached from the glacier during recession on its forefield is possible (Ariebreen).

The surface of 0°C is commonly recognized as integrating the glacial and periglacial environments (e.g., Dobiński, 2006; Fig. 1). In the periglacial environment it is called a permafrost table (PT) and a permafrost base (PB), and their continuation in the glacial environment is respectively: the glacier surface (during the summer season) and the boundary between temperate and cold ice, called the cold-temperate transition surface (CTS; Pettersson *et al*., 2007). This isotherm extending from the permafrost-covered substrate to the glacier marks the boundary of freezing of the polythermal glacier margin, and this location is responsible for the formation of the shear zone and/or the beginning of the penetration of englacial debris bands into the glacier (Etzelmüller *et al*., 2003) and the creation of frozen material, the thickness of which may be up to several metres, in the glacier foot in the cold part of the glacier (King *et al*., 2008).

Dobiński (2006), considering permafrost to be a thermal state of the lithosphere, regarded cold glacier ice as permafrost, as also any form of naturally occurring ice within the lithosphere. He indicated, however, a problem that is associated with the term “cryotic”, which denotes the unfrozen portion of permafrost, and which therefore, cannot refer to glacial ice.

In this situation it seems appriopriate to propose a solution which would expand the definition of permafrost, not least because interest in the relation glacier/permafrost has risen significantly, an ever greater number of publications concerned with temporal and/or spatial dependence and variation in the glacial environment and the main element of the periglacial environment, which is permafrost (Kneisel, 1999; Etzelmüller *et al*., 2003; Kneisel, 2003; Etzelmüller and Hagen, 2005; Haeberli, 2005; Harris and Murton, 2005; Murton, 2005; Waller and Tuckwell, 2005).

There is also a synthetic approach referring to geomorphological processes which, although initiated by the glacier, create forms belonging to the periglacial environment, that may be referred to in terms of the paraglacial concept (Ballantyne, 2002). Yet, the paraglacial concept is a one-sided geomorphological concept, concerning the impact from the glacial environment towards the periglacial one. However, freezing, while being the most characteristic process of the periglacial environment, is also important to the glacial environment. Considering ice, in this context, as a monomineralic rock allows, we consider, a more synthetic approach to the forms and processes of both environments, glacial and periglacial (*cf*., Dobiński, 2011).

The thermophysical freezing process, although essentially changing the thermal structure of the glacier and affecting a number of its properties, has no glacial character because it is influenced by the climate. Such freezing occurs both within and beyond the glacial area, and is particularly characteristic of the periglacial environment (French, 2007). Studying the variation within this process is especially important with reference to the evolution of permafrost. The upward or inward movement of CTS/PB surfaces is mainly related to climate change, i.e. the temperature occurring on the surface of the glacier. It is indicated by a high correlation reaching R^2 = 0.87 which exists between the change in the thickness of the Storglaciären cold layer and the change in temperature BTS (bottom temperature of the winter snow cover). A 1°C increase in BTS will lead to a change in the upper boundary conditions of the cold surface layer, resulting in its thinning of by 3–12 m (Pettersson *et al*., 2007). The temperature of cold ice is also subject to seasonal variability, but only at sub-zero temperatures. In cold ice, just as in permafrost, we can distinguish the location of the zero annual amplitude of seasonal temperature changes (ZAA; *cf*., fig. 5 in Pettersson *et al*., 2007).

At the bottom of a glacier, a complex series of processes takes place (e.g., Boulton and Hindmarsh, 1987; Alley, 1989; Knight, 1989; Piotrowski *et al*., 2001; Waller, 2001; van der Meer *et al*., 2003), however, glaciers and ice sheets having a frozen bottom and lying on frozen ground are rarely studied and poorly understood. The revolution in appreciation of the interaction of glaciers with their soft substrate was limited to warm-based glaciers (Waller, 2001). Given current knowledge in this field, we suggest a return to the concept of Hughs (1973). He treated the concept of permafrost very broadly, defining it as a state which may include cold glaciers, containing H_2O sedimentary material remaining at a temperature below zero, as well as frozen rock devoid of water in any form. In giving a summary of mechanisms of basal ice formation, Knight (1997) enumerated the processes associated with the accretion of ice, diagenesis of ice, entrainment of debris and thickening of a sequence. Not all of these are associated with glacial ice. According to him, the emergence of new regelation or congelation ice under the glacier is also possible. Ice of this type is assimilated into the glacier, together with the weathered material, and can be distinguished from ice formed in the glacier accumulation zone on the basis of isotope studies or by analysing the composition of gases present in the ice (Knight, 1997).

