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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

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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 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 paraglaciacion 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,
overlaid 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 km²) 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 km² of glacier valley (Figure 3). Mass loss was particularly rapid during 1960-1990 when the glacier lost *ca.* 0.02 km² 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; Malecki 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 Elsalva
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 cannibalisation 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 glaciogenic 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.,
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.

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<table>
<thead>
<tr>
<th>Spit</th>
<th>1936</th>
<th>1961</th>
<th>1990</th>
<th>2009</th>
<th>rate of spit expansion [m yr⁻¹]</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>axis length [m]</td>
<td>area [sq m]</td>
<td>axis length [m]</td>
<td>area [sq m]</td>
<td>axis length [m]</td>
</tr>
<tr>
<td>FS1</td>
<td>present</td>
<td>679</td>
<td>210,000</td>
<td>696</td>
<td>18,000</td>
</tr>
<tr>
<td>FS2</td>
<td>embryo formed</td>
<td>270</td>
<td>10,000</td>
<td>431</td>
<td>13,600</td>
</tr>
<tr>
<td>FS3</td>
<td>x</td>
<td>x</td>
<td>315</td>
<td>7100</td>
<td>502</td>
</tr>
</tbody>
</table>

*Table 1. Post-LIA changes (length/area) changes to the Ferdinand Spits. x – landform did not exist in this period so no rate calculated.*
<table>
<thead>
<tr>
<th>Kongsfjorden NW Spitsbergen</th>
<th>Bellsund W Spitsbergen</th>
<th>Hornsund SW Spitsbergen</th>
<th>Sørkappland S Spitsbergen</th>
<th>Petuniabukta Central Spitsbergen</th>
</tr>
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<tbody>
<tr>
<td>Highest observed rates of coastal erosion</td>
<td>up to 30 m (1 m yr⁻¹) [1966-1995]</td>
<td>up to 110 m (1.5 m yr⁻¹) [1936-2007]</td>
<td>up to 46 m (0.9 m yr⁻¹) [1960-1990]</td>
<td>up to 460 m (6.7 m yr⁻¹) [1936-2005]</td>
</tr>
<tr>
<td>Highest observed rates of coastal seaward progradation</td>
<td>up to 120 m (5 m yr⁻¹) [1966-1990] &amp; (4 m yr⁻¹) [2011-2014]</td>
<td>up to 65 m (0.9 m yr⁻¹) [1936-2007]</td>
<td>up to 13 m (0.3 m a⁻¹) [1960-2011]</td>
<td>up to 300 m (20 m yr⁻¹) [1990-2005]</td>
</tr>
<tr>
<td>Highest rates of elongation of coastal landforms (spits)</td>
<td>x</td>
<td>up to 91 m (1.8 m yr⁻¹) [1960-2009]</td>
<td>x</td>
<td>up to 187 m (9.8 m yr⁻¹) FS3 [1990-2009]</td>
</tr>
<tr>
<td>Glacier retreat rates (LT): land-terminated (TW): tide-water and mean distance between snout and coast (SCD)</td>
<td>(LT) Midtreløvenbreen: (10 m yr⁻¹) • SCD: 1200 m</td>
<td>(LT) Scottbreen: (16 m yr⁻¹) [1936-2002] • SCD: 2300 m</td>
<td>(LT) Aribreen: (10 m yr⁻¹) • SCD: 2200 m</td>
<td>(LT) Kambreen: (12 m yr⁻¹) [1900-2005] • SCD: 600 m</td>
</tr>
<tr>
<td></td>
<td>(TW) Kronebreen (150 m yr⁻¹)</td>
<td>(TW) Recherchebreen (27 m yr⁻¹) [1960-2008]</td>
<td>(TW) Hansbreen: (25 m yr⁻¹) [1900-2008]</td>
<td>(TW) Hambergbreen: (160 m yr⁻¹) [1900-2000]</td>
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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 - Ferdinanbreen 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 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 Elsaer delta 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. (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/].