Quantification of historical landscape change on the foreland of a receding polythermal glacier, Hørbyebreen, Svalbard

Marek W. Ewertowski[1], David J.A. Evans[2], David H. Roberts[2], Aleksandra M. Tomczyk[1], Wojciech Ewertowski[1], Krzysztof Pleksot[1]

[1]*Faculty of Geographical and Geological Sciences, Adam Mickiewicz University, Poznań, Poland*
[2]*Department of Geography, Durham University, Durham, UK*

Abstract: The assessment of multi-decadal scale change in a polythermal glacial landsystem in the high-Arctic is facilitated by a quantitative approach that utilises time-series of aerial photographs, satellite images, digital elevation models and field geomorphological mapping. The resulting spatio-temporal analysis illustrates a transition from glacial to proglacial/paraglacial conditions indicating that: (1) the areal coverage of ice between the maximum LIA extent and 2013 decreased from 29.35 km^2 to 16.07 km^2, which is a reduction in the glacierized area in the catchment from 62% to 34%; (2) the ice volume loss in the proglacial area amounted to 214.9 (+-3%) millions m^3, which was attributed mostly to melting of the glacier snout but to a lesser extent the degradation of ice-cored landforms; (3) the transition from areas formerly covered by glacier ice to ice-cored moraines, glacifluvial deposits, and other landforms was the most intense in the period 1990-2013; (4) two end member scenarios (polythermal glacial landsystem domains) evolve during glacier recession, each one dictated by the volume of debris in englacial and supraglacial positions, and include subglacial surfaces (limited englacial and supraglacial debris) related to temperate basal ice, and ice-cored lateral moraines and moraine-mound complexes (significant supraglacial debris accumulations) related to marginal cold based ice. An additional assemblage of geometric ridge networks (discrete or linear englacial and supraglacial debris concentrations) relates to crevasse and hydrofracture infill branching out from an esker complex and is indicative of either surging or later rapid release of pressurised meltwater from temperate to cold based parts of the former glacier snout.

Keywords: Svalbard, glacial geomorphology, polythermal glacial landsystem, GIS, ice-cored moraine, Arctic, paraglacial

1. INTRODUCTION

Glacial forelands exposed due to glacier recession are among the most dynamic landscapes in polar and mountainous areas (e.g. Bennett and Evans, 2012; Bennett et al., 2010; Carrivick and Heckmann, 2017; Staines et al., 2014) and are intensively modified by various geomorphological processes related to glacial retreat, meltwater activity, and the paraglacial adjustment of topography (Ballantyne and Benn, 1994; Carrivick and Heckmann, 2017;

Kellerer-Pirklbauer et al., 2010; Kirkbride and Deline, 2018; Mercier et al., 2009; Rachlewicz, 2009; Rachlewicz, 2010). Widespread ice marginal recession in Svalbard since the end of the Little Ice Age (LIA) has resulted in the exposure of extensive glacier forelands containing a wide range of glacial landforms, many of which continue to evolve well after deglaciation due to their significant ice cores. The resulting landform-sediment associations are representative of the sub-polar glacial landsystem, in many places featuring surge signatures, and their presently dynamic state provides the ideal opportunity to study changes in glacial process-form regimes in response to changes in both climatic conditions and geomorphological processes. Such investigations into the direct connections between processes and forms in contemporary glacial landscapes allows researchers to compile modern analogues for contextualising palaeoglaciological reconstructions (cf. Benn and Lukas, 2006; Boulton, 1972; Evans, 2009). Such investigations also provide a basis for the prediction of future changes in the polar regions. However, most previous quantitative studies on Svalbard have focused on glaciology (e.g. Błaszczyk et al., 2013; Hagen and Liestøl, 1990; Hagen et al., 1993; Jania and Hagen, 1996; Lefauconnier and Hagen, 1991; Małecki, 2013; Małecki, 2014; Małecki, 2016; Moholdt et al., 2010a; Moholdt et al., 2010b; Nuth et al., 2013; Nuth et al., 2010; Rachlewicz et al., 2007; Sund et al., 2009; Ziaja, 2001; Ziaja, 2005) rather than the ongoing transformations of proglacial landscapes (e.g. Bennett et al., 2000; Bernard et al., 2016; Lønne and Lyså, 2005; Lukas et al., 2005; Lyså and Lønne, 2001; Midgley et al., 2013; Schomacker and Kjær, 2008; Sletten et al., 2001; Strzelecki et al., 2015; Zagorski, 2011; Zagorski et al., 2012; Ziaja, 2004; Ziaja and Pipała, 2007). Moreover, quantification of the dynamics of landform evolution on glacier forelands using time-series of digital elevation models or airborne laser scanning has to date been limited (e.g. Bernard et al., 2016; Etzelmüller, 2000a; Etzelmüller, 2000b; Irvine-Fynn et al., 2011; Kociuba, 2014; Kociuba, 2016; Kociuba et al., 2014; Midgley et al., 2018; Tonkin et al., 2016).

This study addresses landform transformations on a multi-decadal time-scale on the foreland of Hørbyebreen, a Svalbard polythermal glacier, in order to assess landscape changes and glacial landsystem evolution over the last 80 years. The main aim of this study is to characterize and quantify the deglacial and postglacial transition from glacial to proglacial (paraglacial) conditions and thereby evaluate spatial and temporal glacial landsystem evolution. The objectives identified to achieve this aim are: (1) to map and quantify landform development; and (2) to evaluate the changing patterns of geomorphological processes responsible for landscape change.

2. STUDY AREA

Svalbard is located in the High-Arctic, where continuous permafrost ranges in thickness from 100 m to 500 m (c.f. Etzelmüller and Hagen, 2005; Humlum et al., 2003). The study area is in the central part of Spitsbergen in the vicinity of Petuniabukta, an area characterised by less ice cover than most coastal zones due to lower precipitation totals. Around Petuniabukta, an area of 26 km^2 has been exposed since the termination of LIA as a result of glacier retreat (Ewertowski and Tomczyk, 2015). Earlier investigations into glaciers and glacial landforms near Petuniabukta provide a substantial background knowledge of both the geomorphology (e.g. Allaart et al., 2018; Evans et al., 2012; Ewertowski, 2014; Ewertowski et al., 2016; Ewertowski et al., 2012; Ewertowski and Tomczyk, 2015; Gibas et al., 2005; Gonera and Kasprzak, 1989; Hanáček et al., 2011; Karczewski, 1989; Karczewski et al., 1990; Karczewski and Kłysz, 1994; Kłysz, 1985; Křížek et al., 2017; Pleskot, 2015; Rachlewicz, 2009; Rachlewicz, 2010; Rachlewicz and Szczuciński, 2008; Stankowski et al., 1989; Strzelecki et al., 2018; Strzelecki et al., 2015; Szuman and Kasprzak, 2010), and glaciology (e.g. Małecki, 2013; Małecki, 2014; Małecki, 2016; Małecki et al., 2013; Rachlewicz, 2009; e.g. Rachlewicz et al., 2007), enabling us to link the results of this study to regional changes in the cryosphere.

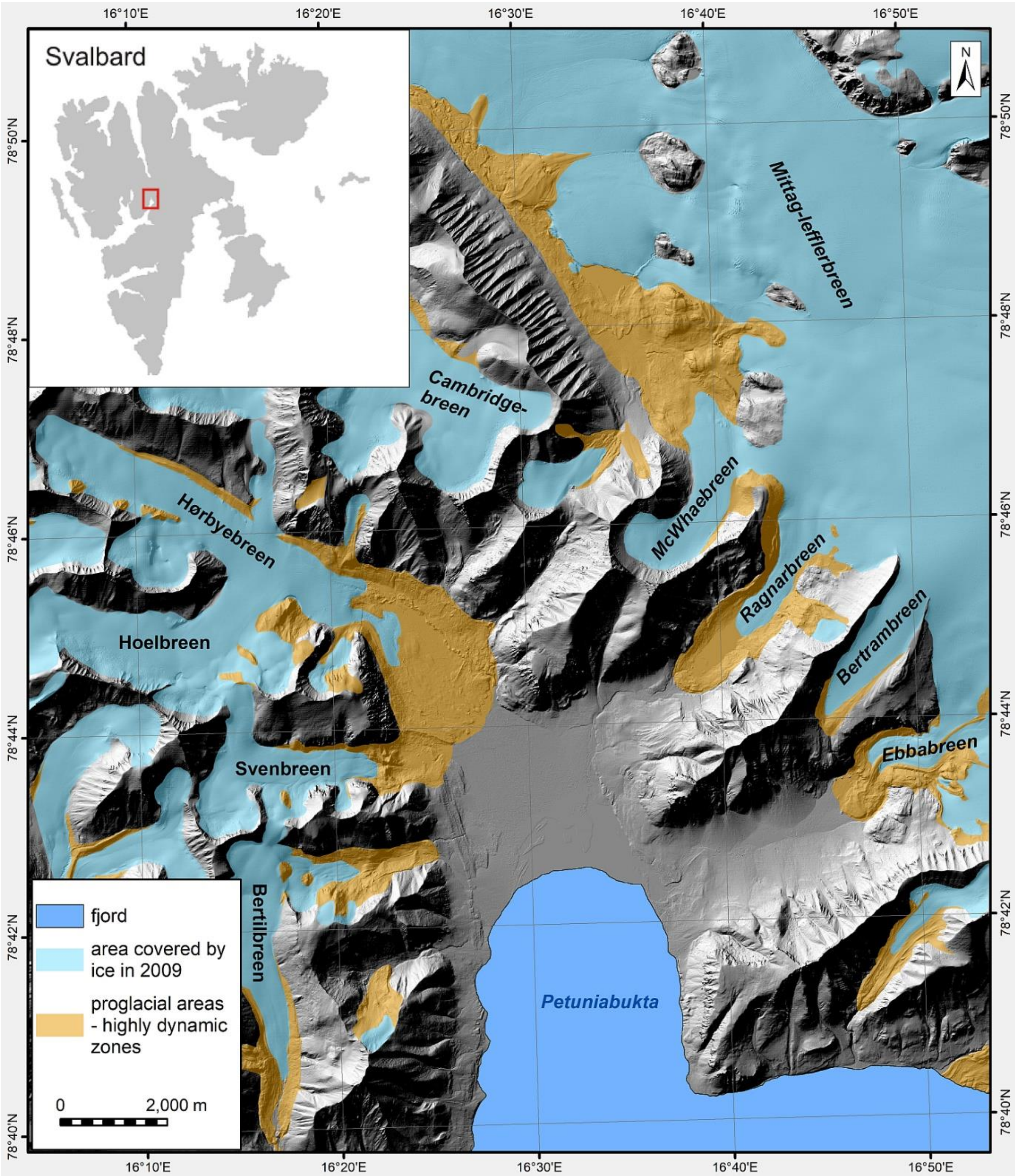

Figure 1. Location of the study area.

