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Dynamics and glacial history of the Drangajökull ice cap, Northwest Iceland

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Leirufjörður valley and Drangajökull ice cap in August 2001, photograph by Oddur Sigurðsson.

Contents

Preface

This PhD project has been carried out as collaboration between the Department of Geology and Mineral Resources Engineering in the Norwegian University of Science and Technology and the Institute of Earth Sciences in University of Iceland, resulting in a joint PhD degree. This thesis contains result of research carried out in the years 2011-2015, including fieldwork both in summers and winters around the Drangajökull ice cap. The thesis is based on five papers listed below as Appendices I-V. The papers are either published, in review or as draft manuscript.

Appendix I
Brynjólfsson, S., Schomacker, A., Ingólfsson, Ó., 2014. Geomorphology of the Drangajökull ice cap, NW Iceland, with focus on its three surge-type outlets. Geomorphology 213, 292-304.

Appendix II
Brynjólfsson, S., Schomacker, A., Guðmundsdóttir, E.R., Ingólfsson, Ó., 2015. A 300-year surge history of the Drangajökull ice cap, northwest Iceland, and its maximum during the ‘ Little Ice Age`. The Holocene 25 (7), 1076-1092.

Appendix III
Brynjólfsson, S., Schomacker, A., Ingólfsson, Ó., Keiding, J.K., 2015. Late Weichselian-Early Holocene glacial history of Vestfirðir peninsula, northwest Iceland, constrained by ^{36}Cl cosmogenic exposure datings. Quaternary Science Reviews (in review).

Appendix IV
Brynjólfsson, S., Schomacker, A., Korsgaard, N.J., Ingólfsson, Ó., 2015. Elevation and volume changes of surge-type glaciers of Drangajökull ice cap, northwest Iceland. (draft manuscript).

Appendix V
Ingólfsson, Ó., Benediktsson, Í.Ö., Schomacker, A., Kjær, K.H., Brynjólfsson, S., Jónsson, S.A., Korsgaard, N.J., Johnson, M., 2015. Surging glaciers in Iceland – research status and future challenges. Earth-Science Reviews (in review).

Abstract

This thesis describes the glacial history, glacier dynamics, sediments and landforms of the Drangajökull ice cap as well as the glacial history and dynamics of the eastern Vestfirðir peninsula in northwest Iceland from the Late Weichselian until present. The aim was to reconstruct and improve the present understanding of the glacial history, surge history and dynamics of the Drangajökull ice cap.

The results reveal a topographically controlled ice sheet which more and less covered the Vestfirðir peninsula during the last glaciation. Cold-based non-erosive sectors of the ice sheet covered most of the mountains while fjords and valleys were occupied by dynamical, warm-based ice. Ice thinning and deglaciation started over the mountain plateaux 26 ka BP; the deglaciation was stepwise and asynchronous, uplands and some valleys were deglaciated 14-15 ka BP while valleys draining the main outlets of Drangajökull were occupied by outlet glaciers until c. 9 ka BP.

The forefield proximal to the present Drangajökull ice cap is characterised by thin, coarse grained till and locally weathered bedrock, except for the sandur covered valley floors. The landforms mapped at the surging outlet glaciers are not unique for surging glaciers, and furthermore the mapped landform assemblage does not resemble landsystem models for surging glaciers.

The surge-type outlets of Drangajökull reached their LIA maximum extent asynchronously during surges ~1700-1846 AD. Review of historical data and geomorphological mapping revealed twice as many surges than previously recorded. The surge interval varies from 10-140 years between and within the outlets. Surges were most frequent during the 19th century and the earliest 20th century. No clear relationship between surge initiation or periodicity and climate could be established. A distinct ice discharge occurs during surges, reflected in 10-30 m surface thinning of the upper reservoir areas and 10-120 m thickening of the receiving areas. During the present quiescent phase, the reservoir areas thicken by c. 0.5-0.7 m a^{-1} and the receiving areas thin by c. 1 m a^{-1}, which might bring the glacier surface to a pre-surge stage in 45-65 years.

Future studies could focus on extensive morphological mapping and direct dating of glacial features, aiming to add further details to the glacial history and test the reconstructions of Drangajökull presented here. Further investigation of the surge-type glaciers, e.g. extensive monitoring of weather and the glacier conditions, geophysical surveys both of the glaciers and their forefields, might also contribute to an improved understanding of the Drangajökull surging glaciers.

Ágrip

Í þessari ritgerð er fjallað um jöklunarsögu, sveiflur, setmyndanir og landmótun Drangajökuls. Einnig er fjallað um jöklunar- og afísunarsögu Vestfjarða, frá seinni hluta síðusta jökulskeiðs til nútíma. Aðalmarkmið rannsóknarinnar var að kanna betur sögu jöklunar og framhlaupa Drangajökuls.

Niðurstöðurnar benda til þess að Vestfirðir hafi að langmestu leiti verið huldir ís á hámarki síðasta jökulskeiðs, íshellan var undir sterkum áhrifum djúpra dala og fjarða Vestfjarðskagans. Þýðjöklar/ísstraumar flæddu tiltölulega hratt eftir fjörðum og dölum og rufu undirlag sitt. Ofarlega í fjallshlíðum eða yfir fjalllendi voru snörp skil í eiginleikum jökulíssins. Þar uppi var gaddjökull ráðandi sem flæddi hægt við innri aflögun og lét undirlag sitt að mestu ósnortið. Jökulhörfun hófst á hæstu fjöllum fyrir 26 þúsund árum, jöklaleysingin var ósamstíga milli svæða, en fyrir um 14-15 þúsund árum voru hærri landsvæði og sumir dalir þegar íslausir á meðan aðrir firðir og dalir máttu þola ágang meginskriðjökla Drangajökuls þangað til fyrir u.þ.b. 9 þúsund árum síðan.

Svæðið umhverfis Drangajökul einkennist af þunnum og grófum jökulruðningi í bland við ísmótaðar og veðraðar klappir en aðliggjandi dalbotnar eru að mestu huldir ármöl og áreyrum. Framhlaupsjöklar Drangajökuls náðu hámarksstærð hver í sínu lagi á árunum 1700-1846. Sögulegar heimildir og kortlagning á landmótunarsvæðum þeirra leiddi í ljós jökulgarða sem voru tvöfalt fleiri en áður skráð framhlaup Drangajökuls. Framhlaupahlé eru mjög óregluleg, 10-140 ár, en framhlaup virðast hafa verið einna tíðust á 19. öld og í upphafi 20. aldarinnar. Ekkert augljóst samband loftlags við eignleika og tíðni framhlaupanna er greinanlegt. Í nýyfirstöðnum framhlaupum Drangajökuls var algengasta yfirborðsþynning söfnunarsvæðanna á bilinu 10-30 m á meðan þykknun leysingasvæðanna var mun breytilegri, eða um 10-120 m. Þrátt fyrir neikvæða meðal afkomu Drangajökuls, það sem af er yfirstandandi kyrrfasa framhlaupsjöklanna, þykkna söfnunarsvæði þeirra að meðaltali um 0.5-0.7 m árlega. Uppbygging söfnunarsvæðanna ásamt um 1 m árlegri meðalþynningu leysingasvæðanna leiðir til brattara yfirborðs jökulsins og á 45-65 árum gæti yfirborðið orðið sambærilegt því sem var fyrir síðasta framhlaup jöklanna.

