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Post-surge geometry and thermal structure of Hørbyebreen, central Spitsbergen

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Abstract: Hørbyebreen surged in the 19th or early 20th century, as suggested by geomorphological evidences and looped medial moraines. In this study, we investigate its wide-spread geometry changes and geodetic mass balance with 1960 contour lines, 1990 and 2009 digital elevation models, in order to define the present-day state of the glacier. We also study its thermal structure from ground-penetrating radar data. Little is known about the glacier behaviour in the first part of the 20th century, but from its surge maximum until 1960 it has been retreating and losing its area. In the period 1960–1990, fast frontal thinning (2–3 m a^{-1}) and a slow mass build-up in the higher zones (~0.15 m a^{-1}) have been noted, resulting in generally negative mass balance (-0.40 ± 0.07 m w. eq. a^{-1}). In the last studied period 1990–2009, the glacier showed an acceleration of mass loss (-0.64 m ± 0.07 w. eq. a^{-1}) and no build-up was observed anymore. We conclude that Hørbyebreen system under present climate will not surge anymore and relate this behaviour to a considerable increase in summer temperature on Svalbard after 1990. Radar soundings indicate that the studied glacial system is polythermal, with temperate ice below 100–130 m depth. It has therefore not (or not yet) switched to cold-bedded, as has been suggested in previous works for some small Svalbard surge-type glaciers in a negative mass balance mode.

Key words: Arctic, Svalbard, glacier surge, glacier geometry changes.

Introduction

A glacier surge is a cyclic rapid transfer of mass from higher to lower elevated zones, accompanied by fast flow period (10–100 times higher than velocity in a quiescent phase) and, most often, a significant front advance. This dynamic instability is often considered as independent of climate fluctuations (Meier and Post 1969; Raymond 1987), but on the other hand it has been documented that it is closely correlated to the cumulative mass gain in the higher zones of the glacier, hence to the mass balance (Dyurgerov *et al.* 1985; Eisen *et al.* 2001). Surges are not a common phenomenon. Only 1% of world glaciers show surge-character (Jiskoot *et al.* 1998) but they are very common in several regions of the world, including Alaska, Pamirs or Iceland among others. Svalbard, a highly glacier-covered Arctic archipelago, is also a well known cluster of surge-type glaciers. According to different authors, they stand for 13% (Jiskoot *et al.* 1998), 36% (Hamilton and Dowdeswell 1996) or as much as 90% (Lefauconnier and Hagen 1991) of the total number of the archipelago's ice masses. The controls and occurrence of Svalbard glacier surges have been previously described by many authors and they differ from those found in the other parts of the world. Their characteristic feature is long duration of both active and quiescent phases and low velocities during the active phase, driven by environmental factors such as climate and geology (*e.g.* Liestøl 1969; Schytt 1969; Hagen 1987; Dowdeswell *et al.* 1991; Lefauconnier and Hagen 1991).

In early 20th century, a sharp climate warming occurred on Svalbard, terminating the Little Ice Age (LIA) period of glacier advances (*e.g.* Nordli and Kohler 2003). The equilibrium line altitude (ELA) has raised, decreasing the mass balance of Svalbard glaciers together with their volume and area. Dowdeswell *et al.* (1995) argues that this climate shift has had even more negative consequences. According to these authors, surge activity has decreased in Svalbard due to negative mass balance mode, which prolongs the quiescence phases or ceases the surge behaviour. They have also found evidence that small glaciers which used to surge in the past may change their thermal structure from warm-based to cold-based by enhanced thinning in all elevation bands, as showed for Scott Turnerbreen. On the other hand, it has been reported that some Svalbard surge-type glaciers are building-up towards a new surge, *e.g.* Kongsvegen (Melvold and Hagen 1998), Hessbreen (Sund and Eiken 2004), Finsterwalderbreen (Hodgkins *et al.* 2007) and many other ice masses, particularly in the eastern and southern Svalbard (Nuth *et al.* 2010; Moholdt *et al.* 2010). Also, many glaciers did surge in the last decades despite the generally unfavourable climate conditions (*e.g.* Hagen 1987; Dowdeswell and Benham 2003; Benn *et al.* 2009).

The purpose of this study is to present geometric evolution of Hørbyebreen, a surge-type glacier in central Spitsbergen, Svalbard Archipelago (Karczewski 1989). With use of maps and digital elevation models, we measured its elevation changes and geodetic balance in order to investigate whether it is building towards a new surge event. Our paper reports its length, area and volume changes since its surge maximum and we seek to link these changes with climate. We also bring new data on thermal structure of a Svalbard surge-type glacier and compare our results to the thermal shift of Scott Turnerbreen reported by Dowdeswell *et al.* (1995).