Hørbyebreen is a valley glacier located at the northern end of Petuniabukta (Figure 1). The glacier consists of two main flow units, the main unit being named Hørbyebreen and a tributary unit named Hoelbreen. Its snout is low-lying and surrounded by a large latero-frontal moraine ridge related to its maximum extent during the LIA. A recent study by Małecki et al. (2013) found that the glacier was up to 170 m thick and is polythermal, with a temperate 40 m thick basal layer beneath 100-130 m of polar ice (Małecki et al., 2013). Based on characteristic features such as a looped medial moraine, Hørbyebreen has been classified as a surge-type glacier (Farnsworth et al., 2016; Gibas et al., 2005; Karczewski, 1989; Małecki et al., 2013) but there have been no direct historical observations of any surges by Hørbyebreen. Hence, other interpretations of the glacier dynamics and associated landsystem signatures have suggested that Hørbyebreen may instead represent a system that dominated by either: 1) overprinted polythermal and surging behaviour; or 2) the build-up and rapid intermittent release of meltwater from the warm-based internal part of a polythermal snout (Evans et al., 2012).

3. METHODS

Landscape changes in front of Hørbyebreen were quantified using time-series of remote sensing data (aerial images, satellite images, digital elevation models [DEMs]) and field geomorphological mapping. Four sets of historical aerial photographs were obtained from the Norsk Polar Institute. In addition, two sets of satellite images were purchased from Digital Globe and Apollo mapping companies. In total, the remote sensing data consist of six sets:

1) Panchromatic oblique frame camera aerial photographs from 1936 and 1938
2) Panchromatic frame camera photographs from 1961 with a ground pixel size of 0.55 m
3) Panchromatic frame camera satellite image from Korona satellite from 1965 and 1966
4) False infrared colour frame camera photographs from 1990 with a ground pixel size of 0.7 m
5) Colour digital camera photographs from 2009 with a ground resolution of 0.4 m
6) Multispectral satellite image from Worldview-2 satellite from 2013 with a ground pixel size 0.5 for panchromatic and 2-m for multispectral bands.

Aerial photographs (sets 2, 3 and 4) contained the full camera information and were processed using photogrammetric software; DEMs and orthophotos were then generated from them. A total of 16 points, scattered around the glacier snout, were measured in 2012 and 2013 with Topcon dGPS and used as ground control points for retrospective georeferencing. An additional 9 points were used as independent control points to assess the quality of the generated DEMs. The root mean square error (RMSE) in 3D space was calculated from elevation differences between the DEMs and independent control points: 0.61 m for 1961, 1.25 m for 1990, and 0.95 m for 2009. Additionally, systematic and random vertical errors were investigated by comparing the 1961 and 1990 DEMs with the 2009 DEM at points where elevation was expected to be stable (i.e. non-glaciated, non-ice-cored and stable rock areas expected to undergo very little transformation). These errors were higher than those calculated from ground-surveyed checkpoints with RMSE of 3.18 for 1961 and 3.40 for 1990; however, their distribution was close to normal, suggesting that errors were random rather than systematic. Satellite images from 2013 were orthorectified using DEM and ground control points and then pansharpened.

Volumetric transformation of the snout area was calculated from DEMs of differences (DoDs), (cf. Wheaton et al., 2010). This involves the subtraction of DEMs from each other, enabling the assessment of spatial and temporal changes, and thus they encompass the two periods 1961-1990 and 1990-2009. Also important to note is that quantification of transformation of ice and landform volume changes was restricted only to the mapped part of the glacier snout and not its entire area. Planimetric transformations were investigated based on all available remote sensing and field-based data, hence they covered a longer period. Two glacial geomorphology maps and results of earlier investigations have previously been published (Evans et al., 2012; Gibas et al., 2005; Karczewski, 1989; Karczewski et al., 1990; Karczewski and Rygielski, 1989; Rachlewicz, 2009; Wojciechowski, 1989), as well as a study concerning changes in glacial geometry (Małecki et al., 2013) and dynamics (Rachlewicz, 2009). Therefore, we present new data (including additional maps), with the focus on changes in landforms within the snout area. Mapping was performed digitally (on-screen vectorisation) and verified during field campaigns. The projection used in the analysis was UTM zone 33N. All elevation values provided in the text and figures are ellipsoidal (not above sea level), unless otherwise indicated.

4. RESULTS AND INTERPRETATIONS – SPATIAL AND TEMPORAL TRANSFORMATIONS OF GLACIER SNOUT AND ICE-MARGINAL (PROGLACIAL) AREA

As a baseline for the results presented here, the 2009 DEM indicates that the elevation of the glacier ranged from 65 to 695 m a.s.l., and the area classified as the glacier foreland ranged from 24 to 111 m a.s.l., although the highest ridges of the lateral moraine reached 320 m a.s.l. The first part of this section provides a brief description of the extent of glacier changes, and this is followed by an overview of the volumetric and planimetric changes in landforms and adjacent glacier snout. We then provide a description of the spatial and temporal pattern of glacial landsystem transformation.

4.1. Glacier changes LIA - 2013

4.1.1. Glacier extent changes 1900(1920) – 2013

The maximum extent of Hørbyebreen has been delimitated using its ice-cored latero-frontal moraine complex (Evans et al. 2012). It is assumed that the distal slope edges of the moraine indicate the approximate maximum position of the ice margin. There are no direct data relating to the timing of this maximum expansion of the snout and therefore we assume that the recession started around 1900 based upon several studies documenting at least some Svalbard glaciers having already retreated from their maximum position during the period 1900-1920 (cf. Lamplugh, 1911; Slater, 1925). The extent of the glacier in 1961, 1990, 2009, and 2013 was delimitated based on remote sensing data and field verification.

The total decrease in the extent of Hørbyebreen for the period 1900 – 2013 was about 13.3 km^2 or almost 50% of the maximum extent (Table 1, Figure 2). The presented values indicate the change in only the clean glacier cover (i.e. ice surface without debris); there is still a large amount of ice buried under the debris of the moraine complex (Evans et al. 2012). Shrinkage in the ice cover led to the detachment of three smaller ice units in valleys to the north and one to the south, which in the past were linked to the main trunk of the snout (Figure 2). Areas provided in Table 1 take into account the total amount of exposed ice cover, including the main bodies of Hørbyebreen and Hoelbreen as well as the smaller units now detached and occupying higher elevation cirque basins. Since the LIA, the total percentage of the catchment occupied by ice decreased from 62% to 34%, which demonstrates the importance of switching from glacial to paraglacial geomorphological processes over a very large area.

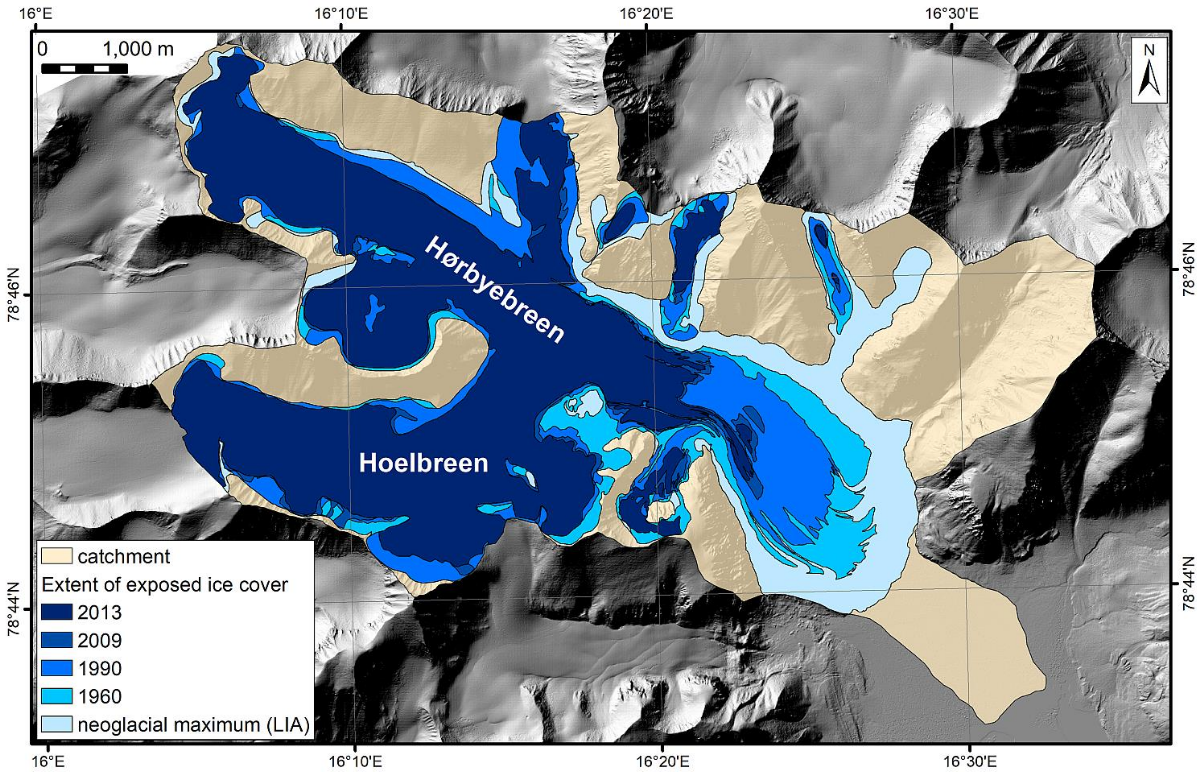

Figure 2. Changes in exposed ice cover from LIA – 2013, Hørbyebreen, Svalbard.

The decrease in ice volume was particularly visible in the recession of the glacier snout (Figure 2), which also varied between the Hoelbreen and Hørbyebreen ice flow units. The distance of Hørbyebreen unit snout retreat from the maximum position to the 2013 glacier margin was 3400 m, representing a mean annual retreat rate of 30 m/year, whereas the Hoelbreen unit snout retreated 2000 m or at around 18 m/year. Recent retreat rates (after 1990) were much faster.