Rannsóknin hefur leitt í ljós allflókið mynstur íseiginleika, sveiflna og jöklunarsögu Drangajökuls. Vitneskju okkar um þessa þætti mætti auka með áframhaldandi kortlagningu, aldursgreiningu jökulmyndaðra landforma og jökulættaðs sets. Þannig mætti sannreyna þær niðurstöður sem hér eru kynntar og skerpa línur varðandi sögu og eiginleika síðasta jökulskeiðs. Eignleikum, eðli og sögu framhlaupsjöklanna í Drangajökli mætti gera betri skil. Gagnlegt gæti verið að beita fleiri aðferðum, t.d. langtíma vöktun á veðri og afkomu jöklanna, jarðeðlisfræðilegar kannanir á íseignleikum, undirlagi jöklanna og landmótunarumhverfi þeirra gætu einnig aukið skilning okkar.

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I want to thank co-authors Esther Ruth Guðmundsdóttir, Niels Jákup Korsgaard and Jakob Kløve Keiding for their analysis and contribution to the papers. Great thanks to Jón Bjarni Friðriksson, Julie Andreassen, Kathrine Naegelie and Sverrir Aðalsteinn Jónsson, who have walked many kilometres across mountains, valleys and glacier rivers carrying backpacks full of rocks and other interesting things in the field. Halldór G. Pétursson, Ívar Örn Benediktsson, and Lena Håkansson have been very supportive and interested in my work, always ready to discuss things and provide help, literature and data that have been needed. The local people that own the abandoned farms and huts in the remote vicinity of Drangajökull are thanked for their great hospitality during the fieldwork. Ásgeir Sólbergsson, Indriði Aðalsteinsson, Þórður Halldórsson, Þröstur Jóhannesson and Sigurður Sigurðsson are thanked for much valuable information and interesting stories from the Drangajökull area. The Icelandic Institute of Natural History kindly provided me a four years leave from my work which enabled all this to happen.

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1. Introduction

1.1 Background

When glaciers grow and decay, they leave certain fingerprints as evidence for their presence. Those fingerprints, such as moraines and distribution of till and erratic boulders are fundamental to configure marginal positions and the size of former ice sheets and glaciers. More detailed information about processes and properties of former glaciers can be established on the basis of thorough mapping and investigation of landforms and sediments. Thermal regime, subglacial characteristics, basal hydrological conditions, ice velocity, flow direction and type of glaciation are among things which can be reconstructed.

In terms of the Drangajökull ice cap, the glacial history, configuration and properties are relatively poorly known from the Last Glacial Maximum (LGM) until the late Holocene. Careful study of those fingerprints along with a good chronological control can be a key to past developments, present conditions and at same time for predictions on the future development of the ice cap.

An ongoing debate is whether or not the Vestfirðir peninsula hosted an independent ice cap with valley glaciers that left relatively large areas ice free, rather than an extensive ice sheet totally covering the area (Þórarinsson, 1937; Hoppe, 1982; Hjort et al., 1985; Andrews et al., 2002; Geirsdóttir et al., 2002; Hubbard et al., 2006; Principato et al., 2006; Norðdahl et al., 2008; Principato, 2008). Furthermore, glacial conditions like, if some sectors of the Icelandic Ice Sheet (IIS) were cold based or not, are little discussed and remain largely unknown.

Based on glacial geological studies which provide indirect dating from sediment cores and a few geological sections, a deglaciation history has been proposed for the Drangajökull ice cap and the Vestfirðir peninsula (Hjort et al., 1985; Sywitisky et al., 1999; Andrews et al., 2000, 2002; Geirsdóttir et al., 2002; Castañeda et al., 2004; Principato, 2003, 2008). However, it suffers from a lack of chronological control (Geirsdóttir et al, 2009), and challenging questions about the configuration, thickness, ice dynamics and thermal conditions remains unanswered. A recent ^{36}Cl nuclide exposure dating of glacial landforms (Principato et al., 2006) and a subsequently established production rate for the ^{36}Cl nuclide in Icelandic basalt (Licciardi et al., 2008) indicate a potential to clarify those problems and significantly improve the glacial history.

Because research on Drangajökull has focused on geomorphology and potential connections between marginal fluctuations and climate, the fluctuations of the three main outlets of the ice cap are relatively well recorded (Thoroddsen, 1933, 1958; Eyþórsson, 1935; Þórarinsson, 1943; Lewis, 1964; John and Sugden, 1962; Sigurðsson, 1998, 2000;

Þrastarson, 2006). However, the surge-type glaciers and their forefields have not been studied in terms of the recorded surge activity. This stimulates several important questions about the main surge characteristics of Drangajökull; is the surge duration, quiescent phase, accelerating and decelerating phase, pattern of ice discharge, sediment-landform assemblage and sediment distribution of Drangajökull surges similar to other surge-type glaciers? If not, what are the factors controlling the different expression of the surge characteristics? Furthermore, do the recorded surges represent the complete surge history of the Drangajökull ice cap?

1.2 Project objectives

The Drangajökull ice cap is drained by three main outlet glaciers that are known to have retreated considerably and surged 3-4 times each since their maximum extent during the Little Ice Age (LIA). Our understanding of the glacial conditions of the Vestfirðir peninsula from the Late Weichselian until the end of the LIA, suffers substantially from a lack of terrestrial data, mainly direct datings of landforms and sediments but also geomorphological data. Due to the location of the ice cap on Vestfirðir, the northwest peninsula of Iceland, it is sensitive to flow variations of either the cold polar water from North or the warmer Atlantic water from South, which converge in the ocean just off the peninsula (Bergþórsson, 1969; Ogilvie, 1984; Eiríksson et al., 2000). Despite substantial fluctuations, at least during the last three centuries, and its uniqueness in terms surge behaviour and glaciological conditions, Drangajökull has received surprisingly little attention from the scientific community. Earlier studies from the area were focused on marginal fluctuations and geomorphology in general (Thoroddsen, 1933, 1958; Eyþórsson, 1935; John and Sugden, 1962; Lewis, 1964; Þrastarson, 2006). In a recent pilot study, Principiato (2003) described numerous sites with glacial landforms and raised beaches. Her pioneering work with cosmogenic exposure dating in Iceland and glacial-geological study around the Drangajökull ice cap (Principato et al., 2006; Principato, 2008; Principato and Johnson, 2009) revealed potentials to significantly improve our knowledge of the glacial conditions and the glacial history in northwest Iceland since the Late Weichselian.
The overall aim of this project was to map, describe and interpret landforms and sediments around Drangajökull, and improve our current knowledge of the glacial conditions and the glacial history of the Drangajökull ice cap.
The specific aims were:

- To reconstruct the Late Weichselian – Early Holocene glacial history and conditions of the Drangajökull ice cap and Vestfirðir peninsula, northwest Iceland.
- To investigate the timing and dynamics of late Holocene surges of Drangajökull outlet glaciers.

- To map and interpret the geomorphology of the modern surging glacier landsystems at Drangajökull and establish a conceptual landsystem model.
- To assess the recent ice volume changes of the Drangajökull ice cap and investigate the coupling between climate and glacier changes.

1.3 Glaciers and glacial history of Iceland

Glaciers cover about 11% of Iceland with the largest ice cap, Vatnajökull, being about 8100 km^2. The larger ice caps are all located in south and central Iceland, except the fifth largest ice-cap, Drangajökull, in northwest Iceland. About 150 valley and cirque glaciers are located in the Tröllaskagi peninsula in central north Iceland, and several small glaciers are located in the east Iceland and the northwest peninsula. All Icelandic glaciers are considered warm based at present. All the major ice caps have surge-type outlet glaciers with over 80 recorded advances (Björnsson et al., 2003; Björnsson and Pálsson 2008; Ingólfsson et al., in review).

