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Ground-penetrating radar (GPR) investigations of a large-scale buried ice-marginal landsystem, Skeiðarársandur, SE Iceland

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The sedimentary record of Icelandic ice-contact environments provides critical insights into past glacier margin dynamics and position, relative sea level, and the geomorphic processes that drive the evolution of proglacial environments. This important archive has been little exploited, however, with most glacier and sea-level reconstructions based on limited sedimentary exposures, coring and surface geomorphic evidence. We report an extensive (42 km of data within a 24-km^2 study area) and deep (reflections recorded at depths up to 100 m) low-frequency (40 and 100 MHz) ground-penetrating radar (GPR) survey of the Sandgígur moraines, SE Iceland. GPR profiles reveal a much larger (67 m high) and extensive (1.25 km wide) buried moraine ridge than that suggested by surface topography (typically 125 m wide and 7 m high). These data reveal that the Sandgígur moraines was deposited during a major Holocene re-advance of Skeiðarárjökull. The moraine ridge is buried by sediments dominated by glacifluvial deposits with an estimated sediment volume of 1.04 km^3. We combine GPR-derived subsurface architecture and the surface morphology to develop a conceptual model detailing the geomorphic evolution of the moraine and surrounding region. These results provide new insights into the Holocene evolution of Skeiðarársandur, identifying the presence of a former major ice-margin position, as well as a past relative sea-level limit. Furthermore, we establish that sediment supply and available terrestrial accommodation space are dominant drivers in the formation and evolution of vast sandar environments.

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Sandur plains (sandar) are depositional systems beyond the margin of a glacier or ice sheet, typically located in the present-day Arctic and common within the Quaternary record of low latitudes. The development and evolution of these depositional systems are primarily controlled by fluctuations in the mass balance of their feeding ice mass (Krigström 1962; Þórarinsson 1972; Maizels 1983). However, the topographic setting of sandar, tectonic processes and changes in global sea level also play an important role in their evolution (Maizels 1993a). Despite considerable research (Krigström 1962; Church 1972; Maizels 1993a; Gomez *et al*. 2000; Marren 2002a, b; Zieliński & van Loon 2002, 2003), the primary controls on the formation and evolution of sandur environments during glaciation and rapid deglaciation remain to be established. The controls of sandur sedimentation and geomorphological reworking can be broken into active or passive factors (Weckwerth *et al.* 2021). Active factors include: (i) changing climate conditions and the resultant glacier margin fluctuations; and (ii) variations in the hydrology of proglacial fluvial systems, especially related to the magnitude and frequency regime of the meltwater system (e.g. flood events vs. non-flood events) (Marren 2005; Carrivick & Heckmann 2017; Strzelecki *et al*. 2018; Kociuba *et al*. 2019; Weckwerth *et al*. 2021). Passive factors influencing sandur sedimentation are generally regarded as: (i) long-term geomorphic processes (e.g. location and distribution of existing proglacial fluvial systems); (ii) calibre of the prevailing sediment (e.g. gravel, sand or clay); and (iii) changes in the geology of the catchment (Weckwerth *et al*. 2021).

Some of the best examples of sandar are those in Iceland. It has been hypothesized that Icelandic sandar comprise thick alluvial successions containing detailed records of the events that contributed to their formation (e.g. Marren 2005). However, surprisingly limited work has been undertaken to analyse the large-scale sedimentary architecture of these systems, and to reconstruct how they have evolved during the Holocene. It can be argued that more is known about the large-scale architecture of palaeo-sandur and glacigenic depositional environments from ancient glaciations (e.g. Ghienne & Deynoux 1998; Ghienne *et al*. 2010; Girard *et al*. 2012) and Quaternary glaciations (Brennand 1994; Brennand & Shaw 1994; Tuttle *et al*. 1997; Winsemann *et al*. 2009, 2018; Lang *et al*. 2021), than is known about contemporary outwash plains. However, inferences from the sedimentary record lack temporal constraint on the processes that have influenced sandur development.

Glacier margin fluctuations are known to generate proglacial fluvial system responses in the form of sandur morphological change. Changes in glacier mass balance influence meltwater hydrology and its ability to modify sandur environments. This, in turn, is dependent on the coupling of the glacial source to the proglacial depositional system (Marren 2005; Weckwerth *et al*. 2011, 2019, 2021; Knight & Harrison 2014, 2018; Carrivick &

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Heckmann 2017; Kociuba *et al.* 2019). Aggradation and a steepening of sandar are seen to occur during periods of glacier margin stability or advance where the glacier is coupled to the sandur (Maizels 1979; Marren 2002b). In contrast, during periods of glacier retreat, where the glacier is decoupled from the sandur plain, vast braided networks transition into singular entrenched channels with increasing terrace formation, as incision dominates (Galon 1973a, b; Klimek 1973; Marren 2002b; Marren & Toomath 2013). The rates of incision and sandur erosion are controlled by extreme precipitation-driven ablation events which impact short-term changes in the meltwater system (Marren 2005; Beylich & Laute 2015; Lane *et al.* 2017; Guillon *et al.* 2018; Kociuba & Janicki 2018; Weckwerth *et al.* 2021). Retreating ice margins can also lead to the development of lateral drainage networks in topographic lows formed following ice retreat from the previously aggraded sandur surface (Price 1969; Bogacki 1973; Galon 1973a, b; Klimek 1973; Gomez *et al.* 2000).

The relationship between ice-margin fluctuations and the related aggradation or incision of the sandur, coupled with the sediment supply of the glacial meltwater systems and occurrence of high magnitude flood events leads to a complex sediment stratigraphy (Maizels 1979, 1983; Marren 2002b, 2005). Early sandur models (e.g. Boothroyd & Nummedal 1978) do not capture this complexity and predominately represent a snapshot of the sandur surface. Over longer time scales this may be repeated vertically. However, fluctuations in ice-margin position, meltwater outlets and sediment supply may lead to variations in the thickness and lateral extent of the overlying deposits (Marren 2004). The inherent relationship between glacier margin fluctuations and sandur formation means that sandur plains have considerable potential to provide insight into past ice mass history, meltwater dynamics, and sedimentation, over recent geological time scales (e.g. the Holocene).

The sandar of Iceland represent the currently unrealized potential to ascertain the controls on large-scale proglacial sedimentation. The terrestrial history of Icelandic glaciations and their impact on the evolution of the surrounding landscape has largely been based on interpretation of recent events from sedimentary exposures and geomorphological features within the near surface (<20 m). Relatively little is known about the longer-term Holocene glacial history of southeast Iceland, particularly the Late Holocene to Little Ice Age (LIA) (AD 1300–1900) glacial history, which remains poorly constrained, and hypothesized neoglacial activity (1.5–4.2 ka) remains to be tested (Gudmundsson 1997; Geirsdóttir *et al.* 2019). The evolution of the ice-marginal environments of Iceland since the LIA has been extensively studied. This Iceland-focused work has driven the development of the active temperate glacial land system (Evans *et al.* 1999; Evans & Twigg 2002; Chandler *et al.* 2020), which has been used to inform interpretation of ice-marginal dynamics and processes for Quaternary glaciation (e.g. Evans *et al.* 1999). Icelandic-derived glacial land system models, however, include little information about the large-scale sedimentary architecture of the proglacial environments. This limits their application to the understanding of the impact of changes in base-level, accommodation space and sediment flux for sandur evolution.

It is widely assumed that Icelandic sandar are the product of multiple high magnitude jökulhlaups over the last ~10 ka following rapid deglaciation from the Younger Dryas maximum (Maizels 1991, 1993a, 1997; Guðmundsson *et al.* 2002). Sandur systems prone to large magnitude episodic jökulhlaups (e.g. Iceland) produce a distinctive facies succession that is dependent upon the nature of the jökulhlaup (e.g. exponentially-rising or linearly-rising stage hydrograph shape) and the available sediment supply (Maizels 1989, 1993b, 1997; Rushmer *et al.* 2002; Roberts 2005; Rushmer 2006, 2007). In contrast, sandar dominated by relatively low magnitude seasonally-produced meltwater (i.e. ablation-controlled) tend to produce a suite of sedimentary facies driven by channel migration and the formation of large gravel bars and in-channel large-scale dunes (Allen 1983; Miall 1985; Bristow 1996; Marren 2005).

The large-scale sedimentary architecture of sandar environments can be characterized, and its potential unlocked, by borehole or geophysical observations (e.g. Kurjański *et al.* 2021). However, such observations constraining the subsurface of Icelandic sandar are sparse and are typically based on shallow near-surface (up to 10 m depth) sediment exposures. A lack of widespread geophysical observations is surprising, given that Icelandic sandar are highly suitable for ground-penetrating radar (GPR) surveys due to limited vegetation coverage, the lack of soil development and the 'GPR-friendly' physical properties of outwash sediments (i.e. sand and gravel; Neal 2004; Reynolds 2011). These factors are conducive to extensive and deep geophysical characterization of the subsurface. To-date, the majority of GPR surveys conducted in Iceland have focused on specific landforms (e.g. Kjær *et al.* 2004; Carrivick *et al.* 2007; Burke *et al.* 2008, 2009, 2010a, b; Blauvelt *et al.* 2020), and surveys targeting the sedimentary architecture of recently emplaced proglacial jökulhlaup deposits (e.g. Russell *et al.* 2001b, 2006; Cassidy *et al.* 2003; Harrison *et al.* 2019). The influence of glacier margin fluctuations, accommodation space and base-level changes on sandur-wide sedimentation cannot be extracted from the spatially restricted nature of existing studies.

Poor knowledge of the large-scale subsurface architecture of sandur environments limits our ability to constrain the driving mechanisms (e.g. glacier growth and decay, accommodation space, sediment supply, relative sea level) in how these vast depositional systems are formed and evolve during glacial cycles. Therefore, detailed analysis of the sedimentary architecture of modern large-scale unconfined outwash systems and how these reflect important drivers such as glacier recession, sea-level change and episodic sediment flux from enormous high-magnitude glacier outburst floods (jökulhlaups) is paramount. Characterization of sandur systems in varying topographic and glaciological settings is crucial to aid the development of more sophisticated models of sandar sediment architecture (Marren 2004).

This paper aims to: (i) evaluate the impact of glacial fluctuations and glacial sedimentation on sandur formation; (ii) establish the primary controls on the development of sandar sedimentary architecture; and (iii) develop a conceptual model for the formation and evolution of sandar environments. To fulfil these aims we characterize the large-scale sedimentary architecture of a large-scale active sandur in southeast Iceland using extensive low-frequency GPR surveys.

Skeiðarársandur and the Sandgígur moraines

Skeiðarársandur, southeast Iceland, is the largest (~1300 km^2) active proglacial outwash system in the world (Fig. 1). The sandur is fed by meltwater systems emanating from the margin of Skeiðarárjökull, a 1400-km^2 surge-type outlet glacier of the Vatnajökull ice cap (Björnsson *et al*. 2003; Roberts 2005; Magnússon *et al*. 2011). Skeiðarársandur is believed to have formed entirely during the Holocene (Maizels 1993a; Guðmundsson *et al*. 2002). During the Last Glacial Maximum (LGM) the majority of Iceland was ice covered, and glaciers likely terminated off the coast of southern Iceland until ~10.3 ka (Norðdahl 1990; Ingólfsson 1991; Maizels 1993a; Guðmundsson *et al*. 2002; Norðdahl *et al*. 2008). The formation of Skeiðarársandur is assumed to be primarily driven by high magnitude jökulhlaups (Þórarinsson 1974; Russell & Marren 1999; Guðmundsson *et al*. 2002; Marren 2002a, b; Russell *et al*. 2006), with large depositional units of outburst flood sediments predicted to represent up to 94% of the total sandur thickness (Maizels 1993a). However, this numerical model derived prediction is poorly constrained by observations beyond historical records (e.g. Ives 2007).

Seismic refraction investigations of Skeiðarársandur suggested a three-layer subsurface sequence of bedrock, overlain by over-consolidated sediments and unconsolidated sediments of a glacifluvial origin (Guðmundsson *et al*. 2002). The volume of sediment above bedrock across the sandar was estimated at 100–200 km^3 (Guðmundsson *et al*. 2002). The uppermost unconsolidated sediments, thought to be Holocene in age, increase in thickness from 80 m in proximal regions to 150 m in distal regions of the sandur, and have an estimated volume of up to 100 km^3, equating to a volumetric sedimentation rate of ~1 km^3 per century (or 0.01 km^3 a^{-1}) (Guðmundsson *et al*. 2002). Such estimations of sediment volume and sedimentation rates at Skeiðarársandur should be viewed with caution, however, due to the limited spatial extent of the seismic data collected and the lack of any borehole constraints on the nature of the sediments (Guðmundsson *et al*. 2002). Jökulhlaup sediment yield during a large outburst flood at Skeiðarársandur in November 1996 was 1.8×10^8 m^3 ($\geq3.1\times10^{11}$ kg) equating to a sediment flux of 1.8×10^6 kg s^{-1} during the 47-hour-long flood (Smith *et al*. 2000; Roberts 2005; Russell *et al*. 2006). The Holocene sedimentation rate proposed by Guðmundsson *et al*. (2002) roughly equates to five 1996-equivalent jökulhlaups per century (Guðmundsson *et al*. 2002; Russell *et al*. 2006).

The topographic relief of the western part of Skeiðarársandur is dominated by two moraine complexes: (i) the large push-moraine complex best defined where it is dissected by the Gígjukvísl river herein referred to as the 'Gígjukvísl' moraines; and (ii) the Sandgígur moraines located 2 km further down-sandur, which comprise a set of linear, ice-margin parallel, isolated ridge fragments (Fig. 1).

Skeiðarárjökull is believed to have reached its Little Ice Age (LIA) maximum position in 1784, 1857, 1871 and 1890–95 (Jóhannesson 1985). The first three of these LIA maximum positions coincide with glacier surges in 1787, 1857 and 1873 (Björnsson *et al*. 2003) with the 1890–95 position corresponding with the maximum LIA extent across the southern and southeastern margins of Vatnajökull (Björnsson & Pálsson 2008). Þórarinsson (1939) suggested that the Sandgígur moraines represent the maximum LIA extent of Skeiðarárjökull in ~1750 (Fig. 1C). Þórarinsson (1939), however, does not provide evidence to support this interpretation, which is at odds with historic accounts suggesting the LIA maximum extent of Skeiðarárjökull in the vicinity of the Gígjukvísl moraines was reached in 1784, 1857, 1871 and 1890–95 (Jóhannesson 1985; Grove 2004). Sedimentary evidence suggests that at least three glacier advances were required to form the Gígjukvísl moraine complex (Molewski & Olszewski 2000) providing further support to the notion that Gígjukvísl moraines represent the maximum LIA extent of Skeiðarárjökull. Given that the Gígjukvísl moraines represent the LIA maximum, the Sandgígur moraines have to be pre-LIA in age. That the Stóralda moraine at nearby Svínafellsjökull is ascribed to a neoglacial (4.5–1.5 ka) ice margin (Gudmundsson 1997) supports the interpretation of the Sandgígur moraines as representing a pre-LIA neoglacial frontal position of Skeiðarárjökull.

The Sandgígur moraines are broken into six segments which have a relatively subtle surface geomorphic expression (typically 125 m wide and 7 m high; Fig. 1C). The total extent of the moraine fragments is 3.5 km with each segment ranging between 150–650 m long. Information regarding the formation of the moraines is limited, with no detailed geomorphic or sedimentological investigations present in the literature.

