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# Snow distribution patterns on Svalbard glaciers derived from radio-echo soundings

Mariusz GRABIEC[1], Dariusz PUCZKO[2], Tomasz BUDZIK[1] and Grzegorz GAJEK[3]

[1] *Wydział Nauk o Ziemi, Uniwersytet Śląski, ul. Będzińska 60, 41-200 Sosnowiec, Poland*
*<mariusz.grabiec@us.edu.pl> <tomasz.budzik@us.edu.pl>*

[2] *Instytut Geofizyki, Polska Akademia Nauk, ul. Księcia Janusza 64, 01-452 Warszawa, Poland*
*<dp@igf.edu.pl>*

[3] *Instytut Nauk o Ziemi, Uniwersytet Marii Curie-Skłodowskiej,*
*ul. Kraśnicka 2cd, 20-718 Lublin, Poland <gajcy@o2.pl>*

**Abstract**: The spatial distribution of snow thickness on glaciers is driven by a set of climatological, meteorological, topographical and orographic conditions. This work presents results of snow accumulation studies carried out from 2006 to 2009 on glaciers of different types: valley glacier, ice plateau and ice cap. In order to determine snow depth, a shallow radio echo-sounding method was used. Based on the results, the following snow distribution patterns on Svalbard glaciers have been distinguished: precipitation pattern, precipitation-redistribution pattern, redistribution pattern and complex pattern. The precipitation pattern assumes that the snow distribution on glaciers follows the altitudinal gradient. If the accumulation gradient is significantly modified by local factors like wind erosion and redeposition, or local variability of precipitation, the accumulation pattern turns into the precipitation-redistribution pattern. In the redistribution pattern, local factors play a crucial role in the spatial variability of snow depth. The complex pattern, however, demonstrates the co-existence of different snow distribution patterns on a single glacial object (glacier/ice cap/ice field).

Key words: Arctic, Spitsbergen, Nordaustlandet, snow accumulation, GPR.

## Introduction

The winter snow cover on glaciers has great importance for their state and for processes occurring on the surface as well as inside the glaciers. Snow accumulation is the key element of glacier mass balance. Due to its high albedo, snow cover reduces melting rate during thaw periods and reduces the energy balance of glaciers. The insulating properties of snow influence the thermal state of glaciers. Water derived from melting snow cover plays a crucial role in glacial hydrology and increases basal sliding. Within the firn zone, melting water percolates and refreezes forming internal accumulation.

The amount of snow accumulated on a glacier's surface is conditioned by a set of factors that can generally be divided into meteorological-climatological and topographic-orographic conditions. Meteorological-climatological factors include: atmospheric circulation that specifies direction of advection and types of air mass; amount of solid precipitation; air temperature, conditioning snow cover stability; and finally wind speed and direction, determining the direction, spatial range and quantity of snow redeposition. However, among topographic-orographic factors the following can be distinguished: altitude, slope exposure to the most frequent direction of air mass advection, orographic barriers that canalize air flow, slope inclination and aspect, distance to natural obstacles, surface roughness and micro- or mesoforms of relief.

Given the large number of factors, an hypothesis of the existence of different patterns of spatial distribution of snow cover in relation to combinations of the conditions can be put forward. The present work is focused on analysis of the spatial variability of snow thickness on Svalbard glaciers of different types: valley glacier terminating in the sea, land-terminating valley glacier, highly elevated ice plateau, and ice cap. These types represent different configurations of topographic-orographic settings of the surface and the surroundings, and are under diverse topoclimatological conditions.

The purpose of this study is to indicate differences between snow distribution patterns on glaciers of Svalbard. The goal will be achieved by the analysis of spatial variability of snow thickness on selected glaciers. Typical snow accumulation patterns will be developed and compared with each other.

**Milestones in snow research on Svalbard glaciers**. — Ritter's measurements carried out within Swedish-Norwegian Arctic Expedition in 1931–32 on glaciers of the Kongsfjord area: austre Brøggerbreen, midre Lovènbreen, Kongsvegen and Holtedahlfonna are considered as the beginning of systematic study on the snow cover of Svalbard glaciers (Ahlmann 1933). The same expedition performed snow studies on ice caps of Nordaustlandet (Ahlmann 1933). The first attempt to determine complex spatial variability of snow accumulation on Svalbard glaciers was undertaken by Mikhaliov and Singer (1975). These authors paid attention to the diminution of snow accumulation towards central and northern Spitsbergen. They also pointed out the importance of exposure of glacier surfaces to air advection from the south-east and south-west, which ensures better snow supply. Essential information on snow distribution on glaciers was provided by the map of mass accumulation at the equilibrium line by Koryakin *et al.* (1985). The map shows maximum accumulation along the western coast of Spitsbergen, whereas the minimum occurs in the north-eastern part of the island. Snow studies on Svalbard in the 1990s were widely performed using shallow radio echo-sounding. Based on the 1997 radar survey, Winther *et al.* (1998) pointed out the eastern sector as the most favourable for snow accumulation, whereas minimum snow thickness was observed in the central part of Spitsbergen. The increase of snow accumulation towards the south was confirmed once again. The project was continued on a greater number of glaciers in 1998–1999, generally supporting earlier results (Sand *et al.* 2003). The most significant achievements in snow research on Svalbard were gathered and summarized by Winther *et al.* (2003). An algorithm for estimation of winter accumulation on Spitsbergen glaciers on the basis of distance to the open sea and basic meteorological data from coastal stations was proposed by Grabiec (2005).

Complex snow measurements on Nordaustlandet are rare. The first map of snow accumulation of the island was prepared from data collected during the International Geophysical Year 1957/1958 by the Swedish Glaciological Expedition to Nordaustlandet (Schytt 1964). A recent map of snow distribution on Austfonna was prepared based on repeated GPR profiling in 1999, 2004 and 2005 (Taurisano *et al.* 2007). According to the map of total accumulation by Schytt (1964) the snow accumulation on Vestfonna increases with elevation a.s.l. much more regularly than on Austfonna, however bias towards the north-east is also observed. Analysis of several dozen snow pits collected during the period 2007–2010 on Vestfonna (Möller *et al*. 2011) proved altitudinal control of snow distribution on the ice cap.

**Geographical settings**. — The snow cover measurements included areas of Svalbard that represent different morphological types of glaciers. The studied areas are located longitudinally from Wedel Jarlsberg Land in the south to Nordaustlandet in the northern part of the archipelago. In particular, the spatial distribution of snow thickness was investigated on a tidewater valley glacier (Hansbreen), a high ice plateau (Amundsenisen), a land-terminating valley glacier (Renardbreen) and a large ice cap (Vestfonna) (Fig. 1).

Hansbreen in the Hornsund area is the southernmost of the glaciers investigated. The glacier is of medium size (56 km$^2$ surface area), with a north-south course and a 4 km front calving into Isbjornhamna (Fig. 2A). The ice divide between Hansbreen and Vrangpeisbreen is at *ca* 490 m a.s.l. The Hansbreen system consists of a main stream and 3 western tributary glaciers. The glacier is bordered on the east by the Sofiekammen massif (highest point Wienertinden – 925 m a.s.l.), forming an orographic barrier for air masses advecting from the east. The mountain range bounding Hansbreen on the west is generally 150–200 m lower (especially in the southern part) than Sofiekammen. The spatial distribution of snow cover on Hansbreen was investigated by classical methods (snow probing along profiles) in 1989 (Grabiec *et al.* 2006), and again in 2003 and 2005 (Grabiec, unpublished data). Earlier studies showed significant asymmetry of snow distribution with the minimum on the eastern side and the maximum on the western side of the glacier (Grabiec *et al.* 2006). The asymmetric snow cover pattern is attributed mainly to snow redeposition towards the west.


Fig.1. Study area. Triangles – studied glaciers (Renardbreen, Hansbreen, Amundsenisen, Vestfonna); circles – meteorological stations at Hornsund, Ny-Ålesund, Svalbard Lufthavn and Sveagruva.

The ice plateau of Amundsenisen, north of Hansbreen, plays a significant role for many glaciers of Wedel Jarlsberg Land as a vast accumulation area. Amundsenisen forms a *ca* 3 km wide and *ca* 12 km long NW-SE strip of *ca* 36 km$^2$ total surface area (Fig. 2B). The plateau is one of the highest ice fields in Svalbard, ranging between 650 and 750 m a.s.l. Amundsenisen is bounded on the north-east by a sharp ridge including peaks over 900 m a.s.l. (Kopernikusfjellet, the highest point, is at 1035 m a.s.l.). The border of Amundsenisen on other sides is not so clear and is formed by a series of nunataks protruding above the outlet glaciers (Nornebreen, Bøygisen and Høgstebreen).

