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Surface and bed morphology of Hansbreen, a tidewater glacier in Spitsbergen

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Abstract: Hansbreen, a medium size tidewater glacier in Southern Spitsbergen (Svalbard) is one of the most intensively studied glaciers in the Arctic. This work presents new digital elevation models of its surface and basal topography based on data collected during GPS/GPR campaigns conducted in the spring seasons of 2005 and 2008, as well as on other recent topographic/bathymetric sources. The mean thickness of the glacier is calculated as 171 m and its volume is estimated to be 9.6 ($\pm$0.1) km^3. The main feature of the bedrock morphology is a vast depression that is overdeepened below sea level and extends as far as 11 km upstream from the glacier front. This depression is divided into four individual basins by distinct sills that are related to the main geological/tectonic features of the area. The bedrock morphology affects considerably the glacier's surface topography. The influence of bedrock and surface relief on the subglacial drainage system geometry is discussed. Vast depressions on the glacier surface favor concentration of meltwater and development of moulin systems.

Key words: Arctic, Svalbard, glaciology, bedrock topography.

Introduction

The reaction of a glacier system to external forcing and processes internal to the glacier depends strongly on the characteristics of the glacier. The size and morphology of the glacier are among the most important factors determining or influencing the response of glaciers to climate warming. While the topography of the glacier surface is an important influence on the distribution of its surface mass balance, the subglacial bedrock topography also plays an important role in controlling mass transfer by glacier flow, the development of the glacier's meltwater drainage

system, and fluctuations of the glacier front. While calving intensity is controlled by the sea water depth at the glacier terminus, bedrock topography has especially important influence on the behavior of tidewater glaciers (Brown *et al.* 1982; Paterson 1994).

Hansbreen in southern Spitsbergen is one of most intensively studied tidewater glaciers in the Arctic (Figs 1, 2). Observations, measurements and experiments conducted on the glacier include mapping of its surface topography and radio-echo sounding of its ice thickness and the subglacial bed topography. Thanks to the proximity of the Polish Polar Station in Hornsund, year-round monitoring of the lower part of the glacier has been conducted for the past two decades.

Knowledge of the glacier's topography is important as a background for further glaciological, topo-climatological and environmental studies. Data on the surface and bedrock topography of Hansbreen have been published for example by Glazovsky *et al.* (1991a, b, 1992), Pälli *et al.* (2003), and Macheret (2006). Some of these sources are out of date due to ongoing deglaciation, while others can be improved upon through the use of more accurate data collection methods.

The primary objective of this paper is to present newly acquired and more precise data on the geometry of Hansbreen, along with analyses of its morphometric properties. The paper describes the glacier's environmental setting (geological structure of the area and climatic parameters) and provides an overview of its current state. It also analyzes the influence of the bedrock topography on the glacier's surface topography, and compares the geometry of Hansbreen with that of other tidewater glaciers in the Svalbard archipelago. It is worth noting that no comprehensive studies of the surface/bedrock morphology and morphometry of Svalbard glaciers have previously been conducted, despite the presentation of some pertinent data in existing glacier inventories (*e.g.* Hagen *et al.* 1993; Błaszczyk *et al.* 2009). Detailed information on glacier geometry and environmental setting is vital for understanding and modeling glacier processes.

Hansbreen and its environmental setting

Hansbreen is located in Southern Spitsbergen and flows from north to south to terminate in Hornsund (Figs 1, 2). It is a medium sized glacier (*c.* 56 km^2) with a *c.* 4 km-wide tongue and a 1.5 km-wide active calving front. The ice divide between Hansbreen and Vrangpeisbreen is well defined and located at *c.* 490 m a.s.l. The boundary between Hansbreen and its eastern neighbor Paierlbreen is more difficult to define due to transfluence of ice from the accumulation field to Kvitungisen (a tributary of Paierlbreen) through a glacial breach. Due to a surge of Paierlbreen in the 1990s, the glacier surface topography in the area was altered. Thus, the location of the ice divide has changed over time and different surface areas have been reported for Hansbreen in different publications.
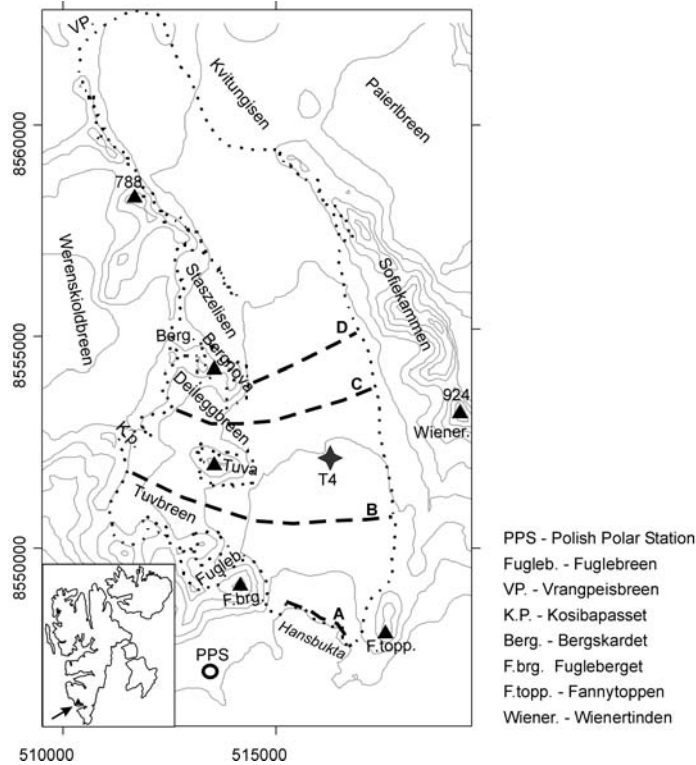
Fig. 1. Study area – Hansbreen and surroundings. The dotted line shows Hansbreen outline and front position in 2004. The dashed lines marked A–D indicate the location of transverse profile lines shown on Fig. 6; T4 – location of mass balance stake No. 4 with permanent GPS monitoring of the glacier movement and automatic weather station (see details in the text).

The Hansbreen system consists of the main trunk glacier and 4 western tributary glaciers: Staszelisen, Deileggbreen, Tuvbreen, and Fuglebreen (Fig. 1). Three small cirque and apron glaciers on the eastern margin fed the main glacier trunk in the past, but are now separated from Hansbreen by ice-cored frontal moraines that merge with the lateral moraines of Hansbreen. Only the meltwater from these small ice bodies feeds the Hansbreen drainage system. These glaciers are not considered further in this paper.

The glacier valley lies between the Sofiekammen ridge (with a highest point at Wienertinden – 924 m a.s.l.) and a zone of mountains on the western side, with a highest elevation at the northern peak of Slyngfjellet (788 m a.s.l.). Two relatively low passes, Kosibapasset and Bergskardet (below 500 m a.s.l), connect to Werenskioldbreen to the west. The eastern mountain ridge has a mean altitude of *c*. 700 m a.s.l. and a relative elevation of about 500 m above the glacier's surface, excluding lower elevation passes. It forms an orographic barrier to air masses advecting from the east. Therefore, a distinct foehn effect can be observed during easterly winds. The mountain range bounding Hansbreen on the western side is

Fig. 2. General view on Hansbreen (oblique picture from airplane). Courtesy of SLICES Project
(Swansea Glaciology Group, Swansea University, UK), July 2004.

generally 150–200 m lower than that on the eastern side (especially in the south-
ern part) and it is dissected by tributary valleys.

The main geological formations of SW Spitsbergen tend to be S-N oriented
and lie parallel to the edge of the continental shelf. Wedel Jarlsberg Land, where
the valley of Hansbreen is located, is underlain by metamorphic and sedimentary
rocks of the Precambrian and Cambrian Hekla Hoek Formation (Birkenmajer
1978a, b, 1990; Czerny *et al*. 1993). During the Caledonian orogeny the Hekla
Hoek Formation was folded and metamorphosed. The Western Spitsbergen Orog-
eny caused strong deformation and faulting, and old Hekla Hoek rocks were thrust
over younger formations and faulted (Harland 1997).