Table. 1. Changes in area of exposed ice cover from LIA – 2013, Hørbyebreen, Svalbard.

year	ice cover [km^2]	Ice cover change [%] LIA = 100%	Ice cover percentage of the catchment area (47.61 km^2) [%]
LIA	29.35	100	62
1960	24.69	84	52
1990	21.67	74	46
2009	16.83	57	35
2013	16.07	55	34

4.1.2. Volume and elevation changes over the glacier snout (1961-2009)

DEM of Differences for the foreland of Hørbyebreen during the periods 1960-1990, 1990-2009 as well for whole studied period (1960-2009) have shown that widespread lowering is mostly related to the melting of the exposed ice surface. Only small areas indicated positive elevation changes. The total net volume of change over the 1961-2009 period from the glacier snout and foreland was about -215 million m^3 (+/- 3%). Annual percentage changes in snout area and volume vary over the study period (Table 2). Spatial variability was different for the two main ice flow units, as indicated in the patterns of elevation changes (Figure 3). The total volume loss from that part of the latero-frontal which was already exposed in 1960 equalled to 1.8 million m^3 (+/- 5%) for the period 1960-2009, which translates to a mean volume loss of 0.15 m/year. This varied significantly, however, from 0 m/year to almost 1.3 m/year depending on local topography and the impact of deglacial/paraglacial processes; i.e. most of the transformation of latero-frontal ice-cored moraine can be attributed to debris flows and backwasting of exposed ice cores (Fig 4 a-c).

Data shown in Table 2 and Figure 3 take into account the propagated errors for subsequent DEMs. The percentage of the Area of Interest with detectable changes varied from 64% to 90%, while errors for volume loss were much smaller (from 3 to 8%) than for deposition (from 62 to 70%) (Table 2). Much larger uncertainty in the measurements of volumes of deposition is related to the fact that the increase in surface elevation was usually small (within the range of DEM errors).

Table 2. Volume changes over the foreland and snout area of Hørbyebreen

Period	Area of loss (m^2)	Area of deposition (m^2)	% of Area of Interest with Detectable Changes (%)	Total volume loss (m^3)	Total volume of deposition (m^3)	Total net volume differences (m^3)	Average thickness of volume loss (m)	Average thickness of deposition (m)
1960 - 1990	4 317 648	17 402	89%	152 906 451 (+/- 5%)	48 610 (+/- 67%)	-152 857 841 (+/- 5%)	35	2
1990 – 2009	2 936 321	163 102	64%	62 203 486 (+/- 8%)	407 124 (+/- 70%)	-61 796 363 (+/- 8%)	21	1
Total period (1960 – 2009)	4 310 904	51 572	90%	215 116 548) +/- 3%)	123 832 (+/- 62%)	-214 992 716 (+/- 3%)	50	2

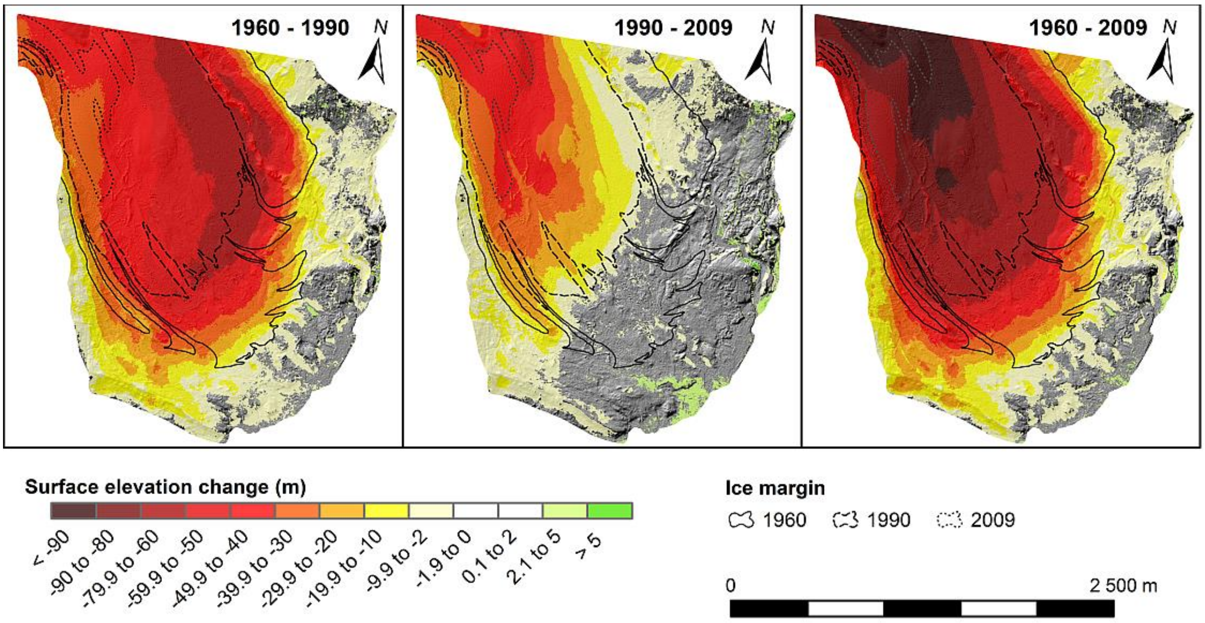

Figure 3. Surface elevation change (m) over the snout and proglacial area of Hørbyebreen.

Between 1960 and 1990, a larger amount of ice was lost from the Hørbye ice flow unit (NE side) compared to a much smaller loss from the Hoel ice flow unit (SW side). Spatial variability of ice loss resulted in an increasing asymmetric transverse snout profile over the study period (Figure 5). Specifically, the NE side of the glacier snout lowered much more rapidly than the SW side. This might relate to various factors such as surge activity and/or high mountain walls that effectively reduce the amount of solar energy reaching the southern part of the glacier. Conversely, in some place heat radiation from dark rocks resulted in separation of clean ice surface from valley sides (Fig. 4d). Spatial variability in glacier snout changes such as this can also be interpreted as the product of an uneven distribution of supraglacial debris. On the Hoelbreen ice flow unit in particular this is manifest in the occurrence of a thickening debris cover that is characterised by prominent longitudinal debris stripes comprising a relatively thin (< 10 cm) layer of angular debris (Fig. 4e). The presence of such a thin layer of debris on the glacier surface can reduce the ice melt significantly over time (Fig. 4f) (cf. Nakawo and Young, 1981; Nicholson and Benn, 2006; Nicholson and Benn, 2013; Östrem, 1959).

Figure 4. (a) Minor exposure of ice-core in the southern lateral moraine resulting from the development of small debris flow. Note the relatively thin cover of debris protecting moraine from melting and indicating that this segment of the moraine did not undergo significant transformations in the past; (b) Cohesive debris flow developed on gentle slopes; (c) Major exposure of ice-cores in the northern lateral moraine. Note that debris cover is about 1.5-2 m thick (person for scale) indicating that this part of the moraine was repetitively transformed by mass movement processes. (d) Retreat of the exposed ice surface from the valley side due to the emission of heatwave radiation from dark surface of rocks. Note that such separation seriously limits delivery of fresh debris from valley sides onto glacier surface; (e) Longitudinal debris stripe – thin and relatively sparse cover of angular debris, which can be traced on the glacier surface, as well as in the deglaciated foreland overlain on other deposits; (f) Section through the ice covered by longitudinal debris stripe. Note that even such thin (< 0.1 m) layer of debris is sufficient to limit ice melting – as a result of differential ablation clean ice surface visible in the background significantly lowered.

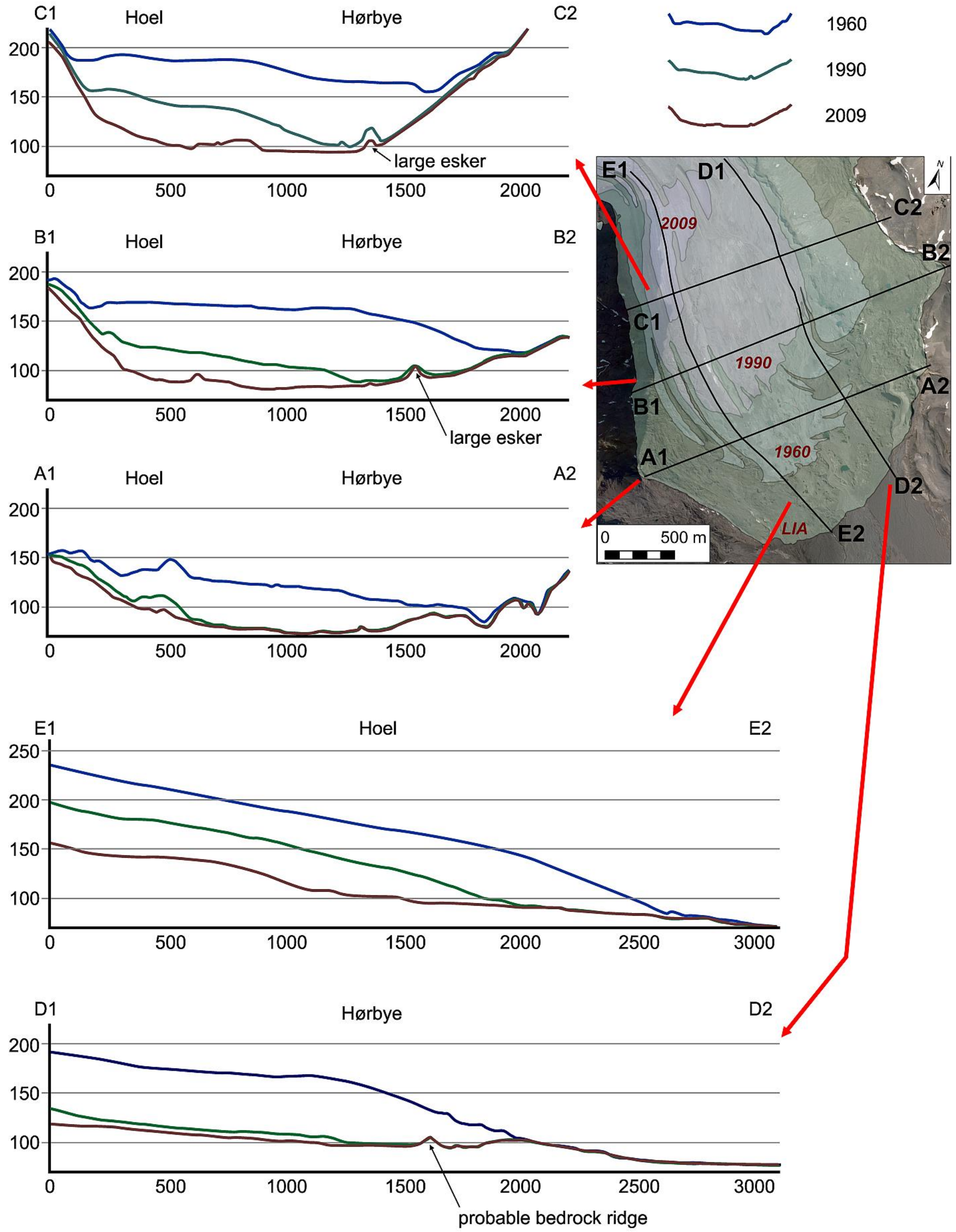

Figure 5. Decrease in surface elevation between 1960 and 2009. Longitudinal profiles demonstrate differences in surface lowering of the Hoel and Hørbye ice flow units, whereas the transverse profiles show the development of the asymmetric profile of the glacier surface.