The possibility of accumulation of regelation or congelation ice occurs not only in the glacier substrate. At a much greater scale it occurs in the form of so-called internal acumulation in the accumulation area of the glacier. This type of accumulation is common. Internal accumulation in other areas may be responsible for 5 to 70% of glacial accumulation (Trabant and Mayo, 1985; Bazhev, 1997; Rabus and Echelmeyer, 1998). These types of ice are typical of the periglacial environment in as much as they are not able to produce a glacier on their own. Ice (snow) accumulation is necessary for this to happen, at least in the initial phase.

The thickness of ice-rich debris layers beneath the Spitsbergen glaciers determined using GPR ranges from 0.5 to 13 m (King *et al*., 2008), Boulton and Dobbie (1993) indicate thicknesses in the range of 4–47 m.

In all the glaciers surveyed, which have a polythermal structure, there can clearly be distinguished a boundary between cold and temperate ice. At the front of the glaciers, where the frozen glacier part ends, sediment layers begin to appear (Figs. 5 and 6). In the substrate we can observe a characteristic horizon, which disappears upwards to the top of the glacier. Analysis of the evolution of the glacier fronts as well as published information allows two different interpretations of these places. One involves decollement of the ice-debris layer beneath the glacier. The decollement constitutes also the boundary between the areas respectively covered and not covered by permafrost. The freezing process is static and the glacier movement does not limit it significantly. The glacier dynamics are limited by the freezing process to a large degree. The increase in the tension in the frozen section of the glacier may cause decollement which does not reach the bottom of permafrost because of insufficient pressure from (weight of) the frozen section of the glacier.

Physical analysis of the thermal structure of glaciers allow us to apply to them a model derived from research on permafrost following unification of the terms (see Dobiński, 2006).

The results obtained show thermal variations of mountain glaciers in recession. The Tatras, mid-latitude mountains, represent a marginal area of glaciation in which a temperate glacieret occurs. A larger area of glaciation together with polythermal glaciers is the subarctic area of Northern Scandinavia, whereas Svalbard has a relatively wide range of glaciation, where large valley glaciers and outlet glaciers are polythermal, while smaller mountain glaciers are entirely frozen. In very general terms one can say that the volume which frozen depends on the amount of snow accumulation. The thermal structure of the glaciers depends on the air temperature which prevails in their environment, and the percolation of water in the firn zone, which is restricted along with global warming by the increasing surface of accumulated ice. The decrease of their range results in the exposure of their forefields, in which there are also visible traces of retreating glaciation in the form of buried and dead ice. In some instances buried ice may have a direct connection to the glacier, although it does not show any signs of movement. Most often there are, however, dead ice blocks with significant englacial debris layers within them. The ice, although of glacial origin, belongs to the periglacial environment.

CONCLUSIONS

1. In all areas studied it was found that glacial processes associated with the accumulation, ablation, and glacier movement overlap with periglacial processes whose main characteristic is freezing, covering both glacial and periglacial environments. The integrating environmental axis in the relation so defined is the perennial MPS. In the glacier this is the boundary of the cold-temperate transition surface (CTS) and on the forefield it is the permafrost base (PB).

2. The degradation of glaciers means that in favourable climatic conditions, i.e. at fairly low average annual air temperatures, the CTS may increasingly include glacier ice and penetrate into the substrate. Where the average annual air temperature is lower than –1°C, the degradation of the glacier can create conditions for permafrost aggradation in its substrate even before its complete melting. The entire glacier is then subject to the freezing process, which in fact is identified as a basic periglacial process.

3. Also under the glacier, and especially under the frozen front of a polythermal as well as a cold glacier, freezing processes that lead to the formation of a debris-laden layer occur. This layer is only partially formed by the effect of glacial processes, as its accumulation (transport) is dependent on the movement of the glacier. Its freezing is, however, caused by the penetration of cold from the atmosphere, or the advance of the glacier on to the area covered by permafrost. In both cases, interaction and overlapping of glacial and periglacial processes occur.

4. Both on the surface and the floor there may be present ice which is not the result of snow accumulation but is a consequence of other processes, such as regelation, congelation or internal accumulation. These processes alone would not allow the emergence of the glacier, and therefore all of them, and the ice created in this way, originally belong to the periglacial environment.