Due to significant role of Amundsenisen as a reservoir of ice mass for the majority of glaciers of the region, snow studies were taken up there as early as in the

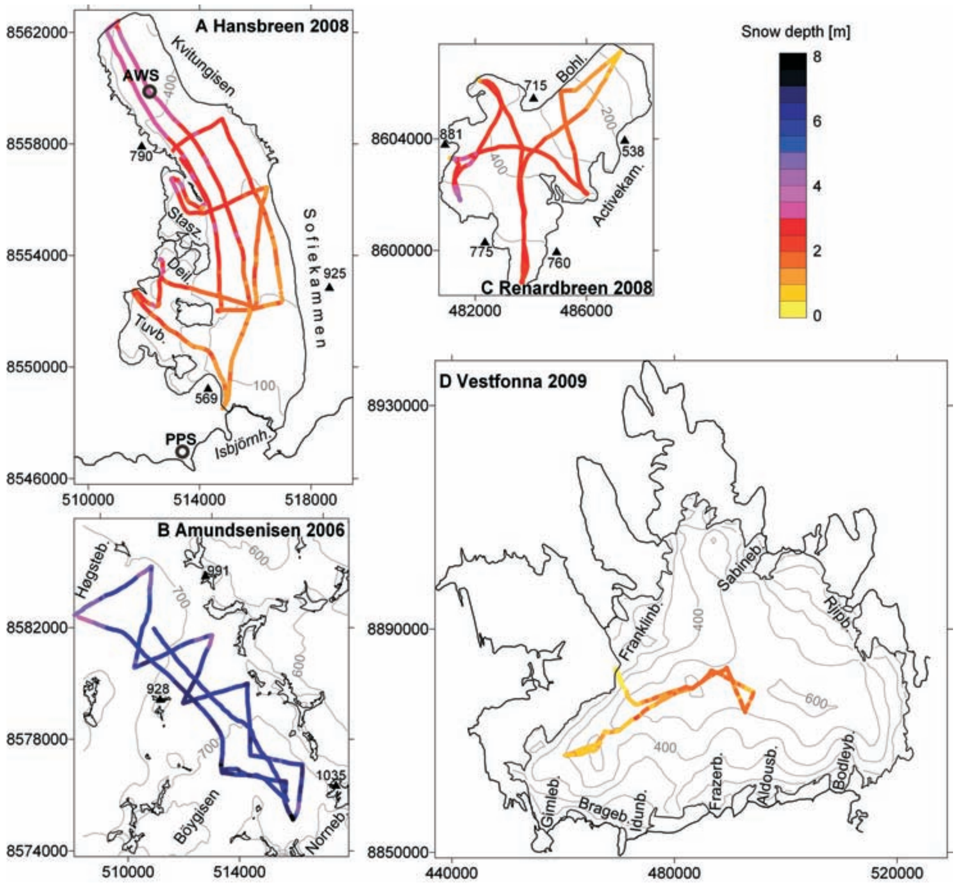


Fig. 2. Snow thickness on glaciers investigated by GPR during 2006–2009. The coordinate system of maps A–C is UTM 33X on WGS84. Map scale for Vestfonna (D) is smaller than for the other glaciers. Vestfonna is located on the border between UTM zones 33X and 35X. To use homogenous coordinates for the entire map (D) we adopted UTM coordinates for the non-standard zone 34X, which is not applied in cartographical works. PPS – Polish Polar Station in Hornsund, AWS – automatic weather station in Hansbreen accumulation zone, Tuvb. – Tuvbreen, Deil. – Deileggbreen, Stasz. – Staszelisen, Isbjörnh. – Isbjörnhamna, Høgsteb. – Høgstebreen, Norneb. – Nornebreen, Bohl. – Bohlinryggen, Activekam. – Activekammen, Franklinb. – Franklinbreane, Sabineb. – Sabinebreen, Rjipb. – Rjipbreen, Bodleyb. – Bodleybreen, Aldousb. – Aldousbreen, Frazerb. – Frazerbreen, Idunb. – Idunbreen, Brageb. – Bragebreen, Gimleb. – Gimlebreen.

1970s. The investigations were related to point measurements of snow water equivalent and internal structure of the snow cover in order to estimate mass balance (Baranowski 1977; Pereyma 1981; Migała *et al.* 1988; Pulina 1991; Jania 1994; Głowacki and Leszkiewicz 1994). The purpose mentioned above was also executed by ice cores (Zagrodnov and Samoilov 1982) and snow pit analyses (systematic monitoring since 1999 by the Institute of Geophysics, Polish Academy of Sciences). The first attempt to determine snow thickness on the plateau using shallow radio echo-sounding was carried out in April 2001 (Melvold 2008).

Renardbreen, a valley glacier of 27 km$^2$ surface area, is located in the Bellsund area (Fig. 2C). The glacier's snout is exposed towards the north-east and the upper part consists of 3 glacial basins. The southern part of the front was calving to Josephbukta until 1990's (Zagórski 2005; Zagórski *et al.* 2008), but the front of Renardbreen changed recently from tidewater to land-terminating as an effect of significant recession. The snout is surrounded on the south-east and north-west by the Activekammen and Bohlinryggen ranges respectively, with peaks reaching 715 m a.s.l. The upper part of the glacier is bounded by a series of ridges of varied relief, cut by numerous passes. The peaks around Renardbreen rise 250 m on average above the glacier surface.

So far, studies of snow cover on Renardbreen are poor and limited to discrete measurements in 1987 (Piasecki 1988), performed in order to estimate mass balance elements. Mass balance monitoring was also carried out in the seasons 2005–2009.

Vestfonna is the northernmost study area. Vestfonna is the second largest ice cap in Nordaustlandet (2455 km$^2$ – Hagen *et al.* 1993). The elevation range of the cap is up to 630 m a.s.l. (Fig. 2D). Many outlet glaciers calve into surrounding seas (Franklinbreane, Sabinebreen, Rjipbreen, Bodleybreen, Aldousbreen, Frazerbreen, Idunbreen, Bragebreen and Gimlebreen). Their termini are grounded.

Initial work on ablation and accumulation on Vestfonna was done in 1931 by the Swedish-Norwegian Arctic Expedition (Ahlmann 1933). Measurements of mass balance elements and stratigraphy of snow-firn pits were executed along a traverse across Nordaustlandet. The next results of snow thermal conditions and accumulation were delivered by the Oxford University Arctic Expedition of 1935–1936 (Glen 1939). Wide snow-cover studies were conducted during the International Geophysical Year 1956–1957 by the Swedish Glaciological Expedition to Nordaustlandet led by Valter Schytt (1964).

**Meteorological and climatological settings**. — Distribution of snow thickness is determined mainly by climatological and meteorological conditions influencing the amount of snow precipitated, and its redeposition. Various areas on Svalbard are under different climatic conditions. It is affected by air masses of contrasting properties, *i.e.* warm air masses advected from the south and south-west, formed by the Icelandic low pressure center, and on the other hand cold masses from the east or north-east related to a high pressure center in the Arctic Ocean or Greenland. Collision of the distinct air masses determines weather variability, as well as thermal, precipitation and wind conditions. The most changeable weather conditions occur in winter, when contrasts between properties of both air masses are the most significant (Hanssen-Bauer *et al.* 1990).

Due to specific light conditions in Svalbard, the types of air mass advection play a crucial role in winter air temperature formation. The air temperature in northern Svalbard, affected more often by cold air masses from the east and north-east, is considerable lower than in the southern part of the archipelago where southern and south-western advection is more frequent. Hence, the difference in

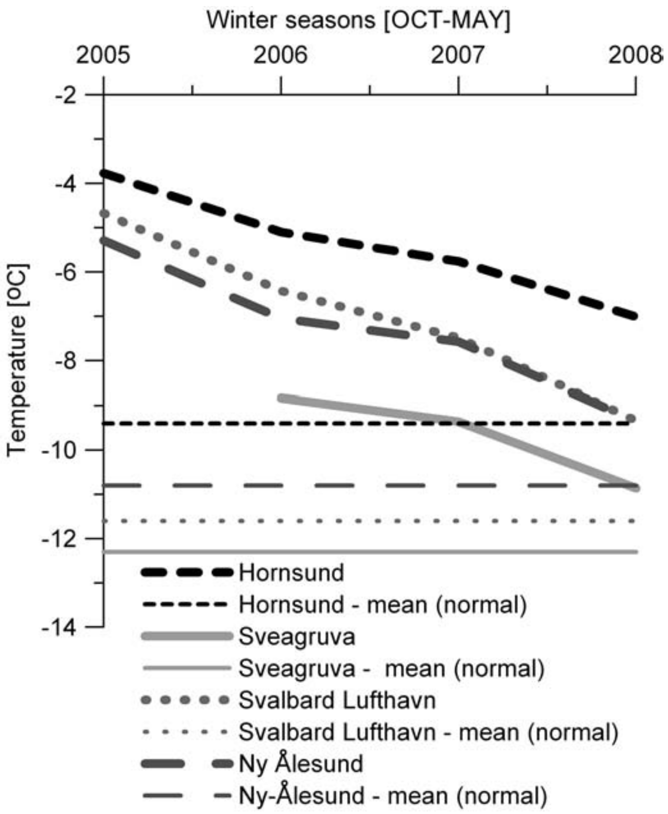


Fig. 3. Mean temperatures of winter seasons (OCT-MAY) 2005/2006–2008/2009, and long-term average winter temperatures over the normal period (1961–1990), at Svalbard meteorological stations.

mean long-term winter air temperature between Hornsund (S Spitsbergen) and Ny-Ålesund (NW Spitsbergen) is 1.4°C (Fig. 3). When comparing the average air temperature of the winter season 2007/2008 on Vestfonna (335 m a.s.l.) and Hansbreen (200 m a.s.l.), the difference is 3.7°C (-12.3°C on Vestfonna after Beaudon and Moore 2009, www.kinnvika.net, and -8.5°C on Hansbreen). Also, due to distance from maritime influence, the interior of Spitsbergen is colder than the coastal areas. The average long-term temperature in the internal stations (Longyearbyen, Sveagruva) is lower by *ca* 1 to 3°C than in coastal stations (Hornsund, Ny-Ålesund) (Fig. 3). The variability of winter weather in Svalbard is expressed in thaws with temperature above 0°C. The thaws are recorded in the snow cover as ice-crusts. These play a very important role in the evolution of snow cover mainly by limiting the deflation of snow.