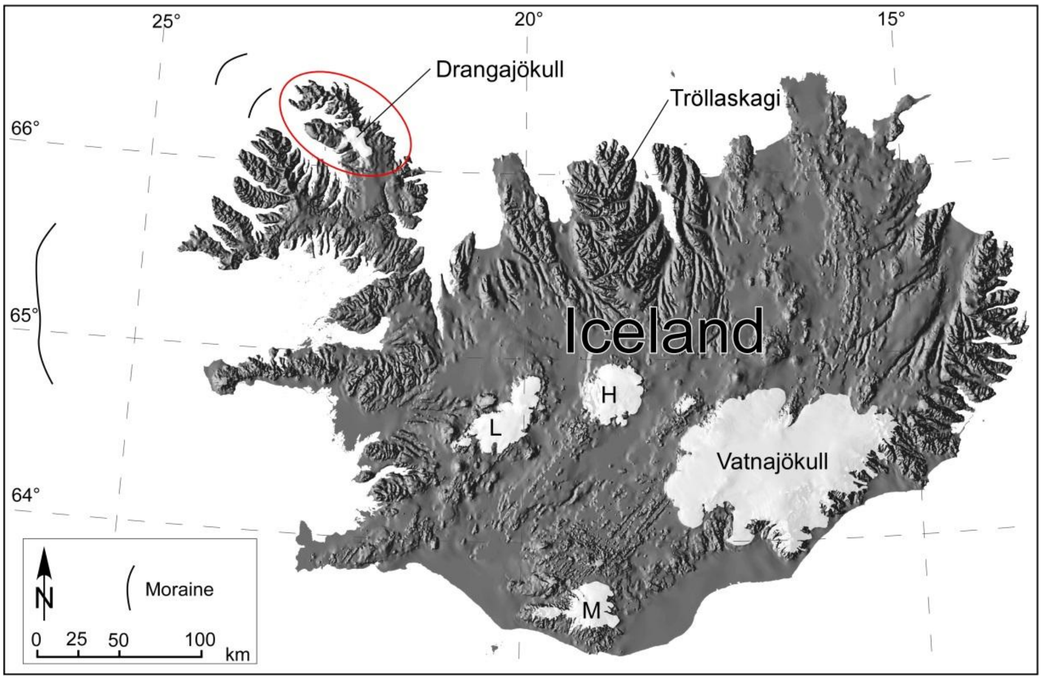

Figure 1. Overview of the major ice caps in Iceland (L=Langjökull, H=Hofsjökull, M=Mýrdalsjökull). Moraines on the shelf of Vestfirðir considered as potential LGM position of the Icelandic Ice Sheet are marked (Ólafsdóttir et al., 1975; Andrews et al., 2002; Gerisdóttir et al., 2002).

The LGM position of the IIS, at about 18-21 ka BP, is not spatially well constrained (Hoppe, 1982; Ingólfsson and Norðdahl, 2001; Geirsdóttir et al., 2007). Marine data and various terrestrial morphological features indicate a 1000-2000 m thick ice sheet over the island, extending to the shelf areas (Ólafsdóttir et al., 1975; Syvitsky et al., 1999; Eiríksson et al., 2000; Andrews et al., 2000; Geirsdóttir et al., 2002; Norðdahl and Pétursson, 2005; Hubbard et al., 2006; Norðdahl et al., 2008). Small ice free areas most likely existed along high coastal mountains in the northwest, north and east Iceland (Einarsson and Albertsson, 1988; Ingólfsson, 1991; Roberts et al., 2007). Contrastingly, other studies suggest local glaciation characterised by individual ice-caps drained by valley glaciers and leaving relatively large ice free areas in the Vestfirðir peninsula (Þórarinsson, 1937; Sugden and John, 1976; Sigurvinsson, 1983; Hjort et al., 1985; Norðdahl, 1990; Rundgren and Ingólfsson, 1999; Andrews et al., 2002; Principato, 2008).

About 14-15 ka BP, the marine based ice sheet broke up and retreated rapidly to present day dry land (Syvitsky et al., 1999; Eiríksson et al., 2000; Andrews et al., 2002; Geirsdóttir et al., 2002). Several moraines and clusters of moraines mapped in the lowland and central Iceland, together with observations of the Vedde and Saksunarvatn tephras indicate a step-wise and rapid terrestrial deglaciation (Norðdahl, 1990; Kalddal and Víkingsson, 1991; Ingólfsson, 1991; Ingólfsson et al., 1997). Recent studies indicate that central Iceland, the highland, was largely deglaciated by 10.2 ka BP (Larsen et al., 2012).

At least some ice caps, e.g. Langjökull (Fig. 1), are considered to have been largely absent during the Holocene Thermal Maximum (HTM) and started to re-grow about 5 ka BP at the onset of Neoglaciation (Larsen et al., 2012; Geirsdóttir et al., 2013). Icelandic glaciers fluctuated during the Neoglaciation and most of them reached their maximum extent during LIA, either in the 18th or 19th century (Grove, 1988; Stötter et al., 1999; Kirkbride and Dugmore, 2006, 2008; Larsen et al., 2011, 2012).

1.4 Glacier surges

Glacier surging is a cyclic flow instability, characterised by switches between phases of fast flow, an active phase, and slow flow, a quiescent phase. Both temperate and polythermal glaciers surge. They cluster in certain areas, indicating rather poorly known environmental and glaciological factors that control their location (Meier and Post, 1969; Þórarinsson, 1969; Raymond, 1987; Sharp, 1988; Jiskoot et al., 2000; Murray et al., 2003; Benn and Evans, 2010; Sevestre and Benn, 2014). However, their occurrence is most common where climatic conditions are bounded by mean annual temperature approximately 0-10°C and annual precipitation of 200-2000 mm (Sevestre and Benn, 2014). The surge phase of temperate glaciers typically last 1-2 years or less, while a 3-10 years surge duration is common for polythermal glaciers in Svalbard. The length of the quiescent phase is more inconsistent within regions, e.g. 30-500 years in Svalbard, 10-140 years in Iceland and, typically 20-40 years elsewhere (Meier and Post, 1969; Sharp, 1988; Hamilton and Dowdeswell, 1996; Jiskoot et al., 2000; Murray et al., 2003; Björnsson et al., 2003). During the active phase, ice is usually transported down-glacier from a reservoir area, down to a receiving area in the lower part of the glacier, with ice velocity often 10-100 times greater than during the quiescent phase (Meier and Post, 1969; Þórarinsson, 1969, Raymond, 1987).

The ice surface morphology change considerably during a surge; draw-down of the reservoir areas and thickening of the receiving areas usually cause a marginal advance and decreased surface gradient of the glacier. During the quiescent phase, snow and ice accumulate in the reservoir area while ice thinning and marginal retreat occur in the receiving area. This contributes to a gradually steeper surface gradient of the glacier, similar to the pre-surge stage. Sufficient development of a steep surface profile in complex combination with other factors, e.g. the substratum condition, hydrological conditions and thermal regime have been considered as critical factors to favour a new surge (Kamb, et al., 1985; Raymond, 1987; Dowdeswell et al., 1995; Eisen et al,. 2001; Murray et al., 2003; Frappé and Clarke, 2007; Sund et al., 2009).

Studying geological fingerprints of surge-type glaciers have resulted in establishment of glacier landsystem models (Evans and Rea, 1999, 2003; Brynjólfsson et al., 2012; Schomacker et al., 2014). They are considered useful to recognise the imprints of surging glaciers, reconstruct past glacier activity and its climate implications (Evans and Rea, 1999, 2003). However, on a timescale of decades or perhaps centuries, surging glaciers are not good measurements on climate fluctuations, because of their non-direct response to climate (Yde and Paasche, 2010; Benn and Evans, 2010; Striberger et al., 2011).