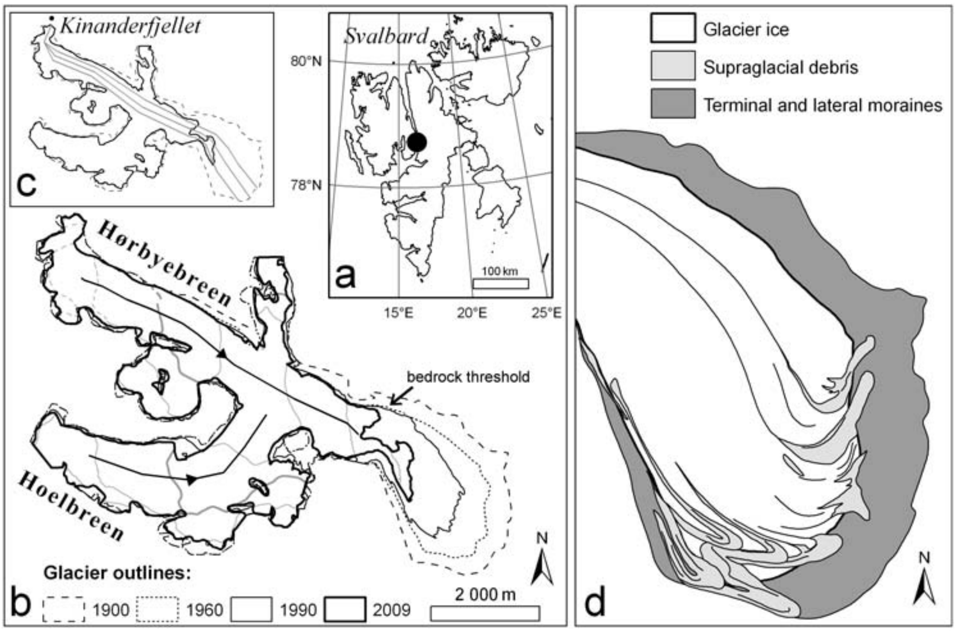

Fig. 1. Location of the studied glacier in Svalbard (**a**) and map of Hørbyebreen glacial system (**b**). Contours are drawn with 100 m interval (thicker contour – 500 m a.s.l.). Solid contours represent 2009 geometry, dashed contours - 1990. Arrow lines – centreline radar profiles (downglacier direction). Map showing profiles used for length measurements (**c**) plotted against Little Ice Age and 2009 outlines (dashed and solid line respectively). Sketch of Hørbyebreen front zone in the 1960's (**d**) with looped medial moraines in the southern part of the snout.

Study area

Hørbyebreen glacial system is one of the largest glaciers (15.9 km^2) in little glacier-covered Dickson Land in central Spitsbergen, Svalbard (Fig. 1a). It consists of two main components: Hørbyebreen and Hoelbreen, confluent in the lower part of the valley (Fig. 1b). Climatic conditions in Svalbard are relatively mild when compared to other High Arctic sites – the mean temperature of the summer period (June–August) at Svalbard Airport (60 km from Hørbyebreen) is 4.9°C (1981–2010), while of the coldest month (February) it is -13.5°C (based on data from Norwegian Meteorological Institute). Dickson Land is characterised by inner-fjord climate type, meaning arid conditions and higher annual thermal contrasts, thus – potentially warmer summer periods (Rachlewicz and Styszyńska 2007; Rachlewicz 2009a; Małecki 2013). Local glaciers may be therefore more exposed to high temperature and receive fewer precipitation, including winter snow. Hence, they are higher elevated when compared to more maritime Svalbard regions, to compensate these unfavourable conditions. Median altitude of Hørbyebreen is 500 m. In previous reports Hørbyebreen system has been expected to be polythermal on the basis of regular icings at its front, forefield geomorphology and character of water discharge (Rachlewicz *et al.* 2007; Rachlewicz 2009a, b; Evans *et al.* 2012). In this study, we investigate its thermal structure with direct geophysical methods.

Hørbyebreen is a surge-type glacial system, but little is known on the timing of the last surge event. First notes about the glacier can be found in Slater (1925), who has visited the site in 1921. In his report, the glacier front is surrounded by crescentic moraines and the ice surface is smooth with well marked debris bands and serpentine flow lines. This description indicates that the glacier must have surged at least ~20 years prior to the Slater's survey, so its cracked surface could be smoothed by melting and the front could retreat. Therefore, we assume the surge to occur in 1900, but it could have been even earlier. A number of post-surge imprints, summarized by Evans and Rea (2003), has been also noted on Hørbyebreen on Norwegian aerial images from 1960, with looped medial moraines among others (Karczewski 1989, Fig. 1d).

The present-day dynamics of Hørbyebreen are relatively high as for a Svalbard glacier of this size. A preliminary study with a limited number of fixed stakes has shown maximal flow velocities on the order of 12 m a^{-1} and an average ~25% velocity increase during summer months (Rachlewicz 2009a). Flowlines observed on its modern surface suggest that most of ice in the frontal zone is delivered from Hoelbreen basin, so it shall be considered as a more active component and a probable surge trigger. Modern geophysical studies indicate no ice-cored moraines in the forefield, what is interpreted as a remnant of the past surge by Gibas *et al.* (2005). With a recent 2009–2011 aerial survey some discussion with the surge origin of the foreland has been however opened by Evans *et al.* (2012), who presented an alternative view of its genesis, partly explained by a *jökulhlaup* event. However, these authors do not negate the surge hypothesis.