Precipitation in the Arctic is rather low due to relatively stable air stratification and low water vapour content in the atmosphere (Hanssen-Bauer *et al.* 1990). Winter precipitation is formed mainly by snowfall, however rain is not excluded. At Hornsund coastal station, liquid precipitation constitutes of *ca* 20% of total

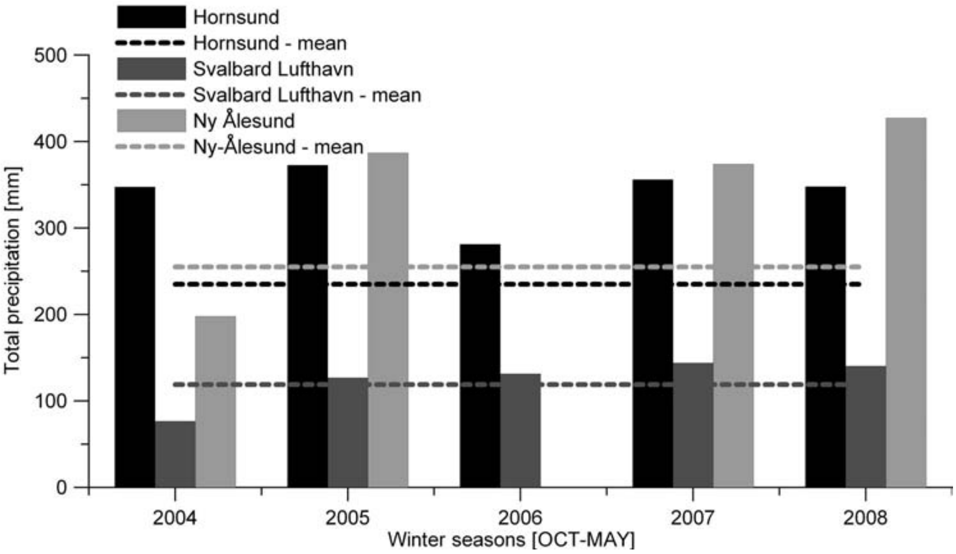


Fig. 4. Total precipitation in winter seasons (OCT-MAY) 2004/2005–2008/2009 and mean long-term sum of precipitation in winter season recorded at Svalbard meteorological stations .

winter (OCT-MAY) precipitation (Łupikasza 2003). The proportion of rain in total precipitation is much lower at higher elevations on glaciers. The spatial variability of precipitation in Svalbard is high (Hagen *et al.* 1993) and it is determined by exposure of slopes towards the direction of advection of humid air masses. Hence, maximum precipitation is related to orographic barriers along the coasts. According to field observations, the winter precipitation increases between 14 and 25% for every 100 m of elevation (Tveit and Killingtveit 1994; Hagen and Lefauconnier 1995). Hagen *et al.* (1993) pointed to predominant eastern advection of air masses as a reason for occurrence of the maximum annual sum of precipitation on the eastern coasts of Spitsbergen. On the other hand, precipitation on the western coast is mostly supplied by western and southern air advection (Førland *et al.* 1997b; Łupikasza 2007). The central part of the island is located within a precipitation shadow. As an example the average long-term sums of winter precipitation at Svalbard Lufthavn are *ca* 50% lower than those observed in coastal stations (Hornsund, Ny-Ålesund) (Fig. 4). The amount of precipitation in Nordaustlandet is comparable to that in the northern limits of Spitsbergen (Hagen *et al.* 1993). The heaviest precipitation and thereby the highest snow accumulation in Nordaustlandet is found in the easterly exposed part of Austfonna (Taurisano *et al.* 2007), whereas the western part and Vestfonna are situated within a precipitation shadow.

The wind field at Svalbard is strongly related to the air circulation pattern. Hence, the most frequent wind directions are from the eastern sector (Markin 1975; Førland *et al.*1997a; Jaedicke and Gauer 2005). Flow directions are modified by local topography. The wind is forced along axes of fiords as well as glaci-

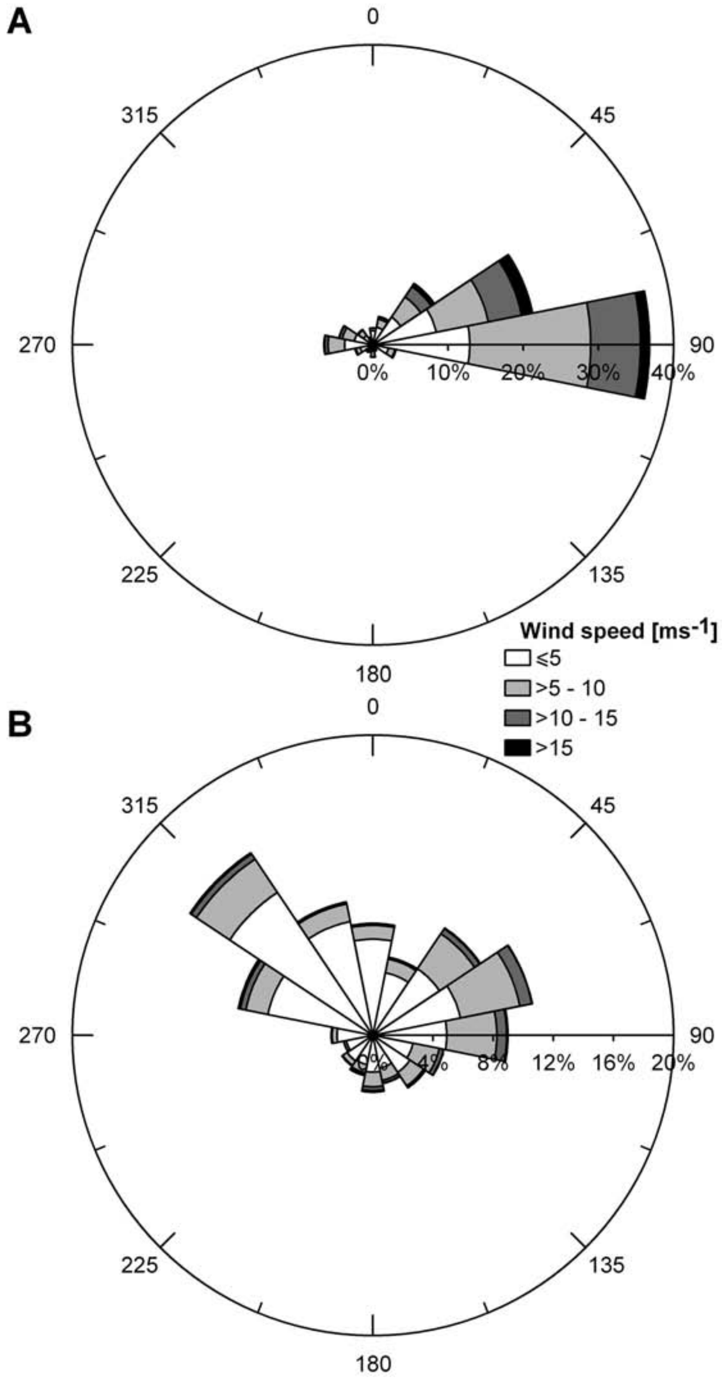


Fig. 5. The distribution of wind directions and wind speed at Hornsund in winter seasons (OCT-MAY) 1978-2011 (**A**), and at Hansbreen in winter seasons 2007/2011 (**B**). Locations shown on Fig. 2A at PPS and AWS, respectively.

Table 1

Dielectric properties and radio wave velocity in selected materials and different snow types (* after Neal 2004 and Moorman 2003)

| Material | Relative dielectric permitivitty ($\varepsilon_r$) | Radio wave velocity ($V$ [m ns$^{-1}$]) |
|---|---|---|
| Air* | 1 | 0.3 |
| Fresh water* | 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 |
| Fresh snow ($\rho$ = 200 kg/m$^3$, w = 0%) | 1.36 | 0.26 |
| Dry snow ($\rho$ = 350 kg/m$^3$, w = 0%) | 1.65 | 0.23 |
| Dry snow ($\rho$ = 500 kg/m$^3$, w = 0%) | 2 | 0.21 |
| Wet snow ($\rho$ = 500 kg/m$^3$, w = 5%) | 2.82 | 0.18 |
| Very wet snow ($\rho$ = 500 kg/m$^3$, w = 10%) | 3.83 | 0.15 |
| Slush ($\rho$ = 500 kg/m$^3$, w = 15%) | 5.06 | 0.13 |

Table 2

Vertical incidence reflection coefficients for some snow – other materials settings

| Settings | Reflection coefficients R (from –1 to +1) |
|---|---|
| Dry snow ($\rho$ = 350 kg/m$^3$, w = 0%) | |
| – Air | 0.12 |
| – Bedrock | -0.27 |
| – Dry clay | -0.22 |
| – Saturated clay | -0.59 |
| – Frozen sediment | -0.27 |
| – Ice | -0.16 |
| Dry snow ($\rho$ = 500 kg/m$^3$, w = 0%) | -0.05 |
| Very wet snow ($\rho$ = 500 kg/m$^3$, w = 10%) | -0.21 |
| Very wet snow ($\rho$ = 500 kg/m$^3$, w = 10%) | |
| – Air | 0.32 |
| – Bedrock | -0.07 |
| – Dry clay | -0.01 |
| – Saturated clay | -0.47 |
| – Frozen sediment | -0.11 |
| – Ice | 0.05 |

ated and unglaciated valleys (Wielbińska and Skrzypczak 1988; Hanssen-Bauer *et al.* 1990; Styszyńska 2007). The Hornsund area is a good example. At the station near the Polish Polar Station, the most frequent eastern winds are additionally strengthened by the E-W course of the fiord, whereas on Hansbreen the valley course forces more frequent winds from NW and ENE (Fig. 5). Additionally, the differences in temperature and humidity in the active boundary layer lead to development of katabatic or foehn winds (Førland *et al.* 1997a; Beine *et al.* 2001). Strong winds are frequently observed in winter seasons (Førland *et al.* 1997a; Styszyńska 2007) especially from eastern directions (Jaedicke and Gauer 2005; Styszyńska 2007).