1.5 Cosmogenic exposure dating

Cosmogenic exposure dating is used to estimate how long a rock surface has been exposed to cosmic radiation on the Earth's surface. It is widely used by geomorphologists to date various geological events, recorded both by sediments and bedrock, e.g. glacier advance and retreat, landslides, and lava flows. The method is particular useful to date glacier fluctuations because glaciers commonly create fresh rock surfaces and landforms by subglacial erosion, which are free of cosmogenic nuclide inheritance (Dunai, 2010).

The main portion of the cosmic nuclides is formed in the surface rock minerals when galactic cosmic rays bombard the Earth's surface. With time the cosmic nuclides build up in minerals exposed to these cosmic rays. Thus, by measuring their concentration in a rock, it can be estimated how long that rock has been exposed (Lal, 1991; Gosse and Philips, 2001; Ivy-Ochs and Kober, 2008; Dunai, 2010).

Several factors can affect the production rate of cosmogenic nuclides in the rocks, e.g. latitude, topographic shielding, position in the Earth's geomagnetic field and atmosphere shielding, i.e. altitude. Those factors can be corrected for by scaling algorithms established by Lal (1991), Stone (2001), Dunai (2001) and Desilets et al. (2006). Erosion, snow cover and exhumation can also affect the cosmogenic nuclide production, by estimating the degree of weathering and potential average snow cover those factors can be approximately corrected for (Dunai, 2010; Balco, 2011).

The most commonly applied nuclides are ^{10}Be, ^{14}C, ^{26}Al and ^{36}Cl and the stable noble gases ^{3}He and ^{21}Ne, each of them applies to specific minerals and elements. ^{36}Cl is the only nuclide that can be applied on Icelandic basalt except if it contains substantial olivine phenocrysts which enable ^{3}He to be applied.

2. Setting

Iceland is located approximately at 63-67 °N and 13-24 °W, near the borders of cool oceanic and atmospheric currents from the north and warmer currents from the south (Fig. 2a). This makes Icelandic glaciers sensitive to climate variations and provides an important arena for studying glacier fluctuations and North Atlantic palaeoclimate (Bergþórsson, 1969; Björnsson, 1979; Guðmundsson, 1997; Eiríksson et al., 2000; Geirsdóttir et al., 2009). This study is focused on the Drangajökull ice cap and its surroundings, located on the eastern Vestfirðir peninsula in northwest Iceland (Fig. 2b).

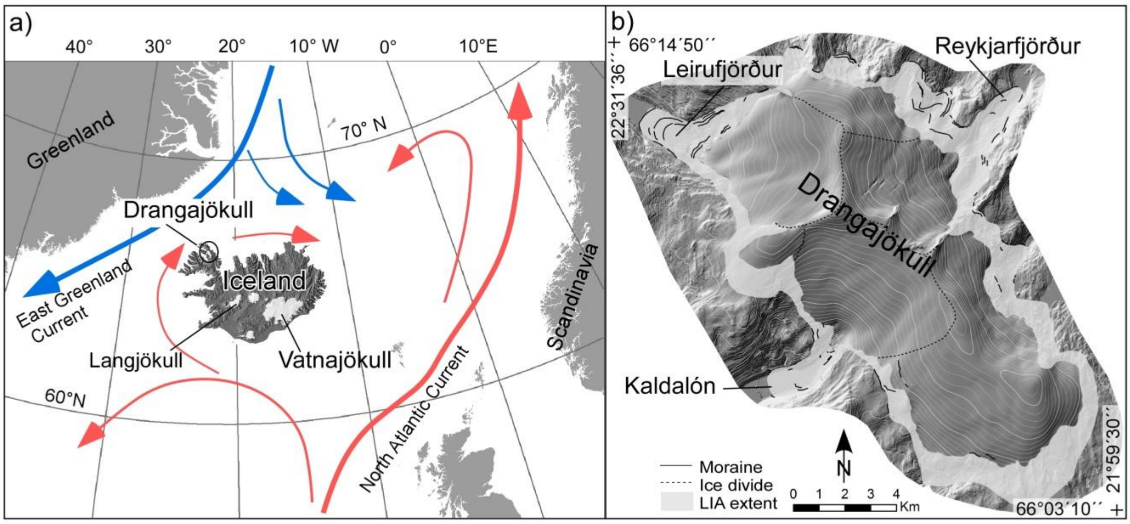

Figure 2. a) Location of Iceland at the boarder of cool and warmer ocean currents centrally in the North Atlantic Ocean (modified from Eiríksson et al., 2000). b) Drangajökull ice cap, its Little Ice Age extent and the three main focus areas Reykjarfjörður, Leirufjörður and Kaldalón.

2.1 Vestfirðir

The Vestfirðir peninsula is located approximately at 66 °N and 23 °W (Fig. 3a). The bedrock consists mainly of 16-7 million years old sub-aerial tholeiitic and porphyritic basalts and some outcrops of olivine basalts. These Miocene flood basalts are interbedded with thin sedimentary layers and occasional volcanoclastic sedimentary horizons (Sæmundsson, 1979; Einarsson, 1991; Kristjánsson and Jóhannesson, 1994; Guðmundsson et al., 1996; Harðarson et al., 1997). The eastern part of Vestfirðir is a relatively large 350-600 m high upland hosting the Drangajökull ice cap (Fig. 3b). In other areas of the Vestfirðir peninsula, the landscape is characterized by glacially eroded steep fjords and valleys, often confined by 500-700 m high basaltic plateaux. The plateaux surfaces are commonly characterized by block fields with 2-4 m wide surface polygons or sorted stripes. Glacial sediments, except for scattered erratic boulders, are absent in the block fields that we explored during field work. However, a diamict, either locally weathered bedrock or till, was observed here and there in between the block field areas. Erratic boulders, often sub-rounded and in some cases relatively fresh looking, occur from lowland locations up to 500-600 m high mountain passes and occasionally up on plateaux and uplands around the ice cap (Paper I and III).

A more alpine landscape, with cirques and short valleys incised between plateaux and horns dominates the Hornstrandir area (Fig. 4d), the northern tip of the Vestfirðir peninsula about 25 km north of Drangajökull. This setting is considered to indicate that the glaciation of the Hornstrandir area was of a local character, rather than covered by the main IIS (Sugden and John, 1976; Símonarson, 1979; Hjort et al., 1985; Principiato et al., 2006).