The glacier surface long profiles lowered significantly during the period 1961-1990, by as much as 100 m for the Hørbye ice flow unit. In 1961, the glacier surface showed a characteristic steep ice front with a mean slope of 11% at the snout contrasting with a flatter up-ice section of 5% (Figure 5). In 1990, the ice profile of the Hørbye ice flow unit lowered significantly, indicating a much larger downwasting than frontal recession (volumetric change rather than aerial change). Between 1990 and 2009, the reduction in elevation of the glacier surface for the Hørbye ice flow unit was much smaller (10 m) than in the previous period, as most of the ice had melted and the glacier bed was exposed. In contrast, the Hoel ice flow unit lowered in a more uniform rate of 50 m in the period 1961-1990 and then another 50 m between 1990 and 2009, but maintaining more or less the same profile shape with a mean slope of 10%.

4.2. Proglacial (ice-marginal) landform changes LIA - 2013

The area covered by clean glacial ice, ice-cored moraines, till plains, glacifluvial deposits, rivers and lakes were quantified for each available period for the proximal part of the glacial system (i.e. between the LIA and the 2013 glacier margins; Fig. 6). The area covered by different surficial units was used as a proxy for identifying the predominant geomorphological processes operating over the foreland during the post-LIA period (Fig. 7).

In 1938, the ice margin was sited just behind the LIA moraine and hence the exposed glacier surface was the most dominant surficial unit (Fig. 7). The whole snout was surrounded by latero-frontal moraines and drainage was not well developed, with the main stream flowing from the NE part of the ice margin. Supraglacial debris in the form of looped stripes was visible over the snout (Fig. 6a). There have been no direct historical observations of the glacier surge that might have been responsible for these looped moraines (cf. Meier and Post, 1969), but their occurrence potentially indicates that the Hoel ice flow unit surged, or at least advanced relatively fast, and displaced the Hørbye ice flow unit towards the NE side of the valley.

In 1960, due to ice melting and decrease in the elevation of the glacier surface, some of the supraglacial debris stripes and medial moraines became more visible (Fig. 6b). An increase in coverage of the unit related to the latero-frontal moraines is also apparent (Fig. 7). Drainage migrated slightly, with the main stream flowing from the northern part of the ice margin breaking through the latero-frontal moraine to produce two narrow corridors.

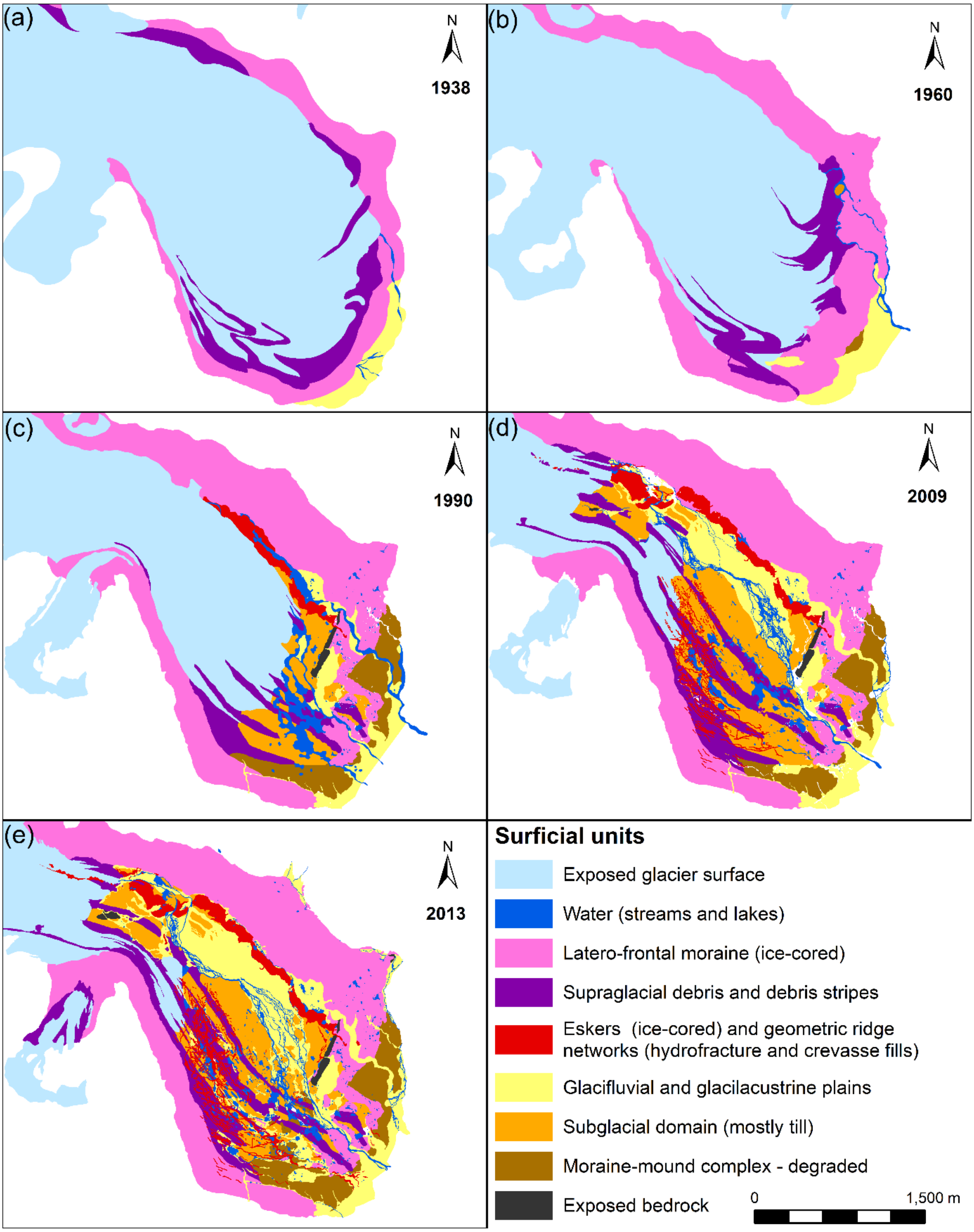

Figure 6. Spatio-temporal changes in dominant surficial units and hence process-form regimes in the snout area and emerging foreland of Hørbyebreen, Svalbard.

In 1990, a further retreat in the glacier margin and lowering of the ice surface resulted in exposure of the glacier bed (mostly till) which was visible between the 1990 ice margin and the latero-frontal moraines (Fig. 6c). Most parts of these areas were, however, flooded by meltwater which was partially blocked by the moraines, creating ponds in front of the ice. Part of the drainage continued along the NE margin using the previously formed corridor, while other meltwater pathways developed a new breach in the central section of the latero-frontal moraine. Part of the latter was probably de-iced over time, as suggested by characteristic landscapes such as kettle-holes and collapsed terrain, and results of previous geophysical investigations (cf. Gibas et al., 2005). In the 1990 imagery a large esker ridge became visible along the NE margin for the first time. The esker was still ice cored and located above the level of the river and outwash, indicating that the esker system was a trace of the previous englacial drainage system.

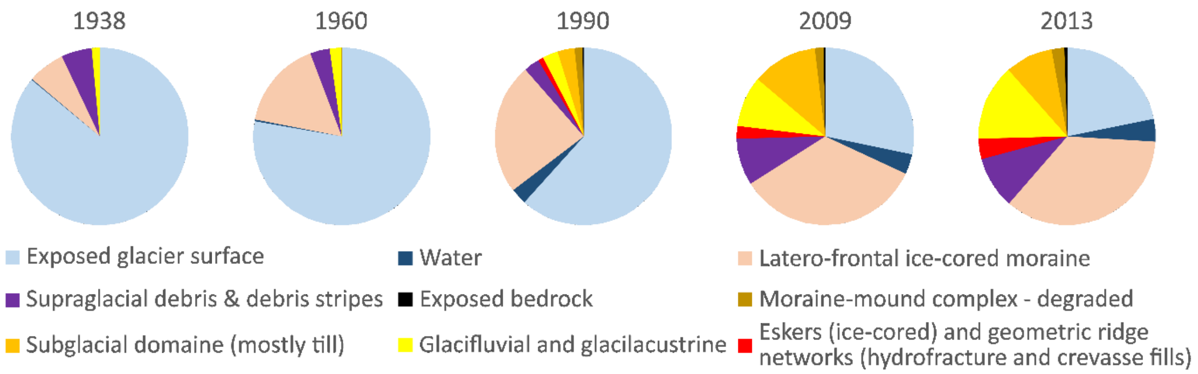

Figure 7. Percentage of different surficial units and hence process-form regimes in the snout and emerging foreland area of Hørbyebreen over time.

In 2009, the ice margin was between 1 and 2 km from its maximum extent at the LIA, and large-scale reorganization of the foreland had taken place (Fig. 6d). The main meltwater portal was still located at the NE part of the margin, but the stream braided through the central part of the foreland. It had two consequences: (1) the NE corridor between the esker and later moraines was abandoned; (2) most of the meltwater went through the central part of the foreland, destroying almost the entire subglacial signature in this area and leaving only small patches of subglacial till plain elevated several meters above the outwash level (Fig. 8a, b). Degradation of the Hoel ice flow unit had a different character, with ice mostly downwasting and developing prominent supraglacial debris stripes overprinted on englacial and subglacial geometric ridge networks (crevasse squeeze ridges, hydrofracture infills and small eskers - Fig. 8c) visible on the SW part of the foreland. Small ponds and streams were also common on this part of the foreland.

