FULL TITLE:

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).

SHORT TITLE: POST-LIA EVOLUTION OF PARAGLACIAL COASTS IN CENTRAL SPITSBERGEN

Mateusz C. Strzelecki[1], Antony J. Long[2], Jerry M. Lloyd[2], Jakub Małecki[3], Piotr Zagórski[4], Łukasz Pawłowski [1], Marek Jaskólski [1]

[1]*Institute of Geography and Regional Development, University of Wroclaw, pl. Uniwersytecki 1, 50-137 Wroclaw, Poland*

[2] *Department of Geography, Durham University, South Road DH1 3LE, Durham, UK*

[3]*Cryosphere Research Department, Adam Mickiewicz University in Poznan, ul. Bogumiła Krygowskiego 10, 61-680 Poznań, Poland*

[4]*Department of Geomorphology, Marie-Curie Sklodowska University in Lublin, Al. Kraśnicka 2cd, 20−718 Lublin, Poland*

ABSTRACT

In Svalbard, the rapid glacier retreat observed since the end of the Little Ice Age (LIA) has transformed the geomorphology and sediment budgets of glacial forelands, river valleys and slope systems. To date, relatively little information exists regarding the impact of such a profound glacial landscape degradation on the evolution of coastal environment. This paper addresses this deficiency by detailing the post-LIA sediment fluxes to the coastal zone in Billefjorden, central Spitsbergen (Svalbard). We analysed the response of the gravel-dominated barrier coast to the decay of Ferdinandbreen, one of the fastest retreating glaciers in the region. Glacier retreat resulted in the development of paraglacial sediment cascade where eroded and reworked glacigenic sediments progressed through alluvial fans to the coast, thus feeding gravel-dominated spit systems in Petuniabukta. We demonstrated the that coastal systems in central Spitsbergen responded abruptly to post-LIA climatic changes. The acceleration of coastal erosion and associated spit development was coincident with rapid climate warming that dates from the 1980's and has been associated with longer ice-free periods and activation of multiple sediment supply sources from the deglaciated landscape. In colder phases of post-LIA period, coastal zone development was subdued and strongly dependent on the efficiency of sediment transport via in a longshore drift. Finally, we discuss the differences in the post-LIA coastal responses between central Spitsbergen and western Spitsbergen highlighting the efficiency of paraglacial sediment delivery from land to the coast controlled by the state of glacial systems, bedrock topography and development of river channels.

INTRODUCTION

Due to its location at the boundary between North Atlantic and Arctic oceanic and atmospheric fronts, the Svalbard Archipelago (Figure 1) is well-placed to study the response of the High Arctic to climate change (D'Andrea *et al.*, 2012). During the last century, the landscapes of Svalbard have experienced a major change from a glacial towards a paraglacial domain as a consequence of widespread glacier retreat and the extensive reworking of glacigenic sediments by non-glacial geomorphological processes (e.g. Mercier *et al.*, 2009; Rachlewicz, 2009; Evans *et al.*, 2012; Małecki *et al.*, 2013; Ewertowski *et al.*, 2016; Strzelecki *et al.*, 2017a). Recent paraglaciaton of Svalbard has been associated with a warming of the climate since the end of the Little Ice Age (LIA), which occurred around AD 1900. Indeed, in recent decades, paraglacial processes have become the most effective geomorphic agent in Svalbard, reducing the impact of direct glacial processes to a secondary role in landscape change. Retreating glaciers have exposed vast areas of fresh and unstable glacigenic sediments that are easily released, eroded, transported and redistributed by processes that include: dead-ice melting (e.g. Ewertowski & Tomczyk, 2015), meltwater streams (e.g. Etzelmüller *et al.*, 2000), jökulhlaups (e.g. Etienne *et al.,* 2008), slope processes (e.g. Tomczyk & Ewertowski, 2017), rock weathering (Strzelecki, 2017), wind action (e.g. Rachlewicz, 2010) and coastal and fjord processes (e.g. Mercier *&* Laffly, 2005; Szczuciński & Zajączkowski, 2012).

To date, coastal change studies in Svalbard have focused on the link between coastal progradation with uninterrupted periods of glaciofluvial sediment supply (e.g. Héquette, 1992; Mercier *&* Laffly, 2005), as well as the significance of episodes of coastal erosion in areas no longer covered or protected by glaciers or sea ice on coastal evolution (e.g Rodzik & Zagórski, 2009; Ziaja *et al.*, 2009; Zagórski *et al.*, 2015). The role of sediment delivery from talus slopes and snow-fed ephemeral streams in controlling coastal evolution has been investigated by Lønne & Nemec (2004). Other studies have drawn attention to the mechanisms that control the development of rocky coasts (e.g. Strzelecki, 2011; Swirad *et al.*, 2017; Strzelecki et *al.,* 2017b); the reworking of glacigenic landforms exposed after the retreat of glaciers by coastal processes (e.g. Zagórski, 2011; Zagórski *et al.,* 2012); the influence of nearshore waters on the development of bottom active layer (e.g. Kasprzak *et al.*, 2017).

The scale of changes observed along relatively exposed coasts of Kongsfjorden, Bellsund, Hornsund and Sørkapp are large and unveiled the effective paraglacial transformation of Svalbard coastal landscape (e.g. Ziaja, 2002; Zagórski *et al.*, 2014; Ziaja & Ostafin 2015; Bourriquen *et al.*, 2016; Grabiec *et al.,* 2017). Concurrently, the response of coastlines in the protected interior of central Spitsbergen (Figure 1), such as in Billefjorden, that experience reduced wave fetch, low rates of precipitation and extended sea ice cover, has received limited attention to date (e.g. Sessford *et al.*, 2015a, b; Strzelecki *et al.*, 2017a; Guégan & Christiansen, 2017).

Set against this context, the overarching aim of this paper is to characterise the response of a gravel-dominated barrier coast, developing in inner-fjord environments of central Spitsbergen, to post-LIA climatic conditions that were characterised by enhanced paraglacial processes, in turn triggered by rapid deglaciation. The secondary aim is to compare the coastal zone changes observed in central part of Spitsbergen with those reported from western and southern Spitsbergen coasts, that evolved in settings influenced by larger glacier systems and storms sourced in the Greenland and Barents Seas.

REGIONAL SETTING

The study area is located in Petuniabukta, at the head of Billefjorden, central Spitsbergen (Figure 1). The area is well known in that it featured in the first systematic analyses of raised shorelines and their associated shells in Svalbard (Balchin, 1941; Feyling-Hanssen, 1955; Feyling-Hanssen & Olsson, 1959; Feyling-Hanssen, 1965). The bedrock geology here and beneath the adjacent fjord comprises a mixture of Precambrian, Devonian and Carboniferous-Permian outcrops (Dallmann *et al*., 2004). As we showed in this study, the bedrock outcrops that cross local coastal plains play and important role in controlling the sediment delivery to the coastal zone. The geomorphology of the valley systems that surround the bay is dominated by periglacial processes (e.g. Uxa *et al.* 2017) and paraglacial sediment reworking of glacial sediments deposited by the Late Weichselian glaciation and the Little Ice Age glacier advance (e.g. Rachlewicz, 2010; Ewertowski 2014; Pleskot 2015). Petuniabukta glaciers were part of a large ice stream complex that drained the Late Weichelseian ice sheet via Billefjorden into Isfjorden (Landvik *et al.*, 1998). Most probably glacier termini at the start of the Holocene were located inland of the present coast shortly after the initial retreat phase during the end of last glaciation (Baeten *et al.*, 2010; Forwick & Vorren, 2010). The mouths of local valleys in the Petuniabukta region were penetrated by marine water and, as relative sea level fell rapidly, spectacular flights of raised beaches developed. One of the most accurate relative sea-level curves for central Spitsbergen (Long et al. (2012)) suggests that in Petuniabukta RSL fell from c. 27 m a.s.l at 9700 cal yr BP and reach close to present sea level by 3100 cal yr BP. The trend in the mid-Holocene RSL data implies that the sea level most probably fell below present level during the late Holocene and later started a slow rise to the present. The RSL rise in the last two millennia was previously suggested by Forman (1990). The recent rise in sea-level was correlated by Feyling-Hanssen (1955) with a glacier re-advance about 2500 years ago based in his observations in Brucebyen, Kapp Napier and Skansbukta (Billefjorden). Further evidence for the late Holocene sea-level rise and stabilization close to present is suggested by the erosion of fan deltas documented in Adventfjorden by (Lønne and Nemec, 2004). This late Holocene RSL rise was most probably controlled by glacio-isostatic depression associated with the collapsing forebulge of the former Svalbard-Barents Sea Ice Sheet (e.g. Lambeck, 1995; Howell *et al.,* 2000; Ingólfsson & Landvik, 2013).

Petuniabukta has a micro-tidal range (*ca*. 1.5 m on spring tides) and a low-energy wave climate that is suppressed by lengthy periods of sea ice cover (up to 7-8 months). During the summer, the coastlines are affected by small icebergs and growlers that are sourced from the Nordenskiöldbreen, the only tidewater glacier in Billefjorden catchment. Wave energy in both bays is limited by the shallow fjord sill (ca. 50 m depth) and a narrow entrance (Szczuciński *et al.*, 2009). Wind conditions are strongly influenced by the surrounding orography and the presence of a large ice-plateau in the NE (Lomonosovfonna). The prevailing winds in Petuniabukta are from the S-SSE (along the fjord axis) and the longest wave fetch potential is from the south. A secondary wind direction is from the NE and represents katabatic winds coming from the valleys of Ragnarbreen and Ebbabreen, two outlet glaciers that drain the ice field (Figure 1). The study area is characterised by one of the driest and warmest climatic conditions among Svalbard regions (e.g. Przybylak *et al.*, 2014). The long-term weather observations at Svalbard Airport station, located ca. 60 km south from the study area, indicate that mean annual air temperature during the post-LIA period (1902-2016) was -5.9 °C and the mean total annual precipitation (1912-2016) was 192 mm (Figure 2). Analysis of weather data shows a cooling of the central Spitsbergen climate in early 20th century and in the 1960s-1970's. During the 1960's cooling the interannual temperature variability was low, whereas in the 1970's phase of cooling the annual variability rose drastically (Marsz *et al.*, 2013). The post-LIA period in Svalbard is also characterised by periods of rapid climate warming (e.g. 1918-1921 and, in particular, during the 1930s-1950s known as a 'the great Arctic warming' (Marsz *et al.*, 2013). The phase of so-called 'modern warming of the Arctic' started in Svalbard in the 1980's. Since the beginning of 21st century there have been several exceptionally warm years in Svalbard (particularly 2012-2016). Variations in air temperatures have also had a strong impact on sea ice conditions. As shown by Macias Fauria *et al.* (2010), a sharp decrease in sea ice cover in the western Nordic Seas in late nineteenth and early twentieth centuries corresponded to the termination of the Little Ice Age. The minima in sea ice extent in the 1920s and 1930s coincided with high air temperatures in the region, whereas the recovery of sea ice conditions in mid-20th century was linked with a general cooling. A decrease in sea ice extent from 1980's until present-day has been associated with the recent rapid warming of the Arctic.

The north part of Petuniabukta is occupied by a muddy tidal flat that is supplied with sediment from the extensive outwash plain that has developed by the Hørbyebreen, Svenbreen and Ragnarbreen rivers (Borówka 1989; Strzelecki *et al.*, 2015). The west and east coasts of Petuniabukta are fringed by gravel-dominated barriers that are separated by short rocky coast sections formed in anhydrite/gypsum and limestone (Strzelecki 2016). The barrier coasts are supplied with sediment mainly from glacial rivers and snow-fed streams that drain extensive talus fans (see Tomczyk & Ewertowski 2017). In addition, solifluction and mass-wasting of uplifted marine deposits are also important sources of coarse clastic materials to the coast. The most distinct feature of eastern coast of Petuniabukta is a spit-platform that developed in the mouth of Ebbaelva throughout the 20th century. It evolved in response to pulses of sediment supplied from a snow-fed alluvial fan delta (Strzelecki *et al.*, 2017a). In contrast to the straight coast formed in the eastern part of Petuniabukta, the western coast is indented with alluvial fan deltas and bedrock outcrops which causes the characteristic headland-bay shape of the coast. The coastal barrier developing in western Petuniabukta widens to the northwest where the coast is supplied by sediment derived from the Elsaelva and Ferdinandelva glacier rivers. This section of coast has a very diverse coastal landscape with numerous deltas, gravel-dominated barriers and barrier-spits, lagoons and relict barriers that are surrounded by a prograding tidal flat (Figure 1).