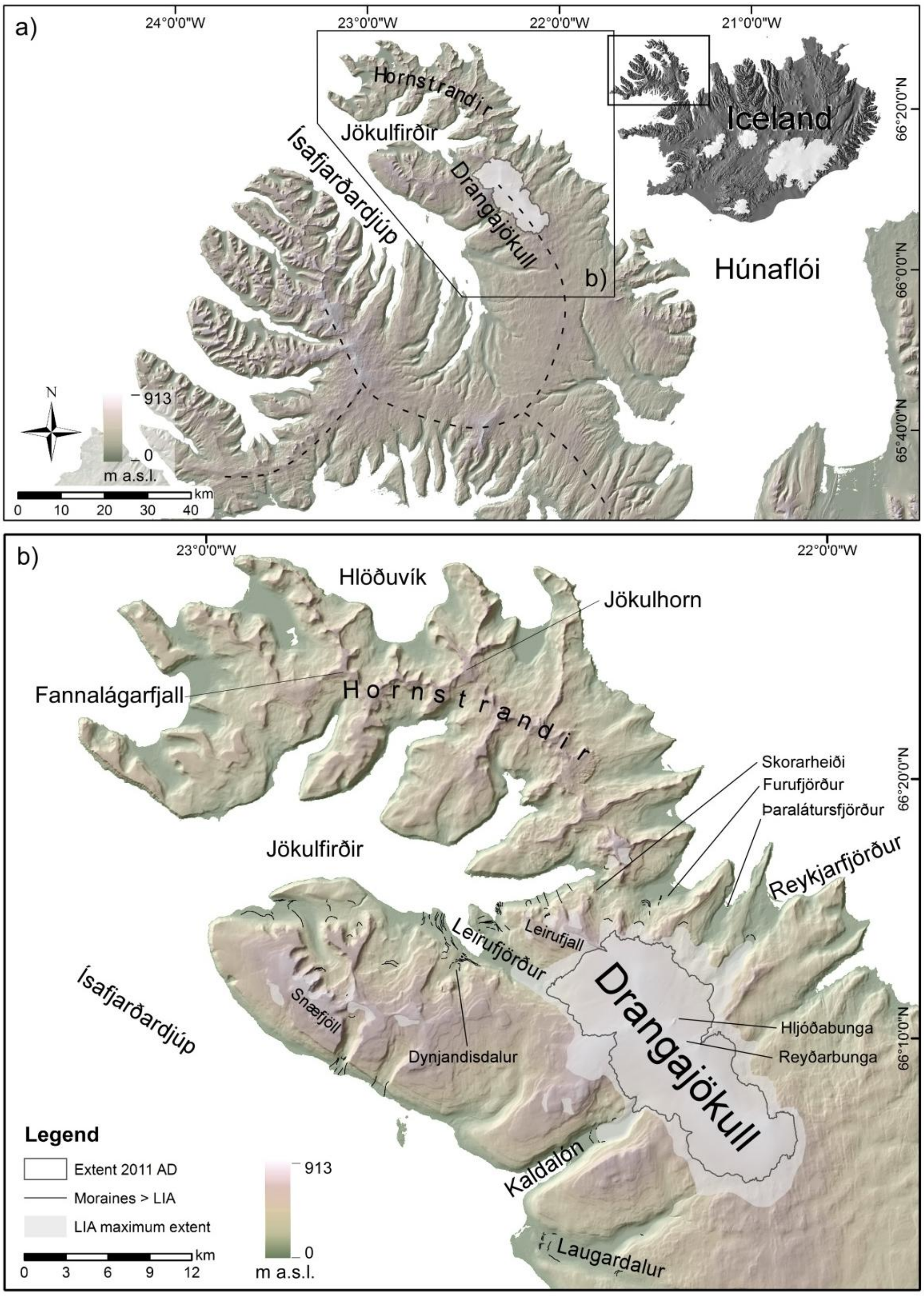

Figure 3. The study area at the Drangajökull ice cap, its surroundings and the Vestfirðir peninsula. The topographical maps are based on a digital elevation model from Loftmyndir ehf. a) Overview of the Vestfirðir peninsula, dashed lines indicate ice divides during LGM (Norðdahl, 1991; Ingólfsson and Norðdahl, 2001). b) Drangajökull and the study area with location of sites discussed in the papers, present and maximum extent during LIA are shown.

2.2 Drangajökull ice cap and its surroundings

The dome shaped Drangajökull ice cap is located approximately 100 to 915 m above sea level (a.s.l.) on the highland plateau of the eastern Vestfirðir peninsula in northwest Iceland (Figs 2 and 3). The ice cap is considered warm based, as all other glaciers in Iceland (Björnsson et al., 2003). A low glaciation limit and an average equilibrium line altitude (ELA) between 550-650 m a.s.l. make the glaciological conditions of Drangajökull unique, as ELA and glaciation limits at other Icelandic ice caps usually lie above 1100 m a.s.l. (Eyþórsson, 1935; Björnsson, 1979; Björnsson and Pálsson, 2008). Those conditions of Drangajökull are considered to be amplified by the proximity to Greenland and the cold polar East Greenland Current (Fig. 2a), and to reflect low summer temperature and high precipitation in the area (Eyþórsson, 1935; Bergþórsson, 1969; Ogilvie, 1984; Eiríksson et al., 2000).

Recent aerial photograph studies and geomorphological mapping suggest that Drangajökull extended over about 190-215 km^2 during the LIA maximum (Sigurðsson et al., 2013; Paper I and II; Fig. 2b). The oldest reliable size estimation of the ice cap gave about 200 km^2 approximately by the end of the LIA (Þórarinsson, 1943, 1958). Recent measurements of its size gave 146 km^2 in 2004 (Sigurðsson et al., 2013) and 142 km^2 in 2011 (Jóhannesson et al., 2013).

At present, the ice cap is mainly drained by three surge-type outlet glaciers, Reykjarfjarðarjökull to the northeast, Leirufjarðarjökull to the northwest and Kaldalón to the west (Figs 2b and 4). These glaciers have been recognised to surge 2-4 times, with surge duration of 4-7 years and 50-140 years surge interval (Eyþórsson, 1963; Sigurðssoon, 1998; Björnsson et al., 2003; Þrastarson, 2006). Currently the glaciers are all in their quiescent phase.