Sedimentary rocks underlying Svalbard surge-type glaciers are considered as their common characteristic (Jiskoot *et al.* 1998). Geology of the Hørbyebreen basin is mostly formed from soft sandstones, while only the lowest part of Hørbyebreen valley is intersected by resistant pre-Devonian metamorphic rocks (Dallman *et al.* 2004). They form a ~100 m high bedrock threshold, surpassed by the glacier front during the last surge and now largely influencing the decay of the present-day terminus (Fig. 1b). The frontal zone of Hørbyebreen, lying below the threshold, is flat and thin so it is subjected to very fast down-wasting. It has mostly easily distinguished edge but several debris bands are present and cover the actual glacier ice.

Data and methods

Glacier geometry information used in our study has been derived from Norwegian aerial surveys and consists of:

- a 1960 map in scale 1: 100 000 and vertical contour spacing of 50 m constructed from photographs in scale 1:50 000 and digitized by Norsk Polarinstitutt (NPI);
- an NPI 1990 digital elevation model (DEM) of 20 m resolution constructed from images in scale 1:15 000;
- a 2009 DEM with pixel size of 20 × 20 m (data missing from the highest zone of Hørbyebreen) constructed in ERDAS Leica Photogrammetry Suite 2010 software from recent aerial NPI images taken at resolution of ~0.5 m in July and August 2009.

Geometry data from 1960 and 2009 have been aligned in x, y and z directions in relation to 1990 DEM using universal co-registration correction described by Nuth and Kääb (2011). In order to investigate geometry changes between 1960 and 1990, the contour line vertices of the1960 map have been transformed to points and each of them has been interpolated into 1990 DEM with bilinear interpolation. In effect, the points contained information about local elevation differences between two periods (*dh*). From these *dh* values, we have interpolated a glacier-wide map of elevation change using three interpolation schemes: kriging, inverse distance weighting and "Topo to Raster" tool (ESRI ArcGIS software). In the latter scheme, we have manually drawn lines of equal elevation change to minimize unrealistic patterns of surface elevation change, which could be easily produced by an automatic interpolation because of sparse contour lines and large *dh* variations at the front. The resulting three rasters of 20 m resolution have been aligned and averaged into the final 1960–1990 map of *dh*. Elevation changes of Hørbyebreen system in the recent 1990–2009 period were quantified by subtraction of a co-registered and aligned 2009 DEM from 1990 DEM. To obtain total volume change of the glacier for both periods (*dV*) cells of *dh* raster were summed. Mean glacier-wide elevation changes (*dH*) were obtained by division of *dV* by average area of the glacier in a given period. To convert *dH* to geodetic mass balance (*B*), with metres of water equivalent as units (w. eq.), we assumed a constant density distribution over the glacier over time (Kohler *et al.* 2007) and multiplied *dH* by a factor of 0.9 (ratio of ice density to water density). Annual retreat rates were calculated from changes of length of the glacier in each period, measured by averaging length of three profile lines extending from Kinanderfjellet to the front (Fig. 1c). Ground penetrating radar (GPR) investigation on Hørbyebreen was carried out on 15/04/2012 during dry-snow conditions. Malå RTA system was used with ProEx control unit and 100 MHz antenna. The antenna was pulled by a snow scooter (driving with average speed of 20–30 km h^{-1}) and the data were collected approximately every 1–2 m. Processing of the data was carried out in ReflexW software and considered static correction, dewow and gain functions. For travel-time/depth conversion, we used wave propagation velocity for ice of 0.170 m ns^{-1}.

We quantified the total error of our geometry change analysis by comparing elevations from all our data sources from non glacier-covered terrain. In the first studied period, we used 18 000 points and in the more recent 35 000 points con-