DATA AND METHODS

To describe and quantify post-LIA coastal zone changes in Petuniabukta we applied geomorphological field observations, differential GPS (DGPS) surveying, and interpretation of aerial images taken by Norwegian Polar institute in years 1936 - 2009. Fieldwork was conducted over five summer seasons between 2008-2012 and additional short visits in 2013, 2014 and 2015. Mapping and analysis of the coastal landforms and the front positions of Ferdinandbreen is based on remote-sensing and published data, with detailed geomorphological mapping completed during each field season. This was combined with field sketches and the interpretation of aerial photographs and old maps, ground-truthed in the field to produce the final geomorphological maps. In order to gain information on coastal zone change and controls of coastal evolution we have analysed the following maps of the central Spitsbergen region: a Cambridge University map of West Spitsbergen based on topographical surveys carried out during Cambridge Expeditions in 1930, 1938 and 1949 by Harland (1949); a geomorphological map of Petuniabukta region, 1: 40,000 by Karczewski *et al*. (1990); a geological map of Billefjorden, 1: 50,000 by Dallman *et al.* (2004); and a map of Hørbyebreen polythermal glacial landsystem by Evans *et al.* (2012). We have also examined sketches and notes on coastal evolution or the state of coastal landforms taken during early expeditions to the central part of Spitsbergen and published by Walton (1922), Slater (1925), Jackson (1931), Balchin (1941) and Feyling-Hannsen (1955).

We surveyed the topography of alluvial fans and coastal spits using a DGPS receiver (horizontal and vertical accuracy - ± 0.02 m). All surveys were tied back to a benchmark established during the 2008 expedition, and elevations refer to height above mean tide level in meters. We compared aerial images taken by the Norwegian Polar Institute (NPI) in 1961, 1990, and 2009 to determine the course of post-LIA coastal evolution during the last century. The basis for comparison was an orthophotomap created from digital aerial images taken in 2009 which were calibrated using ground control points measured with DGPS during the 2010 summer fieldwork. Images from 1961 and 1990 were imported to ArcGIS 9 software, overlain on the 2009 orthophotomap and georectified using a third order polynomial transformation with a total RMSE error of < 0.5 m. The extent of shorelines from 1961, 1990 and 2009 was delimited using the middle of the first, fully emerged beach ridge visible on any image. This procedure minimised the error stemming from different phases of the tidal cycle captured on any individual photographs. Lateral changes in shoreline position that are smaller than 2.5 m are not considered further. This is because it is impossible to distinguish if the visible coastal landforms comprise ephemeral gravel berms or storm ridges, which are currently separated by *ca.* 2 m. The position of LIA frontal moraine and aerial images (1961, 1990 and 2009) were used for establishing the glacier terminus location between the ca. 1900 and 2009 and were compared with data on glacier retreat recently presented by Małecki (2016).

RESULTS

Post-Little Ice Age development of the Ferdinandbreen proglacial zone

Ferdinandbreen is a very small glacier (ca. 1 km^2) which experienced one of the most drastic changes among central Spitsbergen glaciers, retreating ca. 1500 m and loosing ca. 60% of its area since the end of the LIA (Figure 3). At present the glacier is most likely cold-based, although partly temperate conditions at the bed likely existed during the LIA, as suggested at other small ice masses in Svalbard (e.g. Bælum & Benn; 2011; Lovell *et al.* 2015). The Ferdinandbreen foreland is dominated by an ice-cored latero-frontal moraine arc, with hummocky topography inside and superimposed linear debris stripes juxtaposing with crevasse fill ridges and eskers (Evans *et al.*, 2012). Such a glacial landsystem is characterised by large volumes of debris that is frozen at the former cold-temperate transition zone at the glacier snout.

The post-LIA retreat of Ferdinandbreen has exposed *ca.* 2 km^2 of glacier valley (Figure 3). Mass loss was particularly rapid during 1960-1990 when the glacier lost *ca.* 0.02 km^2 of area per year and retreated *ca.* 32.8 m annually. Although the annual recession slowed down to *ca.* 7.4 m yr^{-1} between 1990-2009, the glacier is still one of the fastest retreating in the Petuniabukta region (Małecki 2013; Małecki 2016). The deglacierised part of the valley system is filled with fresh, unstable glacigenic sediments, which are easily reworked by proglacial meltwater streams and modified by paraglacial slope processes (slumping, gullying and debris flows). This paraglacial sediment cascade has been enhanced by melting of ice-cores in glacial landforms (controlled ridges, eskers, lateral moraines) that in turn deliver high rates of sediment delivery to proglacial rivers.

Between 1961-2009 the snout of glacier transformed into a debris covered glacier with a small lake, dividing clean ice from a debris-covered part. Presently, the main supraglacial channels drain into the lake before they form the proglacial stream that flows in the middle of a valley and cuts through ice-cored landforms. Until 1990, Ferdinandelva was partly blocked by *ca*. 250 m-long remnants ice-cored moraine ridge, exposed during the retreat from 1961 position. This forced the stream to shift northwards before it incised through the main LIA moraine. Due to the rapid glacier front recession, the length of the main proglacial channel (between the glacier snout and the gorge in the LIA frontal moraine) has increased from *ca.* 460 m in 1961 to over 1600 m in 2009 (Figure 3). This change has strongly influenced the glaciofluvial sediment supply to adjacent alluvial fans and the coastal zone downstream. In 1960-1990, when sediment 'evacuation' was near the post-LIA maximum, the proglacial streams had a relatively short distance to cover before reaching the alluvial fans. The potential for sediment interception and storage in the small proglacial lake and along the elongated braided channel increased between 1990-2009. This has had a direct influence on the development of alluvial fans and sediment delivery to the barrier coast, as explained below.

Post-Little Ice Age development of Ferdinandbreen alluvial fans

Ferdinand Fan 1 (FF1)

The post-LIA evolution of FF1 is strongly linked with changes in the Ferdinandbreen proglacial drainage network and cessation of proglacial outflow through system of gorges in the moraine belt (Figure 4A). The first gorge in the moraine (MG1) is located in the northern part of a morainic arc, adjacent to the mountain slope and is already cut off from the proglacial river and hangs *ca.* 20 m above the bottom of the present glacier valley. The channel that once originated from the gorge is active only during spring-melt periods, when it drains meltwater from *ca.* 2-3 m deep snowpatches formed in a bedrock gorge (G0), before it spreads across the surface of FF1. The channel incision in the limestone bedrock has created a system of small waterfalls, with well-developed eversion hollows suggesting that it was formerly occupied by a significant stream (Figure 4B).

Remnants of glacigenic sediment covering banks of the bedrock channel suggest that this channel was active during the LIA, when the glacier snout formerly extended out of the valley (see photo of glacier position taken in 1930 - Figure 3) and supplied the growth of *ca.* 100,000 m^2 alluvial fan (FF1). Once the glacier started to retreat from its LIA maximum, the proglacial drainage network shifted towards the modern morainic gorge – MG2 (Figure 4C). The MG2 has been active since at least 1936 and drained proglacial waters in the direction of Ferdinand alluvial fans 2-4 (Figure 5A).

Ferdinand Fan 2 (FF2)

Since the termination of LIA, the Ferdinand Fan 2 has been significantly incised by Ferdinandelva, and by the present-day has lost *ca.* 40% of its (Holocene) maximum area (Figure 5B). The abandoned fan surface spreads between *ca.* 61 m (proximal) and 34 m (distal margin) above mean sea level. The relative sea-level data from the area (Long *et al.*, 2012) suggests that the fan was at sea level at the start of the Holocene and that its formation began just after the retreat of the ice-stream from Petuniabukta. The incision of the fan by the Ferdinandelva progressed as RSL fell rapidly following local deglaciation. The present-day river channel runs from *ca.* 59 m a.s.l. at the gorge in the LIA moraine to *ca.* 31 m a.s.l. where it enters two gorges that are incised in a bedrock ridge (G1 and G2). Based on GIS calculations of fan surface change and field measurements of incision depths we estimated the volume of sediments that had to be eroded by Ferdinandelva and washed away from the FF2 towards Ferdinand Fan 3 and Ferdinand Fan 4 to be *ca.* 210 000 m^3. The most important changes that occurred since the end of the LIA was the enlargement of a ravine in FF2 by lateral erosion of the incised fan deposits and by a shift of the main channel of Ferdinandelva towards gorge G2. This shift happened after 1961, at which date the gorge was still covered by former fan deposits (Figure 5C).

The area eroded from FF2 in years 1961-2009 was *ca.* 23,000 m^2 including 13,000 m^2 of FF2 that blocked access to the bedrock gorge G2. Erosion of FF2 towards the bedrock gorge G2 took place mainly between 1961 and 1990 when the modern river valley has widened to *ca.* 20,000 m^2, whereas between 1990-2009 the river eroded only *ca*. 3000 m^2 of OFF2. Subsequent to fan erosion, Ferdinandelva adjusted its channels to the new enlarged proglacial zone exposed by the retreating glacier. This saw the main channel avulse from the first bedrock gorge G1 towards the second bedrock gorge G2 (Figure 4). This drainage shift initiated the phase of intensive growth of Ferdinand Fan 4 and gradual suppression of the enlargement of Ferdinand Fan 3 (Figure 6).

Ferdinand Fan 3 (FF3)

Ferdinand Fan 3 is the largest fan system along the coast of Petuniabukta and fills the area between the modern barrier coast and the bedrock threshold (Figure 6). A comparison of the FF3 stable surface coverage changes inferred from aerial images indicates that in the last 20 years the fan expanded just 10,000 m^2. This means that sediment accumulation was significantly higher during 1961-1990 period, when the fan expanded from 220,000 m^2 to 300,000 m^2. It is possible to explain this change by the widening of the second gorge (G2) and shifting of main river channels that supplied the fan with sediments between 1961 and 1990 towards the Ferdinand Fan 4 (Figure 6).

The active fan is bordered by remnants of a relict fan that covered *ca.* 800,000 m^2 and has been incised *ca.* 2 to 4 m during the lowering of Ferdinandelva base-level. The relict fan developed on the surface of older Holocene beach deposits. This explains the large accumulations of shells and fragments of whale bones observed in exposed cliff walls. A fragment of the distal margin of old fan is located *ca.* 5 m a.s.l. and suggests that the intensified surface erosion and the accumulation of modern fan began at ca. 4000 cal yr BP (based on the RSL data presented by Long *et al.,* 2012). As in the modern system, the relict fan was bordered by gravel-dominated Ferdinand Old Spits (FOS) whose remnants are located at *ca.* 1.5 m a.s.l. (Figure 7).

Ferdinand Fan 4 (FF4)

Ferdinand Fan 4 is the youngest fan system developed in the area. The phase of rapid growth of fan started in 1960's at the earliest, when Ferdinadelva eroded the remnants of FF2 and reached the bedrock gorge G2 (Figure 5). Before the activation of the new Ferdinandelva channel, which established its way to the coast through G2, the FF4 surface was faintly incised by dozens of ephemeral snowmelt streams that drain from the mountain slopes. These streams drain through another five bedrock gorges (G3-G6) and fed the small fan system (56,000 m^2) that adjoins FF4 from the southwest (Figure 6). Once the glacier river reached the bedrock gorge the erosion through gorge continued and Ferdinandelva incised the surface of older fan that was developing between bedrock ridge and the coast (Figure 5B). Currently, the remnants of the old fan are undergoing erosion during the high spring and summer discharges (Figure 5C). Nowadays FF4 is the main pathway for sediment transport from the Ferdinandbreen valley towards the coast. Between 1990-2009, FF4 grew by *ca.* 50,000 m^2 and currently covers 180,000 m^2. This includes 2400 m^2 that formed after breaching of the coastal barrier and the accumulation of the fan delta in Petuniabukta (Figure 6). The breaching of the coastal barrier by FF4 channels was one of the most important drivers of shoreline change in Petuniabukta during the post-LIA period.

Post-Little Ice Age development of Ferdinand Spit Systems

The formation of new spit systems along FF3 are the biggest change observed along the Petuniabukta barrier coast since the end of the LIA (Table 1). The evolution of three spit systems (FS1, FS2, FS3) have had a profound effect on the evolution of FF3 by blocking its seaward progradation. Analysis of photographs taken by R.M. Jackson in 1930 and an oblique aerial image taken by Norwegian Polar Institute in 1936 suggest that the main body of FS1 had already formed by the 1930s, but did not block the outflow of FF3 channels to Petuniabukta. Between 1961-2009 the spit system extended northwards *ca.* 56 m and its northernmost tip recently reached the main tidal channel in the western part of Petuniabukta tidal flat (Figure 7).