The north-south trend of major ridges and valleys reflects the underlying geo-
logical structure (Fig. 3). The relief developed through fluvial-denudation during the
Tertiary followed tectonic lineaments and lithological boundaries (Jania 1988). The
glaciated valleys are incised into less resistant rock series or located in zones of
overthrusts and faults. The main axes of the glaciated valleys in the Hornsund region
conform with the major tectonic lineaments (Jania 1988). The central section of the
Hansbreen valley is eroded into overthrust phyllites, quartzites, marbles and dolo-
mites of the Deilegga Group, while the northern and western parts are cut into the
Sofiebogen Group (Fig. 3, Birkenmajer 1990; Czerny *et al*. 1993). The axis of the
largest tributary glacier, Tuvbreen, aligns well with the Vimsodden-Kosibapasset
tectonic zone that separates the tectonic block of the Deilegga-Sofiebogen groups
(NE) from that of the Isbjornhamna-Eimfjellet groups (SW) (Fig. 3; Czerny *et al*.
1993). The Isbjornhamna and Eimfjellet groups (mainly metapelites, paragneisses,
calcite-mica schists with marble interlayers and quartzites) form the SW margin of
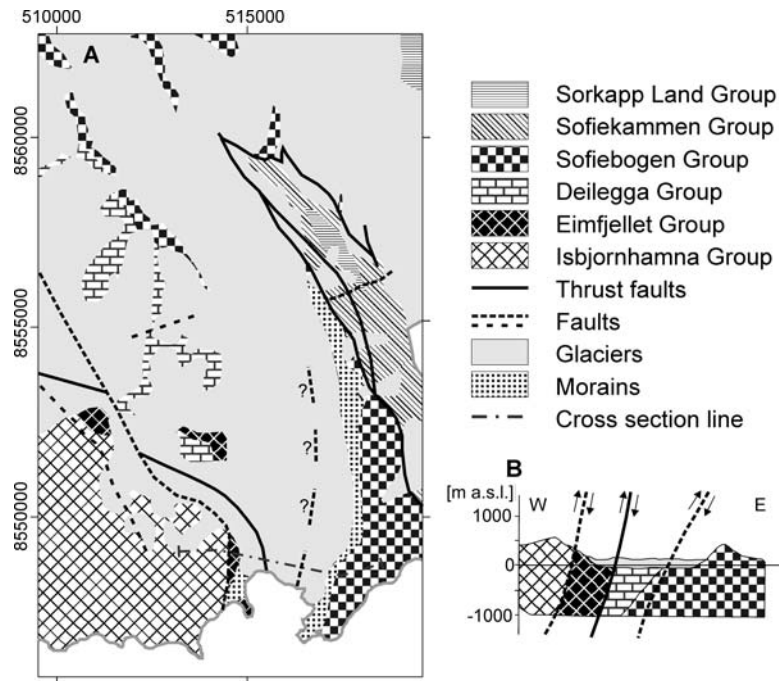
Fig. 3. Map of geological formations and tectonics of Hansbreen and surroundings modified from Birkenmajer (1990) and Czerny *et al.* (1993) (**A**) and geological cross section modified after Birkenmajer (1990) (**B**). The hypothetical fault line underneath the deepest eastern part of Hansbreen is marked by "?".

Hansbreen (Birkenmajer 1990). The orographic setting of the glacier valley clearly relates to the geological structure of the region.

Climatic conditions in the Hornsund area can be characterized using meteorological observations made at the *Stanisław Siedlecki* Polish Polar Station (77°00' N, 15°32' E and 8.1 m a.s.l.) over three decades (1979–2009). The data have previously been processed and interpreted for the period until 2006 (Marsz and Styszyńska 2007). The observations from more recent years are available from the Institute of Geophysics, Polish Academy of Sciences. While the mean annual air temperature was -4.4°C for the period 1979–2006, it reached -2.7°C and -3.5°C in 2009 and 2010 respectively. The maximum mean daily air temperature of +13.5°C occurred on 7 July 2005 and values have been lower in more recent years (+10.8°C on 1 August 2009, +9.8°C 9 July 2010). Nevertheless, the length of the summer period with positive daily air temperatures was 126 days and 130 days in 2009 and 2010 respectively, compared with a mean of 121 days for the period 1979–2010. A very significant positive trend in the mean annual air temperature at Hornsund (0.97°C per decade) was noted over the period of 1979–2010, and similar trends have also been observed for the meteorological stations at Longyearbyen and Ny-Ålesund. Winter warming has contributed more than summer warming to this trend.

Annual precipitation is very variable in Hornsund. The sums of precipitation for 2009 and 2010 (479.7 mm and 448.5 mm) were both higher than a multiyear average of 430 mm. Precipitation in winter makes up *c.* 54% of total annual precipititation. Liquid precipitation constitutes 40% of the mean annual total, while solid and mixed precipitation account for 32% and 28% respectively. A distinct trend towards greater total annual liquid precipitation has been observed, with higher rainfall amounts in the autumn and winter months (October–May) (Łupikasza 2003). Wind direction and speed are very important for redistribution of snowfall on glaciers. At the Hornsund station, the prevailing winds are from the NE, E and SE directions (more than 80%), and the highest wind speeds are associated with winds from these sectors. The most frequent easterly winds are strengthened by the E-W orientation of the fiord. On average, there are 40 days with wind speeds >15 m/s each year. Wind directions on Hansbreen have a significant NW component, with a frequency of *c.* 15% as measured by an automatic weather station in the middle of the ablation zone (stake T4). As a result, the distribution of snow cover on the glacier is highly non-uniform. The minimum snow accumulation is observed in the lowermost reach of the tongue and along the eastern side of the glacier between the calving front and Kvitungisen. This is due to the deflation of snow by strong easterly winds and redeposition towards the west (Grabiec *et al.* 2006, 2011). The mean snow cover thickness measured by high frequency radar sounding in spring 2008 was 2.8 m and observed thicknesses ranged from 0.5 m near the glacier front to 4.5 m in the accumulation area (Grabiec *et al.* 2011).

The annual mass balance of Hansbreen has been measured since 1989. The data have been collected by classic stake readings (11 stakes along the centerline), made twice each year at the end of the accumulation and ablation seasons. Snow accumulation on the glacier surface varies from year to year, but the primary source of variability in the annual surface mass budget of the glacier is the summer ablation, rather than winter mass gain. While the mean annual accumulation in the observation period 1989–2011 was 0.97 m of water equivalent (w.e.), with variability reaching up to 35% from year to year, annual ablation can fluctuate by almost 50% of the mean annual value -1.26 m (w.e.). The mean net surface balance of the glacier was -0.28 m (w.e.) for the 23 year period with extreme values of +0.33 m (w.e.) in 2007/2008 balance year and -1.10 m (w.e.) in 2000/2001 season. The increase of snow accumulation and the decrease of summer melting since 2002 have resulted in slightly less negative balance than reported for period 1989–2001 (-0.38 m w.e., Szafraniec 2002). Due to the redeposition of snow by wind, the equilibrium line runs obliquely to the glacier centerline towards lower elevations in the western part of the glacier. The average equilibrium line altitude (ELA) was *c.* 370 m a.s.l. with significant inter-annual changes within a range from 260 to 500 m a.s.l. No trend has been detected for the time period 1989–2011. Thus, the accumulation area ratio (AAR) varied from 0.69 to almost zero with 0.36 as an average during the period of the measurements. The cumulative net surface

balance of -5.97 m (w.e.), calculated for the period 1989–2011, shows that the glacier has a general mass deficit. If mass loss due to iceberg calving was included in the total glacier mass budget, the total net balance would be significantly more negative.

Annual mass loss in the form of icebergs and frontal melting can be estimated from measurements of the cross sectional area of the actively calving ice-cliff, annual mean flow velocity at the terminus, and annual mean terminus retreat. The ice thickness and the shape of the frontal cross section were obtained from geodetic surveys, radio echo-soundings and measurements of the sea floor bathymetry near the terminus. The average glacier thickness is estimated to be almost 100 m at the terminus. The average retreat rate calculated for the period 1900–2010 is 18 m yr^{-1}, but the calving front retreated more than two times faster (44 m yr^{-1}) in recent years (2005–2010). The glacier flow speed varies in time and space. The mean annual velocity 2007–2008 at the stake T4 located 3.7 km upstream from the front was 55–70 m yr^{-1} (after Błaszczyk et $al.$ 2009). The average annual flow rate at the front (231.5 m yr^{-1}) was calculated on the basis of repeated surveys by terrestrial laser scanning at the end of summer 2009. The movement of Hansbreen is dominated by basal sliding (Jania 1988; Vieli et $al.$ 2004). Short term speedup events with velocities 2–5 times higher than the average are noted after high melting or periods of heavy rain (Vieli et $al.$ 2004).

Using the data on glacier dynamics and front geometry, the calving flux from Hansbreen was estimated. The annual mass loss by calving, the glacier flow velocity and the rate of front retreat vary similarly from year to year. Calculations for the period 2000–2008 show a volume of $c.$ 25.5 10^6 m^3 of lost icebergs annually, making $c.$ -0.41 m yr^{-1} (w.e.) contribution to the overall mass loss. While the mass loss due to calving increased to $c.$ -0.85 m yr^{-1} (w.e.) in the balance year 2008/2009. During the next year, it was lower, $c.$ -0.4 m yr^{-1} (w.e.). Therefore, contribution of calving to the total ablation in the period 2000–2008, balance years 2008/2009 and 2009/2010 was 23%, 35%, and 23%, respectively. Taking into account the average multiyear ablation by calving, the overall net balance of Hansbreen ($c.$ -0.8 m w.e.) was significantly lower than the net surface balance alone.