Winter seasons, when radar soundings of snow cover were carried out (2005/2006–2008/2009), were generally warmer than the long-term average temperature of winter. The winter season 2005/2006 was extremely warm and mean temperatures at Spitsbergen weather stations were 5.6–6.8°C higher than the long-term averages (Fig. 3). But contrasts in temperature between the coast and interior were still preserved. On the other hand, the sums of winter precipitation in the studied seasons were also significantly higher than the long-term averages. In the coastal weather stations, the total winter precipitation was 20–68% higher than the mean, whereas in the interior weather station Svalbard Lufthavn it was 7–21% (Fig. 4) higher than average. The average wind speed at Hornsund in the 2005/2006 season was 7.0 m/s whereas in 2007/2008 it was 6.4 m/s; both are higher than long-term average (6.3 m/s for winter seasons 1978–2011).

## Methods

Ground penetrating radar (GPR) was used to detect snow thickness on glaciers. Radio echo-sounding is based on the reflection and refraction of electromagnetic waves from boundaries between materials of different dielectric properties. The dielectric properties of the materials are defined by relative dielectric permittivity, relating dielectric permittivity of a particular medium to that in a vacuum. The relative dielectric permittivity for selected materials, and different snow types, is shown in Table 1. The dielectric properties of snow are related to liquid water content and are in the range between 1.3 $\varepsilon_r$ and 5 $\varepsilon_r$.

The reflection coefficient characterizes the quantity of energy reflected from boundaries between media. The stronger the contrast in dielectric properties of materials, the higher the reflection coefficient (R) on the boundaries (Neal 2004):

$$R = \frac{\sqrt{\varepsilon_{r2}} - \sqrt{\varepsilon_{r1}}}{\sqrt{\varepsilon_{r2}} + \sqrt{\varepsilon_{r1}}}$$

where $\varepsilon_1$, $\varepsilon_2$ are relative dielectric permittivity of the materials 1 and 2.

The reflection coefficients between snow and selected materials are shown in Table 2. The majority of R coefficients in the table are high enough for easy identification of the boundaries between the mediums. Such properties make GPR an extremely useful tool in the measurement of snow cover thickness. The method gives

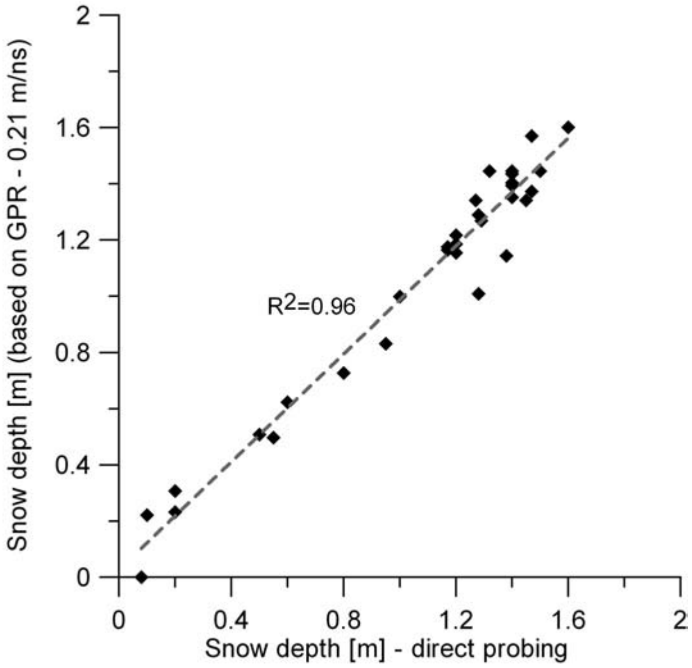


Fig. 6. Comparison of results of snow thickness measurements based on GPR soundings with those of classical probing, Vestfonna 2009.

a two-dimensional picture of spatial variability of the snow cover in a relatively short time of survey.

Impulse radar and antennas of 800 and 200 MHz centre frequency were used for sounding in the common offset mode. Major profiles were collected by a shielded antenna of 800 MHz centre frequency. A 200 MHz unshielded antenna was used for velocity profiling by the common mid-point method and common offset profiling of the snow depth on Amundsenisen. The vertical accuracy of the GPR measurement is calculated as 1/4 of the wavelength propagated by the antenna. The vertical accuracy is 6.5 cm using an 800 MHz antenna, and 26 cm using a 200 MHz antenna. The vertical accuracy of data collected by the antenna of lower frequency is relatively coarse but was used for sounding snow cover over 4 m thick. In such conditions, the vertical accuracy is better than 6.5% of the average snow thickness.

The GPR soundings were performed in spring (April–May) from 2006 to 2009, when winter snow accumulation was near-maximum. In total, a length over 235 km of radar profiles was recorded during the surveys (Hansbreen 64 km, Renardbreen 37 km, Amundsenisen 44.5 km, Vestfonna 90). The GPR antennas were pulled behind a snowmobile and a sledge. The time trigger was used to record 2 traces every second, giving a 1.5–2 m distance between the traces, depending on snowmobile velocity. On Amundsenisen, where the 200 MHz was used, traces were collected every 1 m based on a wheel odometer. The location of every GPR trace was determined by a GPS receiver working in differential kinematic mode. Coordinates were calculated based on GPS reference stations located at fixed points in the glacier's forefield and near the Polish Polar Station in Hornsund.

Common mid-point (CMP) measurement using the 200 MHz antenna was performed in April 2008 on Hansbreen (*ca* 420 m a.s.l.) in order to scale the depth of GPR profiles. Average wave propagation in the snow was determined according to CMP analysis as 0.21 m ns$^{-1}$. This value was adopted for further analysis of GPR data and for recalculating the time scale to depth. Independently of CMP measurement direct snow depth soundings were performed by means of a classic snow probe. The results of snow depth determined by probe soundings are consistent with GPR profiling and 0.21 m ns$^{-1}$ wave propagation velocity. A comparison of snow thicknesses determined by GPR and the classical method is shown in Fig. 6.

## Results

**Spatial variability of snow cover thickness on the glaciers of Svalbard**. — In the case of every glacier studied, the snow distribution is determined by a set of conditions that contribute to develop a characteristic pattern of snow cover thickness.

The GPR survey carried out at the end of accumulation season 2008 on Renardbreen shows a relatively regular increase of snow depth from *ca* 0.5 m in the front, to 4.34 m in the highest areas of the glacier (Fig. 2C). The rate of increase was 0.38 m/100m of elevation. The average snow thickness amounted to 2.12 m and standard deviation 0.58. The snow accumulation gradient was particularly regular below the equilibrium line, up to 450 m a.s.l. This area forms a NE exposed valley isolated by mountain ridges, limiting snow redeposition by predominant east winds. The variability of snow thickness on the glacier over 450 m a.s.l. results from different topographical conditions on accumulation fields (Fig. 7) that influence snow ac-

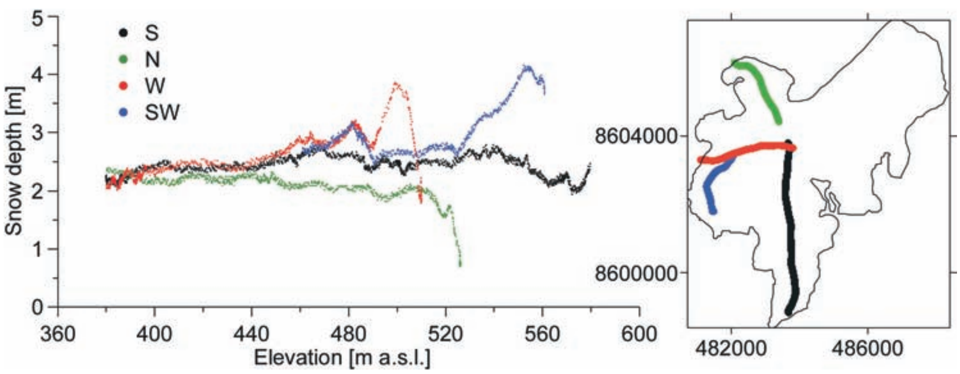


Fig. 7. Relations of snow thickness to altitude recorded in particular profiles above 370 m a.s.l. on Renardbreen in April 2008.

cumulation (niches, surrounding by mountain ridges, passes *etc.*). The highest parts of profiles terminating on passes (profiles W and N) show thin snow cover due to deflation of snow. On the other hand, profiles SW and S end in glacial cirques. The rapid decrease of the snow depth in the upper part of the SW profile results from the steep slope that limits snow deposition due to gravity creeping.