Using archival maps and photographs it is difficult to determine the exact time of spit inception, although it can be assumed that FS2 started to form in the first decades of the 20th century, since the well-establish spit embryo is clearly visible on aerial image from 1936. FS2 was the main barrier system that developed along the western coast of Petuniabukta and constituted its northernmost shoreline until at least 1961.

During the 1960s the coastline had a rather straight edge and the main sediment supply to the barrier was from Elsaelva and ephemeral snowmelt streams from mountains draining throu system of bedrock gorges (G3-G8). Sometime between 1961 and 1990 the spit was cut off from the main barrier that was breached by one of the channels of Ferdinandelva that shifted towards the gorge (G2) and eroded its way to the coast through the remnants of the old fan (Figure 5). By 1990, Ferdinandelva had formed a small delta along the distal margin of FF4 that provided material for development of the third spit (FS3), which started to grow along FS2. The only section of the coast which experienced significant erosion since the end of LIA was a low-lying (*ca.* 0.3 m a.s.l.) relict beach-ridge platform located between the Elsaelva delta and the FF4 fan delta (Figure 8A). The erosion, that in some places reached *ca.* 50 m inland (for 1961-2009 period), led to the removal of *ca.* 1800 m^3 of sediment that was redistributed along the modern barrier (Figure 8B-C).

A reduction in the distance to the modern coast enabled several small snowmelt streams to incise through the weakened barrier and open two new connections between the relict lagoon and the fjord, thus supplying the coast with fine alluvial sediments. Up to this point these sediments would have accumulated behind the barrier (Figure 8D). However, these outlets are ephemeral features that are easily blocked by storm ridges, as has been documented during geomorphological observations after a few stormy days in summer 2009, 2010 and 2014. Therefore, even though by 1990 FS2 was partially cut off from longshore drift, the landform was still able to grow and indeed extended another 126 m by 2009. The reduction of spit area observed between 1990 and 2009 suggests that the landform elongation was partly a product of spit reworking.

The youngest spit, FS3, was fed sediment from the FF4 fan delta. The barrier that in 1990 started to develop from the base of FS2 in 19 years grew *ca.* 190 m and increased in size to *ca.* 30,000 m^2. This growth is the fastest among the three spits, emphasizing the importance of the activation of sediment supply from FF4 and smaller suppliers such as snowmelt streams in last two decades. Remotely-sensed and field-based geomorphological mapping indicates that, at least until 2010, the growth of spits occurred by the formation of recurved hooks (laterals) around the spit terminus that overlapped one another (Figure 7).

DISCUSSION

Climate controls the speed of glacier retreat and associated glacigenic sediment flux to the coast (e.g. Mercier & Laffly 2005; Etienne et al. 2008). In general, more continental climate in central Spitsbergen, than along the W and SW coast of the island (e.g. Rachlewicz

2009; Przybylak et al. 2014; Małecki 2016) force glaciers to retreat higher up into their valleys. As a result of the rapid post-LIA deglaciation, most of Petuniabukta glacier-fed rivers (e.g. Elsaelva, Ferdinandelva, Svenelva, Hørbyelva or Ebbaelva) have now to flow over relatively long distances to reach fjord shorelines (Table 2). What is more, the longer distance between the glacier foreland and the coastal zone induces storage of eroded and transported glacigenic sediments in glacifluvial, fluvial and lacustrine landforms. Intermediate storage of paraglacial sediments across braided river floodplains, in proglacial lakes or alluvial fans (e.g. Ferdinand Fans) is therefore important for coastal evolution. On the NW and SW coasts of Svalbard, coastal morphodynamics are to a greater degree controlled by exposure of glacial landforms during the retreat of tide-water glaciers and the associated rapid delivery of terrestrial sediments through systems of relatively short rivers/streams (e.g. Zagórski et al. 2012; Bourriquen et al., 2016; Grabiec et al., 2017). In such environments, the potential sediment sources (e.g. moraine or outwash plain) can be quickly exhausted due to fluvial and coastal erosion, particularly along more exposed to storms western coasts of Svalbard (Figure 9). In contrast, in Petuniabukta, the high storage capacity of river floodplains and alluvial fans means that the interplay between glacial and coastal zone transformation is more stable. Another important control of sediment storage capacity in Petuniabukta is a bedrock topography. Bedrock steps crossing coastal plain formed system of natural dams that moderate the paraglacial sediment delivery to the coastal zone. During the period of high sediment availability in deglaciated valleys or slopes the local rivers and streams were filling the space between bedrock steps until the erosion of gorge that enabled further sediment release to a lower area. The location of gorges in bedrock not only controlled the formation of alluvial fans but also entailed the location of sediment delivery to coastal system.

Recent accounts of post-LIA paraglacial coastal changes in Bellsund (Figure 9-B); Hornsund (Figure 9-C); Sørkappland (Figure 9-D) indicate that the rates of coast progradation here are slower than erosion (Table 2). In our opinion the erosion and rapid exhaustion of glacigenic sediment sources is intensified by exposure to storms and steep nearshore slopes along the W and SW coasts of Svalbard, reducing the space available for the accumulation of new coastal landforms. The coastal erosion, or the lack of longer-lived recent coastal landforms in sites like Hornsund (SW Spitsbergen) may be also linked with another mechanism. The rapid post-LIA retreat of tide-water glacier systems here has resulted in the lengthening of the fjord shorelines (e.g. the opening of Brepollen), and accelerated tidal pumping between the fjord and the open sea (Figure 9-C). This process may have intensified coastal erosion (Zagórski et al., 2015; Szczuciński et al., 2017). We argue that the dominance of coastal erosion along the western coast of Spitsbergen is therefore a combined effect of limited access to efficient sediment sources and long ice-free seasons, associated with warmer conditions. The increased progradation of Kongsfjorden coast (Figure 9-A) described by Mercier & Laffly (2005) and Bourriquen *et al.* (2016) was associated with uninterrupted sediment supply from the glacial runoff system. It is important to note that in periods of limited sediment supply associated with the migration of river channels, the Kongsfjorden coast experienced net erosion (Mercier & Laffly 2005). In Bellsund, Zagórski et al. (2012) documented that progradation of the outwash plain coast in front of Recherchebreen ceased immediately after a reduction in glaciofluvial sediment supply that was connected to the retreat of the glacier front and opening of a lagoon system that captured most of the freshly released sediments (Figure 9). Elsewhere, the development of coastal systems in previously studied parts of Svalbard (Kongsfjorden; Bellsund; Hornsund and Sørkappland) provide textbook examples of rapid coastal reorganization associated with gaining access to and reworking of glacigenic landforms known from the

Atlantic paraglacial coastal studies (e.g. Taylor *et al.*, 1986; Shaw *et al.*, 1990; Orford *et al.*, 1991; Forbes & Syvitski 1994; Fitzgerald and van Heteren 1999; Hjelstuen et al. 2009; Hein *et al.* 2012, 2014; Forde et al. 2016). The so-called 'lifespan' of a paraglacial coast is therefore dependent on the endurance of the (glacial) sediment source, so that often the phase of intensified growth and migration of paraglacial barriers is abruptly terminated by the exhaustion of sediment supply that occurs when a glacigenic landform is fully eroded (Forbes et al., 1995a).

In Petuniabukta, coastal zone responses to post-LIA landscape transformation occurred in a slightly different way. Coastal change here is predominated controlled by longshore spit extension rather than by gradual progradation of shorelines fed by direct coastal erosion of glacial landforms (Figure 9). Fresh sediments delivered from glacier valleys and reworked sediments ephemerally released from intermediate sediment storage systems (floodplains and alluvial fans) enter relatively low-energy (few storms, longer sea ice period) bays and provide conditions conducive to spit evolution. Another factor which has facilitated the growth of the Ferdinand Spits is a progradation of a tidal flat system that caused nearshore waters to shallow (and created subaqueous spit-platform for spit extension), and which often protected coastal landforms (older spit and barriers surrounded by tidal flat). Such a preservation of spits and barriers is typical not only for Billefjorden, but also other inner-fjord environments of central Spitsbergen and requires further study (e.g. Dicksonfjorden, Van Mijenfjorden or Ekmanfjorden). The post-LIA coastal change in Petuniabukta has revealed interesting scenario of High Arctic paraglacial coastal response, where limited access to glacigenic sediments/landforms for the direct functioning of coastal processes (e.g. erosion of a moraine) is counterbalanced by extended sediment release from floodplains and alluvial fans and, we argue, by lower wave energy caused by a lack of storms and prolonged sea ice conditions that combine to allow preservation of coastal landforms (e.g. spits) for longer periods. This scenario is similar to the fourth stage of fjord evolution model proposed by Syvitski *&* Shaw (1995), when fjords become dominated by paraglacial sedimentation associated with the reworking of glacigenic deposits left by a retreating or fully disappeared glaciers. In the Ballantyne's (2002) concept of paraglaciation, such an evolution of coastal system can be either slowed down or accelerated by sea-level change, and can be dramatically disturbed or even reversed by a climate shift.

We associate the differences in the rate of development of the Ferdinand Spits and local coastal landscape changes with shifts in climatic conditions that occurred in the second half of 20th century in Svalbard (Figure 2). It is important to note that 1960's were the coolest period since the termination of the LIA, and were characterized by cold conditions in all seasons and particularly strong cooling in winters and stable sea ice conditions (e.g. Mahoney *et al.*, 2008; Macias Fauria *et al.*, 2010). In our previous work we have shown that during climate cooling (1960's – 1970's) the sediment supply to Petuniabukta was limited to high magnitude/low frequency pulses associated with snow-melt and summer discharge (Strzelecki *et al.*, 2017a). The Ferdinand Spits were also very sensitive to shifts in climatic conditions. For instance, the development of Ferdinand Spit 1 in the colder phase of post-LIA was much slower (0.6 m yr^{-1}) than between 1990-2009 (2.1 m yr^{-1}), when the recent warming accelerated and sea ice conditions weakened. The climatic warming that continues from the 1980's has extended the duration of ice-free conditions (enabling higher rates of annual longshore sediment transport) and activated the delivery of sediments from valley and slope systems – both factors can explain the faster spit expansion in last 20-30 years (e.g. accelerated growth of Ferdinand Spit 2 from 1990's and almost 10 m per year of Ferdinand Spit 3 extension in years 1990-2009).

This study provides also an interesting example of the impact of relative sea-level change legacy on the recent evolution of barrier coast in central Spitsbergen. Changes in relative sea-level exert a fundamental control on coastline change by limiting or facilitating access to glacial landforms in all types of paraglacial coasts, and by controlling nearshore water depths which in turn impact on wave energy at the coast. The rapid relative sea-level fall observed in Petuniabukta since the onset of the Holocene resulted in cutting off the direct contact of fjord with glacial landforms as early as 9700 cal yr BP (Long *et al.*, 2012), at which time Petuniabukta was filling the entrances to Ebbadalen and Hørbyedalen and reworking the ice-contact deposits produced at that time by the then tide-water Ebbabreen and Hørbyebreen. In the northern part of the bay, the tidal flat started to develop along the distal margin of an outwash plain that was supplied with sediment from the Sven-Hørbye-Ragnarbreen catchments. The drop in the mid Holocene RSL data suggests that after 3100 cal yr BP RSL may have fallen below present sea-level and then started to rise again to reach the present (Long *et al.*, 2012). This period was probably the most important phase for the Ferdinand Fan 2 incision and for establishing the foundations for the modern Ferdinand Fans and spit evolution. As RSL serves as the ultimate base-level and controls the river bed level (Muto 1987), the proposed late Holocene fall in RSL could explain the incision of Ferdinand Fan 2 and the creation of accommodation space for Ferdinand Fan 3 and 4 by dissection and removal of older, uplifted marine deposits on the western coastal lowlands that formed during the early and mid-Holocene. Coastline erosion caused by RSL rise would steepen river gradients, promoting incision. Such incision is seen in many areas by the truncated nature of the lower portion of alluvial fans that formerly were graded to a lower-than-present sea level. Such a sequence of events would place the formation of FOS around the same age as the erosion of small scarp in NE Petuniabukta that divides the Ebba spit-platform and the last late Holocene beach (Ebba LH-1 beach from Long *et al.*, 2012).