Hansbreen has the typical two-layered thermal structure of a polythermal glacier, as can be seen from radar soundings and direct ice temperature measurements (Dowdeswell et $al.$ 1984a; Macheret et $al.$ 1993; Jania et $al.$ 1996; Moore et $al.$ 1999). Firn and ice at the pressure melting point have been noted throughout the body of the glacier in the accumulation zone except for a surface layer $c.$ 15 m thick that is seasonally chilled in winter. In the ablation zone, cold ice overlies temperate ice. The thickness of the cold upper layer varies from more than 120 m close to the equilibrium line to less than 40 m near the terminus. The temperature of the cold ice layer at the standard depth of 10 m was -3.5°C at 324 m a.s.l. (near the ELA), -2.7°C at 127 m a.s.l., and -1.5°C at 60 m a.s.l. in the lower part of the tongue (Jania et $al.$ 1996). The water content of the temperate ice changes with depth and during the ab-

lation season due to meltwater supply to the glacier drainage system (Jania *et al.* 2005). The englacial drainage system of the glacier is channelized in form and mini-channels, probably also existing within the firn area where water percolation plays important role. Beside subglacial channels of dendritic form (Pälli *et al.* 2003) it is likely that a distributed subglacial drainage network occurs in some areas.

Brief overview of previous mapping and radio echo-sounding of Hansbreen

The first known map of Hansbreen is a marine sketch chart of Hornsund at a scale of 1:200 000, without meridians and parallels, prepared during the Austro-Hungarian North Pole Expedition 1872–1874 (Weyprecht 1875). It is worth noting that the majority of the coastal area mapped was given geographical names related to sponsors of the expedition, the Austro-Hungarian emperor's family, leaders and members of the expedition (Norsk Polarinstitutt 1991). The next cartographic works for this region were related to the large scale arc of meridian survey along the eastern coast of Spitsbergen (*c.* 460 km) conducted by the Russian--Swedish expeditions 1899–1903. These were the first surveys to be based on astronomical observations for precise location of the triangulation stations. The terminus position of Hansbreen was marked on the 1:200 000 map of Southern Spitsbergen (Wassiliew 1925).

The first systematic mapping of Spitsbergen was done by Norwegians during 1908–1926 and involved establishment of a triangulation network and the use of plane-table survey and terrestrial photogrammetry (Hoel 1929). The terminus position of Hansbreen was measured in 1918. Norwegian aerial photogrammetric surveys, which obtained oblique aerial photographs (18 × 18 cm), were started in 1936 (Luncke 1936). Hansbreen was within the area covered by photographs taken in 1936, while 1:100 000 maps were published later (Norsk Polarinstitutt 1953). These standard topographic maps of Spitsbergen use the Gauss-Krüger projection and the ED-50 datum. These maps were updated and re-published in the nineteen nineties. Nevertheless, the first topographic map (1:25 000) of the frontal part of Hansbreen was prepared using terrestrial photogrammetry by the German expedition of 1938 on the basis of stereoscopic phototheodolite photos taken from the Tsjebysjovfiellet mountain massif on the opposite side of Hornsund (Pillewizer 1939).

Polish terrestrial photogrammetric surveys of Hansbreen's ice-cliff position were initiated in 1956 and conducted sporadically later (Kosiba 1960). More systematic monitoring of the lower reach of the glacier by this method has been conducted since 1982. The glacier was also covered by black and white aerial photos (23 × 23 cm; scale *c.* 1:50 000) taken by the Norwegian Polar Institute (NPI) in 1960. While the NPI utilized these photographs to update the front positions of tidewater glaciers on the standard 1:100 000 topographic maps, the Institute of Geophysics,

Polish Academy of Sciences used the same photos for preparation of a topographic map of the Hornsund area at a scale of 1:25 000 (Barna 1987). Nevertheless, the accuracy of the maps might be dubious in some areas. Subsequently, a new series of NPI aerial photographs of the area was produced and this covered Hansbreen on 12 August 1990 (false-color IR images of 23 × 23 cm; scale *c.* 1:50 000). A special topographic map of the glacier based upon these images was prepared and published at a scale of 1:25 000 with contour lines every 10 m on land and every 5 m on glaciers (Jania *et al.* 1992). The UTM projection and the ED-50 datum were applied.

The surface topography of Hansbreen was also surveyed using precise differential kinematic GPS methods in 2000, 2005, and 2008. Results from the 2005 and 2008 surveys were used in this study.

The first radio echo-sounding (RES) of Hansbreen was done as part of the Soviet Academy of Sciences program of airborne radio echo-sounding of Svalbard glaciers in 1978–1979 (Macheret and Zhuravlev 1982, 1985). A pulse radar system with a carrier frequency of 620 MHz and signal strength 820 Watts (W) was flown over the glaciers by helicopter (Macheret *et al*. 1985). Hansbreen was among the glaciers surveyed. The flight ran along the axis of the glacier over a distance of 12 km from the glacier front (Kotlyakov and Macheret 1987).

Comparable data were obtained by the Scott Polar Research Institute (SPRI) and NPI airborne RES mission in spring 1980. The 60 MHz center frequency 300 W power radar was mounted beneath an airplane and flew *c.* 14 km over the centerline of Hansbreen (Drewry *et al*. 1980; Dowdeswell *et al*. 1984b). Thanks to lower antenna frequency and low water content of the glacier during the measurements (this radar survey was carried out during the spring whilst the Soviet investigations were mainly made in summer) the signal could penetrate the temperate ice layer and reflect from the bottom of the glacier along 29% of the profile length (Dowdeswell *et al*. 1984b). The thickest ice surveyed was 310 m thick and the glacier bed was far below sea level (Dowdeswell *et al*. 1984b). The internal reflections identified along the profile define the boundary between cold and temperate ice layers (Dowdeswell *et al*. 1984a).

The first ground-based radar soundings by the Institute of Geophysics of the Polish Academy of Sciences were started simultaneously with the Soviet airborne RES in summer 1979. Point soundings every 100 m along two profiles perpendicular to the glacier axis were done with a 440 MHz radar system with 7 W peak power (Czajkowski 1980). The lower profile was located close to the present glacier front and measured a bed at a maximum thickness of *c.* 110 (±15) m. The upper section, located 3 km upglacier, was much deeper with the bed at a maximum thickness of *c.* 230 (±15) m.

Ground-based RES was carried out on Hansbreen in April 1989. The low frequency (2–13 MHZ) radar system with 13 kW pulse power was used to measure ice thickness at 161 discrete points (Glazovsky *et al*. 1991a, 1992). Location of every sounding point was surveyed by 3D theodolite intersection. The radar wave

velocity (RWV) was obtained from velocity profiling using the common mid-point method (Glazovsky *et al.* 1991a, 1992). Results of the RES survey were interpolated and a contour map of bedrock topography was produced (Glazovsky *et al*. 1991a, b, 1992; Macheret 2006). The map shows that 75% of the bed of Hansbreen is located below the sea level (b.s.l.). The bedrock below the main stream consists of three overdeepened basins and the deepest point measured was 102 m b.s.l. The bedrock below the tributary glaciers forms typical hanging cirque valleys. The glacier basin is well separated from Vrangpeisbreen. A subglacial ridge also separates the accumulation area of Hansbreen from Kvitungisen. The ice thickness of main tongue changes from 150–200 m at the terminus up to 400 m 8–10 km from the front (Glazovsky *et al*. 1991a, 1992).

Almost a decade later, in spring 1997 and 1998, another extensive ground based radar survey was carried out on Hansbreen by a Finnish-Polish team (Moore *et al.* 1999; Pälli *et al.* 2003). 25 MHz and 50 MHz antennas were used for deep soundings. The radar system was mounted on sledges and pulled behind a snowmobile. Data were collected at intervals of 1–3 m and positioned by navigational GPS with poor accuracy (up to ±250 m, see Moore *et al.* 1999), controlled and corrected for by navigation between geodetically surveyed stakes (Pälli *et al.* 2003). The radar and surface elevation data were used to construct a bedrock digital elevation model (DEM) on a 157m × 153m grid (Pälli *et al*. 2003). The surface topography was obtained by a combining discrete points located by 3D theodolite intersection with a topographic map from 1990. The results suggest that the area of bedrock below sea level is much less than is shown on the map from the 1989 survey. The main overdeepened basin on this map was separated from the fiord by a pronounced sill elevated more than 50 m above the neighboring basin.