In 2013, ice occupied less than 25% of the foreland (Figure 7), while the ice margin was up to 3 km from its maximum LIA position (Figure 6e). Further migration of the drainage had taken place and had destroyed further subglacial deposits related to the Hørbye ice flow unit. Ice marginal retreat also resulted in the exposure of further segments of the ice-cored esker. On that part of the foreland occupied by the Hoel ice flow unit, more extensive geometric ridge networks were exposed by downwasting ice, together with an increase in the number and size of ponds. Supraglacial debris stripes were still visible and it was possible to trace them in some places as far as the LIA moraines.

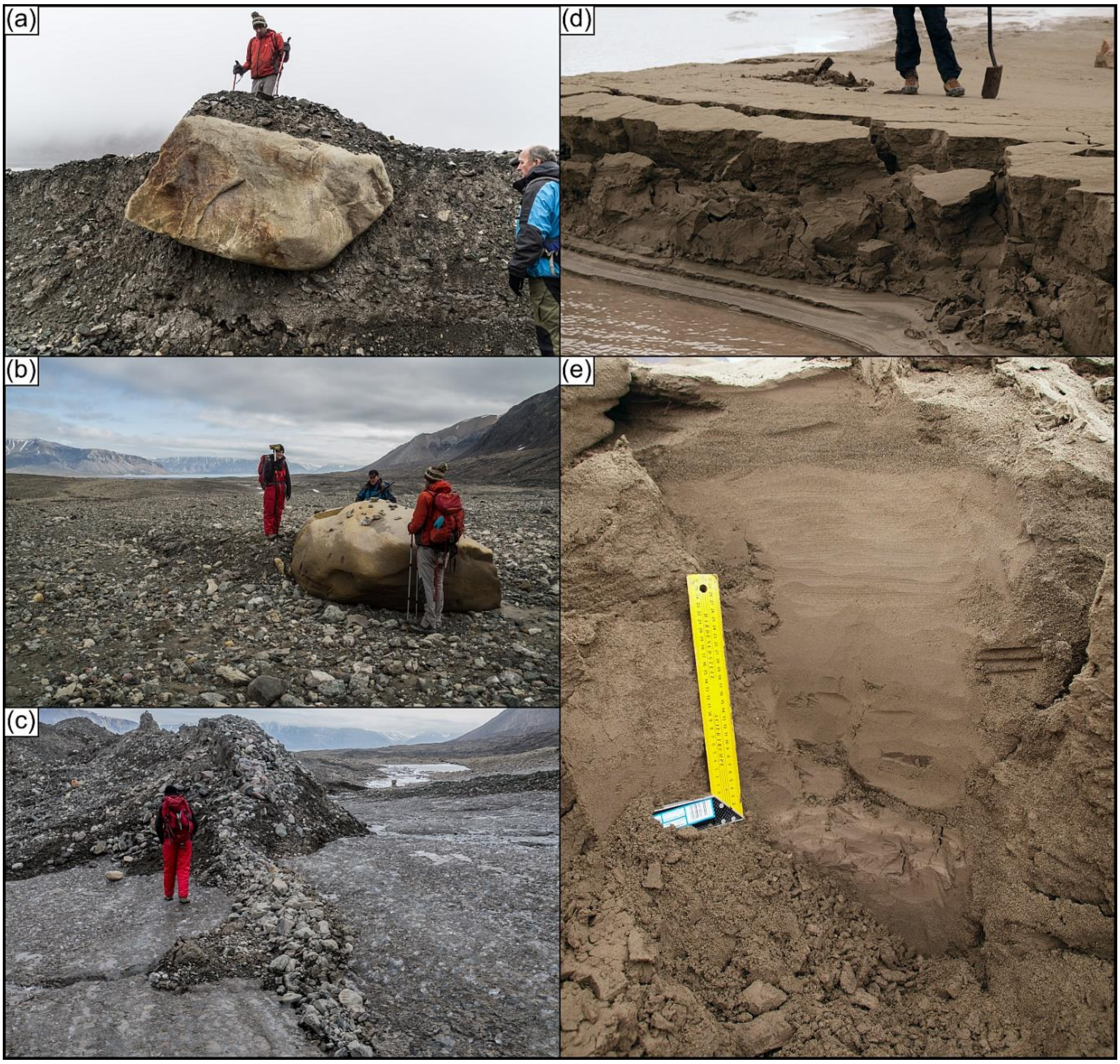

Figure 8. (a) Migration of drainage towards the central part of the foreland lead to the destruction of large areas of subglacial deposits. Only small isolated, elevated fragments of subglacial traction tills remained in the northern part of foreland; (b) Larger areas of subglacial deposits temporarily survived in the central part of the foreland but will be probably removed by meltwater in the future. Note small flute hooked behind the boulder. Former ice flow was from lower-right towards middle-left of the picture; (c) Concentration of debris associated with former englacial channel emerging as the ice surface retreated and downwasted; Note that the ridge visible in the picture is still ice-cored; (d) Fine-grained deposits visible after drainage of a short-lived pond in the southern part of the foreland; (e) section through glacilacustrine deposits accumulated in short-lived pond, which developed after 1990 and drained before 2009.

The time-series of charts depicting change in surficial unit coverage over time, for an area of approximately 10 km^2 (Fig. 7), clearly indicate a decrease in importance of process-form regimes directly related to the ice, along with an increase in the areas covered by glacifluvial deposits, water bodies and ice-cored moraines (Fig. 7). The area covered by the exposed glacier ice diminished from 86% in 1938 to about 20% in 2013. As a consequence of glacier retreat and the evolution of the foreland, areas covered by latero-frontal ice-cored moraines increased steadily from about 7% in 1938 to more than 35% in 2003. In the period since 1990, most of the meltwater has flowed through the foreland and the associated glacifluvial processes have removed a large area of subglacial deposits which have decreased in area from 12% in 2009 to less than 9% in 2013, concurrent with an increase in glacifluvial deposits to almost 14% in 2013.

4.3. Evolution of the drainage network and pro-glacial lakes, 1936 – 2013

Proglacial streams in front of Hørbyebreen varied over time in terms of location and character (Fig. 6). For most years, the main meltwater portal at the ice margin was located at the NE side of the glacier; however, some minor meltwater evacuation portals were also observed in other sections of the margin.

Since 1936, the supraglacial debris cover near the frontal margin of the glacier snout has been increasingly dissected by small englacial and supraglacial channels. As a consequence, a series of small esker-like landforms emerged from the SW margin, indicating the former location of the drainage system to be potentially related to the Hoelbreen ice flow unit. In 1960, minor streams were no longer visible, and only the largest stream flowed from the NE side of the Hørbye ice flow unit. This river was still active in 1990 but it migrated laterally, following the retreat of lateral clean ice margin, and an additional portal was visible where water ponded in the central part of the foreland between the LIA moraines and 1990 margin. Also visible in 1990 was drainage channelled through the LIA moraines with three ravines being occupied by active streams, as well as several smaller abandoned corridors. In 2009, the main meltwater portal was still visible in the northern part of the foreland (more than 3 km from the maximum ice extent); the main river braided into numerous channels through the foreland, whereas the northern corridor was abandoned. At the same time, several minor streams flowed from the Hoel ice flow unit, flowing through a geometric ridge network and thereby creating numerous shallow ponds in which sedimentation rates were high (Fig. 8d) – based on field observations thickness of the lacustrine deposits in small ponds formed after 1990 and drained before 2009 was about 0.8 m (Fig. 8e), which equalled to minimum average rate of approximately 2.8 cm/year. In 2013 some further modifications to the drainage were visible but the main meltwater portal was still located in the north and braiding its way through the foreland, with minor streams and numerous pond to the south.

The clearly developed esker networks reflect a difference in the drainage networks between the Hørbye and Hoel ice flow units. The Hørbye ice flow unit was drained by a single, large channel, which led to the development of a large, ice-cored esker along the N and NE side of the glacier. The Hoel ice flow unit was associated with numerous smaller eskers and geometric ridge networks potentially representing hydrofracture and crevasse infills (Evans et al. 2012).

5. DISCUSSION

5.1. Glacier changes

The historical recession of Hørbyebreen has been far greater than other land-terminating glaciers located in the vicinity of Petuniabukta (Fig. 1) (Małecki, 2016; Rachlewicz et al., 2007), with values of 30 and 20 m/year for the Hørbye and Hoel ice flow units respectively (Fig. 4). The increasing retreat and thinning of the glacier after the LIA indicates that even if the glacier did surge in the past, it is likely not now capable of accumulating enough mass to initiate further surges under current climatic conditions (Małecki et al., 2013). This accelerated retreat has generated a large amount of meltwater, which in turn significantly modified the glacier foreland over time (see Section 4.3 and Fig. 6). The difference in response between the Hørbye and Hoel ice flow units is also clearly marked (Fig. 4). If retreat continues in a similar rate to the present (i.e. 20 – 30 m/year), these two ice flow units will become separate around the year 2080.

5.2. Spatial and temporal changes in the dominant geomorphological processes

The dynamics of the proglacial area of Hørbyebreen illustrate decay in the high-Arctic, polythermal (had probably surged in the past) glacial landsystem. Using different types of surficial units as a proxy for the dominant geomorphological processes, five main process-form regimes were recognized: glacial-related, glacifluvial and glaciolacustrine, downwasting, mass wasting processes, and stabilization. Spatio-temporal analysis over the snout area (approximately 10 km^2) indicates a serious decrease in processes directly related to the glacier, i.e. from more than 90% in 1938 to less than 25% in 2013 (Fig. 9). This glacial-related regime was transformed mostly into one of mass-wasting processes as a result of the development of large ice-cored lateral moraines. Currently, direct glacial processes have a relatively low impact on landscape dynamics in proglacial areas. Rather, most of the transformations are related to:

1) Mass wasting of lateral moraines, where large debris flows develop and lead to repetitive transformation of landform and sediments (Fig. 4a-c);
2) Downwasting of dead-ice buried under supraglacial debris, leading to the emergence of landforms related to the former englacial drainage/crevasse patterns (Fig. 8c);
3) Glacifluvial erosion and deposition which effectively removes traces of other processes and leads to the development of relatively flat, inner outwash plain (Fig. 8a);
4) Localised glaciolacustrine deposition, leading to rapid accumulation of fine-grained sediments (cf. Wojciechowski, 1989) (Fig. 8d, e).