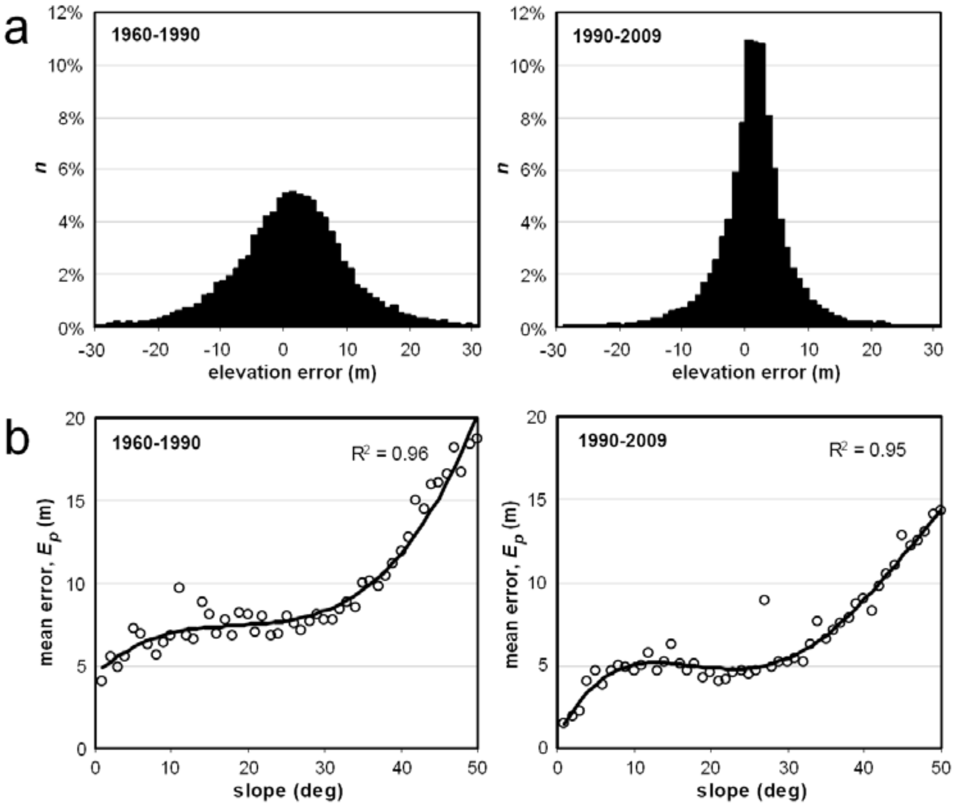

Fig. 2. Errors of geometry datasets used in this study: frequency distribution of elevation differences of co-registered datasets (**a**), note the histograms are inferred from terrain surrounding the glacier, and average elevation error against slope for rock terrain and glacier ice (**b**). For snow-covered surfaces the error is assumed to be twice higher.

taining information about *dh* and slope in relation to 1990 DEM. Glacier ice is here assumed to have similar error characteristics as rocks surrounding the glacier, while errors for snow-covered surfaces are doubled due to low radiometric contrast on aerial images (Nuth *et al.* 2010). Because the glacier is dominated by poorly inclined slopes, conventional elevation error histograms obtained from mountainous terrain surrounding the glacier (Fig. 2a) only partially represent the uncertainties of available datasets. To investigate the actual errors on the glacier, the test-points were divided into groups of 1° inclination and in each cluster we calculated the average root mean square error, representing typical error at a given slope (E_p). E_p plotted against slope shows a clear increasing tendency, well described by third-order polynomial functions (Fig. 2b).

Using a slope histogram of the studied glacial system and the polynomial error functions, we could transform glacier slope values into error estimates for all raster cells lying inside of the glacier boundary. By integration, we obtained a *dh* error estimate representing all points on the glacier (E_g). The final error of our measurement of dH (E_{dH}) has been computed by dividing E_g by a square root of sample size (N).

$$E_{dH} = \frac{E_g}{\sqrt{N}} \qquad (1)$$

The spatial autocorrelation of errors in the available datasets was previously reported and we adopt its scale from the literature to be 1000 m, implying that N becomes the glacier area in km^2 (Nuth *et al.* 2007; James *et al.* 2012). Volume estimate error is calculated by multiplying E_{dH} by the area of the glacier in a studied period.

Results

Area and length changes since the surge maximum. — Little is known about the geometry of Hørbyebreen in the maximum phase of the last surge. The moraine system suggests that the glacier was 10.7 km long and its area was 24.7 km^2. Through all analysed periods Hørbyebreen has been showing a continuous decay, with retreat and area loss rates accelerating with time, and reached peak values after 1990 (Fig. 3). The most recent mean retreat rate of 74 m a^{-1} represents an increase of over three-fold from the 1960–1990 period and over twelve-fold from the earliest 1900–1960 period (Table 1). The reduction of glacier length is high as for a land terminating glacier in Svalbard and exceeded 3,000 m along some transects, with 2 442 m on average. Hence, the area loss has been occurring mostly in the front zone and its total is as high as 36% when compared to the maximum phase area.

Table 1

Geometry and mass balance of Hørbyebreen system in the analysed periods.

	1900–1960	1960–1990	1990–2009
Average (maximum) retreat rate (m a^{-1})	6.4 (6.7)	21.7 (26.3)	74.2 (112.6)
Area change rate (km^2 per decade)	-0.44	-0.98	-1.72
Total elevation change, dH (m)	N/D	-13.5 ± 2.3	-13.5 ± 1.4
Annual elevation change dH/dt (m a^{-1})	N/D	-0.45 ± 0.08	-0.71 ± 0.07
Annual geodetic balance B (m w.e. a^{-1})	N/D	-0.40 ± 0.07	-0.64 ± 0.07
Total volume change, dV (km^3)	N/D	-0.28 ± 0.05	-0.24 ± 0.03
Volume change rate, dV/dt (km^3 per decade)	N/D	-0.09 ± 0.02	-0.12 ± 0.01

Elevation, volume and mass balance changes in 1960–1990 and 1990–2009. — Figure 4 shows maps of the measured surface elevation change. Period 1960–1990 was characterised by a gradual increase of glacier slope, with very fast frontal thinning and slow mass build-up above ~500 m. The overall geodetic balance of Hørbyebreen system was however negative in this period with B = -0.40 ± 0.07 m w. eq. a^{-1} leading to thinning of the glacier by 13.5 ± 2.3 m on av-