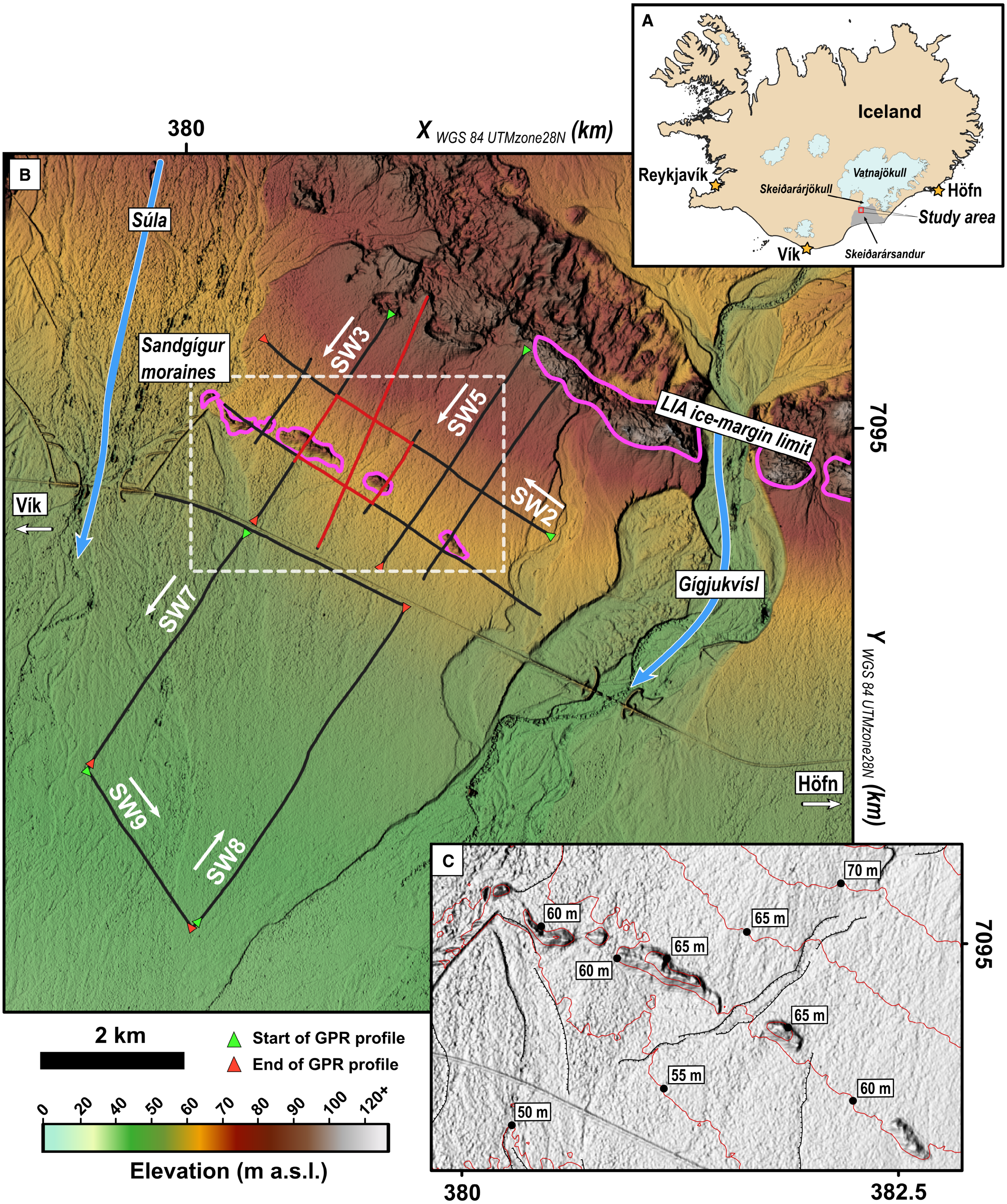


Fig. 1. Location of Sandgígur moraines study area and the extent of the GPR survey. A. Map of Iceland. B. Location of GPR survey in western Skeiðarársandur in the vicinity of the Sandgígur moraines. C. Hillshaded DEM of the Sandgígur moraines. Black lines in (B) highlight low-frequency (40-MHz) GPR profile locations. Red lines in (B) indicate locations where coincident 40-MHz and 100-MHz radar lines were collected. White dashed box in (B) shows the location of (C). Contours displayed at a 5-m interval in (C) and are with reference to elevation in metres above sea level (m a.s.l.). Black dashed lines in (C) indicate obvious terraces. Background DEM and hillshaded DEM in (B) and (C) are ArcticDEM 5×5 m tile products (Porter *et al.* 2018) converted to metres above sea level (m a.s.l.).

Material and methods

A 40-MHz UTSI Groundvue 7 (GV7) low-frequency ground-penetrating radar (GPR) system (Francke & Utsi 2009; Harrison *et al*. 2019) was used to collect radar profiles (~42 km of along-track measurements) of the large-scale subsurface architecture (Fig. 1B). The GV7 has a centre frequency of 40 MHz and an emitted bandwidth of 10–80 MHz. Further technical specifications of the GV7 system are reported in Ross *et al*. (2018). Coincident 100-MHz GPR profiles (using a separate GV7 system) were also acquired collected at specific locations (Fig. 1B) to provide higher resolution subsurface information. Positional data were acquired with a handheld GPS streamed into the GPR recording software. WARR (wide angle reflection and refraction) surveys suggest radio-wave velocities between 0.13–0.19 m ns^{-1} for the sediments at the Sandgígur moraines. We therefore apply a conservative velocity of 0.13 m ns^{-1} for time depth conversion, which corresponds with the maximum velocity for regions comprised of unsaturated sands and gravel (Neal 2004; Reynolds 2011). This is supported by the observation of the spring line intersecting the sandur surface in more medial-to-distal reaches of the sandur plain at an elevation of ~37 m a.s.l., well below the elevation of our survey (50–80 m a.s.l.). Based upon our radio-wave velocity and the centre frequency of the 40 MHz GPR we achieve a theoretical vertical resolution of 0.8125–1.625 m. This limits our ability to resolve sediment structures and/or units less than 1.625 m in thickness.

Radargrams were processed using standard processing steps for low-frequency GPR data (Fig. 2) and interpreted in Reflexw geophysical software (http://www.sandmeier-geo.de/reflexw.html). Topographic information used for elevation correction during the processing of GPR profiles was extracted from ArcticDEM tiles (Porter *et al*. 2018) to radar profiles trace headers, after having been converted from WGS 84 ellipsoid to metres above sea level (m a.s.l.) (Fig. 3). Processed data were then imported into OpendTect seismic interpretation software for 3D visualization (https://www.opendtect.org/). Radar stratigraphical descriptions of radar reflections are taken from Neal (2004). Picks of large-scale bounding surfaces were imported into ArcGIS where the Topo-to-Raster tool was used for digital elevation model (DEM) generation. Topo-to-Raster uses the ANUDEM algorithm, which is considered the most appropriate when aiming to preserve stream networks and ridge lines during interpolation (Hutchinson 1989; Arun 2013). DEM differencing between surface and subsurface elevation models was then conducted and volumetric calculations were performed.

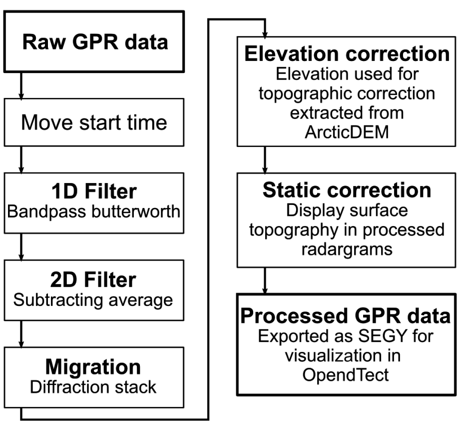


Fig. 2. Processing flow used for 40-MHz GPR data. Processed radargrams displayed in Fig. 3 (uninterpreted) and Fig. 4 (interpreted).

Results

The GPR data (Figs 3, 4) reveal that the Sandgígur moraines have a complex and chaotic stratigraphy, with the large-scale architecture characterized by strong continuous radar reflections (Fig. 4). The radargrams indicate numerous different radar facies that are linked to different depositional styles. Radar reflections were identified at depths of up to 100 m. Radar profiles presented here are representative of the large-scale architectural features and radar facies within the subsurface across the Sandgígur moraines survey area (Fig. 1). Three major contacts and/or boundaries that delineate changes in depositional setting were identified across the GPR survey and are referred to as reflections RS0, RS1 and RS2.

Radar surfaces (RS)

The RS0 reflection is the lowermost major interface that we identify in the GPR data. This reflection could not be traced in all radargrams and is most prominent where it is located at shallower depths in more distal areas of the survey (Fig. 4). In radargrams where RS0 is identifiable it is continuous and coherent. The RS0 reflection reaches a maximum elevation of −30 m a.s.l. (depth of 80 m), and a minimum elevation of −50 m a.s.l. (depth of 90 m) (Fig. 4E). The reflection is of a low amplitude and is subhorizontal in lines collected in a down-sandur (NE–SW) orientation. Lines across the sandur (NW–SE) show that RS0 has a variable morphology. For example, in the most distal radargram (Fig. 4E), RS0 is characterized by a subhorizontal reflection that transitions into a broad (~750 m) trough or depression 1.25–2 km along the radargram. A ~120-m-wide and ~12-m-high ridge is located within the depression 1.4 km along the survey line. The 3D form of both the trough and the ridge are poorly constrained due to a lack of intersecting radargrams.

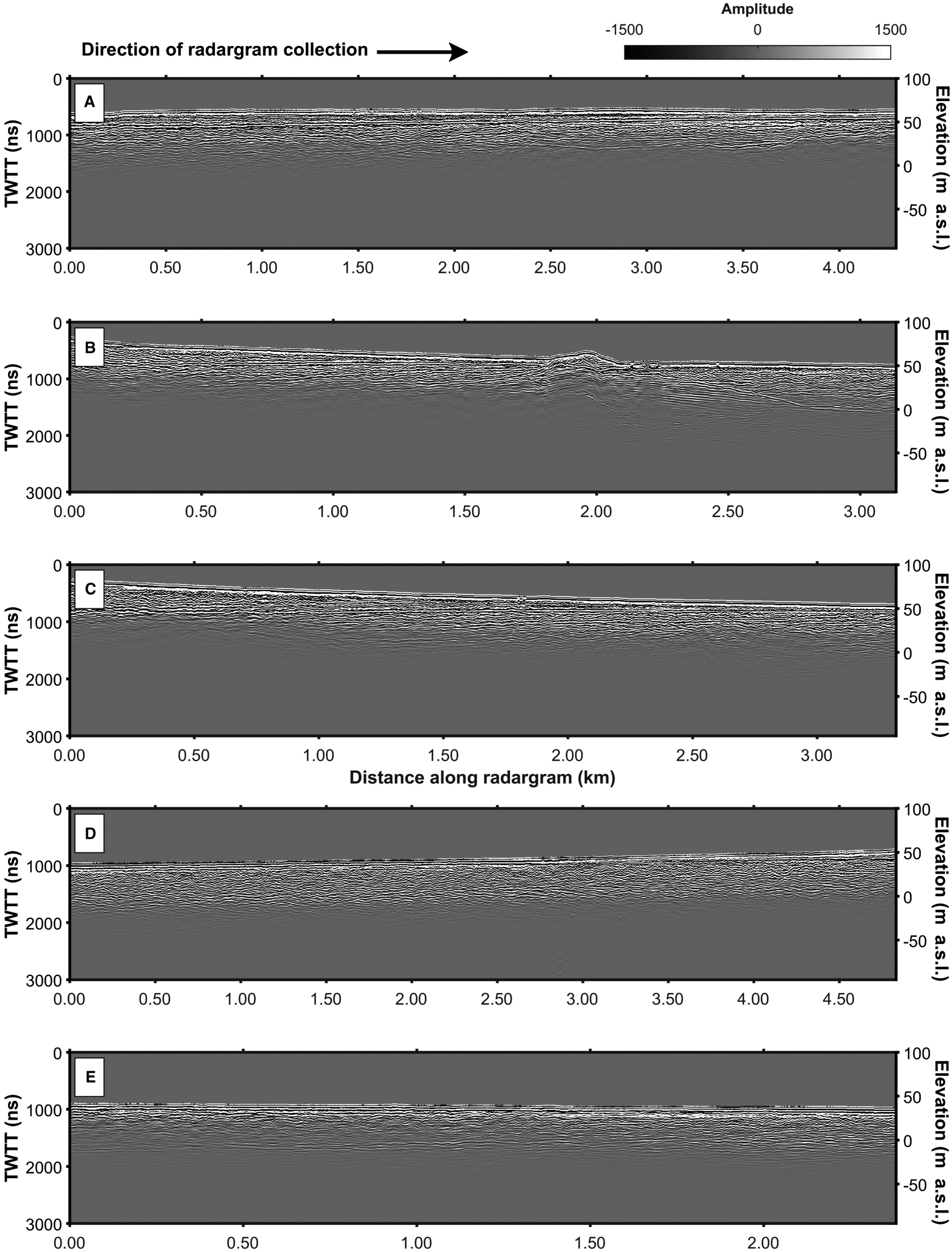


Fig. 3. Processed unpicked radargrams. Profile locations and start and end points can be seen in Fig. 1B. Interpreted radargrams are shown in Fig. 4. A. SW2. B. SW3. C. SW5. D. SW8. E. SW9.