Methods and field data acquisition

The surface and subglacial topographies of the entire Hansbreen basin were studied via a GPS survey in April 2005 and a combined GPS/GPR survey mission performed in April 2008. Surface profiling was done using dual frequency GPS receivers towed across the glacier by snow mobiles. A permanent GPS base station, with 1 Hz data acquisition frequency, located close to the Polish Polar Station at Hornsund was used as the reference station.

The surface topography used in this study is derived mainly from the 2005 GPS survey, which generated a much greater data density than the 2008 survey. The total length of the GPS profiles on Hansbreen and Kvitungisen was 296 km, with an average 128 measurement points per km^2. The results of the kinematic GPS data processing show the accuracy of the elevation survey as ± 1m. When comparing DTMs of the glacier obtained by different methods and in different years, one has to take into account changes in the snow cover thickness. However,

for this study this is not crucial. A geocoded ASTER image from 7th August 2004 was used to delineate the glacier front position.

The elevation of the glacier surface, in a badly crevassed region near the terminus of the glacier, was surveyed by terrestrial laser scanning with a long-range scanner (range up to 6 km). High-resolution topographic data for the frontal part of the glacier were acquired on 29 August and 9 September 2009 (Adamek 2010). These results were combined with the bathymetry map based on data from echo-sounding of the sea bed close to the ice cliff (Giżejewski *et al.* 2010) to determine the ice thickness of the glacier at the actively calving cliff.

The radar data were acquired using a low-frequency ground penetrating radar (GPR). Low-frequency radars are suitable for sounding of polythermal glaciers due to their ability to overcome heavy scattering of electromagnetic waves by water inclusions within a temperate ice layer. The GPR system consists of a 25 MHz unshielded antenna, a control unit and a data acquisition platform. Control unit was mounted on sledge pulled by snow mobile and antennas were pulled behind it at 25–40 km/h speed.

Data were acquired for a network of GPS and radar profiles with a total length of over 100 km. This yielded an average track length density of *c.* 1.8 km km^{-2} with some variability related to the accessibility of different parts of the glacier surface (*e.g.* badly crevassed frontal zone, high and steep cirques). GPR measurements were taken every 0.5 s, equivalent to a distance of 1.5–2 m spacing. Every GPR trace consists of 512 samples in 5314 ns of time window. The vertical resolution of the GPR soundings is assumed to 1/4 of the radar wavelength (1.6 m). The traces were simultaneously positioned by a dual frequency GPS measurements taken in differential kinematic mode (DGPS). The positions were calculated relative to reference data from a GPS base station located in vicinity of Polish Polar Station at Hornsund. The accuracy of the GPS point position is calculated at ±0.1 m.

The processing of the GPR data included the correction of time-zero, the correction and scaling of amplitude, bandpass filtering, and 2D migration. The conversion to depth was done using a radio-wave velocity (RWV) estimated by the CMP (common mid-point) method.

The CMP test was performed on Hansbreen in April 2008 at an elevation of *c.* 190 m a.s.l. near the centreline of the glacier. According to the CMP analysis, the average wave propagation in the glacier was 0.164 m ns^{-1}. This value was adopted for analysis of the GPR data, converting the time scale to a depth scale.

The GPR/GPS data were used to build digital elevation models (DEMs) of the subglacial and surface topographies. Additionally, topographic/bathymetric data from regions outside the glacier were used to improve interpolation of the results. The topography of these unglaciated areas was derived from the 1:25 000 Hansbreen topographic map (Jania *et al*. 1992) and DGPS measurements on rock surfaces deglaciated since 1990. The bathymetry of Isbjornhamna was adapted from the 1:25 000 map of the Hornsund surroundings (Barna 1987) and the bathymetric

map of the recently deglaciated sea floor forefield of the glacier in Hansbukta (Giżejewski *et al.* 2010). Additionally, on recently crevassed and not surveyed area close to the front, the bedrock elevation was gathered from 1989 map in locations of radio echo-sounding survey along profile "A" and in point "AB", for details see Glazovsky *et al.* (1992). The surface and bedrock topography data were checked and gridded using the software package Surfer (Golden Software, Inc.). We used a kriging interpolation algorithm – the geostatistical method of prediction and mapping of spatial variation widely recommended in environmental studies (Rouchani 1996; Wackenagel 2003; Olivier 2010). The ordinary point type kriging was employed without breaklines in order to grid with 100 m spacing. The interpolation process used a maximum radius of data search circle up to 9.8 km and from 8 to 64 data points. All the DEMs were prepared with the same spatial resolution and coordinate system (UTM zone 33X; WGS84 ellipsoid).

Results

The Hansbreen glacial basin is not easy to delineate because it shares an accumulation area with Kvitungisen, a tributary glacier of Paierlbreen (Fig. 1). As a result, several different values for its surface area have been reported in previous papers, *e.g.* 64 km^2 by Hagen *et al.* 1993; 56 km^2 by Szafraniec 2002; 53.0 km^2 by Błaszczyk *et al.* (2009). Błaszczyk *et al.* (2011) reported an area of 54.0 km^2 based upon delineation of the accumulation area using the surface DEM from 1990 and Landsat ETM+ image from 2010. In the current work, the surface relief based on recent DEM was used to help define the outline of the basin in the northeast. The seaward limit was taken as the glacier's terminus position as shown on an ASTER image acquired in 2004. This results in an area for Hansbreen of 56.3 (±0.1) km^2.

The general orientation of the main trunk of the glacier is N-S with a slight easterly deviation in the northern part. The western tributary glaciers flow towards the SE. South-facing slopes account for over 44% of the total glacier surface area, and east-facing slopes account for a further 41% (Fig. 4A). The glacier terminus calves mainly into Hansbukta except along the margins, which terminate on thin glacial sediments deposited over raised marine terraces. The highest DEM node within the glacier outline (664 m a.s.l., see Table 1) is located in the north-western part of the accumulation area of the tributary glacier (Fig. 5A). The average surface elevation of Hansbreen was calculated as 306 m a.s.l., with a standard deviation (s.d.) of ±120 m (Table 1).

Hansbreen is a relatively flat glacier. The median surface slope is 2.6° (Table 1) but the maximum inclination can exceed 30° in the transition zones between the cirques and the surrounding mountain slopes (Fig. 4B). The surface slope along the centerline of the main glacier (1.8°) is significantly less than the slopes along the centerlines of the tributary glaciers (Tuvbreen – 5.6°, Deileggbreen – 3.9°, and
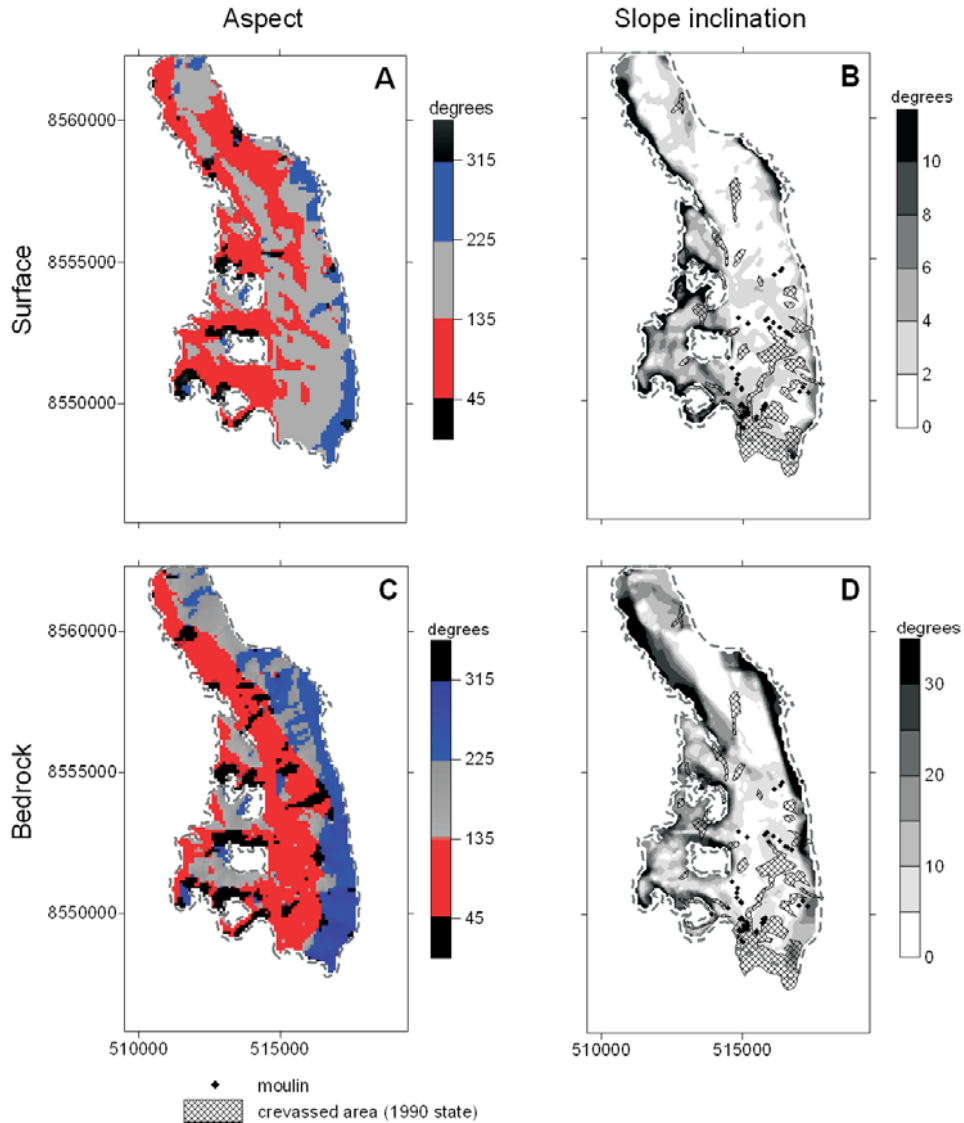
Fig. 4. Aspect and slope inclination of both the surface and bedrock of Hansbreen with marked locations of moulins (black diamonds) and densely crevassed areas (outlined by black checked polygons) according to the 1990 state (Jania *et al.* 1992).