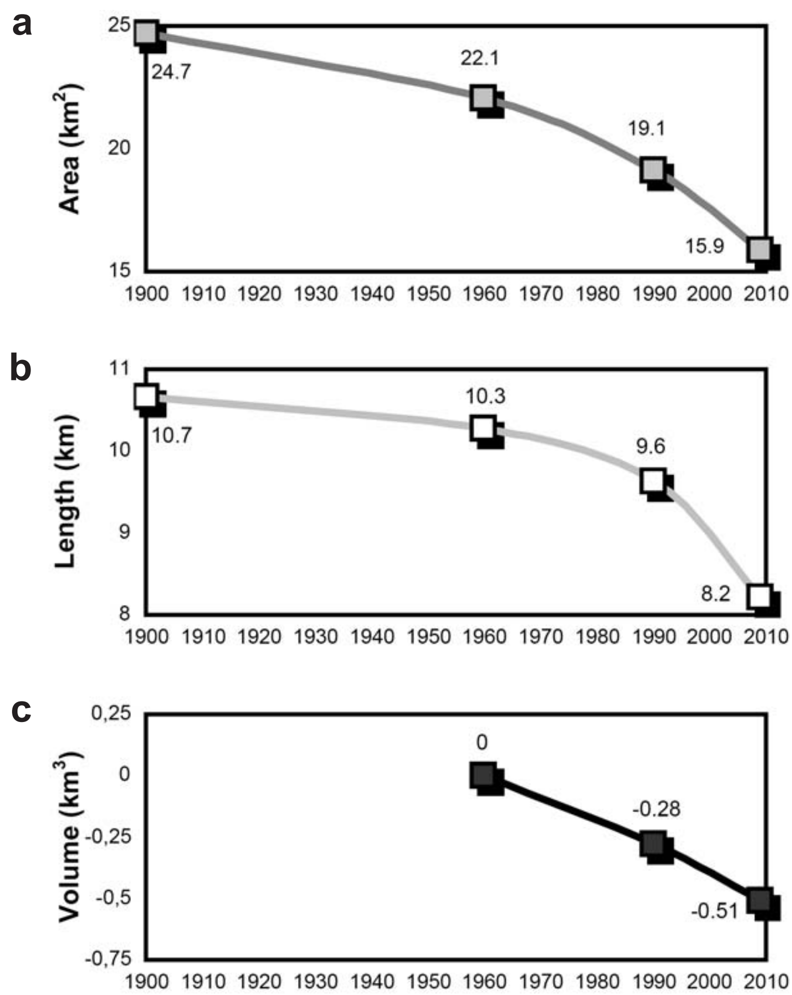

Fig. 3. Changes of Hørbyebreen glacial system: area changes (**a**), length changes (**b**), volume changes (**c**) in relation to year 1960.

erage and giving annual rate of thickness change (*dH/dt*) of -0.45 ± 0.08 m a^{-1}. The total volume loss observed was of 0.28 ± 0.05 km^3.

The elevation changes in the recent studied period were quantified by subtraction of 2009 DEM from 1990 DEM (Fig. 4b). The 2009 DEM is however not complete and elevation data is missing for the uppermost (<500 m a.s.l) zone of Hørbyebreen, accounting for 2.4 km^2 or 15.2 % of its total 2009 area. For this zone, a uniform elevation change observed at Hoelbreen above 500 m a.s.l. has been used to approximate the glacier-wide geodetic balance. In this period, the system has clearly experienced acceleration of thinning rate and has shown almost no build-up on Hoelbreen. We anticipate a similar pattern in the highest reaches of Hørbyebreen. Accounting for the missing data, the overall geodetic balance was