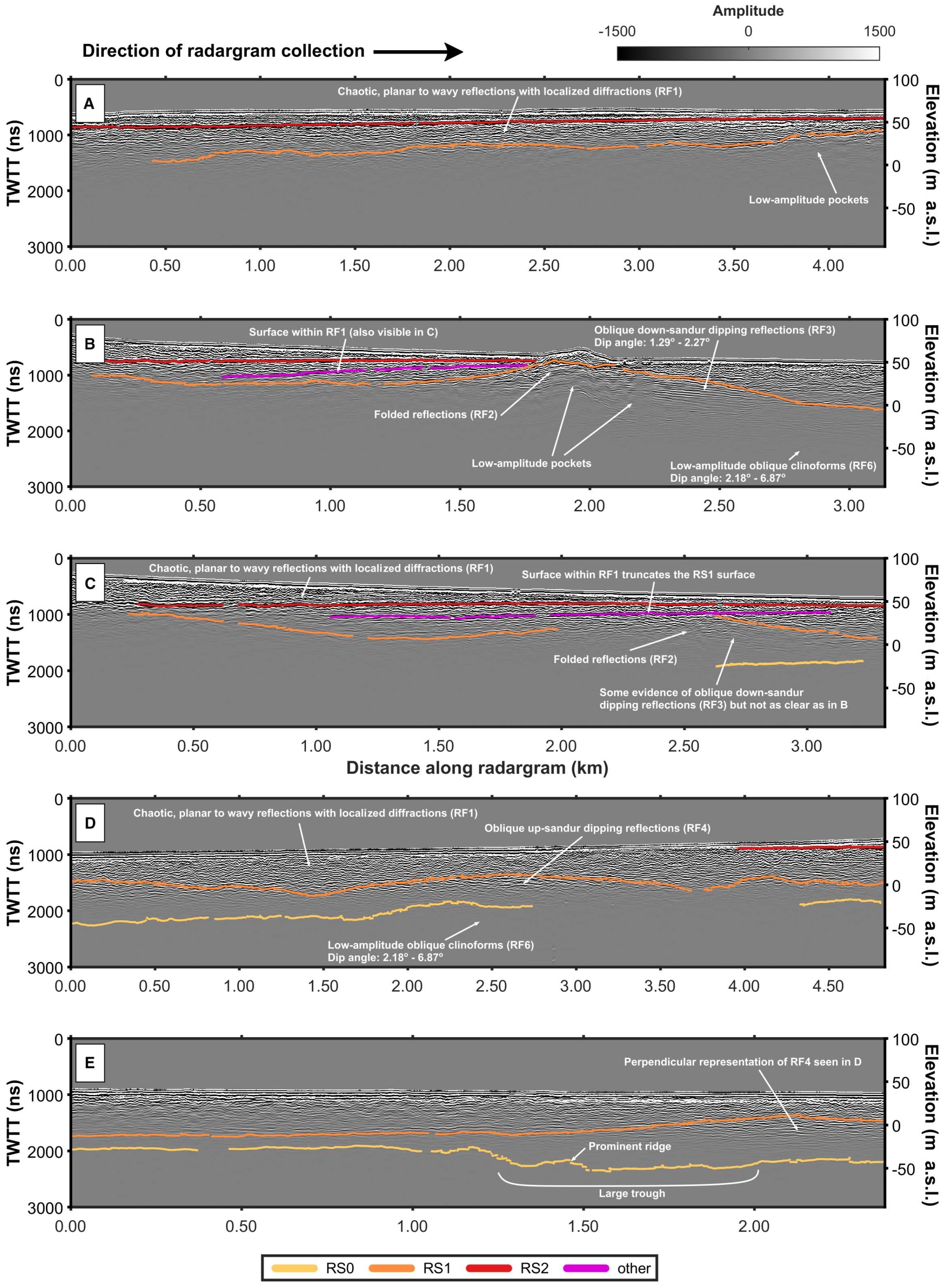


Fig. 4. Radargrams indicating the major bounding reflections within the GPR survey area. Profile locations and start and end points can be seen in Fig. 1B. Unpicked radargrams are shown in Fig. 3. A. SW2. B. SW3. C. SW5. D. SW8. E. SW9.

The most extensive bounding reflection within the survey area is RS1. This reflection was identified in all profiles. The extent and continuity of RS1 vary between profiles, however; it is strongest and most coherent in radargrams near the Sandgígur moraines (e.g. Fig. 4B). This bounding surface reaches a maximum elevation of 52.4 m a.s.l. (depth of ~10 m) and a minimum elevation of −15.4 m a.s.l. (depth of ~59 m). The morphology of RS1 is characterized by a prominent ridge that coincides with the surface of the Sandgígur morainic ridges (Fig. 4B). RS1 is often truncated in radar data acquired between the moraine ridges and is capped by chaotic reflections (Fig. 4C). The peak elevation of RS1 occurs beneath the Sandgígur moraines, lowering to both the north and south of the moraine. Analysis of the radargrams reveals that RS1 is an integral part of the Sandgígur moraines, with the reflection continuing beneath the surface expression of the moraine and extending up and down-sandur beyond it (Fig. 4B). Further down-sandur RS1 varies from a mostly subhorizontal reflection to a more hummocky topography, particularly in the more eastern sections of the radar survey (Fig. 4D).

Reflection RS2 is a coherent, high amplitude, continuous reflector that is traceable in all profiles immediately surrounding the moraine ridges (Fig. 4A–C). The reflector onlaps onto parts of the lower bounding reflection RS1 (Fig. 4B) and in some places extends contiguously down-sandur through gaps in the moraine where the RS1 reflector cannot be traced to the surface (Fig. 4C, D). The reflection intersects the sandur surface ~1 km down-sandur of the Sandgígur moraines (Fig. 4D). RS2 cannot be interpreted as the water table as it truncates lower reflections and where it pinches out to the sandur surface no surface water is present.

Sediment volume and thickness

DEMs of the extensive bounding surfaces RS1 and RS2 provide estimates of sediment thickness and sediment volume for locations surrounding the radar survey. Average sediment thickness above RS1 is 43 m, with a maximum thickness of 68.5 m and minimum of 7.9 m (Fig. 5C). Sediment thickness is similar both up-sandur and down-sandur of the pronounced ridge (e.g. location of the Sandgígur moraines; Fig. 5C). However, up-sandur of the Sandgígur moraines RS1 is located at an elevation of 10–30 m a.s.l., whereas RS1 is mostly below 0 m a.s.l. in the down-sandur direction (Fig. 5). The relief of the subsurface ridge associated with RS1 is 40–50 m and extends in width for ~1.25 km (Fig. 5). The volume of sediment between the sandur surface and the RS1 boundary is approximately 1.04 km^3 over an area of 24.09 km^2 (Fig. 5). This equates to 0.043 km^3 of sediment per 1 km^2.

The average sediment thickness above RS2 is 14.9 m, with a maximum of 50.4 m and a minimum of 1.7 m (Fig. 6). The morphology of RS2 is fan-shaped and generally dips in a NW–SE direction, parallel with the Sandgígur moraines (Figs 4A, 6), with its apex located at or close to the SW2 radargram. The slope of the long profile of the RS2 bounding surface is relatively low (0.15°). The thickness of sediment bounded by RS2 and the sandur surface increases along the long profile of RS2, ranging from 10 m at the fan apex to 17 m at the end of the profile (Fig. 6D). In profiles perpendicular to the Sandgígur moraines (Fig. 6E, F) the thickness of this unit decreases from ~30 m at the LIA limit to ~10 m in the proximity of the Sandgígur moraines. This trend of decreasing sediment thickness in a down-sandur direction is evident from the elevation difference DEM produced for RS2 (Fig. 6C). The volume of sediment above RS2 is calculated at 0.13 km^3 over an area of 9.45 km^2 where the RS2 surface is visible in the radar data (Fig. 6). This equates to 0.014 km^3 per 1 km^2.

The volume estimates we generate are highly dependent on the radar velocity of 0.13 m ns^{-1} being accurate for the sediments around the Sandgígur moraines. Using a radar velocity of 0.08 m ns^{-1}, similar to GPR surveys conducted elsewhere in Skeiðarársandur (Burke *et al.* 2008, 2010a, b), would alter the volume estimates by roughly 38%. For example, at a radar velocity of 0.08 m ns^{-1} sediment volume above RS1 would be 0.64 km^3. Therefore, our estimates of sediment volume above RS1 and RS2 should be taken as maximum estimates.

Radar facies (RF)

Radar facies 1 (RF1). – RF1 (RS1 to RS2) is the most dominant radar facies identified. The entirety of this facies is located above RS1 (Fig. 7). Reflections within RF1 down-sandur of the Sandgígur moraines are generally more chaotic and less continuous (Table 1). Up-sandur of the Sandgígur moraines, reflections are more continuous with fewer diffractions. In radargrams orientated NE–SW (e.g. SW3 and SW4) numerous reflections can be seen onlapping onto the RS1 surface (Figs 4B, 7). Smaller structures (<75 m long and <5 m thick) are present between the more continuous reflections and have a more wavy/undulating geometry (Table 1). However, the low-frequency nature of the 40-MHz GPR system means the true geometry of small-scale structures cannot be fully resolved in these data. Trough-like structures are evident within RF1 and are orientated in both NE–SW and NW–SE directions (Fig. 8). The largest of these trough structures is located down-sandur of the Sandgígur moraines in radargram SW7 (Figs 7, 8). This trough, roughly 700 m wide and 25 m deep (Fig. 8), is orientated transverse to the active fluvial systems of the Súla and Gígjukvísl. Possible trough structures are also present in radargrams SW1 and SW2 and are spatially coincident with gaps in the Sandgígur moraines.

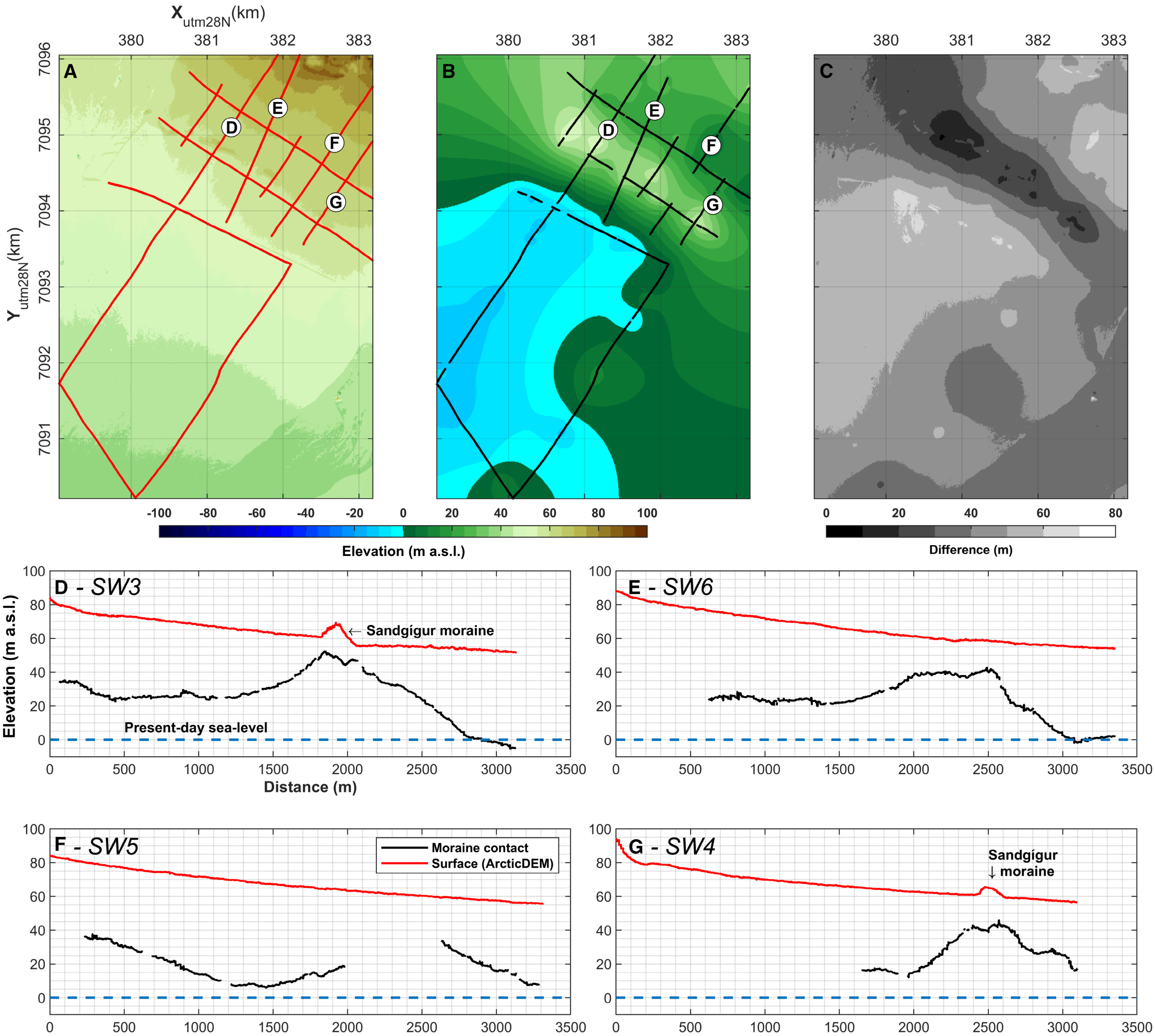


Fig. 5. A. ArcticDEM surface; red lines indicate the location of GPR profiles. B. Interpolated DEM of the RS1 surface. Black lines show the pick locations of RS1 within GPR profiles. C. DEM of difference showing sediment thickness between RS1 and the surface of the sandur. D–G. Elevation profiles of the subsurface ridge that characterizes RS1. Locations of the profiles can be seen in (A) and (B).

Radar facies 2 (RF2). – RF2 is a localized facies that is found beneath the RS1 surface. It is only found directly beneath the Sandgígur moraines (Fig. 7). In most radargrams, RF2 is in contact with the RS1 surface and sandur surface; however, where there are gaps in the Sandgígur moraines the RF2 unit is overlain by a unit of chaotic and planar reflections that can be up to 20 m thick (e.g. Fig. 4B, C). RF2 is characterized by large hummocks and concave reflections and is mostly discontinuous, with offsets apparent between reflections (Table 1). Pockets of low amplitude reflections are located throughout RF2 and in some instances have enhanced levels of noise.

Radar facies 3 (RF3). – Facies RF3 is another localized facies. It is positioned below the RS1 reflection and is only recorded in the SW1 and SW3 radargrams (e.g. Fig. 7). RF3 ranges in elevation from 0 to 35 m a.s.l. (depth of 15–50 m). In down-sandur (NE–SW) orientated radargrams (e.g. Fig. 4B) RF3 is dominated by strong, coherent, oblique reflections (Table 1). These reflections dip at an angle of 1.29–2.27° (Fig. 4B) and are truncated by the RS1 bounding surface. When observing RF3 and its related reflections at the intersection of SW1 and SW3 it can be suggested that the radar facies has a conical 3D form.

Radar facies 4 (RF4). – Radar reflections associated with facies RF4 are mostly confined to the SW8 radargram. This facies exists below the RS1 boundary (Fig. 7), which has a hummocky morphology. Reflec-

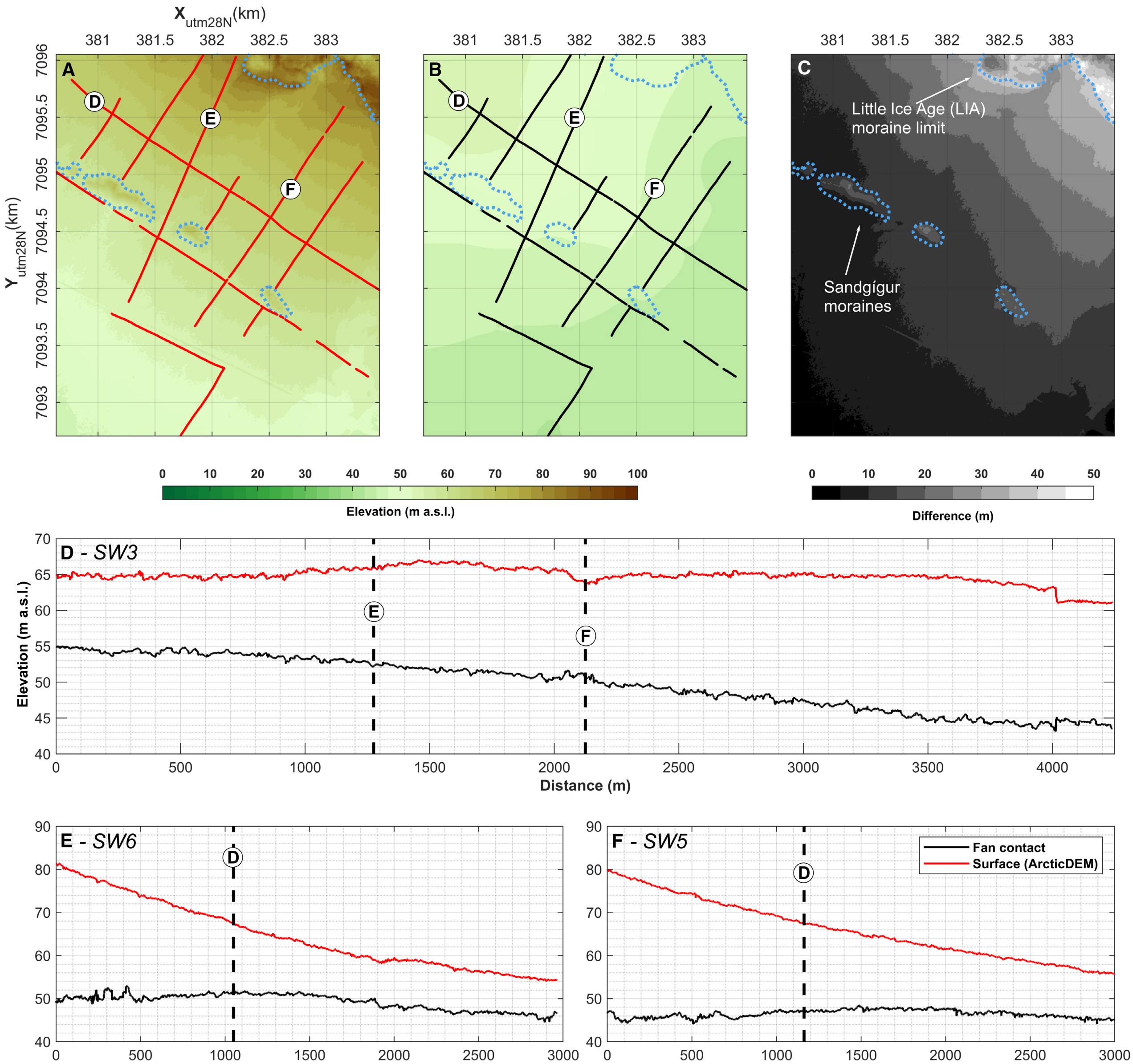


Fig. 6. A. ArcticDEM surface; red lines indicate the location of GPR profiles. B. Interpolated DEM of the RS2 surface. Black lines show the pick locations of RS2 within GPR profiles. C. DEM of difference showing sediment thickness between RS2 and the surface of the sandur. D–F. Elevation profiles of the RS2 surface. Locations of the profiles can be seen in (A) and (B).

tions within this sediment package generally dip up-sandar in a NE–SW orientation (Table 1). Only one line provides a cross-section of this radar facies. In perpendicular orientation RF4 is dominated by discontinuous reflections that have a convex to planar geometry and onlap onto one another.