As suggested by Long *et al.* (2012) it is also possible that some of the late Holocene RSL rise may be the result of local reloading of the Earth's crust due to a late Holocene increase in ice cover in central Spitsbergen. In his study of Cape Napier development in the adjacent Adolfbukta, Feyling-Hannsen (1955) correlated the recent rise in sea-level with glacier readvance about 2500 years ago. This is a timing of the first 'glacial maxima' reached by several small valley glaciers and tide-water glaciers in Isfjorden area that started to form around 3000-4000 years ago, after the potential full melt-out that occurred due to the Younger Dryas-Holocene warming (Svendsen & Mangerud, 1997). It is, however, widely accepted that Spitsbergen glaciers reached their Holocene maximum extent during the second late Holocene 'glacial maxima' that occurred during the Little Ice Age (e.g. Werner, 1993; Mangerud & Landvik, 2007; Majewski, *et al.*, 2009; van der Bilt *et al.* 2015). The local reloading of the Earth's crust during the LIA would then be potentially greater than during the '2500 BP' event. Under this scenario, it is possible that the formation of the FS1 initiated when the sea transgressed the Ferdinand Fans and used the sediments eroded from the shoreface to build a barrier that subsequently migrated onshore and started to extend laterally. This is consistent with one of the common models of barrier formation during RSL rise that suggests that spit elongation is a main driver of barrier onshore development (Orford, 2004).

Studies of mid-latitude paraglacial coasts show a general tendency of inland migration of swash-aligned gravel-dominated barriers with rising sea level (Carter & Orford, 1993; Forbes *et al.*, 1995ab). For drift-aligned barriers, development is largely a function of sediment supply. Reworking of updrift portions of spits can release sediment via cannabilisation of older beach deposits. However, if sufficient sediment is lacking, then drift aligned barriers may breakdown and be redistributed entirely (Orford *et al.*, 1991). However, due to the steep nearshore slope of the fjord, the offshore sediment supply to the spit system may not be effective enough to support spit development.

The analysis of aerial images and geomorphological mapping of Ferdinand Spits revealed the gradual shift of their axes from the NNE (FS1) towards the NE (FS3).

Interestingly, the axes of relict FOS run almost straight on to the N. Such a gradual shift in spit orientation may be associated with wave diffraction and longshore current in shallow waters at the shifting boundary between the FF3 distal margin and the prograding tidal flat (Figure 7). On the other hand, this may also be linked with a transformation from drift-aligned to swash-aligned coastal features that gradually shift their orientation for facing incoming waves (e.g. Orford et al. 2002). According to Orford et al. (1991) and Anthony (2008), changes from drift-aligned to swash-aligned status depend mainly on sediment supply, although change may also be induced by wave climate variations. It may also be argued that changes in the orientation of the axes of the spits may also be related to the initial orientation of the FF3 shoreline. The planform of FS1 has the straightest shape from among the three Ferdinand Spits suggesting the strong influence of longshore drift. According to Carter & Orford (1991) this is a diagnostic feature of a drift-aligned spit formed by high volume but episodic sediment supply, typical of cooler climatic conditions. In contrast, more frequent pulses of sediment supply and longer periods of wave action during warmer phases of the post-LIA periods help explain changes in the orientation and the larger size of the younger spit systems.

CONCLUSIONS

This paper provides new insights into the functioning of High Arctic paraglacial coastal environments of central Spitsbergen that are characterised by sheltered fjord settings and which are supplied by sediments derived from rapidly retreating glacier system and intermediate paraglacial storage systems in the form of floodplain and alluvial fans. Our observations broaden the picture of Svalbard coastal zone previously classified as a ‘stable’ or ‘aggrading’ coastline (Lantuit *et al.,* 2012) and enable us to draw the following conclusions:

- Reconstruction of the post-LIA development of glacial-fed barrier coasts show that since the end of the LIA, the Petuniabukta coast has experienced significant coastal transformation. Climate-induced intensified erosion and transport of sediments from retreating glaciers has led to the degradation of glacial landscape and the formation of extensive alluvial fan systems that feed coastal landforms.
- The post-LIA evolution of alluvial fan systems (Ferdinand Fans) and the spits (Ferdinand Spits) was closely linked to the retreat of Ferdinandbreen. The post-LIA deglaciation and change of glacier snout position forced the shift of Ferdinandelva channels across coastal plain and incision of fan surfaces. Over time, the migration of river channels not only cut off particular fans from glacigenic sediment supply but also moved southward the location of sediment delivery to the fjord what resulted in consecutive formation of three spits.
- The combined action of longshore drift in the fjord and the delivery of reworked glacigenic sediments through a system of alluvial fans resulted in the development of three large spit systems along the alluvial fan shorelines. The rate of spit extension has accelerated over the last 50 years from 0.6 m per year during colder decades (particularly the 1960's, characterised by harsh sea ice conditions and limited delivery of terrestrial sediments) to 9.8 m per year in the recent warming phase (from 1990 onwards associated with longer ice-free conditions and intensified paraglacial sediment transport from slope, valley and glacial sources).
- The migration of modern spits and changes to the orientation of relict coastal barriers suggests the rapid adjustment of landforms to a late Holocene RSL rise associated with local glacier advances around 2500 BP and during the LIA, as well as to RSL fall and potential land uplift following LIA glacial retreat and unloading. RSL fall and the increased elevation of glacier snouts caused the lowering of river base level and the associated incision of alluvial fans and outwash plains. Sediments eroded from those

intermediate sediment storage systems served as the most important source for the development of the barrier coast in Petuniabukta.

- Coastal evolution in these contexts differs from that observed on the western coast of Spitsbergen. The lack of direct access to glacigenic sediment sources is counterbalanced by sediment release from floodplain and fan storage systems to a large degree controlled by bedrock topography (gorges). The limited storm action and relatively stable sea ice conditions mean that the growth and preservation of new coastal landforms is enhanced compared with conditions on the more ice-free and stormy western coasts of Svalbard.

AUTHOR CONTRIBUTIONS

M.C.S. oversaw all aspects of the research, led fieldwork and wrote the manuscript. M.C.S., A.J.L., and J.M.L. guided the intellectual direction of the research. M.C.S., A.J.L., L.P., M.J., and P.Z. assisted with fieldwork. J.M. and P.Z. provided logistical support for the part of fieldwork. J.M. provided glaciological data. All authors reviewed the manuscript.

ACKNOWLEDGEMENTS

This paper is a contribution to the National Science Centre project ‘Model of the interaction of paraglacial and periglacial processes in the coastal zone and their influence on the development of Arctic littoral relief’ (award no. 2013/08/S/ST10/00585 – FUGA Postdoctoral Fellowship). Fieldwork has been partly assisted by National Science Centre SONATA BIS project UMO-2013/10/E/ST10/00166. M.C. Strzelecki has been also supported by the Ministry of Science and Higher Education Outstanding Young Scientist Scholarship, Foundation for Polish Science and Crescendum Est Polonia Foundation. This research contributes to the PAST (Palaeo-Arctic Spatial and Temporal) Gateways Programme. We thank the four anonymous reviewers, the Managing Editor Denis Mercier and the Editor Chris Barrow for their very thoughtful and constructive comments, which significantly improved the manuscript. We are also grateful to Harald Aas from NPI for providing aerial images from Svalbard.

REFERENCES

Anthony EJ. 2008. Gravel beaches and barriers. *Developments in Marine Geology* **4:** 289-324.

Bælum K, Benn DI. 2011. Thermal structure and drainage system of a small valley glacier (Tellbreen, Svalbard), investigated by ground penetrating radar. *The Cryosphere* **5**: 139–149.

Baeten NJ, Forwick M, Vogt C, Vorren TO. 2010. Late Weichselian and Holocene sedimentary environments and glacial activity in Billefjorden, Svalbard. In: Howe JA, Austin WEN, Forwick M, Paetzel M. (Eds.), *Fjord Systems and Archives*. Geological Society, London, Special Publications **344**: 207-223.

Balchin WGV. 1941. The raised features of Billefjord and Sessenfjord west Spitsbergen. *Geographical Journal* **97**: 364-376.

Ballantyne C. 2002. Paraglacial geomorphology. *Quaternary Science Reviews* **21**: 1935–2017.

Błaszczyk M, Jania JA, Kolondra L. 2013. Fluctuations of tidewater glaciers in Hornsund Fjord (Southern Svalbard) since the beginning of the 20th century. *Polish Polar Research* **34**: 327-352.

Borówka M. 1989. The development and relief of the Petuniabukta tidal flat, central Spitsbergen. *Polish Polar Research* **10**: 379–384.

Bourriquen M, Baltzer A, Mercier D, Fournier J, Pérez L, Haquin S, Bernard E, Jensen M. 2016. Coastal evolution and sedimentary mobility of Brøgger Peninsula, northwest Spitsbergen. *Polar Biology* **39**:1689-1698.

Carter RWG, Orford JD.1991. The sedimentary organization and behaviour of drift-aligned gravel barriers. *Coastal Sediments '91*, American Society of Civil Engineers **1**: 934–948.

Carter RWG, Orford JD. 1993 The morphodynamics of coarse clastic beaches and barriers: a short and long-term perspective. *Journal of Coastal Research* SI **15**:158-179.

Dallmann WK., Pipejohn K., Blomeier D. 2004. Geological map of Billefjorden, Central Spitsbergen, Svalbard with geological excursion guide 1 : 50,000: NPI Tematkart, No 36.

D'Andrea WJ, Vaillencourt D, Balascio NL, Werner A, Roof S, Retelle M, Bradley RS. 2012. A mild Little Ice Age and unprecedented recent warmth in an 1800-year lake sediment record from Svalbard. *Geology* **40**: 1007-1010.

Eccleshall SV, Hormes A, Hovland A, Preusser F. 2016. Constraining the chronology of Pleistocene glaciations on Svalbard: Kapp Ekholm re-visited. *Boreas* **45**:790-803.

Etienne S, Mercier D, Voldoire O. 2008. Temporal scales and deglaciation rhythms in a polar glacier margin, Baronbreen, Svalbard. *Norwegian Journal of Geography* **62**: 102 - 114.

Etzelmüller B, Ødegård RS, Vatne G, Mysterud RS, Tonning T, Sollid JL. 2000. Glacier characteristics and sediment transfer system of Longyearbreen and Larsbreen, western Spitsbergen. *Norwegian Journal of Geography* **54**: 157–168.

Evans DJA, Strzelecki M, Milledge DG, Orton C. 2012. Hørbyebreen polythermal glacial landsystem, Svalbard. *Journal of Maps* **8:** 1-11.

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

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

Ewertowski MW, Evans DJA, Roberts DH, Tomczyk AM. 2016. Glacial geomorphology of the terrestrial margins of the tidewater glacier, Nordenskiöldbreen, Svalbard. *Journal of Maps* **12**: 476-487.

Feyling-Hanssen, R.W., 1955, Stratigraphy of the marine Late-Pleistocene of Billefjorden, Vestspitsbergen. *Norsk Polarinstitutt Skrifter* **107**: 1-168.

Feyling-Hanssen RW. 1965. Shoreline displacement in central Vestspitsbergen. *Norsk Polarinstitutt Meddelelser* **93**: 1-5.

Feyling-Hanssen RW, Olsson IU. 1959. Five radiocarbon datings of postglacial shorelines in central Spitsbergen. *Norwegian Journal of Geography* **17**: 122-131.

FitzGerald, D.M., van Heteren, S., 1999. Classification of paraglacial barrier systems: coastal New England, USA. *Sedimentology* **46**: 1083-1108.

Forbes D, Syvitski JPM. 1994. Paraglacial coasts. In: Carter RWG & Woodroffe CD. (eds.) *Coastal Evolution: Late Quaternary Shoreline Morphodynamics.* Cambridge University Press, Cambridge, 373–424.

Forbes DL, Orford JD, Carter RWG, Shaw J, Jennings SC. 1995a. Morphodynamic evolution, self-organisation, and instability of coarse-clastic barriers on paraglacial coasts. *Marine Geology* **126**: 63–85.

Forbes DL, Shaw J, Taylor RB. 1995b. Differential preservation of coastal structures on paraglacial shelves. Holocene deposits of south-eastern Canada. *Marine Geology* **124**:187–201.

Forde TC, Nedimović MR, Gibling MR, Forbes DL. 2016. Coastal Evolution Over the Past 3000 Years at Conrads Beach, Nova Scotia: the Influence of Local Sediment Supply on a Paraglacial Transgressive System. *Estuaries and Coasts* **39**: 363-384.

Forman SL. 1990. Postglacial relative sea-level history of northwestern Spitsbergen, Svalbard. *Geological Society of America Bulletin* **102**: 1580-1590.

Forwick M, Vorren TO. 2010. Stratigraphy and deglaciation of the Isfjorden area, Spitsbergen. *Norwegian Journal of Geology* **90**:163-179.

Grabiec M, Ignatiuk D, Jania JA, Moskalik M, Głowacki P, Błaszczyk M, Budzik T, Walczowski W. 2017. Coast formation in an Arctic area due to glacier surge and retreat: the Hornbreen - Hambergbreen case from Spistbergen. *Earth Surface Processes and Landforms*, doi: 10.1002/esp.4251.