5. Interaction between the glacial and periglacial environments in mountainous areas between 49° and 77°N is complex, and is based on different relations between the MAAT, thermal characteristics of glaciers, and the MPS, which penetrates glaciers and its forefields and constitutes a hub between both environments.

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REFERENCES

ALLEY R. B. (1989) – Water pressure coupling of sliding and bed deformation: II. Velocity-depth profiles. J. Glaciol., **35**: 119–129.

BALLANTYNE C. K. (2002) – A general model of paraglacial landscape response. The Holocene, **12** (3): 371–376.

BARANOWSKI S. (1977) – Subpolarne lodowce Spitsbergenu na tle klimatu tego regionu, SW Spitsbergen. Acta Univ. Wratisl., 393, Stud. Geogr., **31**.

BAZHEV A. B. (1997) – Methods determining the internal infiltration accumulation of glaciers. In: 34 Selected Papers on Main Ideas of the Soviet Glaciology, 1940–1980: 371–381. Moscow.

BJÖRNSSON H., GJESSING Y., HAMRAN S.-E., HAGEN J. O., LIESTOL O., PALSSON F. and ERLINGSSON B. (1996) – The thermal regime of sub-polar glaciers mapped by multi-frequency radio-echo sounding. J. Glaciol., **42** (140): 23–32.

BOULTON G. S. and DOBBIE K. E. (1993) – Consolidation of sediments by glaciers: relations between sediment geotechnics, soft–bed glacier dynamics and subglacial ground-water flow. J. Glaciol., **39**: 26–44.

BOULTON G. S. and HINDMARSH R. C. A. (1987) – Sediment deformation beneath glaciers: rheology and geological consequences. J. Geoph. Res., **92**: 9059–9082.

BRODZIKOWSKI K. and van LOON A. J. (1991) – Glacigenic sediments. Develop. Sediment., **49**. Elsevier, Amsterdam.

DOBIŃSKI W. (2004) – Permafrost in the Tatra Mts.: genesis, features, evolution (in Polish with English summary). Prz. Geogr., **76** (3): 327–343.

DOBIŃSKI W. (2006) – Ice and environment: a terminological discussion. Earth Sc. Rev., **79** (3–4): 229–240.

DOBIŃSKI W. (2011) – Wieloletnia zmarzlina w wybranych obszarach Tatr, Gór Skandynawskich i Spitsbergenu w świetle kompleksowych badań geofizycznych i analiz klimatologicznych. Pr. Nauk. UŚl., **2850**.

DOBROWOLSKI A. B. (1923) – Historia naturalna lodu. Wyd. Kasa im. Mianowskiego, Warszawa.

ETZELMÜLLER B., BERTHLING I. and SOLLID J. L. (2003) – Aspects and concepts on the geomorphological significance of Holocene permafrost in southern Norway. Geomorphology, **52**: 87–104.

ETZELMÜLLER B. and HAGEN J.-O. (2005) – Glacier – permafrost interaction in Arctic and alpine mountain environments with examples from southern Norway and Svalbard. In: Cryospheric Systems: Glaciers and Permafrost (eds. C. Harris and J. B. Murton). Geol. Soc., London, Spec. Publ., **242**: 11–27.

FRENCH H. M. (2007) – The Periglacial Environment. Third edition, Wiley.

GĄDEK B. and GRABIEC M. (2008) – Glacial ice and permafrost distribution in the Medena Kotlina (Slovak Tatras): mapped with application of GPR and GST measurements. Stud. Geomorph. Carpatho-Balcan., **42**: 5–22.

GĄDEK B. and KOTYRBA A. (2003) – Ancient ice in Tatra Mountains? Prz. Geol., **51** (7): 571.

GĄDEK B. and KOTYRBA A. (2007) – Contemporary and fossil metamorphic ice in Medena kotlina (Slovak Tatras), mapped by ground-penetrating radar. Geomorph. Slov. Bohem., **1**: 75–81.

GĄDEK B., RĄCZKOWSKA Z. and ŻOGAŁA B. (2009) – Debris slope morphodynamics as a permafrost indicator in zone of sporadic permafrost, High-Tatras, Slovakia. Zeitschrift Geomorph. N. F., (Suppl. 3), **53**: 1–22.

GRABIEC M., PUCZKO D., BUDZIK T. and GAJEK G. (2011) – The snow distribution patterns on Svalbard glaciers of different types derived from radio-echo soundings. Pol. Polar Res., **32** (4): (in press).