Radar facies 5 (RF5). – Reflections associated with RF5 are located below the RS1 surface (Fig. 7), with an elevation ranging between −20–20 m a.s.l. The thickness of RF5 is not easy to establish due to the lack of a lower bounding surface in most radargrams. However, in radargram SW7 the RF5 facies is uniform and ~10 m thick bounded by RS1 and RS0. RF5 is dominated by a low-amplitude signal with sparse reflections (Table 1). Where reflections are visible, they are moderately continuous but incoherent.

Radar facies 6 (RF6). – Very low amplitude, incoherent, oblique clinoforms (Table 1) can be observed in radargrams SW3 and SW8 at elevations of −30––60 m a.s.l. (Fig. 7). These structures dip at a moderate angle (2.19–6.87°) down-sandur (NE–SW orientation). The basal contact of this facies is not visible. This is likely due to

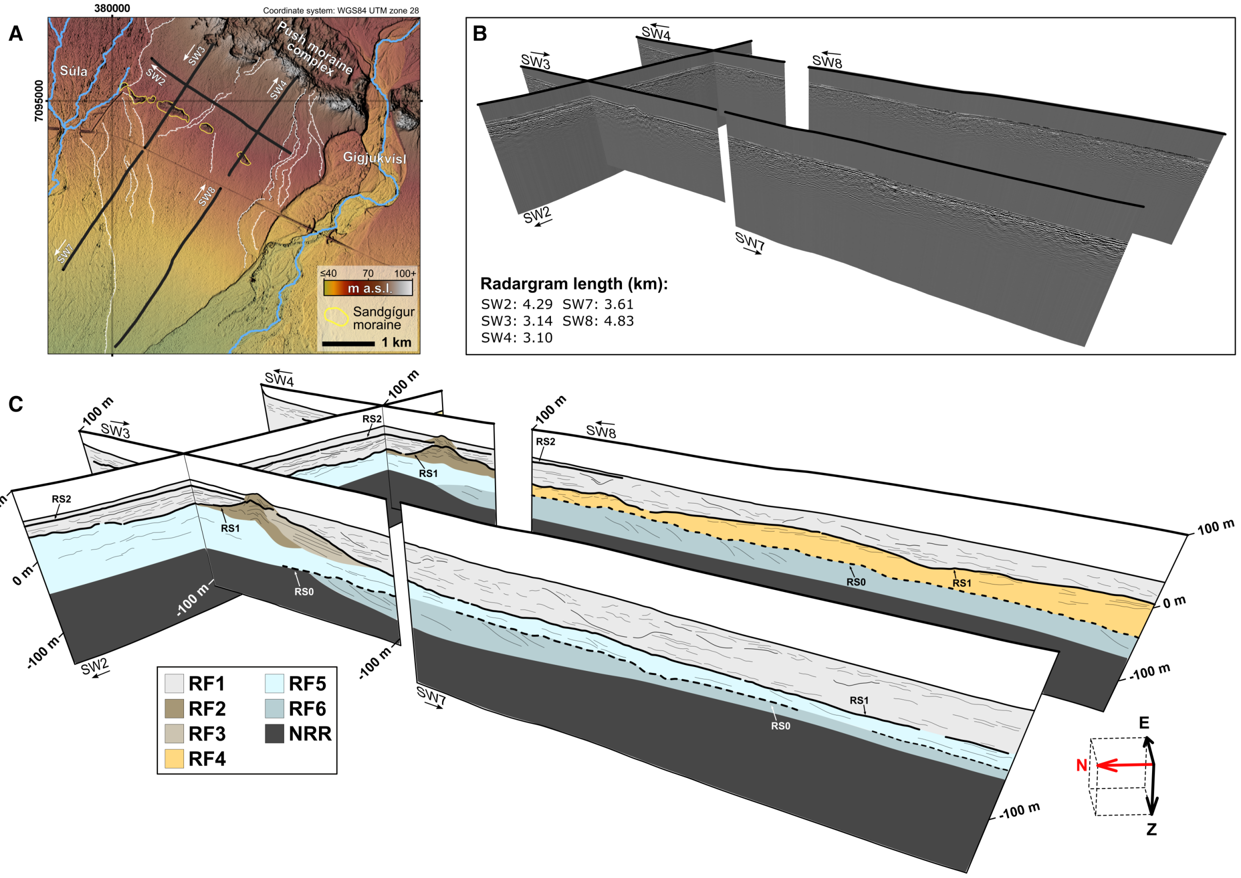


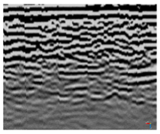
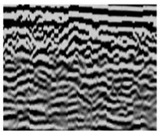
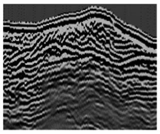
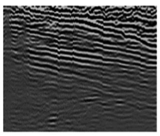
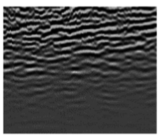
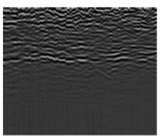
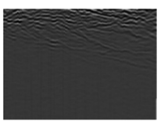
Fig. 7. 3D fence diagram displaying radar facies (RF) for radar data collected at the Sandgígur moraines. A. Location of radar lines. B. Unpicked radargrams. C. Interpretative radar facies panels. Descriptions and interpretations of radar facies can be seen in Table 1.

attenuation of returned radar power as a result of the increased depth of the reflections. It is important to note that our observed dip angles may differ from the true dip of reflections due to changes in water content within the subsurface, especially close to or below present-day sea level.

Radar facies interpretations

RF1 is interpreted as glacifluvial sediments (Table 1) related to outburst flood sedimentation or proximal fan delta deposition (e.g. Jakobsen & Overgaard 2002; Hansen *et al*. 2009). Diffractions within RF1 are caused by the presence of boulders buried in the sandur. The large trough-like structures present within this unit are interpreted as palaeo-channels. RF2 is an indicator of glaciotectonic deformation that is linked to push moraine formation (e.g. Overgaard & Jakobsen 2001; Jakobsen & Overgaard 2002). Large hummocky and concave reflections indicate anticlinal folds and offset reflections that may represent faulting of the sediments (Table 1). The presence of this facies directly beneath the Sandgígur moraines and within the pronounced ridge of RS1 (Figs. 4B, 7) supports this interpretation. Low amplitude lenses present within RF2 represent a homogenous material. This may be buried ice or fine-grained massive deposits. The uniform stacking and oblique nature of the reflections associated with RF3 suggests these are stratified deposits that represent an end-moraine/ice-contact fan (Table 1). The dip angle of the reflections and possible conical 3D-form is comparable to similar end-moraine fans reported elsewhere in Iceland (e.g. Russell *et al*. 2001a; Kjær *et al*. 2004). The interpretation of RF4 is difficult due to the lack of radargrams coincident with the unit, and the depth at which it is recorded. However, RF4 does show radar characteristics similar to the internal structure of drumlins elsewhere in Iceland (e.g. Woodard *et al*. 2020). The low amplitude nature of RF5 makes interpretation difficult. Based upon its reflection characteristics, and location beneath RS1 and the morainic facies (i.e. RF2), however, it could represent either subglacial material or an overridden fluvial deposit. RF6 is interpreted as prograding foresets (Table 1) associated with Gilbert-type delta formation and sub-aqueous deposition (e.g. Hansen *et al*. 2009). It is important to stress, however, that the very low amplitude nature of the RF6 reflections and the depth at which they are recorded mean that we

Table 1. Radar facies descriptions and interpretation.

Radar facies (RF)		Description	Interpretation
 	RF1	Chaotic, planar to wavy/undulating, strong, moderately continuous to discontinuous reflections with localized diffractions.	Massive or layered deposits comprising of channel fills, bedding formed by migrating dunes and bars (sand/gravel/boulders). Hyperbolic diffractions are indication of single boulders. Glacifluvial deposits, outburst flood deposits, proximal (fan) delta deposits. (e.g. Jakobsen & Overgaard 2002; Hansen *et al*. 2009).
	RF2	Mostly large hummocky and discontinuous reflections, presence of some steep, oblique reflections, localized hyperbolic diffractions and localized lenses of low amplitude/noisy reflections.	Deposits consisting of anticlinal folds and faulted/offset reflections indicating deformation. Low amplitude pockets could represent buried glacier ice and/or homogenous deposits (e.g. massive sands). Glaciotectonics, push moraine, tectonized glacifluvial sediments. (e.g. Overgaard & Jakobsen 2001; Jakobsen & Overgaard 2002).
	RF3	Oblique, moderate to low angled, subparallel down sandur dipping continuous reflections.	Stratified deposits (sand/gravel) with a dip angle of 1.29–2.27°. Formation of an ice-contact or end-moraine outwash fan. (e.g. Krzyszkowski & Zieliński 2002; Kjær *et al*. 2004; Lukas 2005).
	RF4	Oblique, low-angled, up-sandar dipping and planar to wavy reflections, moderately continuous with localized diffractions.	Interpretation difficult due to lack of exposure and depth. Reflections show similarity to the geophysical investigations into drumlins elsewhere in Iceland (e.g. Woodard *et al*. 2020). However, subhorizontal to gently up-sandur dipping reflections may also reflect upper stage plane beds formed during critical or supercritical flow regimes (e.g. Alexander *et al*. 2001).
	RF5	Low amplitude chaotic, planar to wavy, moderately continuous to discontinuous reflections.	Low amplitude caused by increased depth and/or materials that attenuate radar signal. Massive or layered deposits. Interpretation difficult due to increased depth.
	RF6	Very low amplitude oblique clinoforms, moderate to high angled, moderately continuous to discontinuous reflections.	Progradational foreset bedding (sand/gravel). Dip angle of 2.19–6.87°. (Fan) delta foresets, Gilbert-type foresets. (e.g. Hansen *et al*. 2009).
No radar reflection (NRR).		No visible reflections.	Signal attenuated due to increased depth.

have lower confidence in this interpretation. Furthermore, the likely increase in groundwater content at depth may have caused reflection distortion within the RF6 radar facies.

Discussion

The landscape evolution of the Sandgígur morainic system can be divided into three specific phases: (i) the pre-moraine landsystem, (ii) formation of the Sandgígur moraines, and (iii) aggradation of the sandur plain and burial of the moraine ridge. Based on the radar-derived sediment architecture and radar facies of the study area we present a schematic model of landscape evolution (Fig. 9) and assess its significance and implications.

Pre-moraine formation

Radar reflections and radar facies located at depth and distal from the present-day ice margin are difficult to confidently interpret due to a combination of their

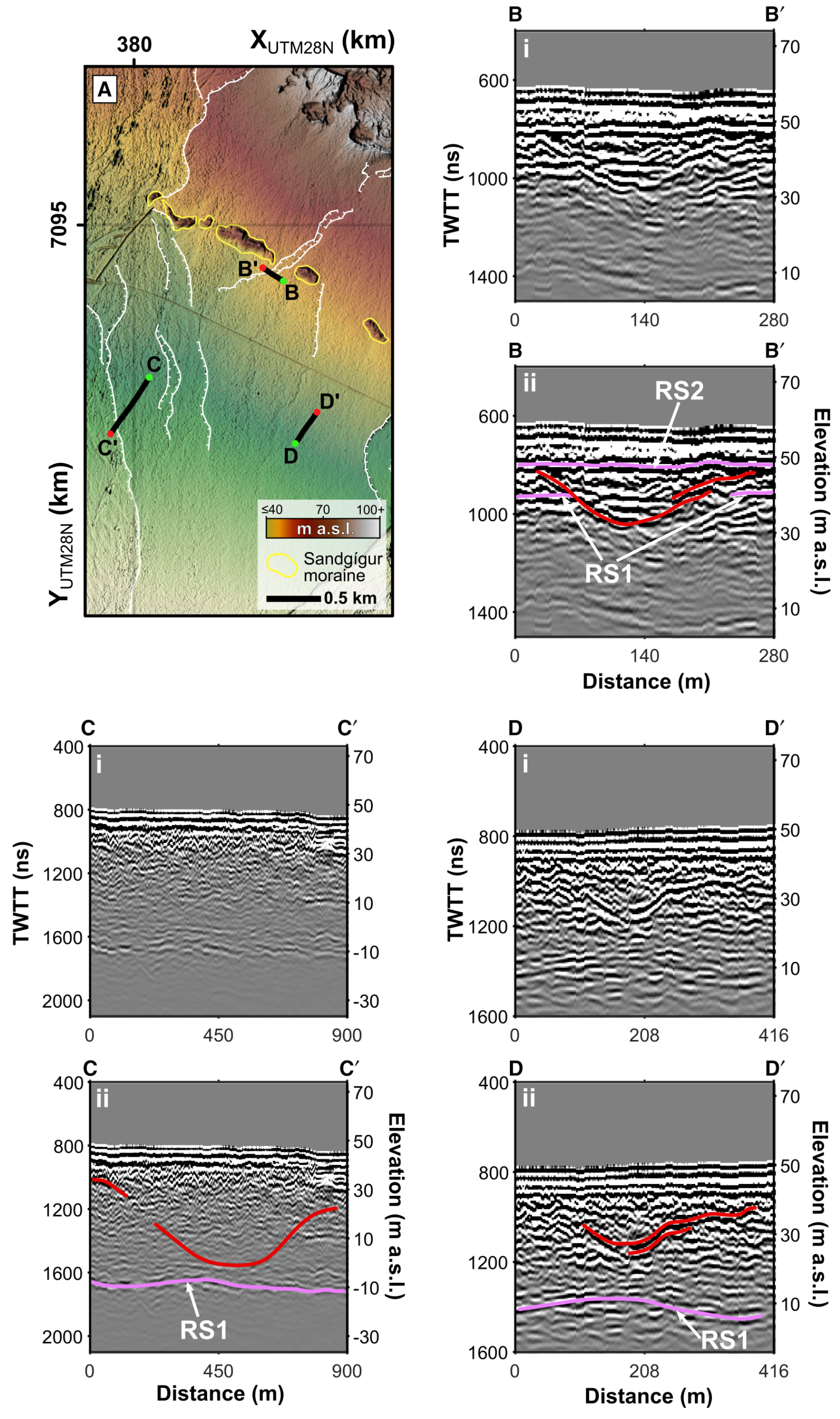


Fig. 8. Examples of large trough structures visible throughout the RF1 radar facies unit. A. Map showing location of trough structures identified in radargrams. B–D. Uninterpreted (i) and interpreted (ii) radargrams showing trough structures. Red lines in (Bii)–(Dii) outline the base of the trough feature.

stratigraphical position and the attenuation of the radar signal with depth. Reflections located at depths of 70–100 m should be taken with caution as this is typically viewed as the limit of GPR penetration depths within sands and gravels (Neal 2004; Reynolds 2011). As such interpretation of these deeper reflections makes reconstruction of events prior to the formation of the Sandgígur moraines less certain. However, in the right conditions broadband GPRs with high pulse repetition frequencies, such as the UTSI GV7, have achieved penetration depths of up to 120 m (e.g. Francke & Utsi 2009). Furthermore, if buried ice is present within the subsurface, penetration depths would be improved due to the increased radio-wave propagation through low loss frozen materials (Neal 2004; Reynolds 2011).