Guégan EBM, Christiansen HH. 2017. Seasonal Arctic Coastal Bluff Dynamics in Adventfjorden, Svalbard. *Permafrost & Periglacial Processes* **28**: 18-31.

Harland WB. 1952. The Cambridge Spitsbergen Expedition, 1949. *The Geographical Journal* **118**: 309-329.

Hein CJ, FitzGerald DM, Carruthers EA, Stone BD, Barnhardt WA, Gontz AM. 2012. Refining the model of barrier island formation along a paraglacial coast in the Gulf of Maine. *Marine Geology* **307–310**: 40–57.

Hein CJ, FitzGerald DM, Buynevich IV, van Heteren S, Kelley JT. 2014. Evolution of paraglacial coasts in response to changes in fluvial sediment supply. *in* Martini IP, Wanless HR, eds., *Sedimentary Coastal Zones from High to Low Latitudes: Similarities and Differences*: Geological Society, London, Special Publications **388**:247 – 280.

Héquette A. 1992. Morphosedimentological dynamics and coastal evolution in the Kongsfjorden area, Spitsbergen. *Polar Geography & Geology* **16**: 321-329.

Hjelstuen BO, Haflidason H, Sejrup HP, Lyså A. 2009. Sedimentary processes and depositional environments in glaciated fjord systems e evidence from Nordfjord, Norway. *Marine Geology* **258**:88-99.

Howell D, Siegert MJ, Dowdeswell JA. 2000. Modelling the influence of glacial isostasy on Late Weichselian ice-sheet growth in the Barents Sea. *Journal of Quaternary Science* **15**: 475-486.

Ingólfsson Ó, Landvik JY. 2013. The Svalbard Barents Sea ice-sheet e Historical, current and future perspectives. *Quaternary Science Reviews* **64**: 33-60

Jackson RM. 1931. A Traverse from Ice Fjord to Wijde Bay, Spitsbergen. *The Geographical Journal* **78**: 277-283.

Karczewski A. (ed.)1990. *Geomorphology – Petuniabukta, Billefjorden, Spitsbergen*. UAM, Poznań, scale 1: 40 000.

Kasprzak M, Strzelecki MC, Traczyk A, Kondracka M, Lim M, Migała K. 2017. On the potential for a reversal of the permafrost active layer: the impact of seawater on permafrost degradation in a coastal zone imaged by electrical resistivity tomography (Hornsund, SW Spitsbergen). *Geomorphology* **293 B:** 347-359. DOI:10.1016/j.geomorph.2016.06.013

Lambeck K. 1995. Constraints on the Late Weichselian ice-sheet over the Barents Sea from observations of raised shorelines. *Quaternary Science Reviews* **14**:1-16.

Landvik JY, Bondevik S, Elverhøi A, Fjeldskaar W, Mangerud J, Salvigsen O, Siegert MJ, Svendsen JI, Vorren TO. 1998. The last glacial maximum of Svalbard and the Barents Sea area: ice sheet extent and configuration. *Quaternary Science Reviews* **17**: 43-75.

Lantuit H, Overduin PP, Couture N, et al. 2012. The ACD coastal database: a new classification scheme and statistics on Arctic permafrost coastlines. *Estuaries & Coasts* **35**: 383-400.

Long, AJ, Strzelecki MC, Lloyd JM, Bryant C. 2012. Dating High Arctic Holocene relative sea level changes using juvenile articulated marine shells in raised beaches. *Quaternary Science Reviews* **48**: 61-66.

Lønne I, Nemec W. 2004. High-arctic fan delta recording deglaciation and environment disequilibrium. *Sedimentology* **51**:553–589.

Lovell H, Fleming EJ, Benn DI, Hubbard B, Lukas S, Naegeli K. 2015. Former dynamic behaviour of a cold-based valley glacier on Svalbard revealed by basal ice and structural glaciology investigations. *Journal of Glaciology* **61**: 309-328.

Macias Fauria M, Grinsted A, Helama S, Moore J, Timonen M, Martma T, Isaksson E, Eronen M. 2010. Unprecedented low twentieth century winter sea ice extent in the Western Nordic Seas since A.D. 1200. Climate Dynamics **34**: 781-795.

Mahoney AR, Barry RG, Smolyanitsky V, Fetterer F. 2008. Observed sea ice extent in the Russian Arctic, 1933–2006. *Journal of Geophysical Research: Oceans* **113**: C11005, doi:10.1029/2008JC004830.

Majewski W, Szczuciński W, Zajączkowski M. 2009. Interactions of Arctic and Atlantic water-masses and associated environmental changes during the last millennium, Hornsund (SW Svalbard). *Boreas* **38**:529–544.

Małecki J. 2013. Elevation and volume changes of seven Dickson Land glaciers, Svalbard, 1960-1990-2009. *Polar Research* **32**:1, 18400, DOI:10.3402/polar.v32i0.18400.

Małecki J. 2016. Accelerating retreat and high-elevation thinning of glaciers in central Spitsbergen: *The Cryosphere* **10**: 1317-1329.

Małecki J, Faucherre S, Strzelecki MC. 2013. Post-surge geometry evolution and thermal structure of Hørbyebreen, central Spitsbergen, Svalbard Archipelago. *Polish Polar Research* **34**:305-321.

Mangerud J, Landvik JY. 2007. Younger Dryas cirque glaciers in western Spitsbergen: smaller than during the Little Ice Age. *Boreas* **36**:278-285.

Marsz AA, Niedźwiedź T, Styszyńska A. 2013. Modern climate changes on Spitsbergen as a basis for determining landscape metamorphosis. [In:] Zwoliński Z., Kostrzewski A., Pulina M. (eds.), *Ancient and modern geoecosystems of Spitsbergen.* Bogucki Wydawnictwo Naukowe, Poznań, ISBN: 978-83-63400-54-5, pp. 391-413.

Mercier D, Laffly D. 2005. Actual paraglacial progradation of the coastal zone in the Kongsfjorden area, western Spitsbergen (Svalbard), *in* Harris C, Murton JB, eds., *Cryospheric systems: glaciers and permafrost*. Geological Society, London, Special Publication **242**: 111–117.

Mercier D, Étienne S, Sellier D, André M-F. 2009. Paraglacial gullying of sediment-mantled slopes: a case study of Colletthøgda, Kongsfjorden area, West Spitsbergen (Svalbard). *Earth Surface Processes & Landforms* **34**: 1772-1789.

Muto T. 1987. Coastal Fan Processes Controlled by Sea Level Changes: A Quaternary Example from the Tenryugawa Fan System, Pacific Coast of Central Japan. *Journal of Geology* **95**:716-724.

Orford J. 2004. Barrier and Barrier Island, *in* Goudie A. ed., *Encyclopedia of Geomorphology*. London, Routledge, p. 59-62.

Orford JD, Carter RWG, Jennings SC. 1991. Coarse clastic barrier environments: evolution and implications for quaternary sea-level interpretation. *Quaternary International* 9: 87–104.

Orford JD, Forbes DL, Jennings SC. 2002. Organisational controls, typologies and time scales of paraglacial gravel-dominated coastal systems. *Geomorphology* **48**: 51-85.

Pleskot K. 2015. Sedimentological Characteristics of Debris Flow Deposits Within Ice-Cored Moraine of Ebbabreen, Central Spitsbergen. *Polish Polar Research* **36**:125–144.

Przybylak R, Araźny A, Nordli Ø, Finkelnburg R, Kejna M, Budzik T, Migala K, Sikora S, Puczko D, Rymer K, Rachlewicz G. 2014. Spatial distribution of air temperature on Svalbard during 1 year with campaign measurements. *International Journal of Climatology* **34**: 3702-3719.

Rachlewicz G. 2009. *Contemporary sediment fluxes and relief changes in high Arctic glacierized valley systems (Billefjorden, Central Spitsbergen*). AMU Press, Poznań: 203 p.

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**: 59–67.

Rachlewicz G, Szczuciński W, Ewertowski M. 2007. Post-“Little Ice Age” retreat rates of glaciers around Billefjorden in central Spitsbergen, Svalbard. *Polish Polar Research* **28**: 159–186.

Rodzik J., Zagórski P., 2009. Shore ice and its influence on development of the shores of south-western Spitsbergen. *Oceanological and Hydrobiological Studies* **38**: 163–180

Sessford E., Strzelecki MC, Hormes A. 2015a. Reconstruction of Holocene patterns of change in a High Arctic coastal landscape, Southern Sassenfjorden, Svalbard. *Geomorphology* **234**: 98-107.

Sessford EG, Bæverfjord MG, Hormes A. 2015b. Terrestrial processes affecting unlithified coastal erosion disparities in central fjords of Svalbard. *Polar Research* **34**:24122, DOI: 10.3402/polar.v34.24122.

Shaw J, Taylor RB, Forbes DL. 1990. Coarse clastic barriers in eastern Canada: patterns of glaciogenic sediment dispersal with rising sea levels. *Journal of Coastal Research* SI **9**: 160–200.

Slater G. 1925. Observations on the Nordenskiöld and neighboring glaciers of Spitsbergen, 1921. *Journal of Geology* **33**: 408–446.

Strzelecki MC. 2011. Schmidt hammer tests across a recently deglacierized rocky coastal zone in Spitsbergen - is there a ‘coastal amplification’ of rock weathering in polar climates? *Polish Polar Research* **32**: 239-252.

Strzelecki MC. 2017. The variability and controls of rock strength along rocky coasts of central Spitsbergen, High Arctic. *Geomorphology* **293 B:** 321-330. DOI:10.1016/j.geomorph.2016.06.014

Strzelecki MC, Małecki J, Zagórski P. 2015. The Influence of Recent Deglaciation and Associated Sediment Flux on the Functioning of Polar Coastal Zone – Northern Petuniabukta, Svalbard, *in* Maanan M, Robin M. eds., Sediment Fluxes on Coastal Areas. *Coastal Research Library* **10:** 23-45.

Strzelecki MC, Long AJ, Lloyd JM. 2017a. Post-Little Ice Age development of a High Arctic paraglacial beach complex. *Permafrost & Periglacial Processes* **28**: 4-17.

Strzelecki MC, Kasprzak M., Lim M, Swirad ZM, Jaskolski M, Pawłowski Ł, Modzel P. 2017b. Cryo-conditioned rocky coast systems: A case study from Wilczekodden, Svalbard. *Science of the Total Environment* **607-608**: 443-453. DOI:10.1016/j.scitotenv.2017.07.009

Svendsen JI, Mangerud J. 1997. Holocene glacial and climatic variations on Spitsbergen, Svalbard: *The Holocene* **7**: 45-57.

Syvitski J, Shaw J. 1995. Sedimentology and Geomorphology of Fjords. *in* Perillo GME. *Geomorphology and Sedimentology of Estuaries*, Amsterdam, Elsevier, p. 113-178.

Szczuciński W, Zajączkowski M. 2012. Factors controlling downward fluxes of particulate matter in glacier-contact and non-glacier contact settings in a subpolar fjord (Billefjorden, Svalbard). In *Sediments, Morphology and Sedimentary Processes on Continental Shelves: Advances in Technologies, Research and Applications*, Li M, Sherwood C, Hill P (eds). International Association of Sedimentologists Special Publications **44**: 369–386.

Szczuciński W, Zajączkowski M, Scholten J. 2009. Sediment accumulation rates in subpolar fjords – Impact of post-Little Ice Age glaciers retreat, Billefjorden, Svalbard. *Estuarine, Coastal & Shelf Science* **85**: 345-356.

Szczuciński W., Moskalik M., Dominiczak A., 2017. Tidal pumping - missing factor in glacial bays evolution*? Geophysical Research Abstracts* **19**, 13516.

Taylor RB, Forbes DL, Carter RWG, Orford JD. 1986. Beach sedimentation in Ireland: similarities and contrasts with Atlantic Canada. *Geological Survey of Canada, Paper* **86-1B**: 55-64.

Tomczyk AM, Ewertowski MW. 2017. Surface morphological types and spatial distribution of fan-shaped landforms in the periglacial high-Arctic environment of central Spitsbergen, Svalbard. *Journal of Maps* **13**: 239-251.

van der Bilt WGM, Bakke J, Vasskog K, D'Andrea WJ, Bradley RS, Ólafsdóttir S. 2015. Reconstruction of glacier variability from lake sediments reveals dynamic Holocene climate in Svalbard. *Quaternary Science Reviews* **126**: 201–218.

Walton J. 1922. A Spitsbergen salt marsh; with observations on the ecological phenomena attendant on the emergence of land from the sea. *Journal of Ecology* **10**: 109-21.

Werner A. 1993. Holocene moraine chronology, Spitsbergen, Svalbard: lichenometric evidence for multiple neoglacial advances in the Arctic. *Holocene* **3**: 128–137.