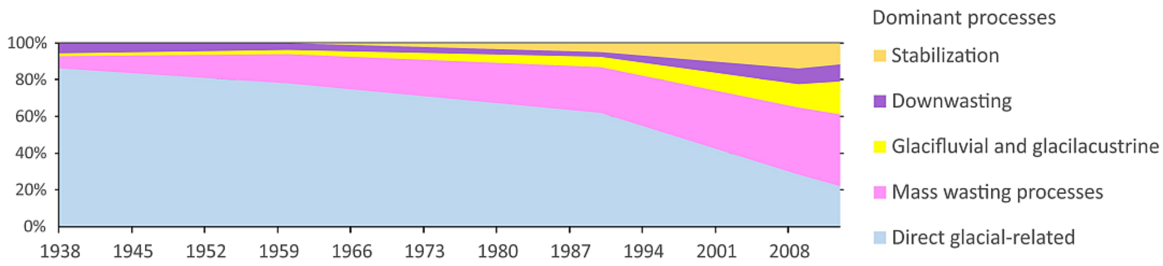

Figure 9. Changes in dominant processes in the snout and emerging foreland areas of Hørbyebreen.

In terms of elevation changes, the average lowering of ice-cored lateral moraines in the period 1960-2009 varied from 0 to 1.5 m/year depending on the local situation. Such values are in the same order of magnitude as those of other Svalbard glaciers (Ewertowski, 2014; Ewertowski and Tomczyk, 2015; Irvine-Fynn et al., 2011; Midgley et al., 2018; Schomacker and Kjær, 2008; Tonkin et al., 2016). The remaining part of the proglacial area (approximately 10% in 2013) was stable over this period and in equilibrium with current climatic and environmental conditions. The relatively small size of the stable area compared to other glaciers in the vicinity (Ewertowski, 2014; Ewertowski et al., 2016; Ewertowski and Tomczyk, 2015) is related to the fact that the drainage of Hørbyebreen has migrated over time and is now impacting upon the whole central part of the foreland, destroying large areas of potentially stable geomorphology.

5.3. Deglaciation scenarios

Glacial retreat has resulted in the emergence and paraglacial transformations of a complex landform assemblage related to a polythermal glacier snout. Based on field observations and the analysis of remote-sensing data, two main domains are identified in the proglacial area of Hørbyebreen (Figure 10):

- **Inner domain related to the temperate ice zone -** In situations where there is a limited amount of englacial or supraglacial debris, the final landscape component in the foreland will be a subglacial till plain (Fig. 8b). Such till plains are relatively stable, as they do not contain dead-ice or steep topography. However, in the case of Hørbyebreen, meltwater has destroyed large areas of till plain relatively quickly (Fig. 8a), leading to the development of low-lying outwash plains filled with sands and gravels.
- **Outer domain related to cold-based snout -** Debris concentrations on the ice surface, mostly related to medial moraines, controlled moraine or lateral moraines, lead to the development of pronounced ice-cored moraine or moraine-mound complexes. Debris thickness is initially thin (< 1 m) but enough nevertheless to protect ice from melting (Fig. 4a) (cf. Nicholson and Benn, 2006; Nicholson and Benn, 2013). Further degradation by debris flows is accelerated by thermal erosion by meltwater and repetitive backwasting processes increase the thickness of the debris cover (Fig. 4c). Ice-cored moraines are the most active in the period shortly following deglaciation; after this initial topographical re-adjustment, thickness of the debris cover is sufficient

to almost completely prevent melting of the ice-cores. Further switching between stable conditions and active degradation depends mostly on local factors (e.g. meltwater, intensive rainfall, over-steepening of topography due to erosion). When stabilised, such ice-cored landforms can be stable component of the landscape (e.g. some fragments of latero-frontal moraine of Hørbyebreen; Fig. 4) over several decades. After final degradation and de-icing, this depositional scenario results in hummocky topography with chaotic hills and ponds.

This basic scheme was complicated by complex interactions between flow units; and potential release of pressurised meltwater from temperate parts of the glacier blocked by frozen snout, which resulted in increasing presence of englacial debris. When debris is concentrated in various structures within the ice (englacial debris) (Fig. 8c) or as relatively thin longitudinal debris stripes on the surface of the glacier (Fig. 4e, f), the resultant geomorphology is a complex pattern of cross-cutting ridges visible in the SW part of the foreland. Such cross-cutting landforms do not necessarily have a good preservation potential because they contain at least a certain amount of dead-ice that is partly protected from melting by debris cover. A cross-cutting planform of ridges favours water ponding, which in turn creates sedimentation traps, collecting fine-grained glacilacustrine material (Fig. 8d, e). Depending on their preservation potential the final product is usually a landscape that contains low-relief linear hummocks and small intervening ponds.

5.4 Glacial landsystem

Our spatial and temporal analysis of landscape transformations at Hørbyebreen reflects the downwasting and recession of a high-Arctic polythermal glacier and hence the evolution of the landsystem signature that is diagnostic of such glaciers. Inherent within this landsystem evolution is an important switch from direct glacial processes to indirect (paraglacial and proglacial) processes, the latter being conditioned primarily by ice-core degradation and the increasing complexity of meltwater drainage networks (Figure 6). Various landform-sediment assemblages can be identified within the landsystem signature (cf. Evans et al., 2012) and can be summarized as outer and inner domains typical of the Svalbard polythermal glacial landsystem (Glasser and Hambrey, 2003). The outer domain relates to the frozen or cold based snout and comprises the arcuate latero-frontal ice-cored moraine and moraine-mound complexes, the latter containing variable amounts of buried ice and relating to the formation of controlled moraine (*sensu* Evans, 2009). The inner domain comprises streamlined (fluted) tills created in the temperate ice zone that previously lay up-ice from the frozen snout. This simple zonation is complicated by the interaction of different glacier flow units, especially the potential surge of the Hoel ice flow unit to form deformed and looped medial moraines, traces of which are still visible in the landscape. The surge may also have been influential in the creation of large ice-cored moraines, hummocky terrain and crevasse squeeze ridges (cf. Evans and Rea, 2003; Evans et al., 2012; Lovell et al., 2018; Lovell et al., 2015; Roberts et al., 2009), the latter emerging within a dense geometric ridge network that appears to be centred on a zone of linear esker ridges. This landform assemblage comprises small eskers, crevasse-squeeze ridges and hydrofracture fills and might more realistically relate to the rapid release of pressurized meltwater from the temperate ice zone of the glacier snout and its injection into the cold based snout ice rather than surge-induced crevassing. A further azonal domain relates to the extensive meltwater reworking of landforms and sediments to produce glacifluvial outwash plains. Finally, the very localized development of glacilacustrine deposits, especially as they accumulate over buried glacier ice, is unlikely to result in a significant azonal domain.

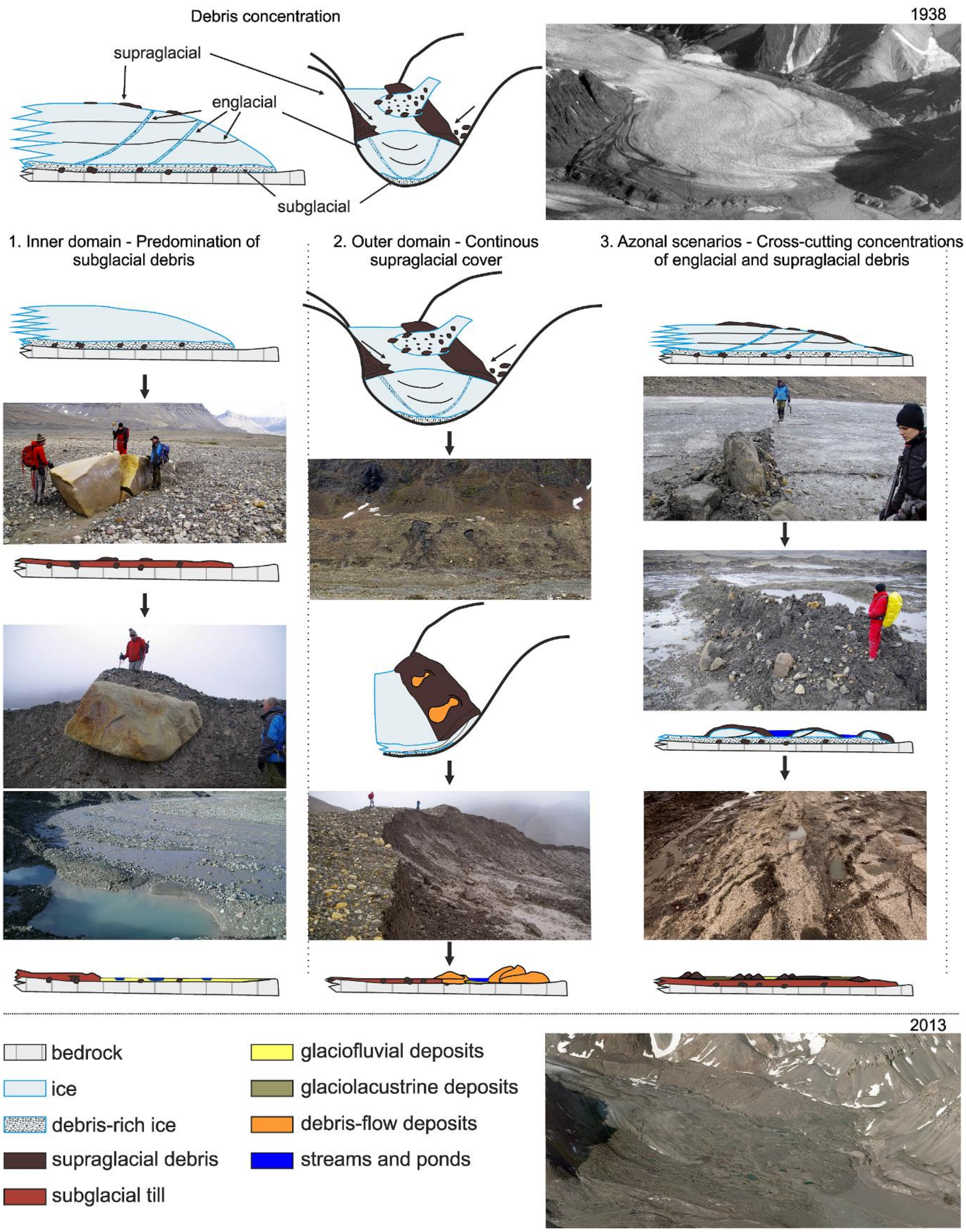

Figure 8. Different scenarios for paraglacial transformation of a complex landform assemblage related to a polythermal glacier snout. Domain zonation and associated debris concentration relates to former distribution of cold-based and temperate ice, and interactions between different flow units. See further explanations in the text.