The effects of wind redeposition are more spectacular on Hansbreen. The mean snow thickness, according to radar soundings from spring 2008, was 2.8 m and ranged from 0.5 m in the frontal part to 4.5 m in the accumulation area. The standard deviation amounted to 0.61 m.

The asymmetric pattern of snow distribution on the glacier was reported earlier by Jania and Pulina (1990) and Grabiec *et al.* (2006). The distribution of snow thickness in spring 2008 (Fig. 2A) did not differ from earlier studies (compare to Grabiec *et al.* 2006). Minimum snow accumulation was observed near the glacier front and along the eastern side of the glacier between the calving front and Kvitungisen. Such snow distribution is caused mainly by deflation of snow by strong eastern winds, and redeposition towards the west. In general, Sofiekammen forms a natural obstacle for the east wind. It may generate a local winter foehn that accelerates snowdrift processes. Winter redeposition episodes form the snow cover on Hansbreen. The asymmetry of snow distribution is exemplified by longitudinal profiles (in elevation range 200–350 m a.s.l.) – eastern, central and western (Fig. 8A) and transversal profiles running at *ca* 200, 300 and 370 m a.s.l. (Fig. 8B). The figures show that asymmetry fades out towards the north and west. While differences of snow depth between the eastern and western profiles at 200 m a.s.l. amounted almost to 1 m, at 350 m a.s.l. the differences were only *ca* 0.25m (Fig. 8A). The E-W gradient of snow accumulation is confirmed by transverse profiles (Fig. 8B). However, it diminishes northward from 3 cm/100 m at 200 m a.s.l. to 2 cm/100 m at 370 m a.s.l. Additionally, the E-W gradient was observed only along one half of the length of the 370 m a.s.l. profile. The decrease of disproportion in the snow thickness towards the north may be related to weaker wind force in a zone where the orographic obstacle (Sofiekammen) is lower.

The effect of snow drifting from the east does not influence the amount of snow on western tributary glaciers as was suggested in earlier studies (Grabiec *et al.* 2006). The snow thickness on lateral glaciers is generally lower than along the central line of the main stream (Fig. 8C). The snow accumulation, as well as the rate of increase with elevation, is lowest in the southernmost, Tuvbreen, whereas the highest values are noted in the northernmost, Staszelisen. The explanation of such phenomenon might be that the northern part of Hansbreen is easily affected by air masses advecting from the east along Kvitungisen, whereas the southern tributary glaciers are within a precipitation shadow.

The GPR measurements in spring 2006 on Amundsenisen gave average snow thickness as 5.7 m (ranging from 3 to 8 m). The standard deviation was 0.56 m. Such a thick snow cover was a result of high precipitation in the

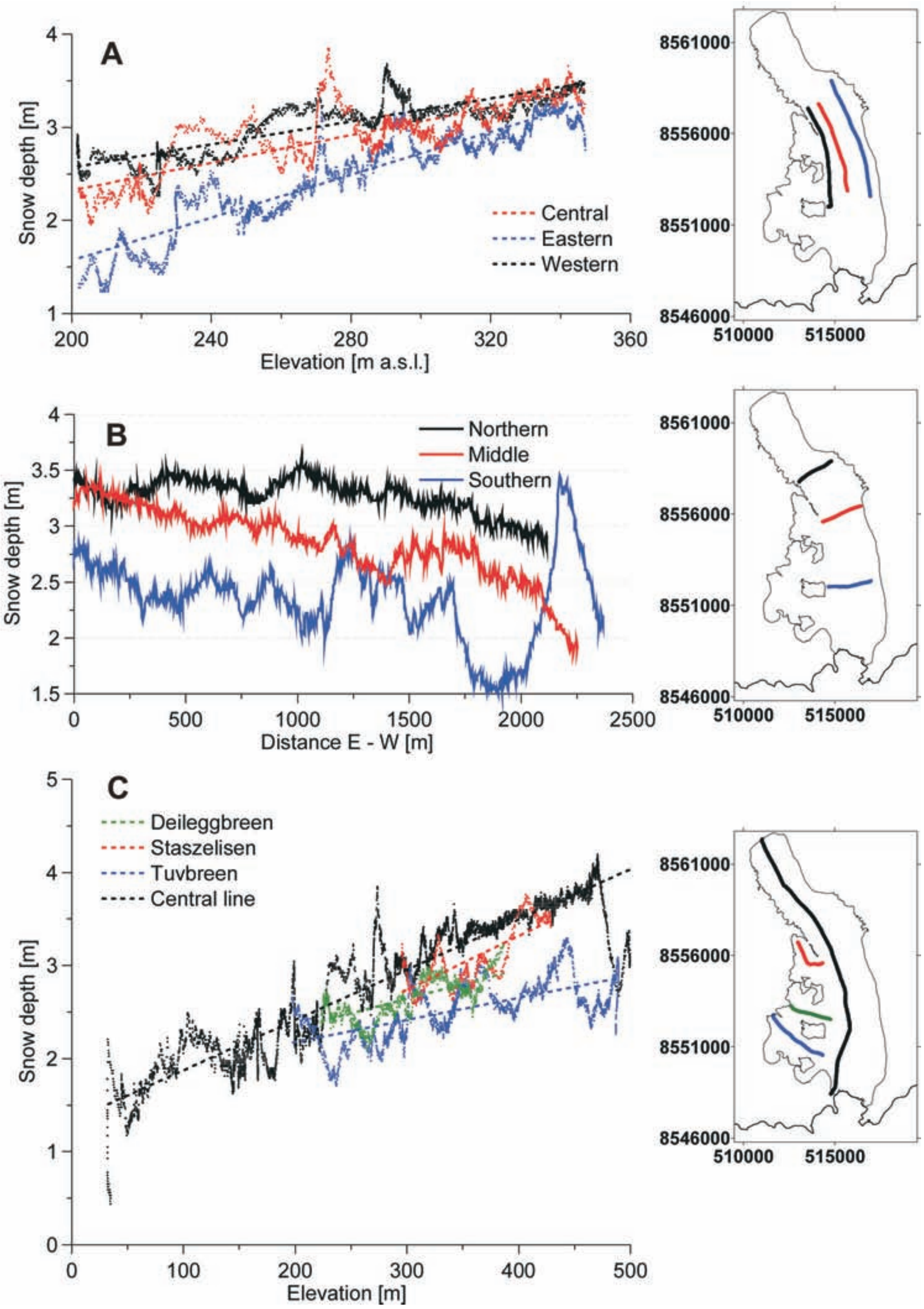


Fig. 8. GPR derived snow thickness and relations to altitude and distance on Hansbreen in April 2008 measured in particular profiles: A – parallel profiles: eastern, central and western in elevation range 200–350 m a.s.l., B – southern, middle and northern profiles, C – profiles along the central line and along tributary glaciers.

2005/2006 winter season. In Hornsund, winter precipitation was 29% higher than the long-term average (Fig. 3). Hence, the winter balance of Hansbreen was 37% higher than the average. It was the highest in the 20-year history of mass balance monitoring of the glacier. The snow depth measured at 440 m a.s.l. on Hansbreen was 4.7 m. Unfortunately, there are not full direct measurements of snow thickness on Amundsenisen. The only snow pit was 2.5 m deep and did not reach the previous summer layer. Based on analysis of a deep snow-firn pit, Baranowski (1977) estimated average winter accumulation on Amundsenisen at 1.56 m w.e. Average accumulation in the winter season 2005/2006 amounted to 2 m w.e., *i.e.* 28% more than the value estimated by Baranowski (1977). So far, the highest winter accumulation on the plateau was noted in 1990/1991 – 1.95 m w.e. (Jania 1994). In the same season, the winter balance on Hansbreen was 11% lower than in the 2005/2006 season.

The above evidence suggests that the results of snow thickness from radar soundings are reliable. However, the season to season variability of mean snow depth on Amundsenisen should be strongly underlined. Based on GPR survey, in April 2001, Melvold (2008) determined the mean snow thickness on the plateau as 1.79 m, *i.e.* 0.61 m w.e. That value is only 38% of the mean winter accumulation estimated by Baranowski (1977). Also in 2000/2001, the winter balance on Hansbreen was one of the lowest during 20 years of observations. It amounted to 0.78 m, 18% lower than the mean long-term value. The spatial variability of snow thickness on Amundsenisen is related to deflation-deposition areas (Fig. 2B). Deflation took place mainly in the NW and NE parts of the plateau, where the situation resembles that on the eastern side of Hansbreen. The snow deposition and maximum snow depth appeared on the opposite side of Amundsenisen, *e.g.* on W and SW parts of the plateau. Deflation of snow, producing lower accumulation, occurred also in transition zones between the plateau and outlet glaciers, especially between Høgstebreen and Nornebreen. A radar profile along the axis of Amundsenisen also shows relations between convex surface mesoforms with snow erosion areas, and concave forms with snow deposition (Fig. 9).