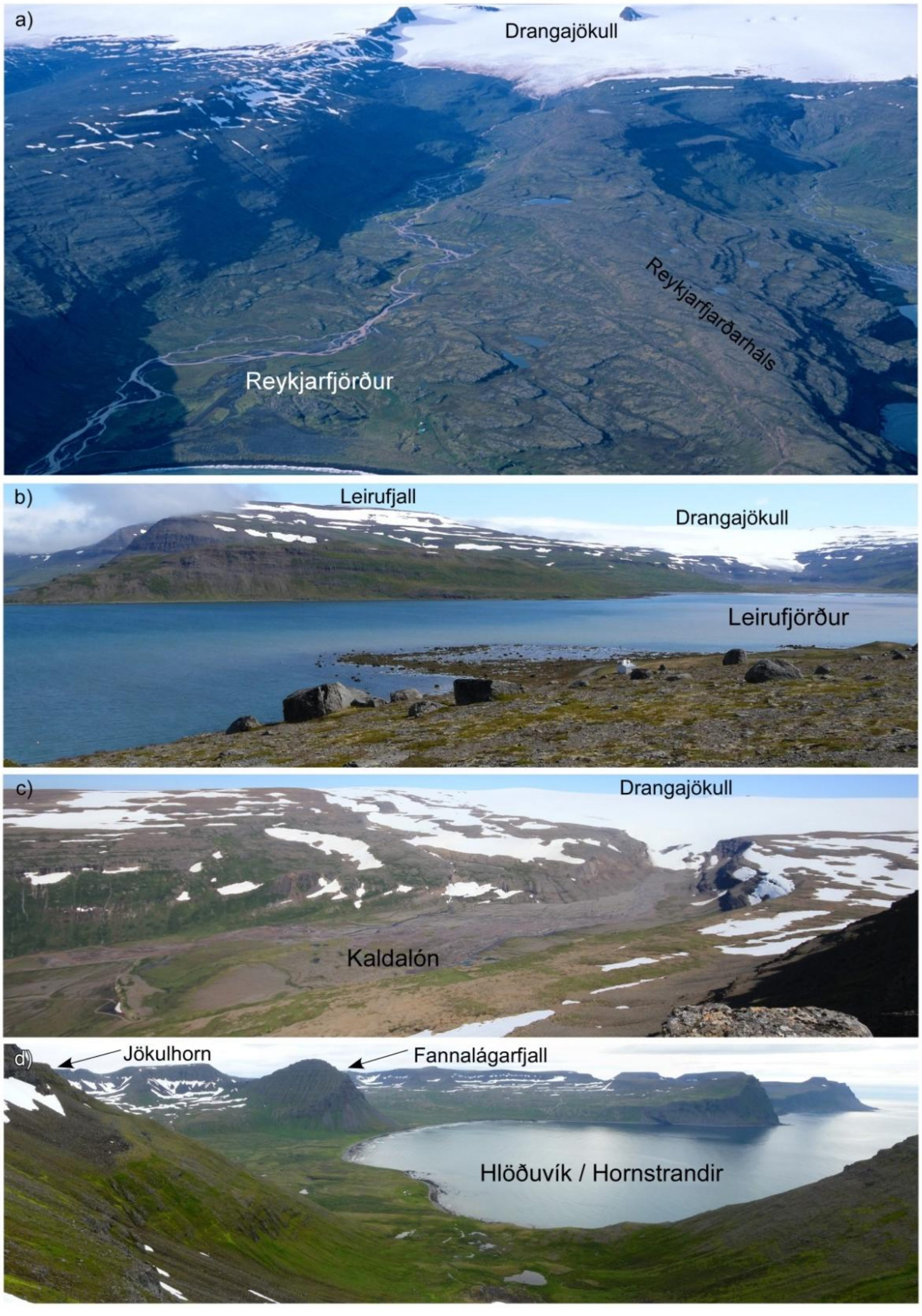

Figure 4. The three main study areas. a) View into Reykjarfjörður towards Drangajökull in the southwest (Oddur Sigurðsson, 2001). b) The east side of Leirufjörður, view towards Drangajökull in southeast, July 2012. c) The inner part of Kaldalón and Drangajökull in the east, July 2013). d) Hlöðuvík in the Hornstrandir area, view towards west, July 2014.

The most prominent geomorphological features around the northern perimeter of Drangajökull are end- and lateral moraines, flutes, outwash plains and periglacial features; such as block field, polygons and sorted stripes on uplands and plateaux above c. 400 m a.s.l. (Paper I; Fig. 4). Several end moraines have been mapped in the forefields of the three surge type glaciers (John and Sugden, 1962; Principato, 2008; Paper I). The outermost moraines indicate maximum extent during the LIA and were formed during surges in the 18th and 19th century (Thoroddsen, 1933, 1958; Eyþórsson, 1935; Þórarinsson, 1943; Björnsson et al., 2003). In Kaldalón, dating of two additional end moraines distal to the LIA maximum moraine have yielded Younger Dryas and Neoglacial ages (Principato et al. 2006; Principato, 2008).

The sediments around Drangajökull, especially in topographically higher areas, are generally a thin sheet of till or locally weathered rocks and often protruded by ice sculpted bedrock. The valleys are the main depocenters partly filled up with fluvial sediments (Paper I). Raised beaches have been recognised from 5 to 48 m a.s.l. on the eastern Vestfirðir (John and Sugden, 1962; Principato, 2008; Brynjólfsson et al., 2014). In the mouth of Kaldalón, they are about 20-30 m a.s.l., 14 m a.s.l. in the mouth of Jökulfirðir and about 5 m a.s.l. in Reykjarfjörður and Leirufjörður (Principato, 2008; Paper II).

2.2 Climate

The present climate of Iceland is classified as cool, temperate maritime with an average annual temperature of 4-5 °C (Einarsson, 1976; Jónsson, 1993; Hanna, et al., 2004). The regional climate of the eastern Vestfirðir peninsula is cool, characterised by a temperature and precipitation gradient from the northeast to southwest. The summers have 6-8 °C average temperature from June-September, and the annual average temperature is 2.5-4 °C (Einarsson, 1976; Jónsson, 1993). About 1100 mm annual average precipitation on the northeast coast and about 580 mm on the west coast of the ice cap, contrast considerably from a modelled 2500-3000 mm annual precipitation on the ice cap (Einarsson, 1976; Crochet et al., 2007). Perennial snow fields occurring on lee sides for the dominating northeasterly winds demonstrate a profound wind effect on the snow accumulation in the area (Paper I).

3. *Methods*

The fieldwork was carried out around the perimeter of the Drangajökull ice cap during the summers 2011-2013 and in the Hornstrandir area, about 20 km north of the ice cap, during the summer 2014. During the fieldwork, base-camps were in the forefields of the three main outlet glaciers of Drangajökull and the small bay Hlöðuvík, in the Hornstrandir area. All these areas are remotely located and were accessed by STOL aircraft on tundra tires or boat, except Kaldalón which can be approached by a normal car. The base data used for all maps and many of the figures in Appendices I-VI, are aerial orthophotographs, provided by Loftmyndir ehf., recorded in 2005 with 0.5 m ground resolution, and a LiDAR (Light Detection and Ranging) derived digital elevation model (DEM) recorded in 2011 with 5 m ground resolution, provided by the Icelandic Met Office (IMO). The ISN93/WGS84 reference system was used for data handling and the final maps. The geographic information system (GIS), ArcMap 10, was used to manually digitise all the mapped features. The drawing software Canvas, version 11, was used to produce sketches, drawings, logs, process figures and make final refinements of the maps. Overview of the main methods and their relationship to the papers is shown in Figure 5.

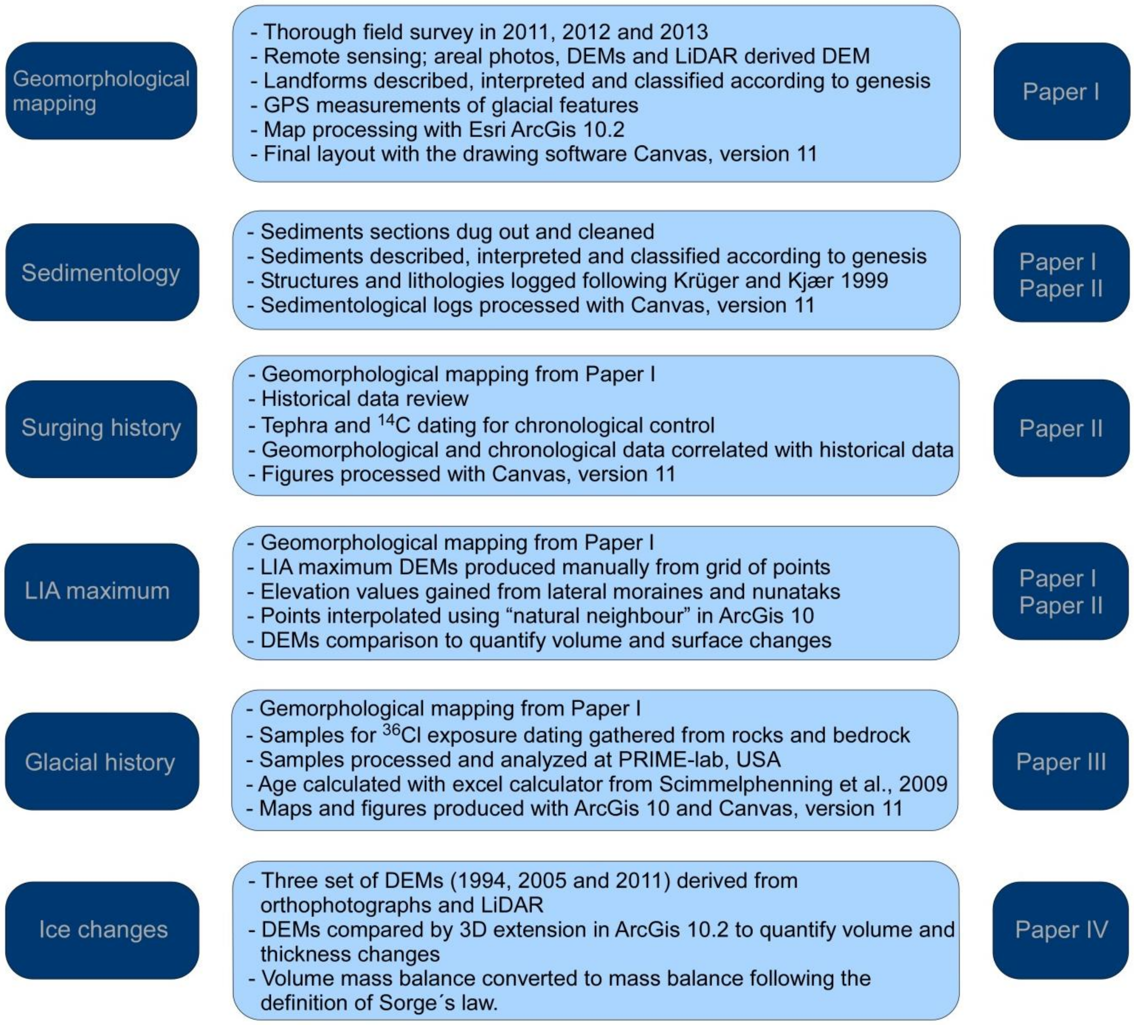

Figure 5. Overview of the main methods and data applied during each part of the work.