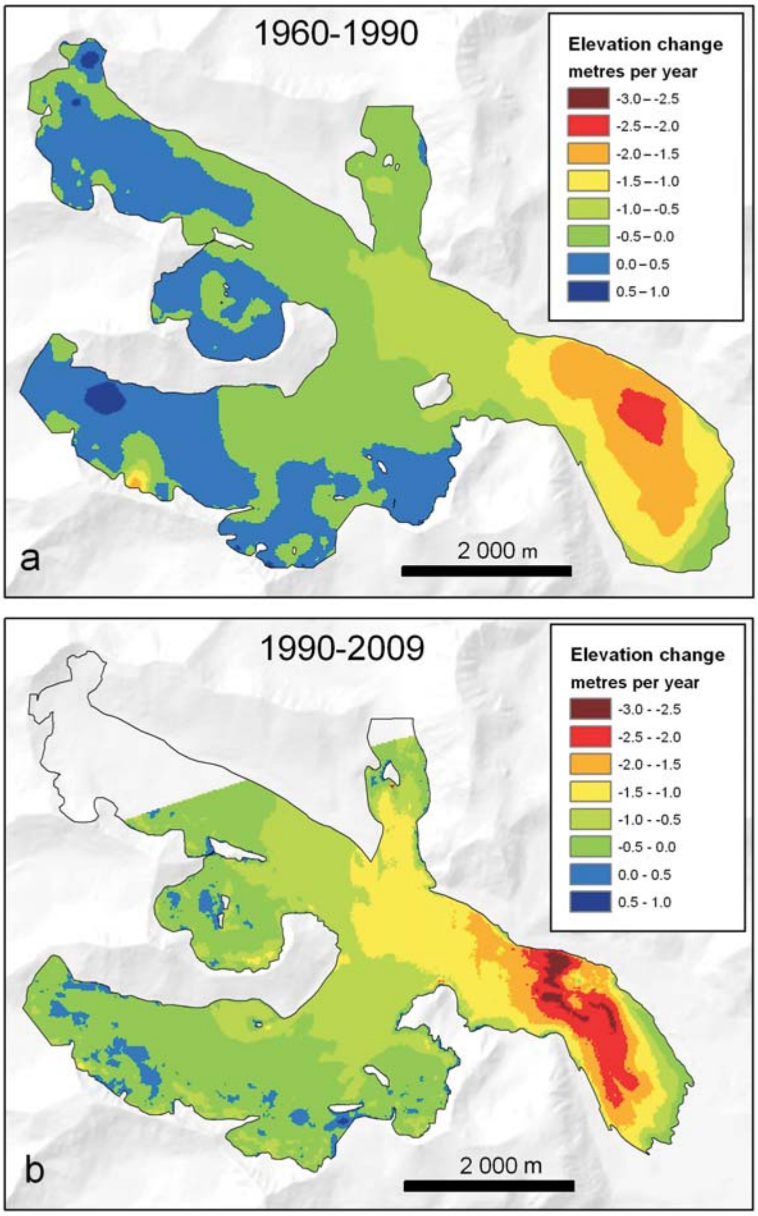

Fig. 4. Mean annual elevation change (*dh/dt*) of Hørbyebreen for the period 1960–1990 (**a**), with an up-build visible in the upper zones and the period 1990–2009 (**b**), with thinning over the entire glacier area.

more negative than in the earlier period, with B = -0.64 ± 0.07 m w.eq. a^{-1}, resulting in dH = -13.5 ± 1.4 m (-0.71 ± 0.07 m a^{-1}). Acceleration of surface lowering led to volume loss rate increase by 26% (-0.09 *vs.* -0.12 km^3 per decade, respectively for 1960–1990 and 1990–2009) and total volume loss in this period of dV = -0.24 ±

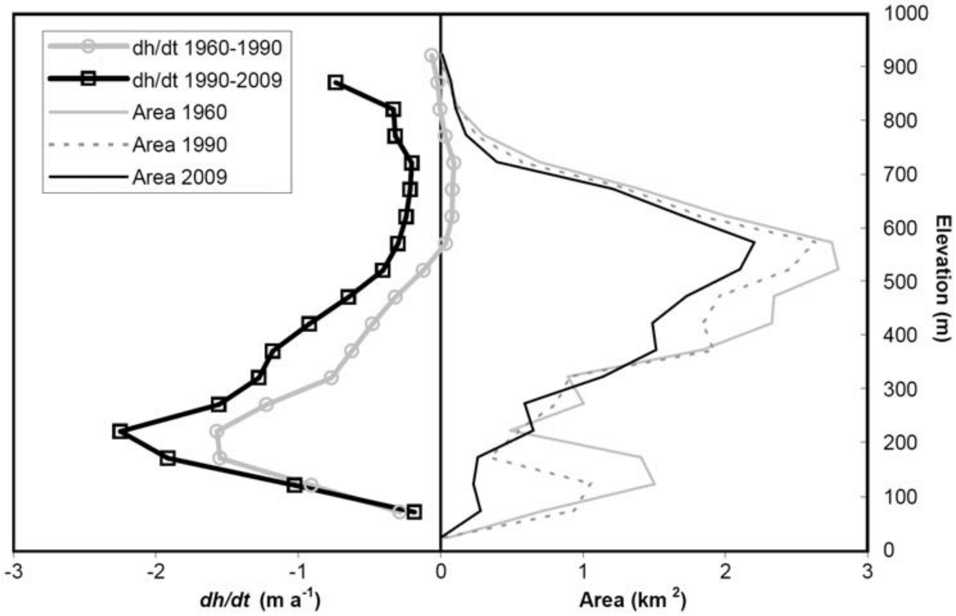

Fig. 5. Annual rates of elevation change, *dh/dt*, averaged along contour lines of Hørbyebreen in periods 1960–1990 and 1990–2009 against its area-altitude distribution (50 m elevation bands) in different years.

0.03 km³. Area-altitude distribution of the system has effectively evolved towards reduction of the surge-generated frontal zone (50–200 m a.s.l.), but negative changes have been occurring in all elevation bands (Fig. 5). Details of the geometry changes are listed in Table 1.

Thermal structure of Hørbyebreen system. — GPR soundings were carried out along centre-lines of the two components, starting from the highest zones at ~620 m a.s.l. to the front zone at the bedrock threshold (Fig. 1b). The results show a clear bedrock reflection all along the profiles and indicate maximum depth reaching 170 m. We interpret the thermal structure of the analysed glacier system as polythermal (Fig. 6). Higher zones of both ice streams are entirely cold and frozen to the bed despite their significant thickness. Central zones of the ice masses are warm-based, as interpreted from "blurred" bedrock reflection (see insets on Fig. 6). Such feature has been also noted on radargrams from nearby polythermal Bertilbreen (Małecki, unpublished results), where direct hydro-thermal investigations confirmed pressure melting-point conditions at the base (Gokhman *et al.* 1982; Zhuravlev *et al.* 1983). In the lowest zones of Hørbyebreen, temperate layer is up to 40 m thick (Fig. 6). Hoelbreen, considered as a more active component of the system, contains relatively small amount of temperate ice, close to the confluence zone with Hørbyebreen stream. In both cases, temperate ice occurs at the base of the thickest zones, beneath approximately 100–130 m depth.