The RS0 surface represents the base of the major reflections and radar units recorded within the GPR survey. RS0 is likely to reflect a major radar-properties (i.e. dielectric) boundary that likely marks a change in depositional setting. The change in reflection characteristics and brightness at depth could indicate that the RS0 surface is a sediment–bedrock boundary. However, similar reflection characteristics can occur between sediment boundaries due to changes in sediment composition, orientation, compaction and the resultant changes in porosity of the substrate (Neal 2004). The

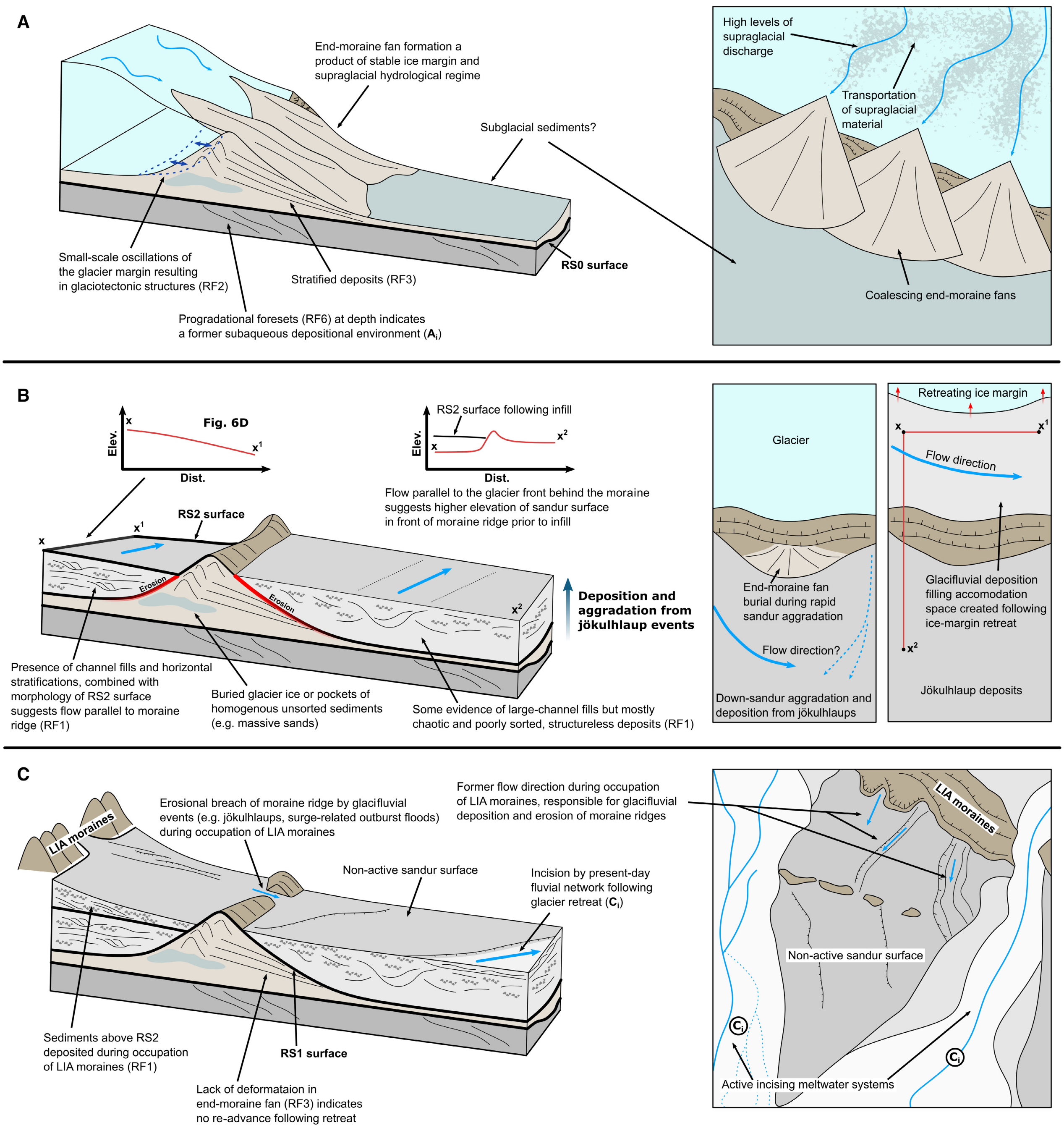


Fig. 9. Schematic model of the evolution of the Sandgígur morainic system from formation (A), to burial (B), and eventual erosional breaches (C).

elevation of RS0 is similar to that of a transition between an unconsolidated sediment package and a unit of over-consolidated sediment reported from seismic refraction surveys (Guðmundsson *et al*. 2002). The presence of reflections below RS0 suggests that it is not bedrock and that it is more likely the over-consolidated sediment unit reported by Guðmundsson *et al*. (2002).

The progradational foresets within the RF6 radar facies inferred from the down-sandur dipping reflections (Table 1) provide evidence of deposition into a body of standing water. Deposition related to the formation of radar facies RF6 (Table 1) in a marine environment is favoured over a proglacial lacustrine environment. Proglacial lake formation would require a topographic high distal of the ice margin (e.g. morainic debris, landslide debris or less commonly glacifluvial sediments) to act as a dam (Carrivick & Tweed 2013), and there is a lack of evidence within the radargrams of a topographic high down-sandur of the reflections within RF6. If the oblique reflections characteristic of RF6 are foresets deposited in a marine environment it would suggest a former relative sea level (RSL) limit of approximately −30 m (excluding any effects of GIA) and a former shoreline located beneath the sandur in the immediate vicinity of the Sandgígur moraines (Fig. 4B). The potential RSL limit indicated by the reflections within RF6 is comparable to the minimum RSL limit during the Holocene regression (8.9 ka BP) in western and south-west Iceland (Norðdahl & Pétursson 2005; Norðdahl *et al*. 2008). This would support the widely held belief (e.g. Þórarinsson 1974; Russell & Marren 1999; Guðmundsson *et al*. 2002; Marren 2002a, b; Russell *et al*. 2006) that the majority of Skeiðarársandur was formed during the Holocene, with development beginning at 8.9 ka BP and persisting until present-day. However, this inference should be taken with caution as the depth of the reflections makes confident interpretation difficult.

RF4 is located down-sandur of the Sandgígur moraines (Fig. 7) but it is difficult to determine its stratigraphical relationship to the moraine ridge. The up-sandur dipping nature of the reflections in the RF4 package and the hummocky nature of the surface above them are similar to those identified within recently formed drumlins elsewhere in Iceland (e.g. Woodard *et al*. 2020). If these are drumlins, they must have formed prior to the formation of the Sandgígur moraines and would signify a former subglacial environment. Alternatively, these features may be representative of large sand waves deposited in a coastal environment (e.g. Allen 1980) or antidune trains associated with high magnitude outwash events with supercritical flow regimes (e.g. Alexander *et al*. 2001; Cassidy *et al*. 2003). Large gravel antidune and/or dune trains preserved in the sedimentary record would require formation and subsequent burial during the same jökulhlaup event to facilitate preservation (Marren & Schuh 2009). Similar features have already been reported in other locations on Skeiðarársandur (Marren & Schuh 2009).

Moraine formation

The RS1 surface is interpreted as a major boundary between: (i) an ice-marginal/subglacial dominated depositional system; and (ii) a glacifluvial proglacial depositional system. The morphology of this surface has a prominent subsurface ridge spatially coincident with the surface expression of the Sandgígur moraines (Fig. 4B). The radar data therefore demonstrate that the Sandgígur moraines continue into the subsurface and are a much larger landform than indicated by the present surface topography; with the majority of the moraine buried beneath the surface of Skeiðarársandur. Morainic features have been identified at depth beneath sandur plains elsewhere in Iceland (e.g. Haraldsson & Palm 1980) and in Greenland (e.g. Storms *et al*. 2012). The maximum height of the Sandgígur moraines, including surface and subsurface expressions and based on the radio-wave velocity of 0.13 m ns^{-1}, is 67 m (Fig. 5). The GPR observations show that the moraine has a maximum width of 1.25 km (e.g. Fig. 4B). These dimensions are comparable to the surface expression of the LIA Gígjukvísl moraines (Fig. 1), suggesting that the Sandgígur moraines represent a glacial terminus position of Skeiðarárjökull of comparable scale and significance. The nature of moraine formation and related processes can be inferred and hypothesized (Fig. 9) from the RF2 and RF3 radar facies. RF3, located on the distal side of the moraine, is interpreted as glacifluvial sediments deposited within an ice-contact outwash fan (Table 1). This suggests that the Sandgígur moraines were formed during a temporarily stationary, or slowly oscillating, ice-marginal phase, as an ice-contact outwash fan of this scale would require a long period (years to decades) of time to form (Krüger 1997; Krzyszkowski & Zieliński 2002; Kjær *et al*. 2004; Lukas 2005; Reinardy & Lukas 2009). The full extent of the ice-contact fan inferred from RF3 is unknown. This is because the reflections associated with the fan are truncated by the high amplitude radar reflection of RS1 (e.g. Fig. 4B). This radar stratigraphical relationship is consistent with the RS1 surface representing an unconformity, and hence evidencing erosion of the most distal sections of the (RF3) fan.

The interpretation of RF3 as an ice-contact fan suggests that deformation structures identified within RF2 could be linked to the collapse of the proximal part of the fan during ice-margin retreat, leaving a proximal rectilinear ice-contact face that creates a marked morphological asymmetry (Lukas 2005). However, the relative symmetry of the moraine ridge recorded by the RS1 surface does not support this. It is more likely that RF2 is evidence of glaciotectonic processes and deformation of the sediment from small oscillations of the glacier margin (Fig. 9); such processes and structures are commonly associated with terrestrial ice-contact fans and dump moraines (Lukas 2005). The large hummocky nature of the reflections and offsets between the reflections within RF2 are comparable in form and scale to folded and faulted sediment units generated by glaciotectonism (e.g. Jakobsen & Overgaard 2002; Benediktsson *et al.* 2010). The absence of deformation within the end-moraine fan unit suggests that there was no major re-advance or reoccupation of the moraine following glacial retreat from its position at the Sandgígur moraines (Lukas 2005) and that fan development was a result of syntectonic deposition. The presence of lenses of low-amplitude reflectivity within the RF2 unit (e.g. Fig 4B) could indicate the presence of buried glacier ice. Radar unit lenses with few and low-amplitude reflections have been interpreted as pockets of sediment-free massive ice (e.g. buried glacier ice; Moorman *et al.* 2003). Buried glacier ice is preserved at depth in sandur in southern Iceland (Everest & Bradwell 2003), and has been identified in more-recent morainic sections at Skeiðarársandur (e.g. Molewski & Olszewski 2000; Russell *et al.* 2001a). The presence of buried ice at depth in the Sandgígur moraines cannot therefore be ruled out.

Moraine development during a glacial surge is another possible glaciological scenario for the formation of the Sandgígur moraines. Skeiðarárjökull is a known surge-type glacier that is documented to have previously surged seven times, the earliest surge recorded occurred in 1787 and the most recent in 1991 (Jóhannesson 1985; Björnsson *et al.* 2003). The surges of Skeiðarárjökull were predominately associated with the western and central portions of the glacier (Björnsson *et al.* 2003) with the limit of the most recent surge (*c.* 1991) marked by a push moraine and an associated fine-grained, stratified outwash fan (Russell *et al.* 2001a; van Dijk & Sigurðsson 2002; van Dijk 2002). Push moraines formed during glacier surges are dominated by heavily deformed and glaciotectonized sediments (Russell *et al.* 2001a; Benediktsson *et al.* 2010), which is consistent with the presence of RF2 within the Sandgígur moraines. The outwash fans associated with the surges of Skeiðarárjökull have similarities to other ice-contact/end-moraine fans in the literature (Russell *et al.* 2001a; van Dijk 2002; van Dijk & Sigurðsson 2002). As a result, RF3 is also consistent with a surge model (Evans & Rea 1999, 2003) for moraine formation (Fig. 4B). However, aside from the potential evidence from the moraine ridge itself, no other evidence suggesting moraine formation by glacial surge is apparent in the GPR data.

Burial of the Sandgígur moraines and sandur aggradation

The large proportion of the Sandgígur moraines that currently lies beneath the sandur surface indicates that there has been significant sandur aggradation post moraine formation. Sediments above the moraine-bounding RS1 surface were deposited after the formation of the Sandgígur moraines (Fig. 9). The estimated volume of the sediment package above the Sandgígur moraines (1.04 km^3), is comparable to the sediment yield from ten 1996-equivalent jökulhlaups (Guðmundsson *et al.* 2002; Snorrason *et al.* 2002). The volume of sediment above the Sandgígur moraines represents ~1% (1.04 km^3) of the Holocene sedimentation (100 km^3) approximated by Guðmundsson *et al.* (2002), accumulating in 1.9% (24.09 km^2) of the total sandur area (1300 km^2) assuming a radio-wave velocity of 0.13 m ns^{-1}.