Table 1. Overview of the aerial photographs and DEMs used in this study

Date	Recorded By	Flying altitude (m a.s.l.)	Cell size (m)	DEM RMS error (m)	Product
29.08. 1994	Landmælingar Íslands	5486	2	1.2	Orthophotograps +DEM
2005	Loftmyndir ehf	c. 3000	0.5	~ 3	Orthophotographs + DEM
20.07. 2011	Icelandic Met Office	c. 2500	5 m	0.5	LiDAR DEM

3.1 Geomorphological mapping

Three detailed glacial geomorphological maps, illustrating all observed glacial related landforms, from each of the surge-type outlet glaciers, Reykjarfjörður, Leirufjörður and Kaldalón, were produced. One large overview geomorphological map presents the main sediments and landforms around Drangajökull ice cap, and a reconstructed maximum extent of the ice cap during the LIA. Initially, preliminary maps were produced by remote studies, based on the orthorectified aerial photographs and the LiDAR derived DEM (Table 1). The initial maps from each area were tested and improved during the fieldwork. Classical methods for mapping glacial geomorphology were applied (Hubbard and Glasser, 2005); landforms and sediments were described, interpreted and classified according to genesis. Boundaries and locations of the mapped features, which were unclear on the aerial photos, were located with accuracy of ±3 m by a hand held GPS, Garmin GPSMAP 62sc. Specific emphasis was on mapping the landforms that are representative for former glacier extension, such as end moraines, lateral moraines, raised beaches and other landforms and sediments relevant for the glacial history. Polygons and polylines were used to delineate the mapped features.

3.2 Surge history

The reconstruction of the surge history in Paper II is established on historical data and the field-based work on geomorphology and sediments. People who lived in the vicinity of the valleys, which the outlet glaciers drain into, were aware of environmental changes and glacial fluctuations, which often were recorded in local annals and church books (Magnússon and Vídalín, 1710; Ólafsson and Pálsson, 1772; Þórarinsson, 1943).

Fluctuations of Drangajökull in the last three centuries were documented in two particular valuable surveys around the ice cap, in 1886-1887 by Thoroddsen (1933, 1958) and in 1931 by Eyþórsson (1935). Their documentations were based on information on glacier fluctuations given by the local people, geomorphology and the location of the glacier at those times. Furthermore, Eyþórsson (1935) installed markers for measurements of frontal fluctuations of the outlet glaciers Reykjarfjarðarjökull, Leirufjarðarjökull and Kaldalónsjökull. Since then the Icelandic Glaciological Society (IGS) has measured the annual frontal fluctuations.

For the reconstruction of the surge history, end moraines together with the historical data were of special interest. The end moraines and the distance between them were surveyed and measured both in the field and on the aerial photos. Some of the moraines were initially of known age and had been related to recorded surges of the outlet glaciers (Þórarinsson, 1969; Björnsson et al., 2003). A relative age of undated moraines were constrained by correlating historical data, moraines of known age, moraines of unknown age and a mean retreat rate of certain periods. Two moraines were directly dated by tephra. Sections providing organic material for ^{14}C dating or tephra for absolute dating are very restricted.

The maximum surge conditions of the glaciers were reconstructed by lateral moraines located on the valley slopes and which correlate with the end moraines that formed during the LIA maximum surge conditions of the glaciers. Lateral moraines are typically formed along valley slopes, demonstrating the ice thickness during glacier advance or still stand (Benn and Evans, 2010). A DEM for each of the surging outlet glacier was produced from a manually made network of points that were given a height value. The spatial distribution of the points, and the maximum area of the surging glaciers during LIA was constrained by the end moraines, lateral moraines and ice divides estimated from shaded relief LiDAR derived DEM (Table 1; Fig. 2b). The altitude values for the points were yielded from the lateral moraines and information about approximately 50 m ice thinning around the nunatak, Reyðarbunga since 1910 (Fig. 3b; Eyþórsson, 1963). The ice thinning was used as a fixed minimum ice thinning for the reservoir area of the present surging glaciers. Finally, the DEMs demonstrating LIA maximum surge conditions were compared with the LiDAR derived DEM from 2011, yielding approximate ice thickness changes, volume changes and aerial changes of the surging outlet glaciers since their LIA maximum conditions until 2011.

3.3 ^{36}Cl exposure dating

In order to obtain deglaciation ages of certain localities around the Drangajökull ice cap, 24 whole rock samples were dated by the ^{36}Cl nuclide cosmogenic exposure dating. A sampling strategy of Ivy-Ochs and Kober (2008) and Balco (2011) was followed. From the northern perimeter of Drangajökull, 21 samples were dated and three from the Hornstrandir area (Figs 3 and 6). Twelve samples were from ice sculpted bedrock, six from moraine boulders and six from erratic boulders. Potential error sources for the age results, such as topographic/snow shielding and erosion, were evaluated in the field and corrected for or avoided if possible. Relative fresh samples, located topographically high, were chosen to avoid errors by snow or sediment cover. Problems because of tip-over of boulders and exhumation (past burial) were avoided by choosing large moraine boulders and erratic boulders preferably located on bedrock or relatively stable blockfield. Finally, topographic shielding was measured by compass and clinometer in the field, and corrected for subsequently.

About 2-5 cm thick samples were hammered and chiselled from the rocks and subsequently crushed to grain size about 147-250 μm in the lab at the Department of Geology and Mineral Resources Engineering at NTNU. All further preparations of the samples, e.g. cleaning, chemical treatment and finally the ^{36}Cl analysis were carried out by the PRIME Lab, Purdue University, Indiana, USA. Finally the exposure ages were calculated using an Excel spread sheet developed for calculating ^{36}Cl ages (Schimelpfenning et al., 2009) and the scaling and shielding factors were calculated by CosmoCalc (Vermech, 2007).