6. CONLUSIONS

This study adds data on quantification of multi-decadal scale changes in proglacial landscapes in the high-Arctic polythermal landsystem by delivering information on planform and volumetric changes in the foreland of Hørbyebreen, Svalbard. Spatio-temporal analysis of the foreland of downwasting and retreating Hørbyebreen indicated that:

1) Area covered by exposed glacial ice between the maximum LIA extent and 2013 decreased from 29.35 km^2 to 16.07 km^2, which is a reduction in the glacierized area in the catchment from 62% to 34%;
2) Total volume loss in proglacial area was 214.9 (+-3%) millions m^3. This was attributed mostly to melting of the glacier snout. Degradation of ice-cored landforms was ten orders of magnitude smaller than of glacier snout.
3) The transition from areas formerly covered by glacier ice to ice-cored moraines, glacifluvial deposits, and other landforms was the most intense in the period 1990-2013.
4) Former glacier thermal regime and associated volume of debris in englacial and supraglacial position resulted in the development of two main domains:
    a. Subglacial surfaces related to the inner, temperate ice zone characterised by former limited englacial and supraglacial debris;
    b. Outer complex of moraine-mounds and arcuate latero-frontal ice-cored moraine related to the former cold-based snout and large concentration of supraglacial debris.
5) Azonal landform assemblages included:
    a. Lopped moraines associated with complex interactions between different flow units, including potential surge of Hoel flow unit;
    b. Complex geometric ridge network in SW part of the foreland indicating former surging behaviour or sudden release of pressurised meltwater from temperate part of the glacier blocked by frozen snout;
    c. Glacifluvial outwash plains deposited due to lateral drainage migration from external part of the landsystem to the central part of the foreland.

Further investigations into the preservation potential of different landform assemblages are necessary to fully quantify the impact of future climatic and environmental changes in the Arctic areas.

References

Allaart, L., Friis, N., Ingólfsson, Ó., Håkansson, L., Noormets, R., Farnsworth, W.R., Mertes, J., Schomacker, A., 2018. Drumlins in the Nordenskiöldbreen forefield, Svalbard. Gff, 1-19.
Ballantyne, C.K., Benn, D.I., 1994. Paraglacial slope adjustment and resedimentation following recent glacier retreat, Fåbergstølsdalen, Norway. Arctic Alpine Res, 255-269.
Benn, D.I., Lukas, S., 2006. Younger Dryas glacial landsystems in North West Scotland: an assessment of modern analogues and palaeoclimatic implications. Quaternary Sci Rev, 25(17-18), 2390-2408.

Bennett, G.L., Evans, D.J.A., 2012. Glacier retreat and landform production on an overdeepened glacier foreland: the debris-charged glacial landsystem at Kvíárjökull, Iceland. Earth Surf Proc Land, 37(15), 1584-1602.

Bennett, G.L., Evans, D.J.A., Carbonneau, P., Twigg, D.R., 2010. Evolution of a debris-charged glacier landsystem, Kviarjokull, Iceland. J Maps, 40-67.

Bennett, M.R., Huddart, D., Glasser, N.F., Hambrey, M.J., 2000. Resedimentation of debris on an ice-cored lateral moraine in the high-Arctic (Kongsvegen, Svalbard). Geomorphology, 35(1-2), 21-40.

Bernard, É., Friedt, J., Tolle, F., Marlin, C., Griselin, M., 2016. Using a small COTS UAV to quantify moraine dynamics induced by climate shift in Arctic environments. Int J Remote Sens, 1-15.

Błaszczyk, M., Jania, J.A., Kolondra, L., 2013. Fluctuations of tidewater glaciers in Hornsund Fjord (Southern Svalbard) since the beginning of the 20th century. Pol Polar Res, 34(4), 327-352.

Boulton, G.S., 1972. Modern Arctic glaciers as depositional models for former ice sheets. J Geol Soc London, 128(4), 361-393.

Carrivick, J.L., Heckmann, T., 2017. Short-term geomorphological evolution of proglacial systems. Geomorphology, 287, 3-28.

Etzelmüller, B., 2000a. On the Quantification of Surface Changes using Grid-based Digital Elevation Models (DEMs). Transactions in GIS, 4(2), 129-143.

Etzelmüller, B., 2000b. Quantification of thermo-erosion in pro-glacial areas - examples from Svalbard. Z Geomorphol, 44(3), 343-361.

Etzelmüller, B., Hagen, J.O., 2005. Glacier-permafrost interaction in Arctic and alpine mountain environments with examples from southern Norway and Svalbard. Geological Society, London, Special Publications, 242(1), 11-27.

Evans, D.J.A., 2009. Controlled moraines: origins, characteristics and palaeoglaciological implications. Quaternary Sci Rev, 28(3-4), 183-208.

Evans, D.J.A., Rea, B.R., 2003. Surging glacier landsystem. In: D.J.A. Evans (Ed.), Glacial Landsystems. Arnold, London, pp. 259-288.

Evans, D.J.A., Strzelecki, M.C., Milledge, D.G., Orton, C., 2012. Hørbyebreen polythermal glacial landsystem, Svalbard. J Maps, 8(2), 146-156.

Ewertowski, M.W., 2014. Recent transformations in the high-Arctic glacier landsystem, Ragnarbreen, Svalbard. Geografiska Annaler: Series A, Physical Geography, 96(3), 265-285.

Ewertowski, M.W., Evans, D.J.A., Roberts, D.H., Tomczyk, A.M., 2016. Glacial geomorphology of the terrestrial margins of the tidewater glacier, Nordenskiöldbreen, Svalbard. J Maps, 12(sup1), 476-487.

Ewertowski, M.W., Kasprzak, L., Szuman, I., Tomczyk, A.M., 2012. Controlled, ice-cored moraines: sediments and geomorphology. An example from Ragnarbreen, Svalbard. Z Geomorphol, 56(1), 53-74.

Ewertowski, M.W., Tomczyk, A.M., 2015. Quantification of the ice-cored moraines' short-term dynamics in the high-Arctic glaciers Ebbabreen and Ragnarbreen, Petuniabukta, Svalbard. Geomorphology, 234, 211-227.

Farnsworth, W.R., Ingólfsson, Ó., Retelle, M., Schomacker, A., 2016. Over 400 previously undocumented Svalbard surge-type glaciers identified. Geomorphology, 264, 52-60.

Gibas, J., Rachlewicz, G., Szczuciński, W., 2005. Application of DC resistivity soundings and geomorphological surveys in studies of modern Arctic glacier marginal zones, Petuniabukta, Spitsbergen. Pol Polar Res, 26(4), 239-258.

Glasser, N.F., Hambrey, M.J., 2003. Ice-marginal terrestrial landsystems: Svalbard polythermal glaciers. In: D.J.A. Evans (Ed.), Glacial Landsystems Arnold, London, pp. 65-88.

Gonera, P., Kasprzak, L., 1989. The main stages of development of glacier margin morphology in the region between Billefjorden and Austfjorden, central Spitsbergen. Pol Polar Res, 10(3), 419-427.

Hagen, J., Liestøl, O., 1990. Long-term glacier mass-balance investigations in Svalbard, 1950–88. Ann Glaciol, 14, 102-106.
Hagen, J., Liestøl, O., Roland, E., Jørgensen, T., 1993. Glacier atlas of Svalbard and Jan Mayen, Norwegian Polar Institute, Oslo.
Hanáček, M., Flašar, J., Nývlt, D., 2011. Sedimentary petrological characteristics of lateral and frontal moraine and proglacial glaciofluvial sediments of Bertilbreen, Central Svalbard. Czech Polar Reports, 1(1), 11-33.
Humlum, O., Instanes, A., Sollid, J.L., 2003. Permafrost in Svalbard: a review of research history, climatic background and engineering challenges. Polar Res, 22(2), 191-215.
Irvine-Fynn, T.D.L., Barrand, N.E., Porter, P.R., Hodson, A.J., Murray, T., 2011. Recent High-Arctic glacial sediment redistribution: A process perspective using airborne lidar. Geomorphology, 125(1), 27-39.
Jania, J., Hagen, J.O., 1996. Mass balance of Arctic glaciers. 5, University of Silesia, Faculty of Earth Sciences, Sosnowiec-Oslo.
Karczewski, A., 1989. The development of the marginal zone of the Hørbyebreen, Petuniabukta, central Spitsbergen. Pol Polar Res, 10(3), 371-377.
Karczewski, A., Borówka, M., Kasprzak, L., Kłysz, P., Kostrzewski, A., Lindner, L., Marks, L., Rygielski, W., Wojciechowska, A., Wysokiński, L., 1990. Geomorphology – Petuniabukta, Billefjorden, Spitsbergen. Adam Mickiewicz University, Poznań.
Karczewski, A., Kłysz, P., 1994. Lithofacies and structural analysis of crevasse filling deposits of the Svenbreen foreland (Petuniabukta, Spitsbergen), XXI Polar Symposium, Warszawa, pp. 123-133.
Karczewski, A., Rygielski, W., 1989. The profile of glacial deposits in the Hörbyedalen and an attempt at their chronostratigraphy central Spitsbergen. Pol Polar Res, 10(3), 401-409.
Kellerer-Pirklbauer, A., Proske, H., Strasser, V., 2010. Paraglacial slope adjustment since the end of the Last Glacial Maximum and its long-lasting effects on secondary mass wasting processes: Hauser Kaibling, Austria. Geomorphology, 120(1), 65-76.
Kirkbride, M.P., Deline, P., 2018. Spatial heterogeneity in the paraglacial response to post-Little Ice Age deglaciation of four headwater cirques in the Western Alps. Land Degrad Dev.
Kłysz, P., 1985. Glacial forms and deposits of Ebba Glacier and its foreland (Petuniabukta region, Spitsbergen). Pol Polar Res, 6, 283-299.
Kociuba, W., 2014. Application of Terrestrial Laser Scanning in the assessment of the role of small debris flow in river sediment supply in the cold climate environment. Annales UMCS, Geographia, Geologia, Mineralogia et Petrographia, 69(1), 79-91.
Kociuba, W., 2016. Assessment of sediment sources throughout the proglacial area of a small Arctic catchment based on high-resolution digital elevation models. Geomorphology.
Kociuba, W., Kubisz, W., Zagórski, P., 2014. Use of terrestrial laser scanning (TLS) for monitoring and modelling of geomorphic processes and phenomena at a small and medium spatial scale in Polar environment (Scott River—Spitsbergen). Geomorphology, 212, 84-96.
Křížek, M., Krbcová, K., Mida, P., Hanáček, M., 2017. Micromorphological changes as an indicator of the transition from glacial to glaciofluvial quartz grains: Evidence from Svalbard. Sediment Geol, 358(Supplement C), 35-43.
Lamplugh, C., 1911. On the shelly moraine of the Sefström Glacier and other Spitsbergen phenomena illustrative of British glacial conditions. P Yorks Geol Soc, 17(3), 216-241.
Lefauconnier, B., Hagen, J.O., 1991. Surging and calving glaciers in eastern Svalbard. Oslo.
Lønne, I., Lyså, A., 2005. Deglaciation dynamics following the Little Ice Age on Svalbard: Implications for shaping of landscapes at high latitudes. Geomorphology, 72(1-4), 300-319.
Lovell, H., Benn, D.I., Lukas, S., Spagnolo, M., Cook, S.J., Swift, D.A., Clark, C.D., Yde, J.C., Watts, T., 2018. Geomorphological investigation of multiphase glacitectonic composite ridge systems in Svalbard. Geomorphology, 300, 176-188.