Staszelisen – 4.2°). Much of the change in slope (2–3°) of the lateral glaciers occurs in the contact area with the main glacier (Fig. 4B). There are several shallow (up to 4 m deep) but vast surface depressions on both the eastern and western sides of the main trunk. In the ablation zone, these are associated with the locations of moulins. Some other minor hollows may be masked by the snow cover, as the geodetic survey was performed in late April, when the snow cover is at its thickest.
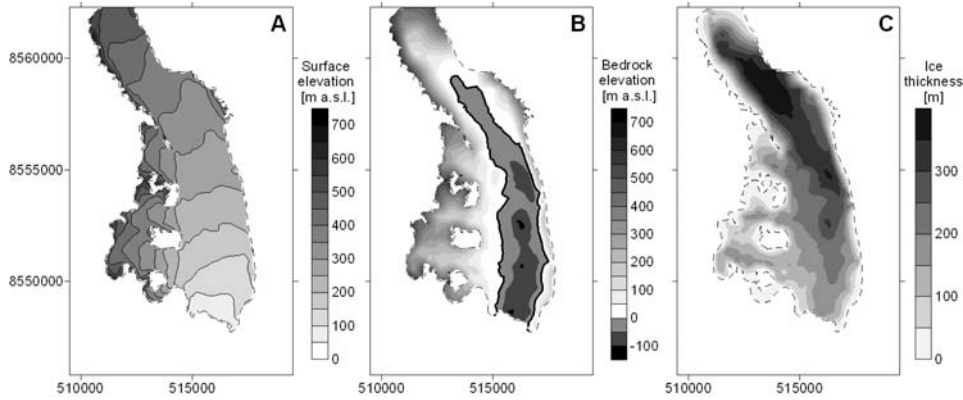
Fig. 5. Maps of surface topography (**A**), bedrock topography (**B**), and ice thickness distribution (**C**) of Hansbreen.

The bedrock surface relief is quite undulating with several large depressions in both the main and tributary branches of the valley (Fig. 5B). Almost 28% of the bedrock surface lies below sea level. The main subglacial depression extends over 11 km upglacier from the present glacier front. The part of the basin that lies below sea level contains several well developed overdeepenings. The deepest point of the bedrock 110 m b.s.l. is located in the depression beneath the main trunk of the glacier. The bedrock beneath the lateral glaciers is more elevated and has the typical hanging valley morphology, with cirques at the valley heads.

The bedrock slope is generally steeper than the surface slope, with a median of 10°. The distribution of slope values reflects the generally U-shaped form of the valley with steep slopes and a rather flat base (Fig. 4D). However, the eastern slopes are much steeper than the western ones. Along the valley floor, the slope increases locally to up to 5–10° in the vicinity of the overdeepenings. The mean bed slope along the central line of the main stream (5°) is less steep than that along the axes of the tributary glaciers (Tuvbreen – 10.8°, Deileggbreen – 13.8°, Staszelisen – 10.3°). The aspect distribution of the glacier bed is shown in Fig. 4C. East facing slopes are most frequent (42%), followed by south and west-facing slopes (24% and 22% respectively).

The mean ice thickness of Hansbreen is 171 (±107 s.d.) m, whereas the maximum thickness was 386 m (Table 1). The thickest ice is located *c.* 11 km from the front, slightly above the average (1989–2011) equilibrium line position. It is possible that the region of thickest ice may extend much further towards the NW, where the bedrock surface was not identifiable in the GPR profiles. The main trunk of the glacier is much thicker than the tributary glaciers. The average thickness along the centerline was calculated as 289 (±79 s.d.) m. Average centerline thicknesses of the tributary glaciers were 118 (±27 s.d.) m for Tuvbreen, 160 (±17 s.d.) m for Deileggbreen, and 141 (±67 s.d.) m on Staszelisen. The total volume of Hansbreen was estimated as 9.6 (±0.1) km³.

Table 1

Basic geometry parameters of Hansbreen

Hansbreen basic geometry parameters	Value
Surface area*	56.3 (±0.1) km^2
Mean surface elevation	306 (±120 s.d.) m a.s.l.
Mean bed elevation	135 (±173 s.d.) m a.s.l.
Min. surface elevation	0 m a.s.l.
Max. surface elevation	664 m a.s.l.
Min. bed elevation	110 m b.s.l.
Max. ice thickness	386 m
Mean ice thickness	171 (±107 s.d.) m
Total ice volume	9.6 (±0.1) km^3
Surface area of the calving front**	*c.* 184200 m^2
Median surface slope inclination	2.6°
Median bed slope inclination	10°
Mean surface aspect	146°
Mean bed aspect	152°

* see comment in the text
** calculated from the data collected in September 2009

Discussion

Comparison of radar data to previous soundings. — The results of recent studies are compared to previous measurements of the glacier thickness. We analyze the data quality so that comparisons of different datasets can take into account differences in the uncertainties associated with the various datasets.

Both the nature of the equipment used in the airborne radar surveys in the 1970s and 1980s and the procedures of data collection and processing influence the accuracy of the radar data. Kotlyakov and Macheret (1987) stated that the differences in ice thickness measurements along repeated profiles in 1970s were up to 60 m with an average 10–30 m. Drewry *et al.* (1980) estimated the vertical error of radar soundings carried out in 1980 to be 1.5% or 10 m whichever is greater. The differences in measured thicknesses at crossing points in a test area were in the range of 4–10 m (Dowdeswell *et al.* 1984a). The radar data obtained from airborne campaigns at Hansbreen in the 1970s and 1980s were displayed on oscilloscopes, recorded on 35 mm film and then digitized manually. Repeat digitization of the same film produced ice thickness estimates with a standard deviation of 2.5–4.7 m, or less than 3% of the ice thickness (Dowdeswell *et al.* 1984a). The ground-based radar surveys in 1979 were performed with an uncertainty of ±15m in ice thickness, arising from errors in oscilloscope reading, uncertainty in the assumed dielectric properties of the ice, and errors in the determination of the ice surface altitude (Czajkowski 1980). Very little is known about the accuracy of the 1989 radar measurements.

However, the rule that the vertical resolution is ¼ of the wavelength emitted suggests that the error might be several meters. Additionally, the analogue oscilloscope display was photographed and then read manually with *c.* 3 m error (Moore 1999). In the 1998 radar campaign, radar frequencies of 25, 50, and 200 MHz were used (Pälli *et al.* 2003) so the vertical accuracy should be comparable with that achieved by the 2008 survey, the error provided by authors is ±5 m.

The RES data from individual surveys also have different navigation and location quality. Airborne navigation was based on Norwegian topographic maps at 1:100 000 or 1:50 000 that show the 1936 geometry of the glaciers. The aircraft were navigated using characteristic topographic features along the centrelines of the glaciers. Dowdeswell *et al.* (1984a) estimated the maximum horizontal positional error to be up to 1 km. The variation of aircraft velocity of *c.* 10% also influences the positioning error of the measurements. No information is provided regarding the error of the vertical coordinate but we suppose that it was obtained from archival topographic maps, and on-board altimeters or radar terrain clearance data. The locations of ground-based radar soundings in 1989 were obtained from theodolite measurements with a mean horizontal error of ±15.4 m (maximum error of ±53.7 m), a mean vertical error ±0.9 m and maximum error of ±1.49 m (Glazovsky *et al.* 1992).