HAEBERLI W. (2005) – Investigating glacier – permafrost relationship in high-mountain areas: historical background, selected examples and research need. In: Cryospheric Systems: Glaciers and Permafrost. (eds. C. Harris and J. B. Murton). Geol. Soc., London, Spec. Publ., **242**: 29–39.

HARRIS C. and MURTON J. B., eds. (2005) – Cryospheric Systems: Glaciers and Permafrost. Geol. Soc., London, Spec. Publ., **242**.

HESS M. (1965) – Piętra klimatyczne w Polskich Karpatach Zachodnich. Zesz. Nauk. UJ, 115, Pr. Geogr., 11, Pr. Inst. Geogr., **33**.

HOLMLUND P. and ERIKSSON M. (1989) – The cold surface layer on Storglaciären. Geogr. Ann., **71A** (3–4): 241–244.

HUGHES T. (1973) – Glacial permafrost and Pleistocene Ice Ages. In: Proc. Second Internat. Conf. Permafrost, North American Contribution: 213–223.

JANIA J. (1988) – Klasyfikacja i cechy morfometryczne lodowców otoczenia Hornsundu, Spitsbergen. Pr. Nauk. UŚl., **910**. Wyprawy Polarne: 12–47.

JANIA J., KOLONDRA L. and AAS H. F., ed. (2002) – Werenskioldbreen and surrounding areas, Spitsbergen, Svalbard, Norway. Ortophotomap 1:25 000. Katowice, Tromsø. Uniwersytet Śląski, Norsk Polarinstitutt.

JANIA J., MOCHNACKI D. and GĄDEK B. (1996) – The thermal structure of Hansbreen, a tidewater glacier in southern Spitsbergen, Svalbard. Polar Res., **15** (1): 53–66.

JANSSON P. (1996) – Hydrology and dynamics of a polythermal valley glacier. Geogr. Ann., **78A** (2–3): 171–180.

KING E. C., SMITH A. M., MURRAY T. and STUART G. W. (2008) – Glacier-bed characteristics of midtre Lovènbreen, Svalbard, from high-resolution seismic and radar surveying. J. Glaciol., **54** (184): 145–156.

KING L. (1986) – Zonation and ecology of high mountain permafrost in Scandinavia. Geogr. Ann., **68A** (3): 131–139.

KNEISEL C. (1999) – Permafrost in Gletschervorfeldern. Eine vergleichende Untersuchung in den Ostschweizer Alpen und Nordschweden. Trierer Geogr. Stud., **22**.

KNEISEL C. (2003) – Permafrost in recently deglaciated glacier forefields – measurements and observations in the eastern Swiss Alps and northern Sweden. Zeitschrift Geomorph. N. F., **47** (3): 289–305.

KNIGHT P. G. (1989) – Stacking of basal debris by layers without bulk freeze-on: isotopic evidence from West Greenland. J. Glaciol., **35** (120): 214–216.

KNIGHT P. G. (1997) – The basal ice layer of glaciers and ice sheets. Quatern. Sc. Rev., **16**: 975–993.

KOSIBA A. (1960) – Some results of glaciological investigations in SW-Spitsbergen. Zesz. Nauk. UWr., Ser. B, **4**.

ŁOZIŃSKI W. (1912) – Die periglaziale Fazies der mechanischen Verwitterung. C.R. XI Internat. Géol. Congr. Stockholm.

MACHERET Yu.Ya., MOSKALEVSKY M.Yu. and VASILENKO E. V. (1993) – Velocity of radio waves in glaciers as an indicator of their hydrothermal state, structure and regime. J. Glaciol., **39** (132): 373–384.

MENZIES J. (1981) – Temperature within subglacial debris – a gap in our knowledge. Geology, **9**: 271–273.

MOORE J. C., PÄLLI A., LUDWIG F., BLATTER H., JANIA J., GĄDEK B., GŁOWACKI P., MOCHNACKI D. and ISAKSSON E. (1999) – High resolution hydrothermal structure of Hansbreen, Spitsbergen, mapped by ground-penetrating radar. J. Glaciol., **45** (151): 524–532.

MOORMAN B. J., ROBINSON S. D. and BRUGESS M. M. (2003) – Imaging periglacial conditions with ground-penetrating radar. Permafrost and Periglacial Process, **14**: 319–329.