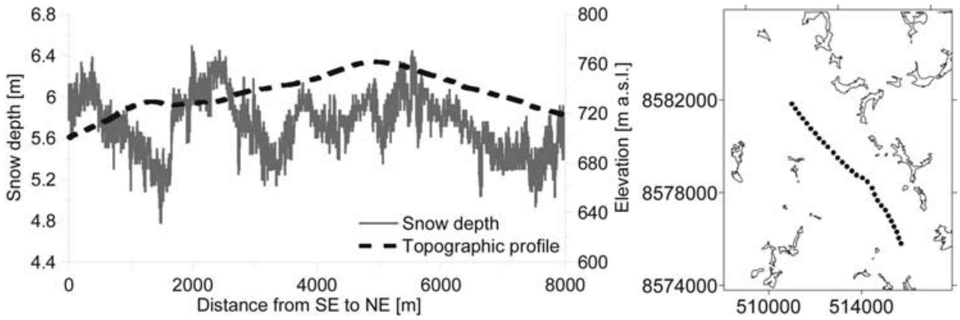


Fig. 9. Variability of snow thickness and elevation changes along the axis of Amundsenisen (April 2006).

A different pattern of spatial variability of snow cover was observed on Vestfonna ice cap. Snow distribution on ice cap slopes is related to a precipitation-elevation gradient, but vast areas on the top of the cap show a distinct snow accumulation pattern, where neither elevation change nor topography of ice-free surroundings play a crucial role. It seems that the air circulation pattern over Nordaustlandet and the mesoscale surface roughness (over horizontal distances ranging from a few hundred meters to a few kilometres) play the main role in snow distribution on the interior of Vestfonna.

The radar soundings performed in spring 2009 were conducted mainly along the central ridge and in a profile from the forefield near Doncker to Ahlmann Summit on the top of the ice cap. The snow thickness ranged between 0 m, in the areas affected by wind erosion, and 2.24 m, with an average value of 1.12 m and standard deviation 0.19 (Fig. 2D). The amount of snow accumulation reflects low precipitation values in that region. Additionally, as mentioned above, Vestfonna is located within the precipitation shadow of Austfonna. Analysis of a GPR profile from the forefield to the top of the ice cap indicated a relatively close relationship between snow accumulation and elevation over the range of altitude 350–600 m a.s.l. Below 350 m a zone of deflation *ca* 2 km wide extended with minimum snow depth 0 m. A narrow (*ca* 0.5 km) zone of redeposition occurred below, directly at the glacier front. As a result of redeposition the snow thickness is the highest (1.3 m) close to the edge of Vestfonna, and remains in the form of a band enclosing most of the ice cap through the whole year (Ahlmann 1933). These are clearly visible on summer satellite images. Snow thickness along the crest of Vestfonna shows distinct variability with a slight increase towards the east.

**Accumulation gradient**. — Despite the differences in spatial patterns of snow accumulation as shown above, the most important factor determining spatial variability of snow thickness is altitude above sea level. The influence of elevation on snow distribution is clearly seen on Renardbreen. The average increase of snow depth with altitude was 0.38 m/100m (Fig. 10A). 63% of variability in snow accumulation was explained by altitude.

The snow thickness – elevation relationship on Hansbreen was less clear. The accumulation gradient changed in different parts of the glacier. The snow thickness gradient calculated along the central line was equal to 0.54 m/100 m of elevation (Fig. 8C). Based on data from all measurement points, the snow depth increased by 0.44 m/100 m of elevation on average and the determination coefficient was $r^2 = 0.61$ (Fig. 10B). The asymmetric accumulation pattern on the glacier was manifested also in the accumulation gradient. Within the elevation range 200–350 m a.s.l. the accumulation gradient along the western profile amounted to 0.62 m/100 m; along the central line 0.74 m/100 m; and along the eastern profile 1.15 m/100 m of elevation (Fig. 8A). The accumulation gradient on each lateral glacier is lower than along the central line (Tuvbreen 0.23 m/100 m, Deileggbreen 0.31 m/100 m, and Staszelisen 0.58 m/100 m of elevation) (Fig. 8C).

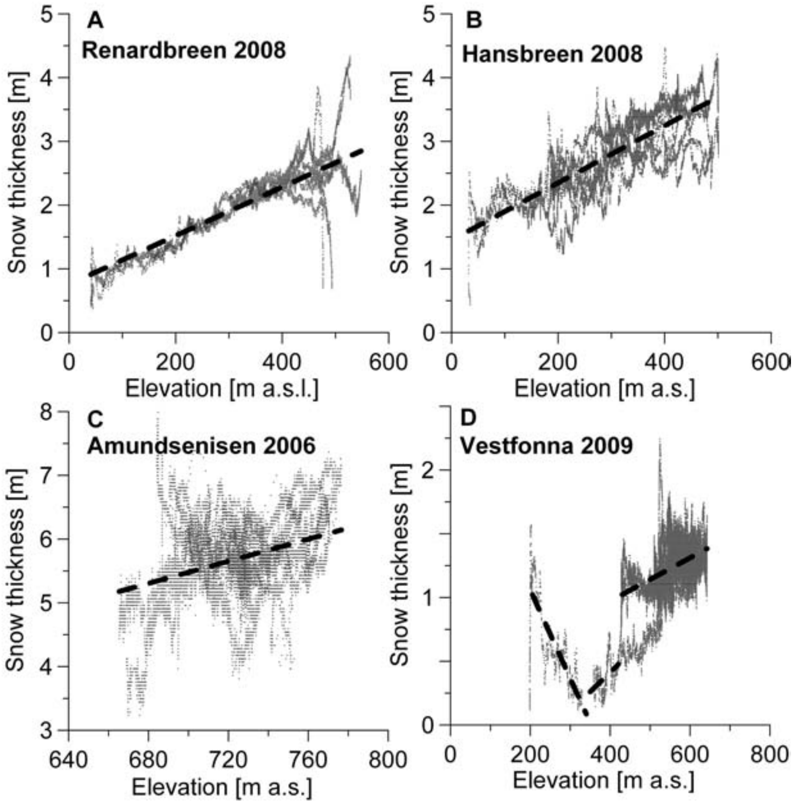


Fig. 10. Relations of snow thickness to altitude on the studied glaciers.

The altitude-dependent rise of snow thickness on Amundsenisen was not so clear due to the relatively narrow range of elevation of the plateau. The relationship became clearer in transition zones between the plateau and outlet glaciers. In the zone 660–780 m a.s.l., the average accumulation gradient was high (0.87 m/100 m), but the determination coefficient was low ($r^2 = 0.13$) (Fig. 10C).

For Vestfonna three types of snow depth-altitude relationships can be distinguished (Fig. 10D). In the lowest (inversion) zone (200–330 m a.s.l.), the snow depth decreased with altitude as an effect of redeposition of snow blown off the slopes above. In this zone, the accumulation gradient was estimated as -0.69 m/100 m of altitude. In the middle zone (330–430 m a.s.l.), snow depth regularly increased at 0.30 m/100 m of elevation. The highest zone of the ice cap (over 430 m a.s.l.) shows a snow distribution pattern similar to that described for Amundsenisen. The snow thickness – altitude relationship was insignificant ($r^2 = 0.16$) and the zonal accumulation gradient was 0.17 m/100 m.

**Normalized directional gradient of snow accumulation**. — In order to evaluate factors determining the spatial variability of snow thickness, the normalized directional gradient of snow accumulation (NDG) in S-N and W-E directions was calculated (Fig. 11). The NDG factor is independent of the accumulation gradient and was calculated as an absolute deviation of snow thickness from the linear regression of snow depth *versus* altitude in S-N and W-E directions. The NDG factor for Vestfonna was calculated based on every record above 430 m a.s.l. Data from lower elevations were excluded due to clearly distinct accumulation conditions on slopes of the ice cap in comparison to the internal part of Vestfonna. On the other glaciers, the NDG was calculated based on observations from the entire elevation range.

Generally there is no one dominant trend of direction of snow thickness change on all studied glaciers. On particular glaciers, different directions of snow thickness change played the main role, as a joint result of regional air circulation conditions, local precipitation pattern and topoclimatic conditions.

The clearest directional trends in snow accumulation were observed on Amundsenisen. According to the NDG analysis, snow thickness on the plateau increases towards the south and east. The NDG factor in the S-N direction was calculated as -0.11 m/km (Fig. $11C_1$), whereas in the W-E direction it was 0.09 m/km (Fig. $11C_2$).

No significant directional trends were found in snow accumulation variability on Vestfonna. The NDG factors are low and equal 0.01 m/km in both S-N and W-E directions (Fig. 11D). In fact, low values of absolute snow depth work against high values for NDG. Slightly better conditions of snow accumulation towards the north and east can be explained by a general trend of precipitation increase towards the eastern side of Nordaustlandet (Taurisano *et al.* 2007).

On the remaining glaciers the directional changes of snow thickness are more complex. The snow conditions on Renardbreen improve from the north and south towards the center of the S-N range. The NDG factor from the south to the center is 0.19 m/km and then -0.13 m/km towards the north (Fig. $11A_1$). The NDG factor in the W-E direction is more homogeneous but only 0.04 m/km (Fig. $11A_2$).