The 1998 radar measurements were triggered every 1 or 3 m using a wheel odometer and positioned using a navigational GPS receiver. The positioning errors provided by Moore *et al.* (1999) for similar location system was up to 250 m and an average of 50–100 m. The accuracy of the vertical coordinates in this dataset is provided by Pälli *et al.* (2003) as ±15 m (combination of GPS error and surface elevation changes between 1990 and 1998), but the nature of the GPS positioning allows them to be treated as at least 1.5 times worse than the error in the horizontal coordinates. Additionally, the radar data were not migrated, which may cause substantial errors in ice thickness estimates in areas of steeply dipping slopes.

Different strategies were used to estimate RWV in each survey. Airborne surveys and the ground survey in 1979 used theoretical values of RWV (the differences with the real state are unknown). It is not clear how the RWV was calculated for processing of the 1998 data. In fact, the velocity profiling (CMP) method was used (P. Głowacki *personal communication*) but, we cannot exclude that the RWV was calculated also from the travel time to the point diffractor and the shape of the resulting hyperbolas, similarly to Moore *et al.* (1999). However this method is accurate only when the positions of traces along the profiles are very precisely located, which may not be the case for this dataset. The 1989 and 2008 radar surveys were analysed using a RWV calculated from a CMP measurement carried out on Hansbreen during the period when the radar soundings were made and those data can be considered the most reliable.

The differences in quality between the different radar datasets prevented us from combining the data into a common dataset in order to build a more comprehensive

model. Only a few points located on recently inaccessible frontal part of Hansbreen were collected from 1989 bedrock map (Glazovski *et. al* 1992) and included in the modeling.

The map of the bedrock topography of Hansbreen presented here is in general agreement with the results of previous surveys in 1989 and 1998 (Glazovsky *et al*. 1992; Pälli *el al.* 2003). It agrees better with the 1989 map than with that based on the 1998 survey. Both the current map and the results of Glazovsky *et al*. (1991a, 1992) and Macheret (2006) show a deep and wide connection between Hornsund and the Hansbreen subglacial valley. The size and location of depressions bounded by bedrock sills in the middle of the main trunk of the glacier, and the bedrock elevation at the Staszelisen entrance, are also in full agreement with the 1989 results. However, while Glazovsky *et al*. (1991a, 1992) state that 75% of the total bedrock surface area is located below sea level, the map presented here suggests that only 28% of the bedrock surface area lies below sea level.

According to Pälli *et al.* (2003), the Tuvbreen and Deileggbreen subglacial valleys do not have a U-shaped cross section. That conclusion is probably dependent on the method of interpolation of the bedrock contours, which did not take into consideration the topography of the unglaciated surroundings. According to Pälli *et al.* (2003), the maximum depth of the central overdeepening does not reach 100 m b.s.l. and the area of overdeepening is less extensive than shown in Glazovsky *et al*. (1992) bedrock model.

The maximum ice thickness of 386 m, based on the 2008 data, is 50–70 m greater than the values obtained from airborne soundings (Macheret and Zhuravlev 1982; Dowdeswell *et al*. 1984a; Kotlyakov and Macheret 1987). However, airborne soundings were carried out only along centerlines and they could miss the thickest part of Hansbreen. On the other hand Glazovsky *et al*. (1992) estimated a maximum ice thickness at 410 m. We do not exclude that such a thickness may be found in the northern part of the central overdeepening, where our soundings did not detect the bedrock due to strong attenuation of the signal.

Taking a parabolic approximation of the glacier valley shape and surface area of 72 km^2, Macheret and Zhuravlev (1982) calculated the total volume of Hansbreen to be 11.91 km^3. The error of the glacier volume estimate was calculated to be 8% of the ice volume *i.e. c.* 1 km^3 (Macheret and Zhuravlev 1982). The accuracy estimate was derived by comparing the parabolic approximation of the volume of selected glaciers with the volume based on ground-based radio echo--soundings. This volume estimate, provided by Macheret and Zhuravlev (1982) for an unspecified reference period, is considerably higher than the 2008 estimate (9.6 (±0.1) km^3). Based on the surface topography from archival topographic maps (Norsk Polarinstitutt 1953) and a recent bedrock elevation model, we calculated that the glacier surface area in 1936 was 59.6 km^2 and the volume was 11.8 (±0.8) km^3. This suggests that the volume of Hansbreen had decreased by *c.* 2.2 (±0.8) km^3 by the beginning of the 21st Century. The Hansbreen volume calculated by

Macheret and Zhuravlev (1982) is close to our results obtained for 1936, but they calculated this value assuming a surface area that was 12 km^2 larger than the value we used. Such an extensive glacier surface area could be obtained for 1936 by including Kvitungisen, eastern and western tributary glaciers as well as the main tongue. Taking into consideration the differences in surface areas adopted by Macheret and Zhuravlev (1982) and in our approach, the Soviet estimate seems to be an underestimate. The lower volume estimate for the glacier calculated in 1982 may be strongly influenced by misinterpretation of 1970s Soviet soundings in which internal reflecting horizons were taken as the bed reflection (Dowdeswell *et al.* 1984a), resulting in underestimation of the ice thickness. In *Glacier Atlas of Svalbard and Jan Mayen* thickness and volume of ice masses were calculated using empirical formula (Hagen *et al*. 1993, p. 31). Hansbreen's volume in 1980 was estimated as 9.6 km^3 based on radio echo-sounding data from the glacier centerline, and an empirical approximation of the glacier surface area as 64 km^2, including the western part of Kvitungisen. The authors do suggest that the volume may have been underestimated because of the quality of the RES data used for the depth calculations. Taking into consideration that the 2004 surface area of the glacier used in our calculations was lower because of the exclusion of Kvitungisen, glacier retreat (*c*. 1.5 km^2 from 1980 to 2004) and surface lowering (*c*. 20 m on average) we conclude that the volume of the glacier estimated by Hagen *et al*. (1993) was indeed too low.

Subglacial topography. — Results of glacier bedrock survey allows to identify the major characteristics of Hansbreen's subglacial topography. Asymmetry of the cross sectional profiles is clearly visible. The deepest part of the valley is locate close to the eastern margin of the glacier and eastern slopes are steeper than the western ones (Figs 5C, 6). Therefore, the valley cross section does not have the typical U-shape profile. This could be explained by the bedrock geology with an overthrust fault located close to the Sofiekammen mountain ridge beneath the glacier, as suggested by Birkenmajer (1990) and shown on his geological cross profile (Fig. 3). The Deilegga Group was probably pushed over the Sofiebogen Group on the eastern side of the glacier.

Similarly, the development of the Tuvbreen valley and the valley of the other western tributary might also be linked to the overthrust zones marked on the same map. The overdeepened basins within the Hansbreen valley floor are separated by transverse bedrock sills, whose location corresponds to pronounced mountains on both sides of the glacier. The lowest one lies between Fugleberget (569 m a.s.l.) and Fannytoppen (391 m a.s.l.), the second one higher up the valley connects Tuva (543 m a.s.l.) with the Wienertinden massif (924 m a.s.l.). The third sill runs obliquely from Bergnova on the western edge of the glacier to the Wienertinden massif on the east (Fig. 1). Bearing in mind the complicated lithology and tectonics of the Hecla Hoek Formation, it is difficult to judge whether all the morphologic features of the glacier bed and relief of the surrounding area are related directly to

the geological structure. Nevertheless, it has been shown that the major geomorphic features of southern Spitsbergen have orientations similar to that of the main tectonic lines in the area (Jania 1988).

Within the glacier's asymmetric bedrock cross profile, a narrow, almost flat, shelf-like area is visible along the western margin. One can interpret this feature as a Holocene raised marine terrace with an elevation corresponding to the 10 m a.s.l terrace outside the glacier (Fig. 6), similar to the terrace on which the Polish Polar Station is located. The upper age limit of the sediments, from which this landform is built, was specified by dating two samples of fossil peat melted out from the glacier ice along a foliation line going from the Tuva nunatak and collected in 1996. They were dated by the ^{14}C method. The resulting uncalibrated dates were 7160 ± 240 BP (Gd-9998) and 5200 ± 240 BP (Gd-13000) (G. Rachlewicz *personal communication* 2010; Oerlemans *et al.* 2011). More organic materials, including a whale rib bone melted out from the lateral land-based part of the glacier front, were found on its western forefield. These evidences prove glacier free conditions in the Hansbreen valley during the Holocene Climatic Optimum. Such relief features were not observed along the eastern steep slope of the valley.