MURTON J. B. (2005) – Ground-ice stratigraphy and formation at North Head, Tuktoyaktuk Coastland, Western Arctic Canada: a product of glacier-permafrost interactions. Permafrost and Periglacial Processes, **16**: 31–50.

MURTON J. B., WHITEMAN C. A., WALLER R. I., POLLARD W. H., CLARK I. D. and DALLIMORE S. R. (2005) – Basal ice facies and supraglacial melt-out till of the Laurentide Ice Sheet, Tuktoyaktuk Coastlands, Western Arctic Canada. Quatern. Sc. Rev., **24**: 681–708.

NAVARRO F., GRABIEC M., PUCZKO D., JONSELL U. and NAWROT A. (2008) – Internal structure of Ariebreen, Spitsbergen, from radio-echo sounding data. In: The Dynamics and Mass Budget of Arctic Glaciers. Extended abstracts: 78–81. Workshop and GLACIODYN (IPY) Meeting, 29–31 January 2008, Obergurgl, Austria.

NEAL A. (2004) – Ground-penetrating radar and its use in sedimentology: principles, problems and progress. Earth Sc. Rev., **66**: 261–330.

PÄLLI A., MOORE J. C., JANIA J., KOLONDRA L. and GLOWACKI P. (2003) – The drainage pattern of Hansbreen and Werenskioldbreen, two polythermal glaciers in Svalbard. Polar Res., **22** (2): 355–371.

PATERSON W. S. B. (1994) – The physics of glaciers. Third Edition. Pergamon.

PETTERSSON R., JANSSON P. and HOLMLUND P. (2003) – Cold surface layer thinning on Storglaciären, Sweden, observed by repeated ground penetrating radar surveys. J. Geoph. Res., **108** (F1): 6004.

PETTERSSON R., JANSSON P., HUWALD H. and BLATTER H. (2007) – Spatial pattern and stability of the cold surface layer of Storglaciären, Sweden. J. Glaciol., **53** (180): 99–109.

PĘTLICKI M., LAPAZARAN J., NAVARRO F., GŁOWACKI P. and MACHÍO F. (2008) – Ice volume changes of Ariebreen, Spitsbergen, during 1936–1990–2007. In: The Dynamics and Mass Budget of Arctic Glaciers. Extended abstracts: 89–92. Workshop and GLACIODYN (IPY) Meeting, 29–31 January 2008, Obergurgl, Austria.

PIOTROWSKI J. A., MICKELSON D. M., TULACZYK S., KRZYSZKOWSKI D. and JUNGE F. W. (2001) – Were deforming beds beneath past ice sheets really widespread? Quatern. Internat., **86**: 139–150.

RABUS B. T. and ECHELMEYER K. A. (1998) – The mass balance of McCall Glacier, Brooks Range, Alaska, U.S.A.; its regional relevance and implications for climate change in the Arctic. J. Glaciol., **44** (147): 333–351.

ROBINSON E. S. and CORUH C. (1988) – Basic exploration geophysics. Wiley, Chichester.

SZAFRANIEC J. (2002) – Influence of positive deree-days on the surface ablation of Hansbreen, Spitsbergen glacier. Pol. Polar Res., **23** (3–4): 227–240.

SZPONAR A. (1975) – The marginal zone of the Arieglacier. In: Results of Investigations of the Polish Scientific Spitsbergen Expeditions: 1970–1971, **1** (ed. A. Jahn). Acta Univ. Wratisl., **251**: 127–138.

TRABANT D. C. and MAYO L. R. (1985) – Estimation and effects of internal accumulation on five glaciers in Alaska. Ann. Glaciol., **6**: 113–117.

Van der MEER J. J. M., MENZIES J. and ROSE J. (2003) – Subglacial till: the deforming glacier bed. Quatern. Sc. Rev., **22**: 1659–1685.

Van EVERDIGEN R. O. (1998) – Multi-language glossary of permafrost and related ground-ice terms. Definitions.

WALLER R. I. (2001) – The influence of basal processes on the dynamic behaviouir of cold-based glaciers. Quatern. Internat., **86**: 117–128.

WALLER R. I. and TUCKWELL G. W. (2005) – Glacier – permafrost interactions and glaciotectonic landform generation at the margin of the Leverett Glacier, West Greenland. In: Cryospheric Systems: Glaciers and Permafrost (eds. C. Harris and J. B. Murton). Geol. Soc., London, Spec. Publ., **242**: 39–51.

WASHBURN A. L. (1973) – Periglacial processes and environments. Edward Arnold, London.