Zagórski P. 2011. Shoreline dynamics of Calypsostranda (NW Wedel Jarlsberg Land, Svalbard) during the last century. *Polish Polar Research* **32**: 67-99.

Zagórski P, Gajek G, Demczuk P. 2012. The influence of glacier systems of polar catchments on functioning of the coastal zone (Recherchefjorden, Svalbard). *Zeitschrift für Geomorphologie* Suppl. **56**: 101-122.

Zagórski P, Strzelecki MC, Rodzik J. 2014. Processes controlling the past and recent evolution of coastal environments in the southern Bellsund, Svalbard. In Migala K, Owczarek P, Kasprzak M, Strzelecki MC (eds.), *New perspectives in polar research,* Institute of Geography and Regional Development, University of Wrocław, 205-230.

Zagórski P, Rodzik J, Moskalik M, Strzelecki MC, Lim M, Błaszczyk M., Promińska A, Kruszewski G, Styszyńska A, Malczewski A. 2015. Multidecadal (1960–2011) shoreline changes in Isbjørnhamna (Hornsund, Svalbard). *Polish Polar Research* **36**: 369–390.

Ziaja W. 2002. Changes of the landscape structure of Sørkappland. In Sørkappland landscape structure and functioning (Spitsbergen, Svalbard), ed. W. Ziaja, and S. Skiba, 18–50. Krakow: Wydawnictwo UJ.

Ziaja W, Maciejowski P, Ostafin K. 2009. Coastal Landscape Dynamics in NE Sørkapp Land (SE Spitsbergen), 1900–2005. *AMBIO* **38**: 201-208.

Ziaja W, Lisowska M, Olech M, Osyczka P, Węgrzyn M, Dudek J, Ostafin K. 2011. *Transformation of the natural environment in Western Sørkapp Land (Spitsbergen) since the 1980s*. Jagiellonian University Press, Cracow, 92 p.

Ziaja W, Ostafin K. 2015. Landscape–seascape dynamics in the isthmus between Sørkapp Land and the rest of Spitsbergen: Will a new big Arctic island form? *AMBIO* **44**: 332-342.

	1936	**1961**		**1990**		**2009**		**rate of spit expansion [m yr^{-1}]**		
Spit		**axis length [m]**	**area [sq m]**	**axis length [m]**	**area [sq m]**	**axis length [m]**	**area [sq m]**	**1961-1990**	**1990-2009**	**1961-2009**
FS1	**present**	**679**	**210,000**	**696**	**18,000**	**735**	**17,000**	**0.6**	**2.1**	**1.2**
FS2	**embryo formed**	**270**	**10,000**	**431**	**13,600**	**557**	**13,000**	**5.6**	**6.6**	**6**
FS3	**x**	**x**	**x**	**315**	**7100**	**502**	**10,000**	**x**	**9.8**	**x**

Table 1. Post-LIA changes (length/area) changes to the Ferdinand Spits. x – landform did not exist in this period so no rate calculated.

	Kongsfjorden **NW Spitsbergen**	**Bellsund** **W Spitsbergen**	**Hornsund** **SW Spitsbergen**	**Sørkappland** **S Spitsbergen**	**Petuniabukta** **Central Spitsbergen**
Highest observed rates of coastal erosion	up to 30 m (1 m yr^{-1}) [1966-1995]	up 110 m (1.5 m yr^{-1}) [1936-2007]	up to 46 m (0.9 m yr^{-1}) [1960-1990]	up to 460m (6.7 m yr^{-1}) [1936-2005]	up to 50 m (1.04 m yr^{-1}) [1961-2009]
Highest observed rates coastal seaward progradation	up to 120 m (5 m yr^{-1}) [1966-1990] & (4 m yr^{-1}) [2011-2014]	up to 65 m (0.9 m yr^{-1}) [1936-2007]	up to 13 m (0.3 m a-1) [1960-2011]	up to 300 m (20 m yr^{-1}) [1990-2005]	up to 48 m (0.96 m yr^{-1}) [1961-2009]
Highest rates of elongation of coastal landforms (spits)	x	up to 91 m (1.8 m yr^{-1}) [1960-2009]	x	x	up to 187 m (9.8 m yr^{-1}) FS3 [1990-2009]
Glacier retreat rates **(LT): land-terminated** **(TW): tide-water** **and** **mean distance between snout and coast (SCD)**	(LT) Midtre Lovénbreen: (10 m yr^{-1}) • SCD: 1200 m (TW) Kronebreen (150 m yr^{-1})	(LT) Scottbreen: (16 m $yr^{-1)}$ [1936-2002] • SCD: 2300 m (TW) Recherchebreen (27 m yr^{-1}) [1960-2008]	(LT) Ariebreen: (10 m yr^{-1}) • SCD: 2200 m (TW) Hansbreen: (25 m yr^{-1}) [1900-2008]	(LT) Kambreen: (12m yr^{-1}) [1900-2005] • SCD: 600 m (TW) Hambergbreen: (160 m yr^{-1}) [1900-2000]	(LT) Hørbyebreen: (10.6 m yr^{-1}) [1900-2009] • SCD: 3800 m Svenbreen: (11.7 m yr^{-1}) • SCD: 3500 m Ferdinandbreen: (14.2 m yr^{-1}) • SCD: 3100 m Elzabreen: (11.3 m yr^{-1}) • SCD: 3200 m Ragnarbreen: (14.7 m yr^{-1}) • SCD: 5700 m

Table 2. Regional characteristics of coastal changes and glacier retreat rates observed in various regions of Spitsbergen: CENTRAL (study site – Billefjorden); WESTERN (Kongsfjorden – Mercier & Laffly, 2005; Bourriquen et al., 2016; Bellsund – Zagórski, 2011; Zagórski, et al. 2012; Hornsund - Zagórski et al., 2015); and SOUTHERN (Sørkappland – Ziaja et al.,2009, 2011).

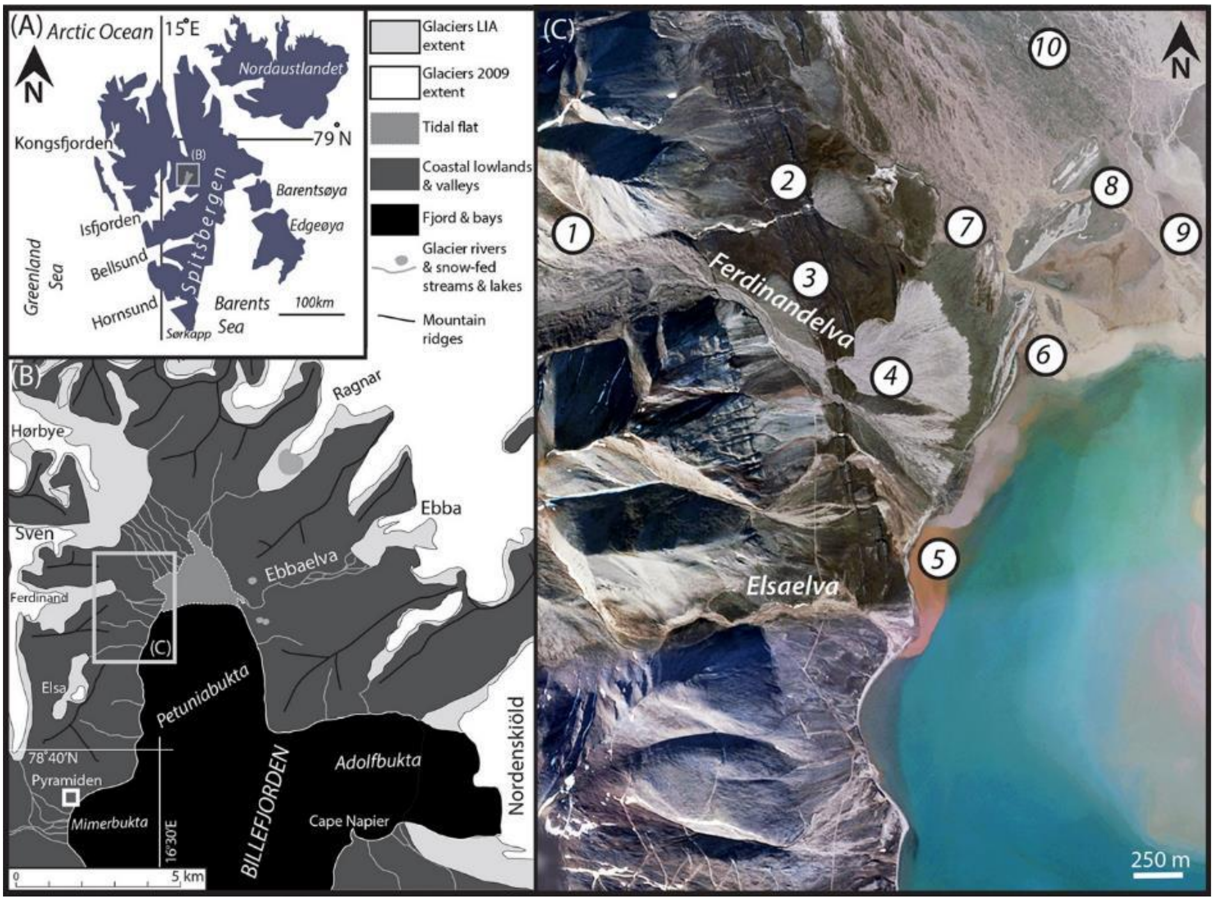

Figure 1. Regional setting. (A) Svalbard Archipelago, (B) Study site: Petuniabukta, Northern Billefjorden, central part of Spitsbergen; (C) Major landforms analyzed in this study: 1 - Ferdinandbreen proglacial zone; 2,3,4 –alluvial fans (Ferdinand Fans) formed between glacier valley and shoreline; 5 - gravel-dominated barrier coast between Elsaelva and Ferdinandelva deltas; 6 - three large spits developed during the post-LIA period (Ferdinand Spits); 7 – uplifted spits (Ferdinand Old Spits); 8 – gravel-dominated remnants of older spits or barrier islands, currently modified by aeolian processes; 9 - prograding front of tidal flat; 10 – outwash-plain formed by glacier rivers draining Svenbreen, Horbyebreen and Ragnarbreen and supplying tidal flat with sediments.

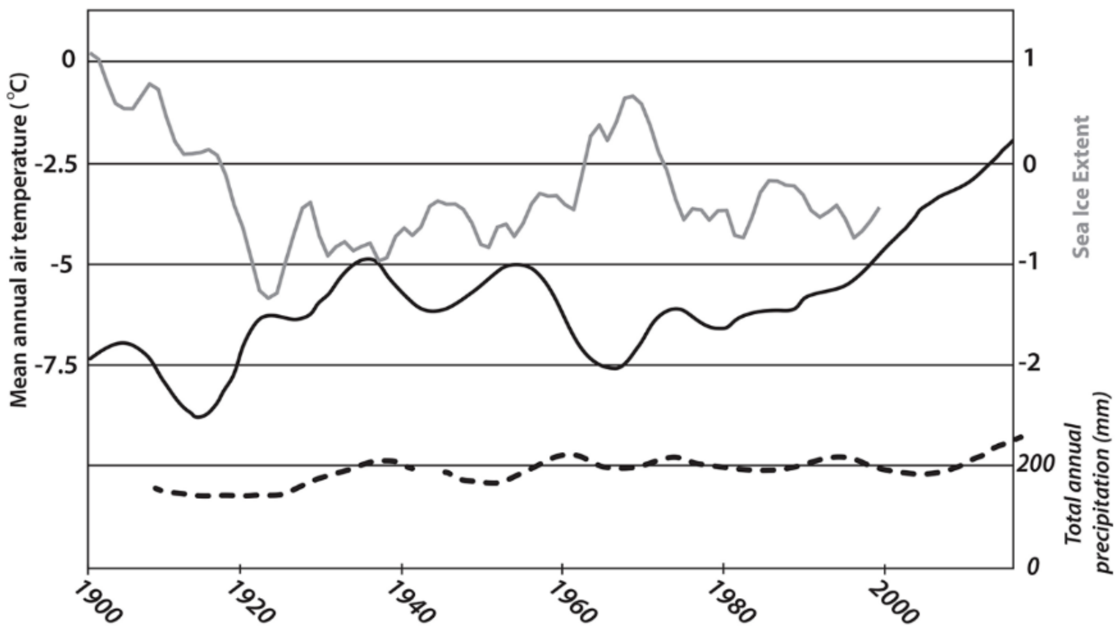

Figure 2. Post-LIA climatic conditions in Svalbard. Grey line - multiproxy reconstruction of sea ice extent in the Western Nordic Seas (1900-1997) modified after Macias Fauria et al. (2010). Black line - the annual mean temperature at the Svalbard Airport Station in years 1902-2016. Black dashed line – the annual total precipitation at Svalbard Airport Station in years 1912 – 2016. Data: Norwegian Meteorological Institute (2017). Air temperature and precipitation in Svalbard, annual mean. Environmental monitoring of Svalbard and Jan Mayen (MOSJ). URL: http://www.mosj.no/en/climate/atmosphere/temperature-precipitation.html