West-east asymmetry of snow thickness is strong on Hansbreen (Fig. $11B_2$). From the centre of the W-E range, the NDG factor toward the east is very strong and amounts to -0.29 m/km. Toward the west, the dispersion of deviation of snow thickness around the regression line is high, and the average NDG factor is 0.07 m/km. This value is related to western tributary glaciers of dominant W-E course. This high scatter is associated with different accumulation conditions on the lateral glaciers (different elevation, location, topographic conditions, distance to the open sea, *etc.*). W-E directional changes show the most favourable conditions for snow accumulation along the central line and western part of the main glacial stream. Accumulation conditions are poorer towards the eastern side of the glacier and on lateral glaciers on the western side of Hansbreen. Additionally, the accumulation conditions on particular tributary glaciers are very diverse. In the S-N direction the NDG factor was only 0.04 m/km, with high scatter around the regression line (Fig. $11B_1$).

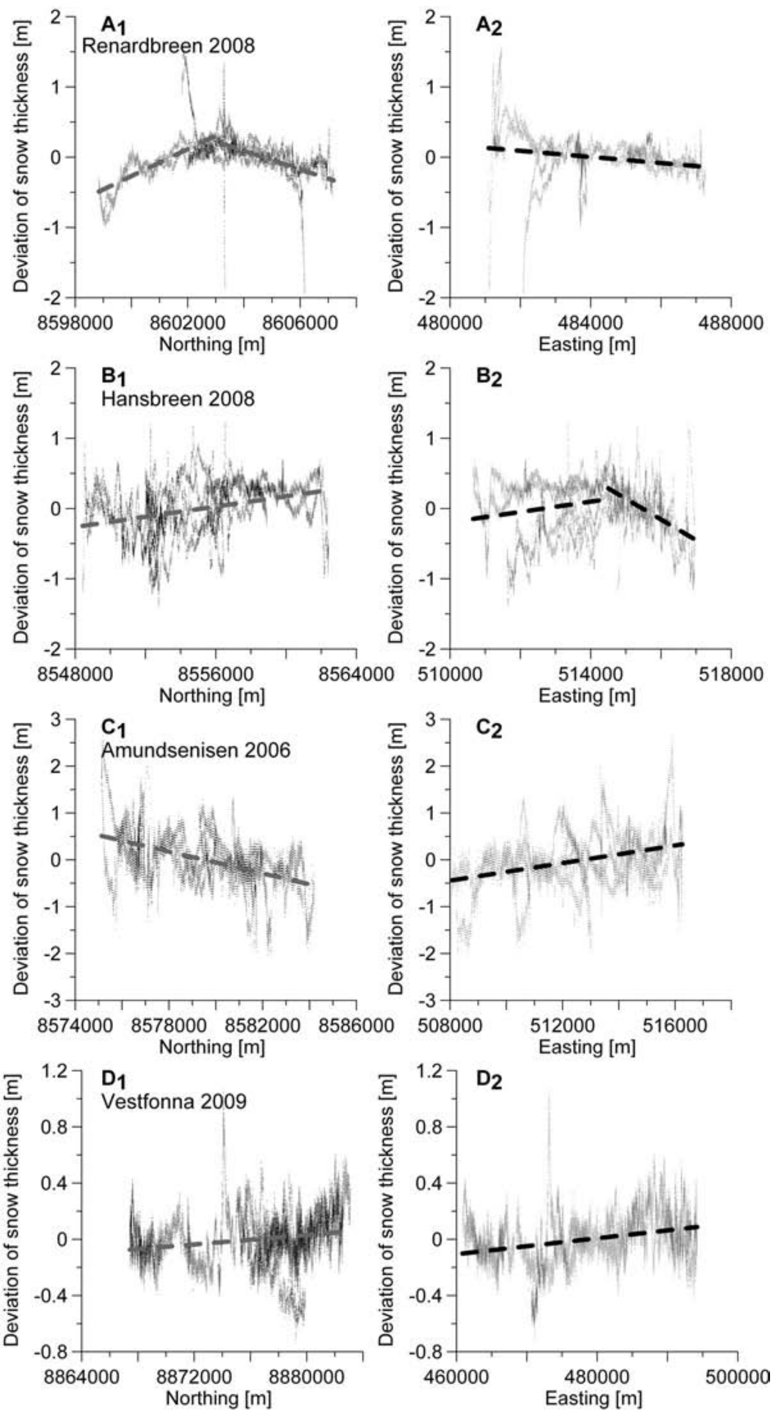


Fig. 11. Normalized directional gradients of snow accumulation (S-N in left column; W-E in right column). Coordinate systems: UTM 33X on wgs84 (**A**–**C**), and non-standard UTM 34X on wgs84 (**D**).

## Discussion

This report presents the spatial distribution of snow thickness on four Svalbard glaciers of different types. The snow distribution on studied glaciers is an effect of numerous factors of different significance.

The amount of snow is controlled by solid precipitation distribution that is the subject of regional trends and local topoclimatic effects. Regional precipitation variability is controlled by distance to the open water as a source of moisture. This trend on Svalbard is manifested in diminution of total precipitation toward the interior (Hagen *et al.* 1993). Such a pattern may be clearly observed on extensive glacier areas like Vestfonna. Despite the limited dataset, eastward increase of the NDG factor corresponds to the regional precipitation pattern on Nordaustlandet (Taurisano *et al.* 2007).

On smaller glaciers, local topoclimatic conditions may play a more important role in precipitation distribution. The most distinct orographic effect is the increase of precipitation with altitude that reflects in snow accumulation gradient. On Renardbreen and Hansbreen the accumulation gradient was respectively 0.38 and 0.44 m/100 m of altitude. The values compare with 0.10 m w.e/100 m obtained for austre Brøggerbreen (Hagen and Lefauconnier 1995). However, topographic influence on the precipitation field is also related to location of orographic barriers in relation to the direction of humid air mass advection. Ideally, maximum orographic precipitation falls as the convergence effect on windward slopes, *i.e.* along the western coast mostly on westerly-facing slopes. On Hansbreen and Renardbreen, the most favourable snow accumulation conditions appear on the western or central parts of the glaciers, sheltered by the N-S oriented ridges. Similarly, Førland *et al.* (1997b) recorded a higher sum of precipitation on sheltered austre Brøggerbreen than on slopes exposed to moist western advection. Such a phenomenon was explained by a combination of two processes related to orographic precipitation: spillover and seeder/feeder (Førland *et al.* 1997b), however, in winter the first process seems to be more important. The spillover effect is controlled by continuation of moist air mass lifting on the leeward side as well as by wind induced transfer of snowflakes during the fall from clouds to the ground. The spillover transfer of droplets at moderate wind speed is 2–4 km, whereas snowflakes may be transferred much more effectively (Førland *et al.* 1997b).

When the snow is deposited on the glacier's surface it becomes the subject of wind redeposition that is one of the most effective processes of reshaping the initial precipitation pattern of snow distribution. Wind drift starts when a wind-shear velocity exceeds the threshold shear velocity and the snow is available to transport (Liston and Sturm 1998). The direction of redeposition is related to the direction of wind which is capable to transport the snow particles. The strongest eastern winds on Svalbard favour westward snowdrift. Even on the glaciers of N-S axis like Hansbreen, where valley course produces more frequent winds from NW, easterly

(ENE) winds remain very strong (Fig. 5B). Snow is eroded from convex and deposited on concave forms on the glacier's surface. Deflation along the eastern parts of N-S trending glaciers (for example Hansbreen) is controlled by:

– eastern wind frequently exceeding threshold shear velocity. Wind speed at the leeward foot of the massif is considered to be equal to the wind velocity at the windward side of the obstacle (Walter *et al.* 2004). A modelling study from central Spitsbergen shows that high wind speed often occurs on the leeside behind large mountains (Jaedicke and Gauer 2005);
– location beyond the zone of snow deposition on the leeside of the massif, as an effect of rapid drops of wind speed just below the crest;
– slightly convex shape of glacier's surface towards the west in the ablation area (perpendicular to the course of the main tongue). In consequence, snow deposition appears on the western side of the glacier;
– possible foehn-like wind blowing from eastern slopes. Some indirect evidence of a foehn, such as typical clouds, were observed over Sofiekammen in the Hornsund area. However, due to relatively low elevation range (*ca* 150–200 m), the foehn effect is not clearly indicated in meteorological records. Temperature rise and decrease of humidity are not so distinct.

The next set of factors controlling snow distribution on glaciers is related to spatially varied snow properties that make the snow available to transport. In the case of a vegetation-free area, those properties are determined by changes in density and microrelief.

Processes that may effectively change the properties of snow density are mid-winter thaws and liquid precipitation. Both of them are thermally controlled and their influence decreases towards higher altitude. The effect of those processes on snow cover is also similar. Snow thickness may decrease slightly due to superficial ablation and densification, but water supplied by rain or melting refreezes forming ice crusts effectively protecting the surface against deflation. According to our data, there is no evidences of diminishing snowdrift effects towards lower elevation as an effect of better snow cover protection by ice crusts. Hence, snow properties seem not to be crucial for general snow distribution, as most snow is redeposited right after a snowfall. On the contrary, in more cases the rain or thaw preserve the effects of redeposition. Some role in snow distribution is played also by avalanches and snow drift supply from outside the glacier. Both of them are considered as marginal on the studied glaciers. The role of avalanche and external snowdrift in distribution of snow accumulation is much significant on small cirque or valley-cirque glaciers, especially on their peripheries.

The significance of particular factors determining spatial variability of snow accumulation varies within and between each glacier, hence a number of snow distribution patterns can be distinguished. The precipitation gradient pattern is the classical type of spatial distribution of snow cover on glaciers. The snow accumulation increases regularly with altitude at a given precipitation – elevation gradient.