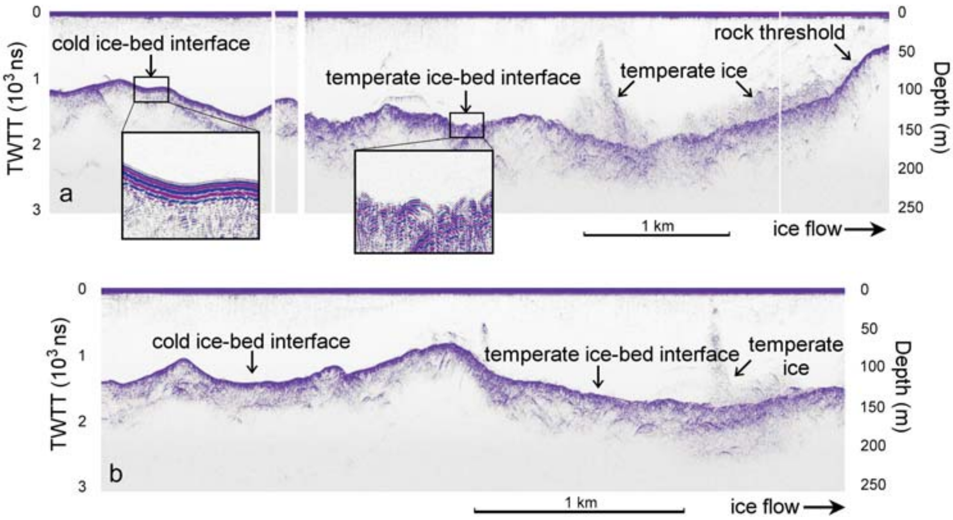

Fig. 6. Centre-line GPR profiles of Hørbyebreen (a) and Hoelbreen (b). The highest zones of glaciers are to the left. Vertical white stripes on Hørbyebreen profile indicate lack of data. TWTT – two-way travel time.

Discussion

The post-surge evolution of the glacier geometry seems to be dominated by decay of the flat, surge-generated front zone. Until year 1960, the front has been steadily retreating, with its further acceleration with time and thinning of the foot by up to 2–3 m annually. Even though the overall geodetic balance between 1960 and 1990 was negative (-0.40 m w. eq. a^{-1}), surface elevation analysis indicates increasing slope of the glacier surface by a slow mass build-up in the reservoir area above 500 m. Thickening rates were low, typically 0.1–0.2 m a^{-1}, and locally reached 0.8 m a^{-1} as the maximum. Such behaviour is typical for Svalbard surge-type glaciers in a quiescence phase and a similar pattern was found on *e.g.* Kongsvegen and Finsterwalderbreen (Melvold and Hagen 1998).

Slow build-up rates found in Svalbard are linked to low precipitation and accumulation rates in the archipelago, which prolong the duration of a surge cycle (Dowdeswell *et al.* 1991; Melvold and Hagen 1998). Duration of quiescent phases observed in Svalbard were reported to be 50 years for Tunabreen (Liestøl 1993), 70 years for Hambergbreen and Hessbreen (Liestøl 1993; Sund and Eiken 2004), 85 years for Bakaninbreen (Murray *et al.* 1998), 80–100 years for Usherbreen (Hagen 1987), 90–130 years for Bjuvbreen (Hamilton 1992), more than 100 years for Recherchebreen, Fridtjovbreen (Liestøl 1993) and Perseibreen (Dowdeswell and Benham 2003) and as much as 370–510 for Bråsvellbreen (Solheim 1991). Comparison of these long cycles to glaciers situated in more precipitous regions as

Medvezhiy Glacier in Pamir (9–14 years, Dolgoushin and Osipova 1975) and Variegated Glacier in Alaska (17–20 years, Kamb *et al.* 1985) clearly shows the role of cumulative mass gain in surge-cycle dynamics. Build-up rates observed on Hørbyebreen between 1960 and 1990 were very low even as for Svalbard, i.e. ~0.15 m a^{-1} *vs.* ~0.50 m a^{-1} on Kongsvegen or Finsterwalderbreen (Melvold and Hagen 1998). Because build-up rates and quiescence phase duration somewhat reflect the local climate, we expect Hørbyebreen to show full surge-cycle length of certainly more than 100 years, with a restriction that climatic conditions are favourable for mass gain in the reservoir area.

After 1990, summer air temperature in Svalbard has been rising even faster than in the previous decades. As calculated by James *et al.* (2012), the positive temperature trend since 1990 has been 0.07°C a^{-1}, being much higher than trend averaged for 1960–2009 (0.02°C a^{-1}). The potential increase of melt is not balanced by higher snow accumulation. Winter precipitation and accumulation on glaciers have shown no tendency or even a certain decrease recently (Førland and Hanssen-Bauer 2003; James *et al.* 2012). Hence, we explain the decrease of Hørbyebreen mass balance to -0.64 m w. eq. a^{-1} and significant thinning of the reservoir area observed after 1990 (Figs. 2 and 3) by climatic factors, similarly as increasing rates of retreat, area loss and volume loss (Fig. 4). Similar changes are observed also on other Svalbard glaciers (*e.g.* Kohler *et al.* 2007; Barrand *et al.* 2010; James *et al.* 2012) and will have further impacts, such as change in general energy balance conditions of the archipelago, reduction of high-albedo snow areas supporting further warming (*e.g.* Lemke *et al.* 2007) and contribution to the sea-level rise by freshwater production from melting glaciers (*e.g.* Meier *et al.* 2007).