The majority of the sediment above RS1 is interpreted as glacifluvial in origin (Fig. 7). Glacifluvial deposition is the dominant process of sedimentation at Skeiðarársandur (e.g. Þórarinsson 1974; Boothroyd & Nummedal 1978; Guðmundsson *et al.* 2002) and GPR reflection characteristics within RF1 (Table 1) are consistent with sedimentary architectures (e.g. chaotic and planar to wavy moderately continuous to discontinuous reflection geometries) formed by glacifluvial deposition reported in other geophysical investigations (e.g. Beres *et al.* 1995, 1999; Jakobsen & Overgaard 2002; Hansen *et al.* 2009). The stratigraphy down-sandur of the moraine ridge (i.e. RF1) is dominated by chaotic reflections with numerous diffractions. Chaotic reflection geometries such as these suggest the presence of reworked and poorly sorted, structureless and matrix supported sediments indicative of deposition from sediment-rich flows during the rapid rising stage of a jökulhlaup (Brennand 1994; Russell & Knudsen 1999; Marren & Schuh 2009). During rapid rising-stage flood flows in an unconfined proglacial setting there is insufficient time for sediment sorting and development of well-defined bedding to occur (Church & Gilbert 1975; Smith 1986; Miall 1992; Hiscott 1994; Todd 1996; Russell & Marren 1999; Russell & Knudsen 2002; Russell *et al.* 2002). This results in high rates of aggradation and deposition, and the presence of a poorly sorted structureless deposit (Rushmer 2006, 2007). Rapid emplacement of sediment during a jökulhlaup is the most likely depositional mechanism of this facies due to the burial and preservation of the interpreted ice-contact/end moraine fan (RF3). Some erosion of the fan is evident (e.g. high amplitude nature of the RS1 reflection and truncation of reflections within the fan body). However, the majority of the fan is preserved at depth beneath the interpreted jökulhlaup deposits. The lack of any clear boundaries in the sediments that characterize RF1 makes it difficult to determine whether this unit is the product of a single or multiple jökulhlaup.

Glacifluvial sediments on the up-sandur side of the moraine ridge below the RS2 surface (i.e. RF1) are typically characterized by less chaotic and more moderately-continuous, planar-to-wavy radar characteristics, than that seen in down-sandur radargrams. Small channel fills and reflections linked to migrating bedforms characterize the unit below RS2 and are predominately orientated parallel (NW–SE) to the Sandgígur moraines (Fig. 9B). This suggests that sediment was deposited by meltwater systems that flowed parallel to the palaeo-ice-margin in an ice-proximal depression (Fig. 9B), which is similar to the present-day drainage system of Skeiðarárjökull. The dip angle (0.15°) and orientation of RS2 (e.g. Fig. 6D) indicate that sediment was deposited from flows running parallel to the Sandgígur moraines. The presence of a lateral drainage network is common in landsystem models for retreating ice margins (Price 1969; Bogacki 1973; Galon 1973a, b; Klimek 1973; Gomez *et al*. 2000). Meltwater drainage parallel to the ice margin requires the lowering of the glacier bed below the elevation of the ice-proximal sandur, thus creating a topographic discontinuity that prevents meltwater from draining onto the sandur surface and diverts meltwater to flow parallel to the ice margin (Gomez *et al*. 2000). This indicates that the burial of the up-sandur portion of the moraine ridge occurred after a period of down-sandur aggradation once the glacier had retreated from the Sandgígur moraines (Fig. 9B).

Sedimentation above RS2 is thought to have occurred when Skeiðarárjökull reached its LIA maximum (Figs 1C, 9C). The surface geomorphology suggests that fans and fluvial channels emanated from the LIA push moraines and continued to just beyond the Sandgígur moraines. In addition, the tapering thickness of sediment above RS2 is consistent with outwash fan formation (Fig. 6E, F). Where reflections are visible within the sediments above RS2 there is evidence of some down-sandur dipping reflections proximally, and more planar continuous reflections distally. This is consistent with outwash fan architecture (Krzyszkowski & Zieliński 2002). The push moraines associated with the LIA margin (Fig. 1) are believed to be formed as a result of multiple glacier surges (Molewski & Olszewski 2000; Björnsson *et al*. 2003). Large sediment-rich outburst floods have been known to occur at the end of surge events as a result of the reorganization of the subglacial drainage network (Kamb *et al*. 1985; Björnsson 1998; Russell *et al*. 2001a; van Dijk 2002; van Dijk & Sigurðsson 2002; Eisen *et al*. 2005). Furthermore, the post-surge quiescent phase can allow for the development of supraglacially-fed outwash fans (Russell *et al*. 2001a; van Dijk 2002; van Dijk & Sigurðsson 2002). The sediments above RS2 could therefore represent surge related outwash fans, which are contrasting to jökulhlaup-related outwash fans due to the absence of both heavily kettled topography and large-scale bar and channel patterns (Russell *et al*. 2001a). The presence of surge related outwash fans in front of the LIA moraines has already been reported from a large, exposed sediment section at the Gígjukvísl gap (Russell *et al*. 2001a). The sediments above RS2 and the potential surge outwash fans at the Gígjukvísl gap section are separated by the incised Gígjukvísl meltwater system and it is therefore plausible that they may have previously been connected. Outwash fans related to surge events would likely be regularly spaced along the ice margin and form laterally extensive and coalesced belts of outwash fans (Russell *et al*. 2001a). If the fans above RS2 are surge-related it would add further support to the suggestion that glacial surges are a significant control in the development and formation of glacifluvial landforms and deposits (Russell *et al*. 2001a; van Dijk 2002; van Dijk & Sigurðsson 2002).

The discontinuous surface expression of the Sandgígur moraines and coincident subsurface features suggest that the moraines have been breached and eroded by later outwash floods. The erosion of the Sandgígur moraines likely occurred during the period of sedimentation above RS2 and during occupation of the LIA moraines. However, some sections of the moraine may have been eroded prior to the formation of the RS2 surface, as the moderately continuous reflection in the SW5 radargram below RS2 can be seen to truncate a buried portion of the Sandgígur moraines (RS1) at depth (Fig. 4C). Therefore, timing of the moraine breaches cannot be conclusively determined. Historic accounts suggest that jökulhlaups in the 18th century eroded portions of the Sandgígur moraines (Grove 2004). It is thought that these outburst floods were caused by the drainage of the Núpslón ice-dammed lake (Þórarinsson 1939; Grove 2004). This scenario is plausible in leading to sedimentation behind the Sandgígur moraines and the erosional breaches. However, historical accounts show that there have been numerous high-magnitude jökulhlaups that have inundated the entire sandur (Þórarinsson 1974) that would have been capable of eroding portions of the Sandgígur moraines.

Implications for sandar evolution

Sandar sedimentary architecture is largely controlled by variations in the glacier margin position (e.g. Marren 2004). Fluctuations of the ice margin influence the regions at which sandar aggradation or incision can take place (Fig. 10). During periods of glacier advance and stability the ice margin is coupled to the extramarginal sandur environment and there is an abundant sediment supply (Fig. 10), promoting continued aggradation beyond the morainic limits. Where climate-driven low-magnitude high-frequency meltwater regimes dominate stratified ice-contact fans are formed (e.g. Fig. 9A). However, a switch to a high-magnitude low-frequency dominated meltwater regime, for example as a result of increase in jökulhlaup prevalence, can result in major down-sandur aggradation of thick jökulhlaup units (e.g. Fig. 9B; Marren 2005). As the glacier retreats accommodation space is created, leading to the activation of intramarginal geomorphic activity and aggradation during periods of increased sediment supply (e.g. jökulhlaup events). Sediment aggradation within a confined sandur region by ice-margin parallel glacifluvial activity in an ice-proximal depression, following retreat from the

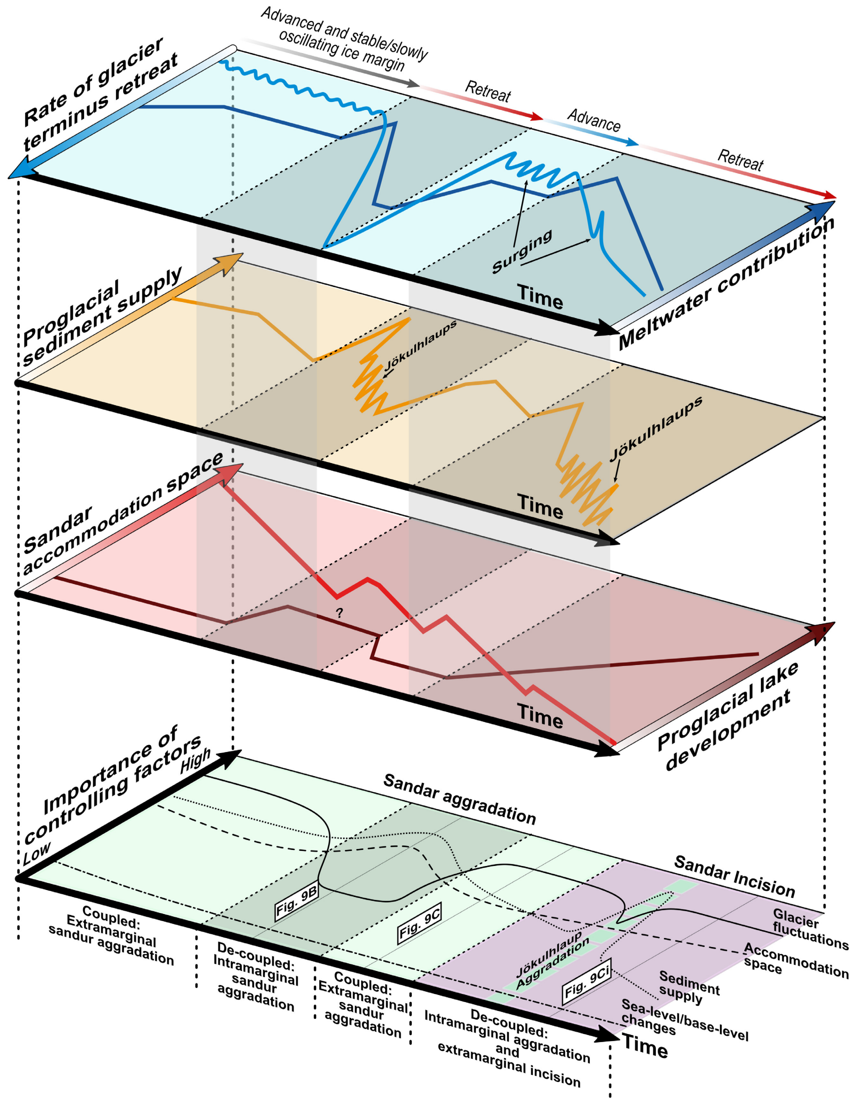


Fig. 10. Conceptual diagram showcasing the dominant controls on the evolution of sandar environments and how they vary in response to glacier margin advance and retreat.

Sandgígur moraines (Fig. 9B), is coincident with landsystem models for retreating ice margins (e.g. Price 1969; Bogacki 1973; Galon 1973a, b; Klimek 1973; Gomez *et al*. 2000).

Sediment supply to the proglacial environment peaks during the initial stages of ice-margin retreat (Antoniazza & Lane 2021) promoting aggradation within the confined intramarginal zone (Fig. 10). Furthermore, during periods of rapid glacier recession sediment flux (Fig. 10) to the proglacial outwash environment increases (Marren & Toomath 2013). This has important implications for how sandar evolve during periods with varying rates of glacier margin recession. As a result of higher sediment supply during rapid glacier retreat (e.g. Marren & Toomath 2013), accommodation space between the newly established ice margin and the coupled end-moraine and sandur plain is filled. This leads to the sandur backfilling into available accommodation space as the ice margin continues to retreat (Fig. 9B). However, it is important to note that this will only occur where there is a high sediment flux available to the proglacial environment and/or sediment-laden jökulhlaups are prevalent. This process of backfilling and aggradation into accommodation space left by the ice mass following retreat can lead to ice-margin parallel draining networks that create complexity within the stratigraphy compared to what may be evidenced in the surface or shallow subsurface of the sandar (e.g. Marren 2004). However, when the rate of retreat reduces, particularly when combined with proglacial lake development and a continued reduction in accommodation space, incision will dominate within the extramarginal (unconfined) sandur environment (Marren & Toomath 2013; Fig. 10). This is predominately the result of the lowering of the proglacial long profile, due to the decrease of meltwater outlet elevation (e.g. Marren & Toomath 2013), and the reduction of sediment supply to the extramarginal sandur due to ice -marginal proglacial lake systems acting as an efficient sediment trap (e.g. Carrivick & Tweed 2013).

The burial of the Sandgígur moraines by thick and extensive glacifluvial deposits demonstrates that high rates of sedimentation, likely the result of jökulhlaup events, are a dominant driver in the formation and aggradation of Skeiðarársandur. Sea level in southern Iceland is known to have been relatively stable since 6 ka BP (Norðdahl & Pétursson 2005). Given the relative stability of the sea level we can infer that in this study base level is not a dominant control. Instead, system-scale sandur evolution is driven by ice-proximal changes. High sedimentation rates and aggradation depths above the Sandgígur moraines (i.e. average sediment thickness of 43 m) suggest that sediment supply and the available terrestrial accommodation space are more dominant drivers of sandur evolution (Fig. 10).

The identification of sediment supply as a primary control on sandur formation is consistent with numerical modelling of the Holocene evolution of Skeiðarársandur (e.g. Maizels 1993a). It is likely that large increases of sediment supply are a result of an increased magnitude and frequency of jökulhlaup drainage over the sandur. The presence of an unstable proglacial fluvial system as a result of these extreme flood events promotes geomorphic activity and sedimentation in both extra- and intramarginal portions of the sandur environment (Carrivick & Heckmann 2017; Weckwerth *et al.* 2021), dependent on the state of the glacier margin at the time of sedimentation (Maizels 1979; Marren 2002a, b, 2005; Russell *et al.* 2006) (Fig. 10). Drainage from ice-dammed lakes, in the form of jökulhlaups, is known to increase in frequency during periods of ice-margin retreat (Þórarinsson 1939; Evans & Clague 1994). Therefore, glacier margin retreat from its position at the Sandgígur moraines and thinning may have contributed towards the formation and drainage of ice-dammed lakes in the region (e.g. Þórarinsson 1939; Björnsson 1992; Grove 2004; Roberts *et al.* 2005). Furthermore, it is believed that the main calderas of Grímsvötn became active after the 1783 Laki eruption thus increasing the frequency of volcanogenic jökulhlaups in the region that could deliver sediment to the entire sandur (Þórarinsson 1974; Guðmundsson 1992).

Conclusions

- We identify a large, partially buried, moraine system attributed to the Holocene or neoglacial, terminus position of Skeiðarárjökull outlet glacier. The Sandgígur moraines are an isolated remnant of a much more extensive moraine system that would have continued eastwards across Skeiðarársandur.
- Jökulhlaup-related glacifluvial sediments dominate the sediment architecture around and overlying the Sandgígur moraines. Erosional breaches of the Sandgígur moraines were the result of either jökulhlaup flows emanating from the large LIA push moraines, drainage of ice-dammed lakes in the region, or large volcanically-induced jökulhlaups that inundated the entire sandur (e.g. Þórarinsson 1939, 1974; Grove 2004; Roberts *et al.* 2005).
- Our GPR data indicate that in proglacial areas prone to high levels of sediment supply, and rapid sedimentation, there is potential for ice-marginal landforms to be preserved in the subsurface. This has important implications for the preservation potential of sandur systems in the geological record. However, preservation potential of ice-marginal landforms remains low where sedimentation has not sufficiently buried the landform.
- The stratigraphical architecture of Skeiðarársandur beneath the Sandgígur moraines system includes reflections indicative of progradational foresets (RF6). This suggests deposition into a body of standing water, such as the sea, at an elevation of approximately −30 m a.s.l., prior to subaerial moraine formation and glacifluvial deposition.

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Author contributions. – DH and NR collected the GPR data used in this study. DH processed and analysed the data. DH wrote the paper with input from all co-authors.