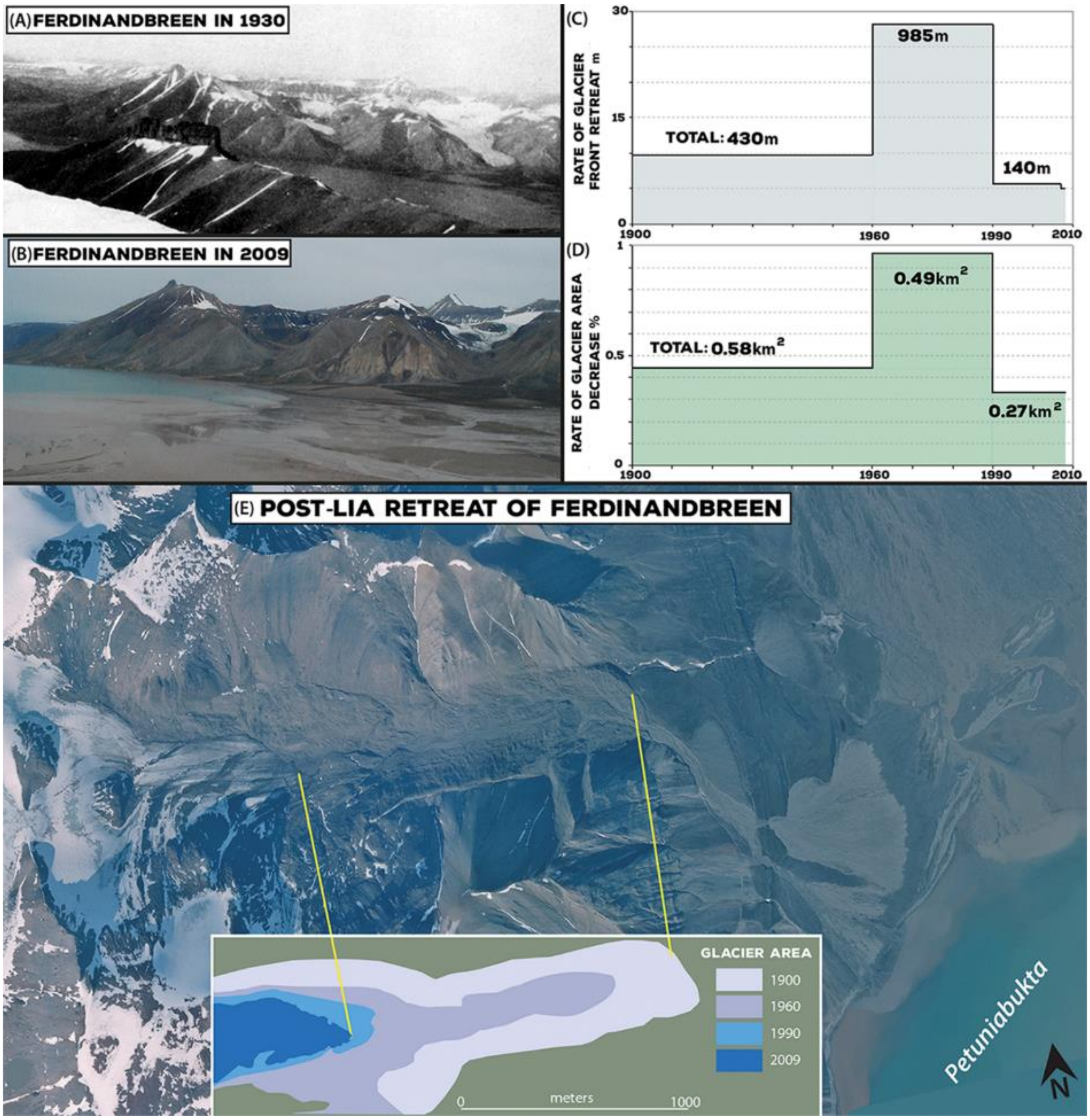

Figure 3. Post-Little Ice Age retreat of Ferdinandbreen. (A) Ferdinandbreen in 1930. Glacier extent was close to the maximum reached during the LIA. Image by Jackson (1931); (B) Ferdinandbreen in 2009; (C) Rate of glacier front retreat since the end of LIA (1900); (D) Rate of glacier area decrease retreat since the end of LIA (1900) modified after Małecki (2016); E) Exposure of valley system by retreating Ferdinandbreen during the post-LIA period. Background: Aerial image taken in 2009 by Norwegian Polar Institute.

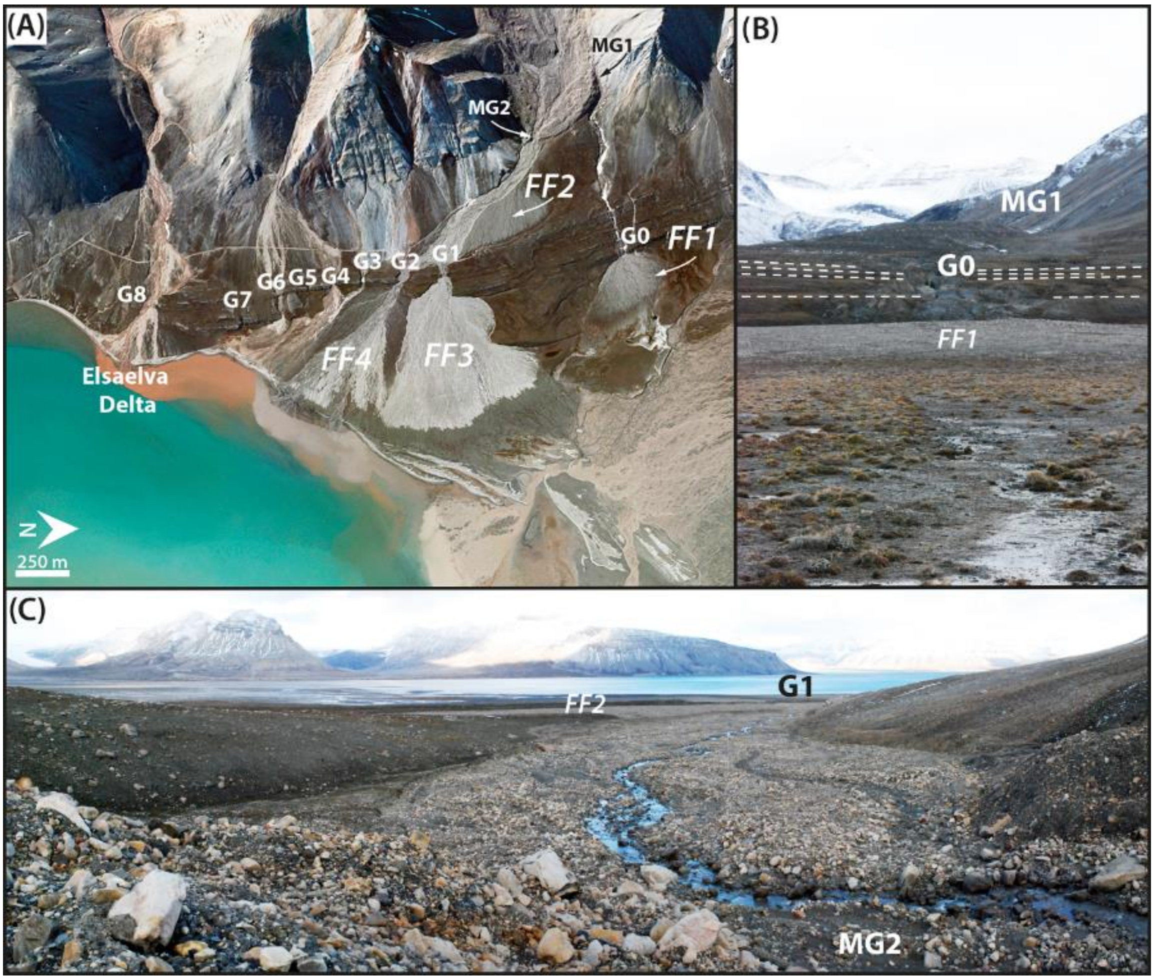

Figure 4. (A) The controls of post-LIA evolution of Ferdinand Fans 1 and 2 (FF1 and FF2) dominated by migration of Ferdinandelva channels and river incision in fan deposits. Sediment supply to the coastal zone between Ferdinandelva and Elsaelva deltas is dominated by snow-melt streams draining through system of bedrock gorges (G2-G8). (B) Area of abandoned Ferdinand Fan 1 supplied by snow-melt streams draining bedrock gorge G0; (C) Present-day river channel eroding moraine gorge (MG2) and draining towards Ferdinand Fans 3-4 trough remnants of Ferdinand Fan 2 and first bedrock gorge G1.

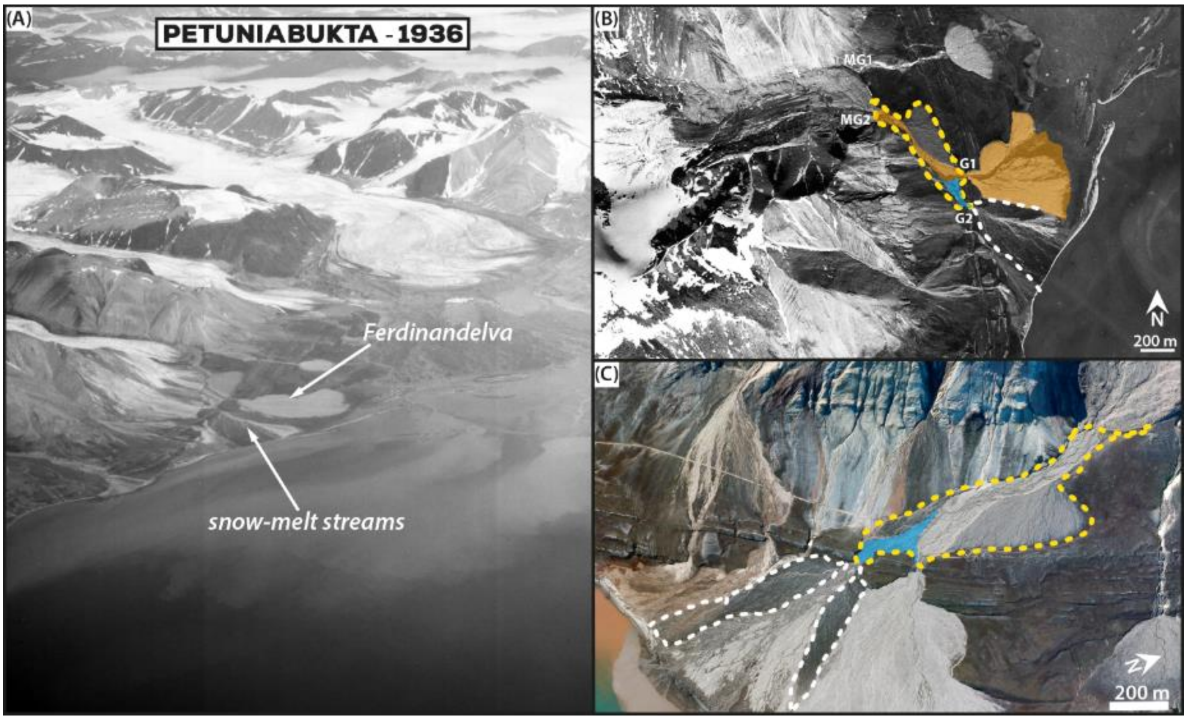

Figure 5. Stages of Ferdinand Fans development during the post-LIA period (A) 1936 – Ferdinandbreen LIA moraine is already breached by river that erodes through remnants of Ferdinand Fan 2 towards the gorge G1 and supplies development of FF3. The area currently occupied by FF4 was supplied by numerous snow-melt streams. Background: Aerial image taken in 1936 by Norwegian Polar Institute; (B) Evolution of channel networks and fan since 1961: yellow dashed line – total area of FF2 which was eroded and incised over the last century; orange zone – area of active Ferdinandelva channel that incised Ferdinand Fan 2 deposits and supplied development of FF3; blue zone – area eroded by Ferdinandelva between 1961-1990 providing the access to bedrock gorge G2 and intensified development of FF4; white dashed line – remnants of old fan eroded and incised during the formation of FF4. Before Ferdinandelva reached the gorge G1 the old fan was supplied by snowmelt-stream from local mountain slopes. Background: Aerial image taken in 1936 by Norwegian Polar Institute; (C) Present-day sediment cascade between glacier and coast. Major channel of Ferdinandelva flows through gorge G2 and supplies development of FF4.