Influence of bed and surface morphology on water drainage system. — The shape of a glacier's surface is controlled by a combination of the orography of the area, the bed topography, the accumulation/ablation distribution, the rheological properties of ice, and the dynamics of the glacier. The subglacial topography may influence the shape of a glacier's surface by controlling the basal shear stress and longitudinal stress in the ice overlying the inclined bed (Knight 1999) as well as the patterns of subglacial drainage and basal friction. A comparison of the average surface and bedrock slopes (dH/dD) along the longitudinal profile of Hansbreen is shown in Fig. 7. A close relationship between the bed and surface slopes is evident over the lowermost part of the profile where the bed topography is relatively rough and the overlying ice is relatively thin (less than 250 m). This is also the case in the uppermost section of the profile, between 12 and 14 km of the profile, where the overlying ice is thicker than 300 m. However, the bed elevation rises rapidly considerably influence on the surface slope. In the middle part of the valley, the relationship between surface and bedrock topography is less clear. Surface topographic features along the glacier centerline show an upglacier shift relative to those on the bed by *c.* 0.5 km. Preliminary results on phase shift suggest a prevailing contribution of basal sliding to glacier movement (Raymond and Gudmundsson 2005). Detailed analysis of the dynamics of Hansbreen and the relationship between its surface and bedrock topographies is the subject of another study in progress.

Glacier topography is an important factor for the development of a glacier's meltwater drainage system. Both the bedrock and surface relief and the thermal state and flow conditions determine where crevasses develop through the development of tensile stresses within the ice (Paterson 1994). While the supraglacial
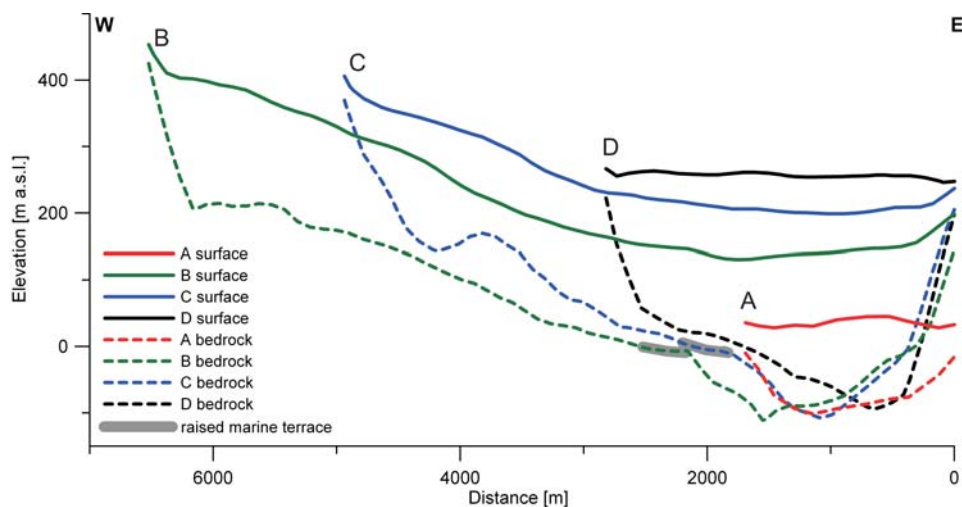
Fig. 6. Transverse cross sections of Hansbreen, for location see Fig. 1. Note exaggeration of the vertical scale. A cross section of the actively calving ice cliff of Hansbreen surface elevation surveyed by the terrestrial laser scanning in September 2009 (data kindly provided by A. Adamek and J. Krawiec) and sea depth soundings (Giżejewski *et al.* 2010).

drainage pattern is determined by the surface topography, the intensity of melting and the permeability of snow/firn, crevasses and shear zones are needed to allow the development of englacial conduits (Röthlisberger and Lang 1987; Van der Veen 1998). Formation of moulins requires a sufficient water supply to a crevasse to drive deeper penetration of a water filled fissure and prevent water from ponding and freezing within a cold ice layer (Holmlund 1988; Van der Veen 2007; Benn *et al.* 2009). The glacier topography is crucial for both these factors. Crevasses are formed in those parts of the glacier where longitudinal tensile stresses develop as ice flows over the lee sides of the bedrock sills and exceed the internal yield strength of the ice. Sustained delivery of meltwater to a crevasses demands concentration of supraglacial drainage from a large area of the glacier surface. When the crevassed areas of Hansbreen identified from the 1990 aerial photos are marked on the topographic map and overlaid on the bedrock elevation model or the slope map, it is clear that their location coincides closely with the positions of the riegels that bound overdeepenings in the bed topography (Fig. 4). The moulins mapped on the topographic map (Jania *et al.* 1992; Kolondra and Pulina 1998) and checked in the field are located on much gentler slopes. These moulins were developed by supraglacial streams flowing into crevasses. They are preserved in areas where there is no active crevassing thanks to the heat carried by water flow into the englacial system and the viscous energy released from melted water (Holmlund 1988; Benn *et al.* 2009). When a new crevasse forms upglacier from the moulin, it intersects the supraglacial stream and captures the water drainage. When the new moulin forms, the old one becomes inactive and closes up over time.
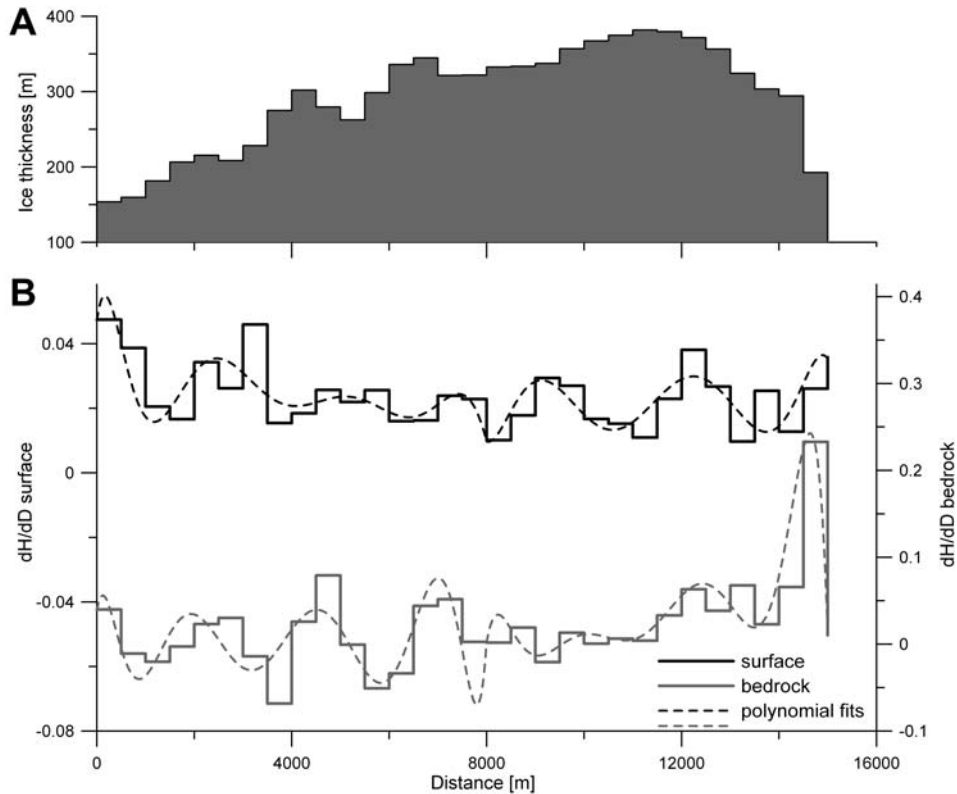
Fig. 7. Ice thickness (**A**), surface and bedrock slope dH/dD (**B**) averaged for 0.5 km segments along Hansbreen centerline starting from the front. Note a different scale of dH/dD axes for the bedrock and surface. The polynomial fits were used for a better interpretation of the graph. Each plot consists of two polynomial fits for distance ranges 0–8000 m and 8000–15000 m respectively.

The glacier geometry also controls the subglacial water drainage pattern. The form of the drainage system and the locations of individual drainage elements are controlled by the form of the subglacial hydraulic potential surface. The hydraulic potential is calculated on the basis of the bedrock elevation and the thickness of overlying ice. According to data from pressure sensors beneath Hansbreen (G.K.C. Clarke *personal communication* 2009) and earlier observations of water level or pressure in moulins (Schroeder 1995; Vieli *et al.* 2004; Benn *et al.* 2009), the subglacial conduits beneath the studied glacier are pressurized, at least temporarily during episodes at the beginning of intense melting period (mainly in June and July). Hence in such conditions, the basal drainage system is determined by the surface to-pography. According to reconstructions of the Hansbreen drainage pattern based on hydraulic potential gradients, the subglacial water paths form a dendritic system with one main outflow located slightly to the east of the center of the glacierfront (Pälli *et al.* 2003). The western tributary glaciers are drained by major conduits run-ning generally along the axis of the valleys with branches reaching into the upper

parts of the valleys (Grabiec *unpublished results*). The existence of subglacial overdeepenings requires the pressure of the water at the ice/bedrock interface to be similar or even higher to the ice overburden pressure to allow the water to escape downglacier or into the basal sediments.