References

Alexander, J., Bridge, J. S., Cheela, R. J. & Leclair, S. F. 2001: Bedforms and associated sedimentary structures formed under supercritical water flows over aggrading sand beds. *Sedimentology 48*, 133–152.

Allen, J. R. L. 1980: Sand waves: a model of origin and internal structure. *Sedimentary Geology 26*, 281–328.

Allen, J. R. L. 1983: Studies in fluviatile sedimentation: Bars, bar-complexes and sandstone sheets (low-sinuosity braided streams) in the brownstones (L. Devonian), Welsh borders. *Sedimentary Geology 33*, 237–293.

Antoniazza, G. & Lane, S. N. 2021: Sediment yield over glacial cycles: a conceptual model. *Progress in Physical Geography: Earth and Environment 45*, 842–865.

Arun, P. V. 2013: A comparative analysis of different DEM interpolation methods. *The Egyptian Journal of Remote Sensing and Space Science 16*, 133–139.

Benediktsson, Í. Ö., Schomacker, A., Lokrantz, H. & Ingólfsson, Ó. 2010: The 1890 surge end moraine at Eyjabakkajökull, Iceland: a re-assessment of a classic glaciotectonic locality. *Quaternary Science Reviews 29*, 484–506.

Beres, M., Green, A. G., Huggenberger, P. & Horstmeyer, H. 1995: Mapping the architecture of glaciofluvial sediments with three-dimensional georadar. *Geology 23*, 1087–1090.

Beres, M., Huggenberger, P., Green, A. G. & Horstmeyer, H. 1999: Using two- and three-dimensional georadar methods to characterize glaciofluvial architecture. *Sedimentary Geology 129*, 1–24.

Beylich, A. A. & Laute, K. 2015: Sediment sources, spatiotemporal variability and rates of fluvial bedload transport in glacier-connected steep mountain valleys in western Norway (Erdalen and Bødalen drainage basins). *Geomorphology 228*, 552–567.

Björnsson, H. 1992: Jökulhlaups in Iceland: prediction, characteristics and simulation. *Annals of Glaciology 16*, 95–106.

Björnsson, H. 1998: Hydrological characteristics of the drainage system beneath a surging glacier. *Nature 395*, 771–774.

Björnsson, H. & Pálsson, F. 2008: Icelandic glaciers. *Jökull 58*, 365–386.

Björnsson, H., Pálsson, F., Sigurdsson, O. & Flowers, G. E. 2003: Surges of glaciers in Iceland. *Annals of Glaciology 36*, 82–90.

Blauvelt, D. J., Russell, A. J., Large, A. R. G., Tweed, F. S., Hiemstra, J. F., Kulessa, B., Evans, D. J. A. & Waller, R. I. 2020: Controls on jökulhlaup-transported buried ice melt-out at Skeiðarársandur, Iceland: implications for the evolution of ice-marginal environments. *Geomorphology 360*, 107164, https://doi.org/10.1016/j.geomorph.2020.107164.

Bogacki, M. 1973: Geomorphological and geological analysis of the proglacial of Skeiðarárjökull, Central western and eastern sections. *Geographia Polonica 26*, 57–88.

Boothroyd, J. C. & Nummedal, D. 1978: Proglacial braided outwash: a model for humid alluvial-fan deposits. *Fluvial Sedimentology 5*, 641–668.

Brennand, T. A. 1994: Macroforms, large bedforms and rhythmic sedimentary sequences in subglacial eskers, south-central Ontario: implications for esker genesis and meltwater regime. *Sedimentary Geology 91*, 9–55.

Brennand, T. A. & Shaw, J. 1994: Tunnel channels and associated landforms, south-central Ontario: their implications for ice-sheet hydrology. *Canadian Journal of Earth Sciences 31*, 505–522.

Bristow, C. S. 1996: Reconstructing fluvial channel morphology from sedimentary sequences. *In* Carling, P. & Dawson, M. (eds.): *Advances in Fluvial Dynamics and Stratigraphy*, 351–371. Wiley, Chichester.

Burke, M. J., Woodward, J. & Russell, A. J. 2010a: Sedimentary architecture of large-scale, jökulhlaup-generated, ice-block obstacle marks: examples from Skeiðarársandur, SE Iceland. *Sedimentary Geology 227*, 1–10.

Burke, M. J., Woodward, J., Russell, A. J. & Fleisher, P. J. 2009: Structural controls on englacial esker sedimentation: Skeiðarárjökull, Iceland. *Annals of Glaciology 50*, 85–92.

Burke, M. J., Woodward, J., Russell, A. J., Fleisher, P. J. & Bailey, P. K. 2008: Controls on the sedimentary architecture of a single event englacial esker: Skeiðarárjökull, Iceland. *Quaternary Science Reviews 27*, 1829–1847.

Burke, M. J., Woodward, J., Russell, A. J., Fleisher, P. J. & Bailey, P. K. 2010b: The sedimentary architecture of outburst flood eskers: a comparison of ground-penetrating radar data from Bering Glacier, Alaska and Skeiðarárjökull, Iceland. *Bulletin of the Geological Society of America 122*, 1637–1645.

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

Carrivick, J. L. & Tweed, F. S. 2013: Proglacial lakes: character, behaviour and geological importance. *Quaternary Science Reviews 78*, 34–52.

Carrivick, J. L., Pringle, J. K., Russell, A. J. & Cassidy, N. J. 2007: GPR-derived sedimentary architecture and stratigraphy of outburst flood sedimentation within a bedrock valley system, Hraundalur, Iceland. *Journal of Environmental & Engineering Geophysics 12*, 127–143.

Cassidy, N. J., Russell, A. J., Marren, P. M., Fay, H., Knudsen, O., Rushmer, E. L. & van Dijk, T. A. G. P. 2003: GPR derived architecture of November 1996 jökulhlaup deposits, Skeiðarársandur, Iceland. *Geological Society Special Publication 211*, 153–166.

Chandler, B. M. P., Evans, D. J. A., Chandler, S. J. P., Ewertowski, M. W., Lovell, H., Roberts, D. H., Schaefer, M. & Tomczyk, A. M. 2020: The glacial landsystem of Fjallsjökull, Iceland: spatial and temporal evolution of process-form regimes at an active temperate glacier. *Geomorphology 361*, 107192, https://doi.org/10.1016/j.geomorph.2020.107192.

Church, M. 1972: Baffin Island sandurs: a study of Arctic fluvial processes. *Bulletin of the Geological Survey of Canada 216*, 1–208.

Church, M. & Gilbert, R. 1975: Proglacial fluvial and lacustrine environments. *In* Jopling, A. & McDonald, B. C. (eds.): *Glaciofluvial and Glaciolacustrine Sedimentation*, 22–100. *SEPM Society for Sedimentary Geology Special Publication No. 23*. Tulsa, Oklahoma.

van Dijk, T. A. G. P. 2002: *Glacier surges as a control on the development of proglacial fluvial landforms and deposits*. Ph.D. thesis, Keele University. 342 pp.

van Dijk, T. A. G. P. & Sigurðsson, O. 2002: Surge-related floods at Skeiðarárjökull Glacier, Iceland: implications for ice-marginal outwash deposits. *In* Snorrason, Á., Finnsdottir, H. P. & Moss, M. (eds.): *The Extremes of the Extremes: Extraordinary Floods*, 193–198. IAHS Special Publication, Reykjavík, Iceland.

Eisen, O., Harrison, W. D., Raymond, C. F., Echelmeyer, K. A., Bender, G. A. & Gorda, J. L. D. 2005: Variegated Glacier, Alaska, USA: a century of surges. *Journal of Glaciology 51*, 399–406.

Evans, S. G. & Clague, J. J. 1994: Recent climatic change and catastrophic geomorphic processes in mountain environments. *Geomorphology 10*, 107–128.

Evans, D. J. A. & Rea, B. R. 1999: Geomorphology and sedimentology of surging glaciers: a landsystems approach. *Annals of Glaciology 28*, 75–82.

Evans, D. J. A. & Rea, B. R. 2003: Surging glacier landsystem. *In* Evans, D. J. A. (ed.): *Glacial Landsystems*, 259–288. Taylor and Francis, New York.

Evans, D. J. A. & Twigg, D. R. 2002: The active temperate glacial landsystem: a model based on Breiðamerkurjökull and Fjallsjökull, Iceland. *Quaternary Science Reviews 21*, 2143–2177.

Evans, D. J. A., Lemmen, D. S. & Rea, B. R. 1999: Glacial landsystems of the southwest Laurentide ice sheet: modern Icelandic analogues. *Journal of Quaternary Science 14*, 673–691.

Everest, J. & Bradwell, T. 2003: Buried glacier ice in southern Iceland and its wider significance. *Geomorphology 52*, 347–358.

Francke, J. & Utsi, V. 2009: Advances in long-range GPR systems and their applications to mineral exploration, geotechnical and static correction problems. *First Break 27*, 85–93.

Galon, R. 1973a: Geomorphological and geological analysis of the proglacial area of Skeiðarárjökull: central section. *Geographia Polonica 26*, 15–56.

Galon, R. 1973b: A synthetic description of deposits and landforms observed on the proglacial area of Skeiðarárjökull. Conclusions with regard to the age of the deposits and the way in which deglaciation is proceeding. *Geographia Polonica 26*, 141–150.

Geirsdóttir, Á., Miller, G. H., Andrews, J. T., Harning, D. J., Anderson, L. S., Florian, C., Larsen, D. J. & Thordarson, T. 2019: The onset of neoglaciation in Iceland and the 4.2 ka event. *Climate of the Past 15*, 25–40.

Ghienne, J. F. & Deynoux, M. 1998: Large-scale channel fill structures in Late Ordovician glacial deposits in Mauritania, western Sahara. *Sedimentary Geology 119*, 141–159.

Ghienne, J. F., Girard, F., Moreau, J. & Rubino, J. L. 2010: Late Ordovician climbing-dune cross-stratification: a signature of outburst floods in proglacial outwash environments? *Sedimentology 57*, 1175–1198.

Girard, F., Ghienne, J. F. & Rubino, J. L. 2012: Channelized sandstone bodies ('cordons') in the Tassili N'Ajjer (Algeria & Libya): snapshots of a Late Ordovician proglacial outwash plain. *Geological Society London Special Publications 368*, 355–379.

Gomez, B., Smith, L. C., Magilligan, F. J., Mertes, L. A. K. & Smith, N. D. 2000: Glacier outburst floods and outwash plain development: Skeiðarársandur, Iceland. *Terra Nova 12*, 126–131.

Grove, J. M. 2004: Icelandic glaciers and sea ice. *In* Grove, J. M. (ed.): *Little Ice Ages: Ancient and Modern*, 13–60. Routledge, London.

Gudmundsson, H. J. 1997: A review of the Holocene environmental history of Iceland. *Quaternary Science Reviews 16*, 81–92.

Guðmundsson, M. T. 1992: *The crustal structure of the subglacial Grimsvotn Volcano, Vatnajökull, Iceland, from multiparameter geophysical surveys*. Ph.D. thesis, University College London. 231 pp.
Guðmundsson, M. T., Bonnel, A. & Gunnarsson, K. 2002: Seismic soundings of sediment thickness on Skeiðarársandur, SE-Iceland. *Jökull 51*, 53–64.
Guillon, H., Mugnier, J. L. & Buoncristiani, J. F. 2018: Proglacial sediment dynamics from daily to seasonal scales in a glaciated Alpine catchment (Bossons glacier, Mont Blanc massif, France). *Earth Surface Processes and Landforms 43*, 1478–1495.
Hansen, L., Beylich, A., Burki, V., Eilertsen, R. S., Fredin, O., Larsen, E., Lyså, A., Nesje, A., Stalsberg, K. & Tønnesen, J. F. 2009: Stratigraphic architecture and infill history of a deglaciated bedrock valley based on georadar, seismic profiling and drilling. *Sedimentology 56*, 1751–1773.
Haraldsson, H. & Palm, H. 1980: A seismic investigation in the Markarfljót sandur area, southern Iceland. *Striolae 2*, 45 pp.
Harrison, D., Ross, N., Russell, A. J. & Dunning, S. A. 2019: Post-jökulhlaup geomorphic evolution of the Gígjökull Basin, Iceland. *Annals of Glaciology 80*, 127–137.
Hiscott, R. N. 1994: Loss of capacity, not competence, as the fundamental process governing deposition from turbidity currents. *Journal of Sedimentary Research 64*, 209–214.
Hutchinson, M. F. 1989: A new procedure for gridding elevation and stream line data with automatic removal of spurious pits. *Journal of Hydrology 106*, 211–232.
Ingólfsson, O. 1991: A review of the Late Weichselian and Early Holocene glacial and environmental history of Iceland. *In* Maizels, J. K. & Caseldine, C. (eds.): *Environmental Change in Iceland: Past and Present*, 13–29. *Glaciology and Quaternary Geology 7*. Springer, Dordrecht.
Ives, J. D. 2007: *Skaftafell in Iceland: a thousand years of change*. 256 pp. Ormstunga, Reykjavík.
Jakobsen, P. R. & Overgaard, T. 2002: Georadar facies and glaciotectonic structures in ice marginal deposits, northwest Zealand, Denmark. *Quaternary Science Reviews 21*, 917–927.
Jóhannesson, H. 1985: þættir ur sogu Skeiðarajökuls. The advances and retreats of Skeiðarárjökull glacier in southeast Iceland in the last 250 years. *Náttúrufræðingurinn 54*, 31–45.
Kamb, B., Raymond, C. F., Harrison, W. D., Engelhardt, H., Echelmeyer, K. A., Humphrey, N., Brugman, M. M. & Pfeffer, T. 1985: Glacier surge mechanism: 1982–1983 surge of variegated glacier, Alaska. *Science 227*, 469–479.
Kjær, K. H., Sultan, L., Krüger, J. & Schomacker, A. 2004: Architecture and sedimentation of outwash fans in front of the Mýrdalsjökull ice cap, Iceland. *Sedimentary Geology 172*, 139–163.
Klimek, K. 1973: Geomorphological and geological analysis of the proglacial area of the Skeiðarárjökull: extreme eastern and western sections. *Geographia Polonica 26*, 89–113.
Knight, J. & Harrison, S. 2014: Mountain glacial and paraglacial environments under global climate change: lessons from the past, future directions and policy implications. *Geografiska Annaler, Series A: Physical Geography 96*, 245–264.
Knight, J. & Harrison, S. 2018: Transience in cascading paraglacial systems. *Land Degradation & Development 29*, 1991–2001.
Kociuba, W. & Janicki, G. 2018: Effect of meteorological patterns on the intensity of streambank erosion in a proglacial gravel-bed river (Spitsbergen). *Water 10*, 320, https://doi.org/10.3390/w10030320.
Kociuba, W., Janicki, G. & Dyer, J. L. 2019: Contemporary changes of the channel pattern and braided gravel-bed floodplain under rapid small valley glacier recession (Scott River catchment, Spitsbergen). *Geomorphology 328*, 79–92.
Krigström, A. 1962: Geomorphological studies of sandur plains and their braided rivers in Iceland. *Geografiska Annaler 44*, 328–346.
Krüger, J. 1997: Development of minor outwash fans at Kötlujökull, Iceland. *Quaternary Science Reviews 16*, 649–659.
Krzyszkowski, D. & Zieliński, T. 2002: The Pleistocene end moraine fans: controls on their sedimentation and location. *Sedimentary Geology 149*, 73–92.
Kurjański, B., Rea, B. R., Spagnolo, M., Cornwell, D. G., Howell, J., Comte, J. C., González Quirós, A., Palmu, J. P., Oien, R. P. & Gibbard, P. L. 2021: Cool deltas: Sedimentological, geomorphological and geophysical characterization of ice-contact deltas and implications for their reservoir properties (Salpausselkä, Finland). *Sedimentology 68*, 3057–3101.
Lane, S. N., Bakker, M., Gabbud, C., Micheletti, N. & Saugy, J. N. 2017: Sediment export, transient landscape response and catchment-scale connectivity following rapid climate warming and Alpine glacier recession. *Geomorphology 277*, 210–227.
Lang, J., Le Heron, D., Van den Berg, J. & Winsemann, J. 2021: Bedforms and sedimentary structures related to supercritical flows in glacigenic settings. *Sedimentology 68*, 1539–1579.
Lukas, S. 2005: A test of the englacial thrusting hypothesis of 'hummocky' moraine formation: case studies from the northwest Highlands, Scotland. *Boreas 34*, 287–307.
Magnússon, E., Björnsson, H., Rott, H., Roberts, M. J., Pálsson, F., Gudmundsson, S., Bennett, R. A., Geirsson, H. & Sturkell, E. 2011: Localized uplift of Vatnajökull, Iceland: Subglacial water accumulation deduced from InSAR and GPS observations. *Journal of Glaciology 57*, 475–484.
Maizels, J. K. 1979: Proglacial aggradation and changes in braided channel patterns during a period of glacier advance: an Alpine example. *Geografiska Annaler, Series A: Physical Geography 61*, 87–101.
Maizels, J. K. 1983: Lichenometry and paleohydrology of terraced sandur deposits, Sólheimajökuir. *Geomorphological and Environmental Studies in Southern Iceland, Report of Aberdeen University Iceland Expedition*, 40–50.
Maizels, J. K. 1989: Sedimentology, paleoflow dynamics and flood history of jökulhlaup deposits: paleohydrology of Holocene sediment sequences in Southern Iceland sandur deposits. *Journal of Sedimentary Petrology 59*, 204–223.
Maizels, J. K. 1991: The origin and evolution of Holocene sandur deposits in areas of jökulhlaup drainage, Iceland. *In* Maizels, J. K. & Caseldine, C. (eds.): *Environmental Change in Iceland: Past and Present*, 267–302. *Glaciology and Quaternary Geology 7*. Springer, Dordrecht.
Maizels, J. K. 1993a: Quantitative regime modelling of fluvial depositional sequences: application to Holocene stratigraphy of humid-glacial braid-plains (Icelandic sandurs). *Geological Society London Special Publications 73*, 53–78.
Maizels, J. K. 1993b: Lithofacies variations within sandur deposits: the role of runoff regime, flow dynamics and sediment supply characteristics. *Sedimentary Geology 85*, 299–325.
Maizels, J. K. 1997: Jökulhlaup deposits in proglacial areas. *Quaternary Science Reviews 16*, 793–819.
Marren, P. M. 2002a: Fluvial-lacustrine interaction on Skeiðarársandur, Iceland: implications for sandur evolution. *Sedimentary Geology 149*, 43–58.
Marren, P. M. 2002b: Glacier margin fluctuations, Skaftafellsjökull, Iceland: implications for sandur evolution. *Boreas 31*, 75–81.
Marren, P. M. 2004: Present-day sandurs are not representative of the geological record. Sedimentary Geology 152, 1–5 (2002). Discussion. *Sedimentary Geology 164*, 335–340.
Marren, P. M. 2005: Magnitude and frequency in proglacial rivers: a geomorphological and sedimentological perspective. *Earth Science Reviews 70*, 203–251.
Marren, P. M. & Schuh, M. 2009: Identifying jökulhlaup deposits in the sedimentary record. *In* Burr, D. M., Carling, P. A. & Baker, V. R. (eds.): *Megaflooding on Earth and Mars*, 225–242. Cambridge University Press, Cambridge.
Marren, P. M. & Toomath, S. C. 2013: Fluvial adjustments in response to glacier retreat: Skaftafellsjökull, Iceland. *Boreas 42*, 57–70.
Miall, A. D. 1985: Architectural-element analysis: a new method of facies analysis applied to fluvial deposits. *Earth Science Reviews 22*, 261–308.
Miall, A. D. 1992: Alluvial deposits. *In* Walker, R. G. & James, N. P. (eds.): *Facies Models: Response to Sea Level Change*, 119–142. Geological Association of Canada, Stittville.
Molewski, P. & Olszewski, A. 2000: Sedimentology of the deposits and their palaeomorphological significance in the Gígjukvísl River Gap, marginal zone of the Skeiðarárjökull, Iceland. *Polish Polar Studies, XXVII International Polar Symposium, Torun*, 235–258.
Moorman, B. J., Robinson, S. D. & Burgess, M. M. 2003: Imaging periglacial conditions with ground-penetrating radar. *Permafrost and Periglacial Processes 14*, 319–329.