As indicated by previous studies the low build-up rates and frequency reduction of surge events in Svalbard are linked to a climate change at the end of the LIA (Dowdeswell *et al.* 1995). It seems probable that the post-1990 anomalous summer temperature rise may cause further surge activity decrease in Svalbard. As we have shown, even relatively high elevated Hørbyebreen may not be able to build-up towards a new surge under present climatic conditions. Due to negative mass balance in its higher zones, the glacier can not reach enough ice thickness in the reservoir area, which seems to be crucial for triggering a thermally controlled surge, a model which is most likely valid for Svalbard glaciers (Fowler *et al.* 2001).

Svalbard ice masses smaller than ~3 km^2 have been mostly classified as entirely (or almost entirely) cold, *e.g.* Scott Turnerbreen (Hodgkins *et al.* 1999), Longyearbreen, Larsbreen (Etzelmüller *et al.* 2000), Tellbreen (Bælum and Benn 2011) or Ariebreen (Machío *et al.* 2007; Nawrot 2011). Larger valley glaciers contain also temperate basal ice *i.e.* Austre Brøggerbreen (Björnsson *et al.* 1996), Werenskioldbreen (Pälli *et al.* 2003), Bertilbreen (Gokhman *et al.* 1982) or Svenbreen (Małecki 2013). In the third group of glaciers, temperate ice at the bottom, as well as in their accumulation zones has been found *i.e.* Erikbreen (Ødegård *et al.* 1992), Midre Lovénbreen, Kongsvegen, Uvérsbreen (Björnsson *et al.* 1996) or Hansbreen (Jania *et al.* 1996; Pälli *et al.* 2003). Radar investigations performed on the Hørbyebreen surface confirm previous speculations that the glacier is polythermal. The structure of Hørbyebreen is interpreted as similar to the second group mentioned above – the glacier shows temperate ice along considerable length of its centrelines, beneath 100–130 m thick cold surface layer, with front zone frozen to the bed. No evidence of temperate firn has been found in its highest reaches, what may be explained by negative mass balance over the whole glacier in the last decades. Moreover, formation of thick superimposed ice on glaciers in this region of Spitsbergen makes their surface poorly permeable to meltwater, thus refreezing and consequent heating of ice by latent heat release, are vastly limited (Małecki 2013).

Dowdeswell *et al.* (1995) conclude that Svalbard surge-type glaciers may turn their polythermal structure to cold if they experience a prolonged lack of mass build-up, as they show for a small and shallow glacier Scott Turnerbreen. At least 5 decades of continuous thinning of the glacier system analysed in this paper were not sufficient to complete such a transformation, what seems to result from considerable ice thickness of Hørbyebreen (up to 170 m) and Hoelbreen (150 m). However, further intensive thinning will certainly modify and limit the extent of temperate ice, but it will most likely remain warm-based until it is less than ~100 m thick, as observed on other glaciers in the study area, *e.g.* on Bertilbreen (Zhuravlev *et al.* 1983) or Svenbreen (Małecki 2013).

Conclusions

From our study we draw the following conclusions:

- Hørbyebreen surged in the late 19th or early 20th century, creating an extensive flat front zone with looped medial moraines. In the period 1960–1990, it has been steepening its profile by fast down-wasting of its front zone (thinning up to 2–3 m a^{-1}) and slow build-up in the reservoir zone above 500 m (~0.15 m a^{-1}). Such behaviour is typical for Svalbard surge-type glaciers in a quiescent phase. The mass balance of the glacier was generally negative in this period (-0.40 m w. eq. a^{-1}), resulting in volume decrease and area loss.
- Considering high elevation of Hørbyebreen, the observed build-up rates in the period 1960–1990 were very low even as for Svalbard, reflecting arid quasi-continental climatic conditions in Dickson Land.
- After 1990 the glacier has shown a decrease in mass balance (to -0.64 m w. eq. a^{-1}, 1990–2009). Negative elevation changes in the reservoir area on the order of -0.25 m a^{-1} were also observed. We link this shift to post-1990 summer temperature rise, not accompanied by an increase in winter accumulation. We state that under present conditions Hørbyebreen is not able to build-up towards a new surge.
- Recent summer temperature anomalies may further reduce the surge activity of Svalbard glaciers, because intensified melting may prolong their quiescent

phases. Moreover, warmer summer periods cause an acceleration of rates of retreat, area loss and volume loss, giving a feedback to further warming by changing the general energy balance of the archipelago and contributing to sea-level rise by enhanced freshwater release.

- The studied glacial system is polythermal. Both Hørbyebreen and Hoelbreen are entirely cold in their upper zones, but warm-based downglacier from their mid-sections. Up to 40 m thick temperate ice has been found in their lower sections. Temperate core is situated in the deepest zones of both ice masses, approximately under 100–130 m thick cold ice layer. No evidence for temperate firn zone was found.
- Svalbard glaciers, which at present are not able to build-up towards a surge may still remain warm-based if they are thick enough. In case of Hørbyebreen (with maximum thickness of 170 m) and Hoelbreen (150 m), ~100 years since the last surge and at least five decades of continuously negative mass balance were not enough to transform into a cold-based glacier.

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