Lovell, H., Fleming, E.J., Benn, D.I., Hubbard, B., Lukas, S., Rea, B.R., Noormets, R., Flink, A.E., 2015. Debris entrainment and landform genesis during tidewater glacier surges. J Geophys Res-Earth, 120(8), 1574-1595.

Lukas, S., Nicholson, L.I., Ross, F.H., Humlum, O., 2005. Formation, Meltout Processes and Landscape Alteration of High-Arctic Ice-Cored Moraines—Examples From Nordenskiold Land, Central Spitsbergen. Polar Geography, 29(3), 157-187.

Lyså, A., Lønne, I., 2001. Moraine development at a small High-Arctic valley glacier: Rieperbreen, Svalbard. J Quaternary Sci, 16(6), 519-529.

Małecki, J., 2013. Elevation and volume changes of seven Dickson Land glaciers, Svalbard, 1960–1990–2009. Polar Res, 32, 18400.

Małecki, J., 2014. Some comments on the flow velocity and thinning of Svenbreen, Dickson Land, Svalbard. Czech Polar Report, 4(1), 1-8.

Małecki, J., 2016. Accelerating retreat and high-elevation thinning of glaciers in central Spitsbergen. The Cryosphere, 10(3), 1317-1329.

Małecki, J., Faucherre, S., Strzelecki, M.C., 2013. Post−surge geometry and thermal structure of Hørbyebreen, central Spitsbergen. Pol Polar Res, 34(3), 305-321.

Meier, M.F., Post, A., 1969. What are glacier surges? Canadian Journal of Earth Sciences, 6(4), 807-817.

Mercier, D., Etienne, S., Sellier, D., Andre, M.F., 2009. Paraglacial Bullying of sediment-mantled slopes: a case study of Colletthogda, Kongsfjorden area, West Spitsbergen (Svalbard). Earth Surf Proc Land, 34(13), 1772-1789.

Midgley, N.G., Cook, S.J., Graham, D.J., Tonkin, T.N., 2013. Origin, evolution and dynamic context of a Neoglacial lateral-frontal moraine at Austre Lovénbreen, Svalbard. Geomorphology, 198, 96-106.

Midgley, N.G., Tonkin, T.N., Graham, D.J., Cook, S.J., 2018. Evolution of high-Arctic glacial landforms during deglaciation. Geomorphology, 311, 63-75.

Moholdt, G., Hagen, J.O., Eiken, T., Schuler, T.V., 2010a. Geometric changes and mass balance of the Austfonna ice cap, Svalbard. The Cryosphere, 4(1), 21-34.

Moholdt, G., Nuth, C., Hagen, J.O., Kohler, J., 2010b. Recent elevation changes of Svalbard glaciers derived from ICESat laser altimetry. Remote Sens Environ, 114(11), 2756-2767.

Nakawo, M., Young, G.J., 1981. Field experiments to determine the effect of a debris layer on ablation of glacier ice. Ann Glaciol, 2, 85-91.

Nicholson, L., Benn, D.I., 2006. Calculating ice melt beneath a debris layer using meteorological data. J Glaciol, 52(178), 463-470.

Nicholson, L., Benn, D.I., 2013. Properties of natural supraglacial debris in relation to modelling sub-debris ice ablation. Earth Surf Proc Land, 38(5), 490-501.

Nuth, C., Kohler, J., Konig, M., von Deschwanden, A., Hagen, J.O., Kaab, A., Moholdt, G., Pettersson, R., 2013. Decadal changes from a multi-temporal glacier inventory of Svalbard. Cryosphere, 7(5), 1603-1621.

Nuth, C., Moholdt, G., Kohler, J., Hagen, J.O., Kääb, A., 2010. Svalbard glacier elevation changes and contribution to sea level rise. Journal of Geophysical Research: Earth Surface, 115(F1), F01008.

Östrem, G., 1959. Ice melting under a thin layer of moraine, and the existence of ice cores in moraine ridges. Geografiska Annaler, 41(4), 228-230.

Pleskot, K., 2015. Sedimentological characteristics of debris flow deposits within ice−cored moraine of Ebbabreen, central Spitsbergen. Pol Polar Res, 36(2), 125-144.

Rachlewicz, G., 2009. Contemporary Sediment Fluxes and Relief Changes in High Artic Glacierized Valley Systems (Billefjorden, Central Spitsbergen). Adam Mickiewicz University, Poznań.

Rachlewicz, G., 2010. Paraglacial modifications of glacial sediments over millennial to decadal time-scales in the high Arctic (Billefjorden, central Spitsbergen, Svalbard). Quaestiones Geographicae, 29(3), 59-67.

Rachlewicz, G., Szczuciński, W., 2008. Changes in thermal structure of permafrost active layer in a dry polar climate, Petuniabukta, Svalbard. Pol Polar Res, 29(3), 261-278.

Rachlewicz, G., Szczuciński, W., Ewertowski, M.W., 2007. Post−"Little Ice Age" retreat rates of glaciers around Billefjorden in central Spitsbergen, Svalbard. Pol Polar Res, 28(3), 159-186.

Roberts, D.H., Yde, J.C., Knudsen, N.T., Long, A.J., Lloyd, J.M., 2009. Ice marginal dynamics during surge activity, Kuannersuit Glacier, Disko Island, West Greenland. Quaternary Sci Rev, 28(3-4), 209-222.

Schomacker, A., Kjær, K.H., 2008. Quantification of dead-ice melting in ice-cored moraines at the high-Arctic glacier Holmströmbreen, Svalbard. Boreas, 37(2), 211-225.

Slater, G., 1925. Observations on the Nordenskiöld and neighboring glaciers of Spitsbergen, 1921. The Journal of Geology, 408-446.

Sletten, K., Lyså, A., Lønne, I., 2001. Formation and disintegration of a high-arctic ice-cored moraine complex, Scott Turnerbreen, Svalbard. Boreas, 30(4), 272-284.

Staines, K.E.H., Carrivick, J.L., Tweed, F.S., Evans, A.J., Russell, A.J., Jóhannesson, T., Roberts, M., 2014. A multi-dimensional analysis of pro-glacial landscape change at Sólheimajökull, southern Iceland. Earth Surf Proc Land, n/a-n/a.

Stankowski, W., Kasprzak, L., Kostrzewski, A., Rygielski, W., 1989. An outline of morpho-genesis of the region between Horbyedalen and Ebbadalen, Petuniabukta, Billefjorden, central Spitsbergen. Pol Polar Res, 10, 749-753.

Strzelecki, M.C., Long, A.J., Lloyd, J.M., Małecki, J., Zagórski, P., Pawłowski, Ł., Jaskólski, M.W., 2018. The role of rapid glacier retreat and landscape transformation in controlling the post-Little Ice Age evolution of paraglacial coasts in central Spitsbergen (Billefjorden, Svalbard). Land Degrad Dev, 29(6), 1962-1978.

Strzelecki, M.C., Małecki, J., Zagórski, P., 2015. The Influence of Recent Deglaciation and Associated Sediment Flux on the Functioning of Polar Coastal Zone–Northern Petuniabukta, Svalbard, Sediment Fluxes in Coastal Areas. Springer, pp. 23-45.

Sund, M., Eiken, T., Hagen, J.O., Kaab, A., 2009. Svalbard surge dynamics derived from geometric changes. Ann Glaciol, 50(52), 50-60.

Szuman, I., Kasprzak, L., 2010. Glacier ice structures influence on moraines developement (Hørbye glacier, Central Spitsbergen). Quaestiones Geographicae, 29(1), 65-73.

Tonkin, T.N., Midgley, N., Cook, S.J., Graham, D.J., 2016. Ice-cored moraine degradation mapped and quantified using an unmanned aerial vehicle: A case study from a polythermal glacier in Svalbard. Geomorphology, 258, 1-10.

Wheaton, J.M., Brasington, J., Darby, S.E., Sear, D.A., 2010. Accounting for uncertainty in DEMs from repeat topographic surveys: improved sediment budgets. Earth Surf Proc Land, 35(2), 136-156.

Wojciechowski, A., 1989. Sedimentation in small proglacial lakes in the Hörbyebreen marginal zone, central Spitsbergen. Pol Polar Res, 10(3).

Zagorski, P., 2011. Shoreline dynamics of Calypsostranda (NW Wedel Jarlsberg Land, Svalbard) during the last century. Pol Polar Res, 32(1), 67-99.

Zagorski, P., Gajek, G., Demczuk, P., 2012. The influence of glacier systems of polar catchments on the functioning of the coastal zone (Recherchefjorden, Svalbard). Z Geomorphol, 56, 101-121.

Ziaja, W., 2001. Glacial recession in Sørkappland and central Nordenskiöldland, Spitsbergen, Svalbard, during the 20th century. Arctic, Antarctic, and Alpine Research, 36-41.

Ziaja, W., 2004. Spitsbergen landscape under 20th century climate change: Sørkapp Land. AMBIO: A Journal of the Human Environment, 33(6), 295-299.

Ziaja, W., 2005. Response of the Nordenskiold Land (Spitsbergen) glaciers Grumantbreen, Habergbreen and Dryadbreen to the climate warming after the Little Ice Age. Ann Glaciol, 42(1), 189-194.

Ziaja, W., Pipała, R., 2007. Glacial recession 2001-2006 and its landscape effects in the Lindströmfjellet-Håbergnuten mountain ridge, Nordenskiöld Land, Spitsbergen. Pol Polar Res, 28(4), 237-247.