Neal, A. 2004: Ground-penetrating radar and its use in sedimentology: principles, problems and progress. *Earth Science Reviews 66*, 261–330.

Norðdahl, H. 1990: The Weichselian and Early Holocene deglaciation history of Iceland. *Jökull 40*, 27–50.

Norðdahl, H. & Pétursson, H. G. 2005: Relative sea-level changes in Iceland: new aspects of the Weichselian deglaciation of Iceland. *In* Caseldine, C., Russell, A., Harðardóttir, J. & Knudsen, O. (eds.): *Iceland: Modern Processes and Past Environments*, 25–78. Elsevier, Amsterdam.

Norðdahl, H., Ingólfsson, Ó., Pétursson, H. G. & Hallsdóttir, M. 2008: Late Weichselian and Holocene environmental history of Iceland. *Jökull 58*, 343–364.

Overgaard, T. & Jakobsen, P. R. 2001: Mapping of glaciotectonic deformation in an ice marginal environment with ground-penetrating radar. *Journal of Applied Geophysics 47*, 191–197.

Porter, C., Morin, P., Howat, I., Noh, M. J., Bates, B., Peterman, K., Keesey, S., Schlenk, M., Gardiner, J., Tomko, K., Willis, M., Kelleher, C., Cloutier, M., Husby, E., Foga, S., Nakamura, H., Platson, M., Wethington, M. Jr, Williamson, C., Bauer, G., Enos, J., Arnold, G., Kramer, W., Becker, P., Doshi, A., D'Souza, C., Cummens, P., Laurier, F. & Bojesen, M. 2018: ArcticDEM. https://doi.org/10.7910/DVN/OHHUKH. Harvard Dataverse, V1.

Price, R. J. 1969: Moraines, sandar, kames and eskers near Breiðamerkurjökull, Iceland. *Transactions of the Institute of British Geographers 46*, 17–43.

Reinardy, B. T. I. & Lukas, S. 2009: The sedimentary signature of ice-contact sedimentation and deformation at macro- and micro-scale: a case study from NW Scotland. *Sedimentary Geology 221*, 87–98.

Reynolds, J. M. 2011: *An Introduction to Applied and Environmental Geophysics.* 696 pp. Wiley-Blackwell, Chichester.

Roberts, M. J. 2005: Jökulhlaups: a reassessment of floodwater flow through glaciers. *Reviews of Geophysics 43*, 1–21.

Roberts, M. J., Palsson, F., Guðmundsson, M. T., Bjornsson, H. & Tweed, F. S. 2005: Ice-water interactions during floods from Grænalón glacier-dammed lake, Iceland. *Annals of Glaciology 40*, 133–138.

Ross, N., Sole, A. J., Livingstone, S. J., Igneczi, A. & Morlighem, M. 2018: Near-margin ice thickness from a portable radar: implications for subglacial water routing, Leverett Glacier, Greenland. *Arctic, Antarctic and Alpine Research 50*, S100007, doi: https://doi.org/10.1080/15230430.2017.1420949.

Rushmer, E. L. 2006: Sedimentological and geomorphological impacts of the jökulhlaup (glacial outburst flood) in January 2002 at Kverkfjöll, Northern Iceland. *Geografiska Annaler, Series A: Physical Geography 88*, 43–53.

Rushmer, E. L. 2007: Physical-scale modelling of jökulhlaups (glacial outburst floods) with contrasting hydrograph shapes. *Earth Surface Processes & Landforms 32*, 954–963.

Rushmer, E. L., Russell, A. J., Tweed, F. S., Knudsen, O. & Marren, P. M. 2002: The role of hydrograph shape in controlling glacier outburst flood (jökulhlaup) sedimentation. *In* Dyer, F., Thoms, M. C. & Olley, J. M. (eds.): *The structure, function and management implications of fluvial sedimentary systems*, 305–313. IAHS Publication 276. IAHS press, Wallingford.

Russell, A. J. & Knudsen, Ó. 1999: Controls on the sedimentology of the November 1996 jökulhlaup deposits, Skeiðarársandur, Iceland. *In* Smith, N. D. & Rogers, J. (eds.): *Fluvial Sedimentology VI*, 315–329. Blackwell Science, Oxford.

Russell, A. J. & Knudsen, Ó. 2002. The effects of glacier outburst flood flow dynamics on ice-contact deposits: November 1996 jökulhlaup, Skeiðarársandur, Iceland. *In* Martini, I. P., Baker, V. R. & Garzon, G. (eds.): *Flood and Megaflood Deposits: Recent and Ancient Examples*, 67–83. *Special Publication No. 32, International Association of Sedimentologists.* Blackwell Science, Oxford.

Russell, A. J. & Marren, P. M. 1999: Proglacial fluvial sedimentary sequences in Greenland and Iceland: a case study from active proglacial environments subject to jökulhlaups. *In* Jones, A. P., Tucker, M. E. & Hart, J. K. (eds.): *The Description and Analysis of Quaternary Stratigraphic Field Sections (Technical Guide 7)*, 171–208. Quaternary Research Association, London.

Russell, A. J., Knight, P. G. & van Dijk, T. A. G. P. 2001a: Glacier surging as a control on the development of proglacial, fluvial landforms and deposits, Skeiðarársandur, Iceland. *Global and Planetary Change 28*, 163–174.

Russell, A. J., Knudsen, Ó., Fay, H., Marren, P. M., Heinz, J. & Tronicke, J. 2001b: Morphology and sedimentology of a giant supraglacial, ice-walled, jökulhlaup channel, Skeiðarárjökull, Iceland: implications for esker genesis. *Global and Planetary Change 28*, 193–216.

Russell, A. J., Roberts, M. J., Fay, H., Marren, P. M., Cassidy, N. J., Tweed, F. S. & Harris, T. 2006: Icelandic jökulhlaup impacts: implications for ice-sheet hydrology, sediment transfer and geomorphology. *Geomorphology 75*, 33–64.

Russell, A. J., Tweed, F. S., Knudsen, Ó., Roberts, M. J., Harris, T. D. & Marren, P. M. 2002: Impact of the July 1999 jökulhlaup on the proximal River Jökulsá á Sólheimasandi, Mýrdalsjökull Glacier, southern Iceland. *In* Snorrason, Á., Finnsdottir, H. P. & Moss, M. (eds.): *The Extremes of the Extremes: Extraordinary Floods*, 249–254. IAHS Special Publication, Reykjavík, Iceland.

Smith, G. A. 1986: Coarse-grained nonmarine volcaniclastic sediment: terminology and depositional process. *Geological Society of America Bulletin 97*, 1–10.

Smith, L. C., Alsdorf, D. E., Magilligan, F. J., Gomez, B., Mertes, L. A. K., Smith, N. D. & Garvin, J. B. 2000: Estimation of erosion, deposition, and net volumetric change caused by the 1996 Skeiðarársandur jökulhlaup, Iceland, from synthetic aperture radar interferometry. *Water Resources Research 36*, 1583–1594.

Snorrason, Á., Jónsson, P., Sigurðsson, O., Pálsson, S., Árnason, S., Víkingsson, S. & Kaldal, I. 2002: November 1996 jökulhlaup on Skeiðarársandur Outwash Plain, Iceland. *In* Martini, I. P., Baker, V. R. & Garzon, G. (eds.): *Flood and Megaflood Processes and Deposits: Recent and Ancient Examples*, 55–65. *Special Publication of the International Association of Sedimentologists 32*. Blackwell Science, Oxford.

Storms, J. E. A., de Winter, I. L., Overeem, I., Drijkoningen, G. G. & Lykke-Andersen, H. 2012: The Holocene sedimentary history of the Kangerlussuaq Fjord-valley fill, West Greenland. *Quaternary Science Reviews 35*, 29–50.

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

Þórarinsson, S. 1939: The ice-dammed lakes of Iceland with particular reference to their values as indicators of glacier oscillations. *Geografiska Annaler 21*, 216–242.

Þórarinsson, S. 1972: Observations on the drainage and rates of denudation in the Hoffellsjökull District. *In* Embleton, C. (ed.): *Glaciers and Glacial Erosion*, 119–129. The Macmillan Press, London.

Þórarinsson, S. 1974: *Vötnin Strið: Saga Skeiðarárhlaupa og Grimsvatnagosa*. 254 pp. Bókaútgáfa Menningarsjóðs, Reykjavík.

Todd, S. 1996: Process deduction from fluvial sedimentary structures. *In* Carling, P. & Dawson, M. (eds.): *Advances in Fluvial Dynamics and Stratigraphy*, 299–350. Wiley, Chichester.

Tuttle, K. J., Øoustmo, S. R. & Andersen, B. 1997: Quantitative study of the distributary braidplain of the Preboreal ice-contact Gardermoen delta complex, southeastern Norway. *Boreas 26*, 141–156.

Weckwerth, P., Przegiętka, K., Chruścińska, A., Woronko, B. & Oczkowski, H. L. 2011: Age and sedimentological features of fluvial series in the Toruñ basin and the Drwęca valley (Poland). *Geochronometria 38*, 397–412.

Weckwerth, P., Sobota, I. & Greń, K. 2021: Where will widening occur in an outwash braidplain? A new approach to detecting controls on fluvial lateral erosion in a glacierized catchment (north-western Spitsbergen, Svalbard). *Earth Surface Processes & Landforms 46*, 942–967.

Weckwerth, P., Wysota, W., Piotrowski, J. A., Adamczyk, A., Krawiec, A. & Dąbrowski, M. 2019: Late Weichselian glacier outburst floods in North-Eastern Poland: landform evidence and palaeohydraulic significance. *Earth Science Reviews 194*, 216–233.

Winsemann, J., Hornung, J. J., Meinsen, J., Asprion, U., Polom, U., Brandes, C., Bußmann, M. & Weber, C. 2009: Anatomy of a subaqueous ice-contact fan and delta complex, Middle Pleistocene, North-west Germany. *Sedimentology 56*, 1041–1076.

Winsemann, J., Lang, J., Polom, U., Loewer, M., Igel, J., Pollok, L. & Brandes, C. 2018: Ice-marginal forced regressive deltas in glacial lake basins: geomorphology, facies variability and large-scale depositional architecture. *Boreas 47*, 973–1002.

Woodard, J. B., Zoet, L. K., Benediktsson, Í. Ö., Iverson, N. R. & Finlayson, A. 2020: Insights into drumlin development from ground-penetrating radar at Múlajökull, Iceland, a surge-type glacier. *Journal of Glaciology 66*, 822–830.

Zieliński, T. & van Loon, A. J. 2002: Present-day sandurs are not representative of the geological record. *Sedimentary Geology 152*, 1–5.

Zieliński, T. & van Loon, A. J. 2003: Pleistocene sandur deposits represent braidplains, not alluvial fans. *Boreas 32*, 590–611.