Due to the simplicity of its assumptions, this pattern is often used for winter mass balance modelling on glaciers (*e.g.* Grabiec 2005). Regular increase of snow thickness occurs in valley or outlet glaciers of spindle-like shape. The elevation range of the glacier should be considerable in order to define a clear relation between precipitation and elevation. A single accumulation field, or a complex of accumulation fields of similar topographic conditions (altitude, slope aspect, surface and surrounding topography), is also favourable for such regular distribution of accumulation. The tongues of glaciers of that type usually have a linear course, and they can terminate either on land or in water. Local topography and topoclimatic conditions play a secondary role in this pattern of snow thickness distribution. Topographic conditions have minor significance on glaciers with axes parallel to the most frequent wind direction (Grabiec *et al.* 2006). In such cases snow redeposition does not disturb the precipitation gradient pattern of snow accumulation very much. The regular snow distribution pattern often occurs only on the lower and middle parts of a glacier. The influence of topography and topoclimate is more evident in the highest parts, close to the mountain slopes. Within this study, such a pattern of snow distribution was observed on Renardbreen (Figs 2C, 10A). The general course of the axis is SE-NW, which is close to the most frequent wind direction from the eastern sector. A similar snow distribution structure can be also expected on other Svalbard glaciers, such as Aavatsmarkbreen (Grabiec *et al.* 2006) in Oscar II Land, the lower and middle part of midre Lovènbreen (Hagen and Liestol 1990) in the Kongsfjord area, and the lower and middle zones of Fridtjovbreen (Troickij 1991, 1996) in the Bellsund area.

A much more frequent type of snow accumulation distribution on Svalbard glaciers is the precipitation-redistribution pattern. The precipitation-elevation gradient still plays a significant role but the pattern is considerably modified by local accumulation conditions. The pattern is represented by glaciers of almost all types with a quite high range of elevation (a few hundred meters). Snow thickness at a particular point comes from the snowfall modified by topographic and orographic conditions as well as wind-induced snow redistribution. Areas of snow deflation and deposition can be distinguished on the glacier surface. The precipitation-redistribution pattern of snow accumulation is favoured by glaciers of complex accumulation fields and tongues that change direction. Such circumstances result in very variable conditions of snow accumulation (slope aspect, elevation, surface topography, topography and orography of the surroundings). Exposure of particular parts of a glacier to the most frequent wind direction and to advection of mild air masses that ensure high precipitation are important factors determining a precipitation-redistribution pattern. Hansbreen is an example of such pattern (Fig. 2A). The axis of the glacier runs perpendicularly to the strongest eastern wind and to the advection of air masses that give effective winter precipitation. The deflation process from the eastern part of the glacier was described by Grabiec *et al.* (2006). On the eastern side of Hansbreen, a strong wind deflates a considerable amount of snow and redeposits it in the central and western part of the tongue, some 2 km away. The process may be accelerated by a foehn-like wind flowing down from Sofiekammen. Measurements carried out in spring 2008 excluded the influence of drifted snow on snow thickness extending as far as the western tributary glaciers, as had been suggested in previous work by Grabiec *et al.* (2006). Snow accumulation on lateral glaciers was considerably lower in comparison to the center line or western side of Hansbreen. This is related to availability or deficit of precipitation in accumulation areas isolated by mountain ridges (Fig. 8C). Tributary glaciers are sheltered from western advection of moist air masses, whereas the western and central sides of the main tongue are favourable for snowfall as result of a spillover effect (Førland *et al.* 1997b).

A similar snow accumulation pattern has been described on many Spitsbergen glaciers, *e.g.* on austre Grønfjordbreen (Troickij 1991, 1996), vestre Grønfjordbreen (Solovyanova and Mavlyudov 2007), Bogerbreen, Bertilbreen (Gus'kov 1983; Gus'kov and Troickij 1984, 1985, 1987), Longyearbreen (Gus'kov 1983; Gus'kov and Troickij 1984) and Finsterwalderbreen (Hodgkins *et al.* 2005). Annual recurrence of the accumulation pattern is typical for these examples.

Most small mountain glaciers, with irregular distributions of accumulation and relatively low influence of altitudinal gradient, can also be included in the group with a precipitation-redistribution pattern of snow distribution. Uneven and variable winter precipitation is largely redeposited, driven by local topoclimatic conditions. Accumulation from precipitation is quite frequently supplied *via* avalanches and snow drift from outside glaciers. On these glaciers, the accumulation pattern that occurs in one season will not necessarily repeat in next season. Examples of small glaciers with irregular winter accumulation are quite numerous on Svalbard, *e.g.* Daudbreen (Gus'kov 1983; Gus'kov and Troickij 1984), Vøringbreen (Gus'kov 1983; Gus'kov and Troickij 1984, 1985, 1987; Solovyanova and Mavlyudov 2007), Aldegondabreen (Solovyanova and Mavlyudov 2007) and Waldemarbreen (Grześ 1996; Grześ and Sobota 1998, 1999, 2000; Sobota 2002, 2005, 2007a, b; Sobota and Grześ 2006).

Ice fields or plateaus surrounded by nunataks or mountain ridges show a redistribution pattern of snow accumulation. A classical precipitation gradient pattern cannot form due to the relatively small elevation range or to its modification by local conditions. Hence the spatial variability of snow accumulation is determined mainly by the topography of the glacier surface or its surrounding, affecting the topoclimatic conditions, especially wind and precipitation fields. Favourable conditions for snow accumulation are observed within concave mesoforms of relief or on the leeward side of orographic dams. However, less favourable accumulation conditions, where snow is deflated, occur on convex forms of relief or on windward slopes. In this work, the redistribution pattern of snow accumulation is represented by Amundsenisen (Fig. 2B). A similar pattern of snow distribution is expected on other Svalbard ice fields, *e.g.* Holtedahlfonna, Isachsenfonna, Løvenskioldfonna, Lomonosovfonna, Hellefonna, Gruvfonna, Ursafonna, Filchnerfonna and Sørkappfonna, but not on their outlet glaciers.

Different types of snow accumulation pattern occur in particular elevation zones of Vestfonna (Figs 2D, 10D). From the lowest elevation to the top, the accumulation patterns are as follow: inversion, precipitation and redistribution. The inversion pattern is related to the low peripheral area of deposition of snow deflated from the ice cap, especially from steep slopes. The deposition area is located at the foot of the ice cap. This zone was previously described by Ahlmann (1933). The snow thickness in this area decreases with increasing altitude, and reaches a minimum in the zone of complete snow deflation down to the bare ice. The inversion pattern of accumulation is topographically determined, *i.e.* can arise below a steep slope, within a concave area. The snow accumulation conditions in the upper zones of Vestfonna are similar to the redistribution and precipitation gradient patterns described above. In the case of megaforms like Vestfonna, the spatial distribution of snow accumulation is additionally modified by regional climatic factors such as the dominant type of air circulation. On Vestfonna, the influence of regional factors is manifested in the increase of snow accumulation towards the east, an effect of the most frequent advection of humid air masses from the east. Similar conditions of snow accumulation were observed on the largest ice cap of Nordaustlandet *i.e.* Austfonna (Schytt 1964; Taurisano *et al.* 2007). Analogous patterns of snow distribution are expected on other ice caps on Svalbard with comparable topographic conditions, *e.g.* Åsgårdsfonna, Balderfonna on Western Spitsbergen, Vegafonna, Glitnefonna on Nordaustlandet, Edgeøyjøkulen, Digerfonna, Storskavlen on Edgøya, and Barentsjøkulen on Barentsøya.

## Conclusions

Based on the analysis of spatial changes of snow thickness on different type of Svalbard glaciers, various patterns of snow accumulation distribution have been distinguished.

- Precipitation pattern (Renardbreen is an example) – snow thickness strongly depends on elevation, giving a simple altitudinal gradient. This pattern may occur on outlet or valley glaciers showing straight, elongated shapes. Glaciers oriented parallel to the most frequent wind direction and/or direction of mild air mass inflow also favour the precipitation pattern of snow distribution.
- Precipitation-redistribution pattern (*e.g.* Hansbreen) – the snow accumulation shows an altitudinal gradient but it is modified by local topoclimatological conditions. The correlation between snow thickness and altitude is weaker than in the precipitation pattern. This pattern is typical for most Svalbard glaciers of mountain, valley and outlet types. Conditions favouring occurrence of the precipitation-redistribution pattern of snow distribution are: complex shape of gla-

cier (more than one accumulation field, curving course of tongue), location of the glacier perpendicular to the dominant wind direction and/or advection of moist air masses.

- Redistribution pattern (*e.g.* Amundsenisen) – the snow distribution is independent of elevation gradient but complex local conditions are the most important. Among local topoclimate-related conditions, the most significant are wind-induced redeposition of snow and distribution of solid precipitation determined by local topography or orography. This pattern of snow accumulation may occur in Svalbard on ice fields or plateaus surrounded by nunataks or mountain ridges.
- Complex pattern (*e.g.* Vestfonna) – co-existence of the above-mentioned accumulation patterns, *e.g.* in different elevation zones. A complex pattern of snow distribution can occur on ice caps where a precipitation or precipitation-redistribution pattern dominates on the slopes, whereas in relatively flat upper areas the redistribution pattern of snow accumulation dominates.

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