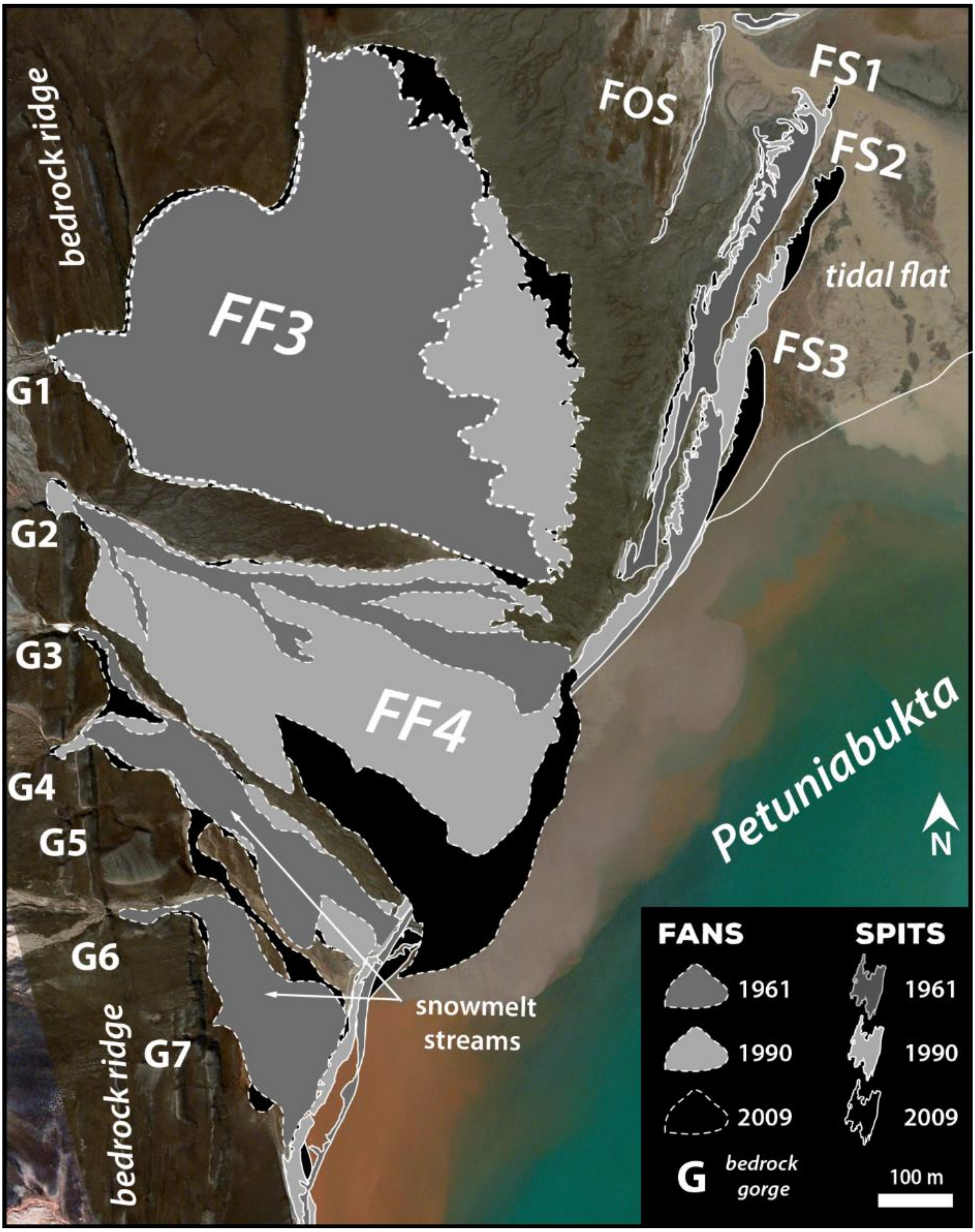

Figure 6. Post-LIA evolution of Ferdinand Fan 3 and 4 (FF3 and FF4) and associated spit systems (Ferdinand Spits FS1, FS2, FS3 and uplifted spits FOS) to a large degree controlled by system of bedrock gorges (G1-G7). Dark grey areas – areas of covered by fans and spits in 1961; Light grey areas – enlargement of fans and spits areas by 1990; Black areas - enlargement of fans and spits areas by 2009.

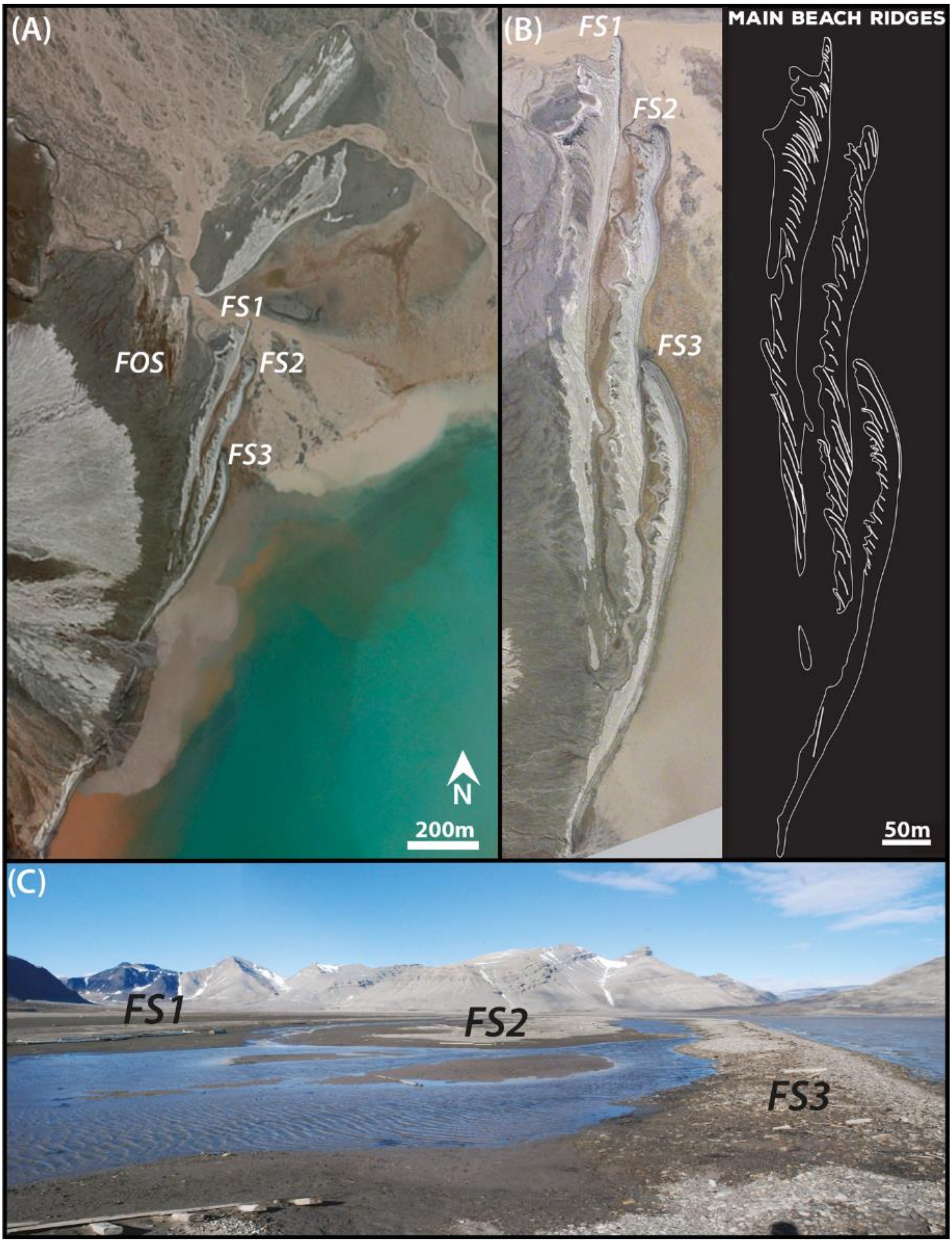

Figure 7. (A) Coastal landscape in NW Petuniabukta dominated by Ferdinand Spit systems. Modern spits: FS1, FS2, FS3 and uplifted spits FOS. Background: Aerial image taken in 2009 by Norwegian Polar Institute; (B) Difference in orientation of Ferdinand Spits and the spatial configuration of the main beach-ridges forming spit hooks; (C) Panoramic view of Ferdinand Spits in 2014. FS1 and FS2 are cut off wave action and their surfaces

are reshaped by aeolian process and limited tidal action. FS3 migrates towards NE following the prograding tidal flat.

Figure 8. (A) The erosion dominated section of Petuniabukta barrier between Elsaelva delta and Ferdinand Fan 4. (B) Modern sand-dominated barriers and spits supplied in sediments from eroded beach-ridge plain; (C) uplifted beach-ridge plain under erosion; (D) Panoramic view of coastal lowland with a barrier-lagoon system between two river deltas (Ferdiandelva and Elaselva).

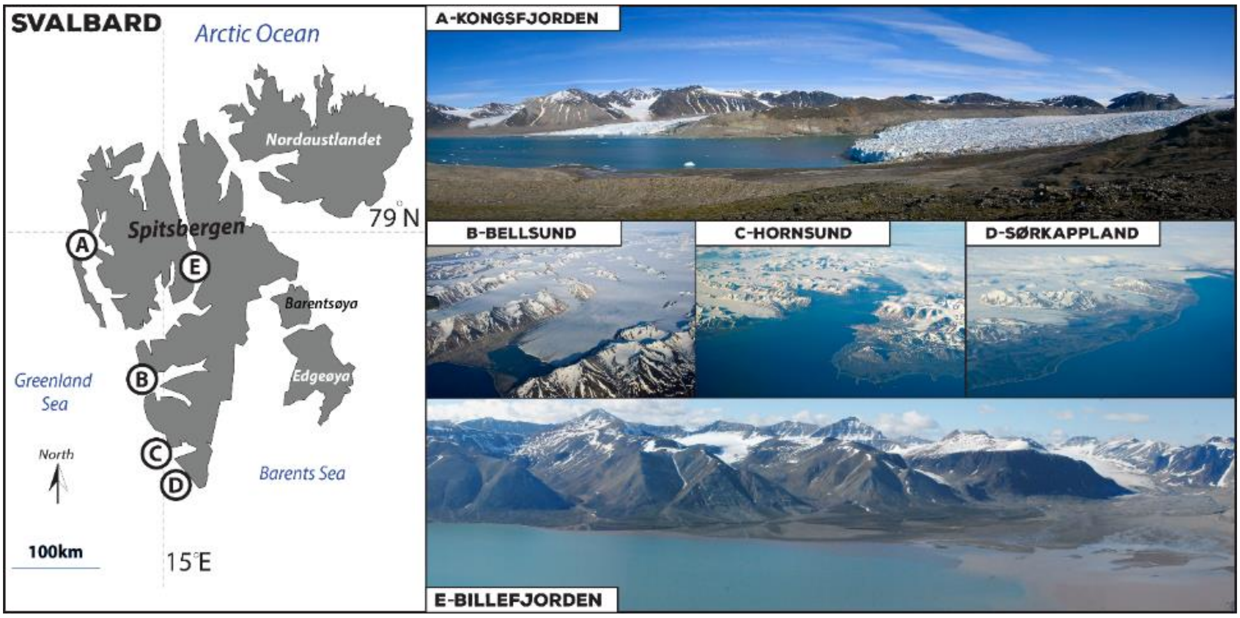

Figure 9. Diversity of paraglacial coastal systems in Spitsbergen. A – coastal zone exposed by post-LIA retreat of Conwaybreen and Kongsbreen in Kongsfjorden; B – Outwash-plain coasts and lagoon developed by retreat of Recherchebreen in southern Bellsund; C – rocky and gravel-dominated barrier coasts exposed during the post-LIA deglaciation in Hornsund; D – barrier coast along western coast of Sørkappland fed by short glacier rivers or direct erosion of glacial landforms left on shore during the retreat; E – mosaic of accumulative coastal systems in northern Billefjorden (study area). Note that western coasts are dominated by presence of tide-water glaciers, relatively short distance between glacier snouts and fjord shorelines and exposure to storm waves from Greenland Sea. Images of glaciers A-D taken by Jürg Alean and Michael Hambrey [https://www.swisseduc.ch/glaciers/].