Previous studies show that water can be trapped in cavities between the glacier sole and the basal till within subglacial overdeepenings (Hooke and Pohjola 1994). When the adverse slope is sufficiently steep, supercooling may occur and conduits might be blocked by frazil ice or anchor ice (Hooke and Pohjola 1994; Cook *et al.* 2006). The water pressure has to rise to overcome the closure of the conduits and escape down-glacier (Röthlisberger 1972). That also stimulates the formation of subglacial channels around the overdeepening (Dow *et al.* 2011), the development of englacial drainage above an overdeepening (Hooke and Pohjola 1994) and/or a distributed drainage system (Röthlisberger 1968; Paterson 1994). In consequence, the pressure of water filling the basin increases up to the threshold value that allows the water to escape downglacier through, above or around the riegel. This could suggest that distributed drainage system plays an important role when subglacial water pressure is close to the overburden pressure. The subglacial water pressure and glacier flow velocity react immediately to the increase of meltwater or rain water supply to the drainage system (Vieli *et al.* 2004). Thus, bedrock and surface topography influence sensitivity of glacier flow rate to changes in water input into the drainage system.

Comparison with other tidewater glaciers. — Having in mind that Hansbreen is one of he most-studied tidewater glaciers in Spitsbergen, an important issue is to assess how its geometry is unique or similar to that of other glaciers on the island. The centreline bedrock and surface profiles of Hansbreen were compared to those of 12 other Spitsbergen tidewater glaciers. For the selected glaciers radio echo--sounding data are available for longitudinal profiles from the calving front to the ice divide or upper limit. The data were collected for the following glaciers: Negribreen, Fridtjovbreen, Hornbreen, Hambergbreen, Paulabreen (Dowdeswell *et al.* 1984b), Borebreen, Tunabreen (Bamber 1987), Chydeniusbreen (Bamber 1989), Recherchebreen, vestre Torellbreen, Comfortlessbreen (Macheret and Zhuravlev 1985), and Kongsvegen (Björnson *et al.* 1996). All of them are polythermal glaciers with the upper cold ice layer. The average length of the centerline of the studied glaciers is 18.5 km (between 9.3 km for Hornbreen and 33.3 km for Negribreen), whereas the length of Hansbreen is 15.7 km. The surface as well as base data were averaged on a distance up to 17 km from the front and compared to Hansbreen as shown on Fig. 8. In the majority of glaciers, overdepenings of their bed are characteristic in longitudinal profile.

Both the surface and bedrock profiles of Hansbreen lie within one standard deviation range of the longitudinal elevation profiles of all the glaciers examined. The surface profile of Hansbreen is very close to the mean and its elevation range and average inclination may be defined as typical. However, the bed profile lies on a com-
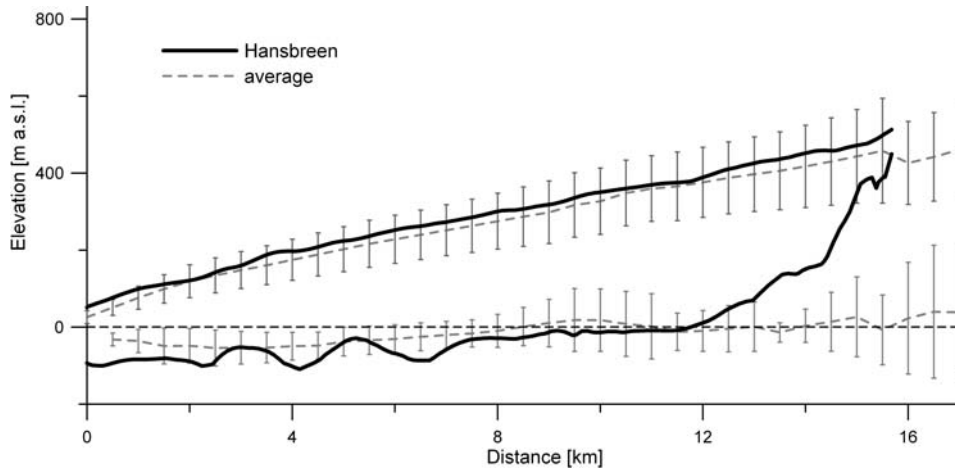
Fig. 8. Longitudinal profile of the bedrock and surface of Hansbreen presented on the basis of averaged data from other larger tidewater glaciers in Svalbard, see text for references. Vertical bars show one standard deviation of the surface as well as bedrock elevation of Svalbard glaciers.

parable elevation but is rougher in comparison to average from other glaciers. The minimum bedrock elevation along the centerline on Hansbreen was found to be 110 m b.s.l. whereas on the others the minimum elevation lay between 21 m b.s.l. (Fridtjovbreen) and 135 m b.s.l. (Comfortlessbreen) with an average of 63 m b.s.l. Hence, Hansbreen has a relatively low minimum subglacial bedrock elevation.

An important characteristic of Hansbreen's bed is the substantial sills that separate consecutive depressions that are overdeepened below sea level. Similarly, beneath some glaciers (*e.g.* Borebreen and Comfortlessbreen) the bedrock in the frontal part is also very rough and the overdeepenings are separated by clear riegels. On average, 67% of the bedrock profiles of the glaciers taken for comparison lie below sea level (from 41% of Paulabreen to 96% of Tunabreen). At Hansbreen, 75% of the length of the longitudinal profile is overdeepened below sea level.

Bedrock topography and retreat rate. — Tidewater glaciers with fronts located on elevated rock bars or shoals formed of glacial sediments are characterized by relatively low frontal velocities and limited frontal fluctuations (*e.g.* Recherchebreen 2005–2006, with an average flow velocity of 30 m a^{-1}, see Błaszczyk *et al.* 2009). Hansbreen is currently calving into relatively deep water (*c.* 90 m in the deepest point) but the Hansbukta is only *c.* 20 m deep *c.* 0.8 km down-fiord from the current front position The front retreated into deeper water just *c.* 30 years ago, hence, the presence of a distinct sill closing the mouth of the valley can explain the relatively low mean retreat rate of Hansbreen in the 20th century. Between 1936 and 2000, the surface area of Hansbreen shrunk by 1.7%, whereas the combined area of all the glaciers emptying into Hornsund shrunk by 8% (Jania *et al.* 2003).

However, when Hansbreen retreated into deeper water, it underwent a dramatic retreat. The retreat rate of Hansbreen in the period 2005–2010 was more than 2 times faster than an average for the period 1900– 2010. Such a rapid retreat has also been observed at other glaciers calving into deep water, such as Columbia Glacier in Alaska (Krimmel 2001; O'Neel *et al*. 2005; Pfeffer 2007), Blomstrandbreen in NW Svalbard (Svendsen *et al*. 2002) and Hornbreen in Hornsund, which retreated from the Treskelen Peninsula after the Little Ice Age (Błaszczyk *et al*. 2011).

Conclusions

The results of the present study provided a more accurate and complete picture of the surface and bedrock topography of Hansbreen. The morphology and morphometric features of Hansbreen mapped in this study are in general agreement with previous works. The total volume of the glacier has been calculated as 9.6 (±0.1) km³ within its area in 2004.

Hansbreen's geometry has developed within an orographic setting strongly related to the geological structure of the metamorphic rocks of the Precambrian and Cambrian Hecla Hoek Formation. The major morphologic features, such as the glacier valley direction and the location of its deepest parts likely reflect the location and orientation of tectonic lineaments such as overthrusts and distinct faults.

The most significant characteristic of the glacier bedrock topography is the presence of overdeepened basins along 75% of the longitudinal profile. Such feature of sublacier topography could facilitate significant/faster front retreat during next decades when ice cliff will contact with deeper sea water in overdeepenings.

A clear asymmetry in the shape of the transverse bedrock profile has been noted, with the deepest parts close to the eastern lateral slope. This feature seems to be controlled by the geological structure. Hanging tributary cirque glaciers are a distinct feature of the glacier's western margin. Both the shape of the valley and ice inflow from the western side shift the main stream of the glacier flow towards its eastern border. This asymmetry disappears in the lower reaches of the glacier and in the main accumulation area.

The main morphometric features of the glacier surface reflect the bedrock morphology. Depressions in the bedrock separated by well pronounced sills have a great influence on the pattern of supraglacial, englacial and subglacial drainage of the glacier. Reproduction of subglacial depressions on the ice surface provides favorable conditions for the focused supply of meltwater to the subglacial drainage system. In consequence, these basins play a clear role in glacier dynamics, in flow velocity and in the domination of flow by basal sliding, as suggested by Jania (1988) and Vieli *et al.* (2004).

Comparison of Hansbreen's geometry with the longitudinal profiles of other Spitsbergen tidewater glaciers shows more similarities than differences. Over-

deepenings of bedrock along glacier valley axes are characteristic feature for the region. The glacier also has a similar hydrothermal structure to other glaciers.

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