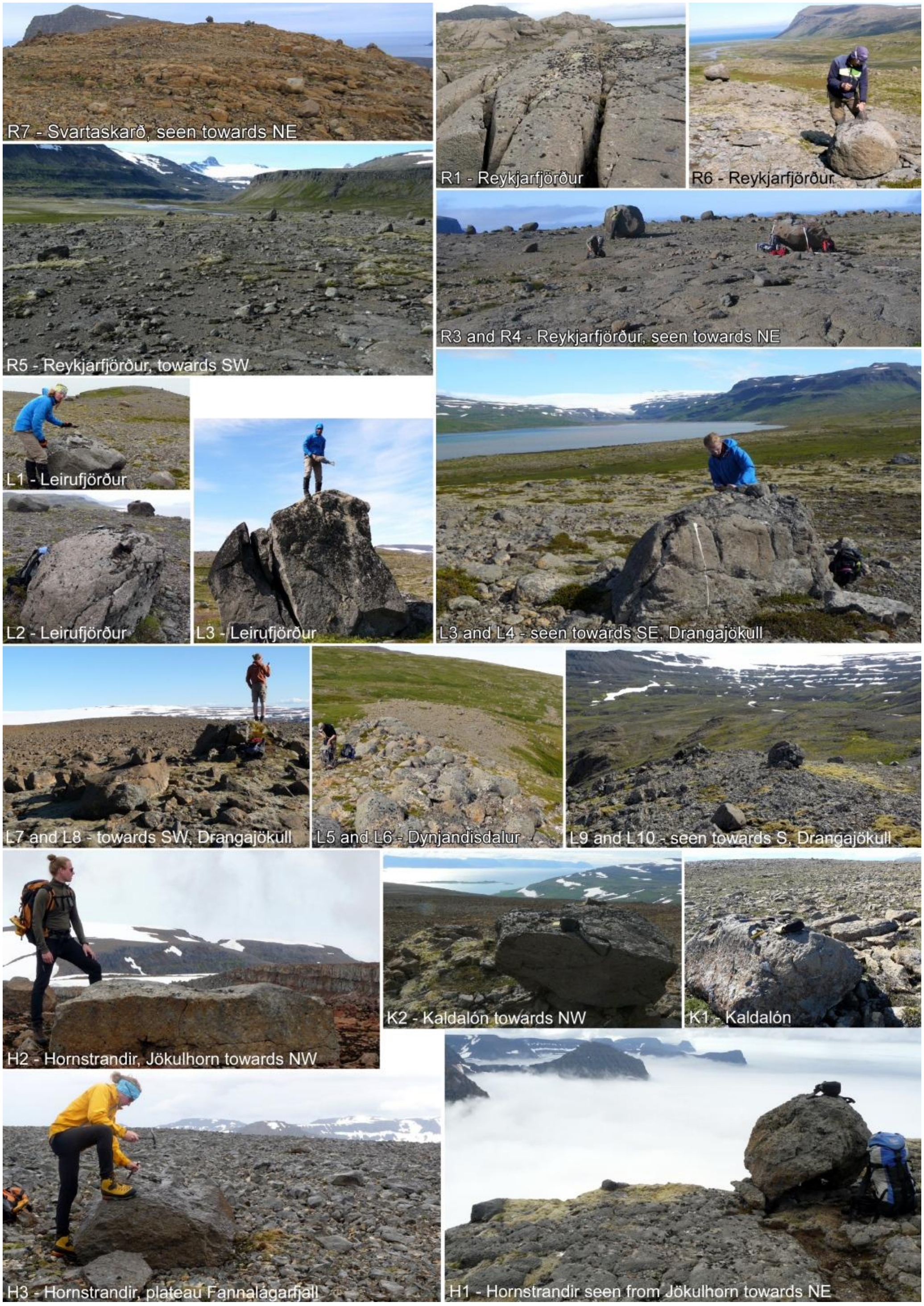

Figure 6. A mosaic figure showing selected localities sampled for cosmogenic exposure dating. The sample numbers and names of the localities are indicated (see Paper III).

3.4 Sedimentology and stratigraphy

Sedimentology and stratigraphy was investigated in surface pits and river cut sections around the perimeter of Drangajökull. Sediments were described and interpreted according to their genesis (Evans and Benn, 2004; Benn and Evans, 2010) and subsequently recorded on geomorphological maps and sedimentological logs or section drawings. Several river cut sections were cleaned, using shovels and spades, in order to study their sedimentology and stratigraphy. For the sketches and logs, we used the data chart of Krüger and Kjær (1999) and its symbols and codes for sedimentary structures and lithologies.

3.5 Ice surface elevation and volume changes

Three set of DEMs, from 1994, 2005 and 2011 (Table 1), which pre-date and post-date the recent most surges of the Drangajökull outlet glaciers were used to quantify surface elevation and volume changes of Drangajökull. Aerial photographs from 1994 supplied by the National Land Survey of Iceland were orthorectified and used to produce a DEM with 2 m ground resolution by stereophotogrammetry. Unfortunately, the southern perimeter of Drangajökull, and parts of Kaldalónsjökull, are missing from the 1994 aerial photograph survey. The LiDAR derived DEM was used to measure time-homologous points outside the ice cap for ground control of the 1994 DEM. Because the LiDAR DEM was measured in the middle of the summer, on the 20^{th} of July 2011, there are still several snow fields in the forefield of the ice cap. The 2005 DEM was supplied by Loftmyndir ehf. and has a RMS error c. 3 m (Table 1). The geographical system ArcGIS10.2 was used to analyse and handle the DEMs, a 3D extension to ArcGIS was used to calculate DEM of differences (DoDs). The DoDs yielded surface elevation changes of the glacier which occurred in the years that elapsed between measurements of each DEM. To obtain volume changes of a specific area, the mean surface elevation change was simply multiplied with the area of study. To enable better comparison of the Drangajökull changes with other glaciers, an average mass balance was calculated from the mean surface elevation changes.

The definition of Sorge's law allows the volume changes to be converted to change of mass. The mean thickness changes were multiplied with a constant density of the ice volume (Bader, 1954; Paterson, 1994). The ice mass at Drangajökull was assumed to have a density of 917 kg/m^3. However, because the density of the lost and gained mass is not known, the applied density is a source of errors in the mass balance calculations, and they should be regarded as estimations.

4. *Summary of papers*

This thesis consists of five papers (I-V). Table 2 gives an overview of the contributions from the authors of these papers.

Table 2. Contributors to the papers presented in Appendix I-V.

Task	Paper I	Paper II	Paper III	Paper IV	Paper V
Logistics and preparation	S. Brynjólfsson A. Schomacker	S. Brynjólfsson A. Schomacker	S. Brynjólfsson A. Schomacker		
Fieldwork	S. Brynjólfsson A. Schomacker J.B. Friðriksson J. Andreassen K. Naegeli	S. Brynjólfsson A. Schomacker J.B. Friðriksson J. Andreassen K. Naegeli	S. Brynjólfsson A. Schomacker J.B. Friðriksson J. Andreassen K. Naegeli		
Geomorpho-logical mapping	S. Brynjólfsson A. Schomacker	S. Brynjólfsson			
Aerial photos and DEMs		S. Brynjólfsson		S. Brynjólfsson N.J. Korsgaard	
Calculation of exposure age			S. Brynjólfsson		
Sedimento-logical logs and descriptions		S. Brynjólfsson			
Review of historical literature		S. Brynjólfsson			
Tephra analysis, geochemical analysis		E.R. Guðmundsdóttir	J.K. Keiding		
Data interpretation	S. Brynjólfsson A. Schomacker Ó. Ingólfsson	S. Brynjólfsson A. Schomacker Ó. Ingólfsson	S. Brynjólfsson A. Schomacker Ó. Ingólfsson	S. Brynjólfsson A. Schomacker N.J. Korsgaard Ó. Ingólfsson	All
Preparation of text/figures	S. Brynjólfsson	S. Brynjólfsson	S. Brynjólfsson	S. Brynjólfsson	Ó. Ingólfsson Í.Ö. Benediktsson S. Shchomacker

4.1 Paper I

In this study, we aimed to define the landform-sediment assemblage of the three surging outlet glaciers of Drangajökull ice cap. Furthermore, we aimed to map the geomorphology, sediments and the extent of the ice cap since its maximum during the LIA. Valleys, fjords and ice margins were surveyed by walking around the northern perimeter of the ice cap, with focus on the three surging glaciers during the summer 2011, 2012 and 2013. Landforms and sediments were described, interpreted and classified according to genesis.
Three detailed geomorphological maps, one from each surging glacier, were produced to demonstrate the landsystems of the surge-type glaciers and one larger overview map (Fig. 7) was produced to present the most pronounced features around the whole ice cap and its maximum extent during LIA (Appendix VI). Aerial photographs and a LiDAR derived DEM were used as base data for the mapping procedure.

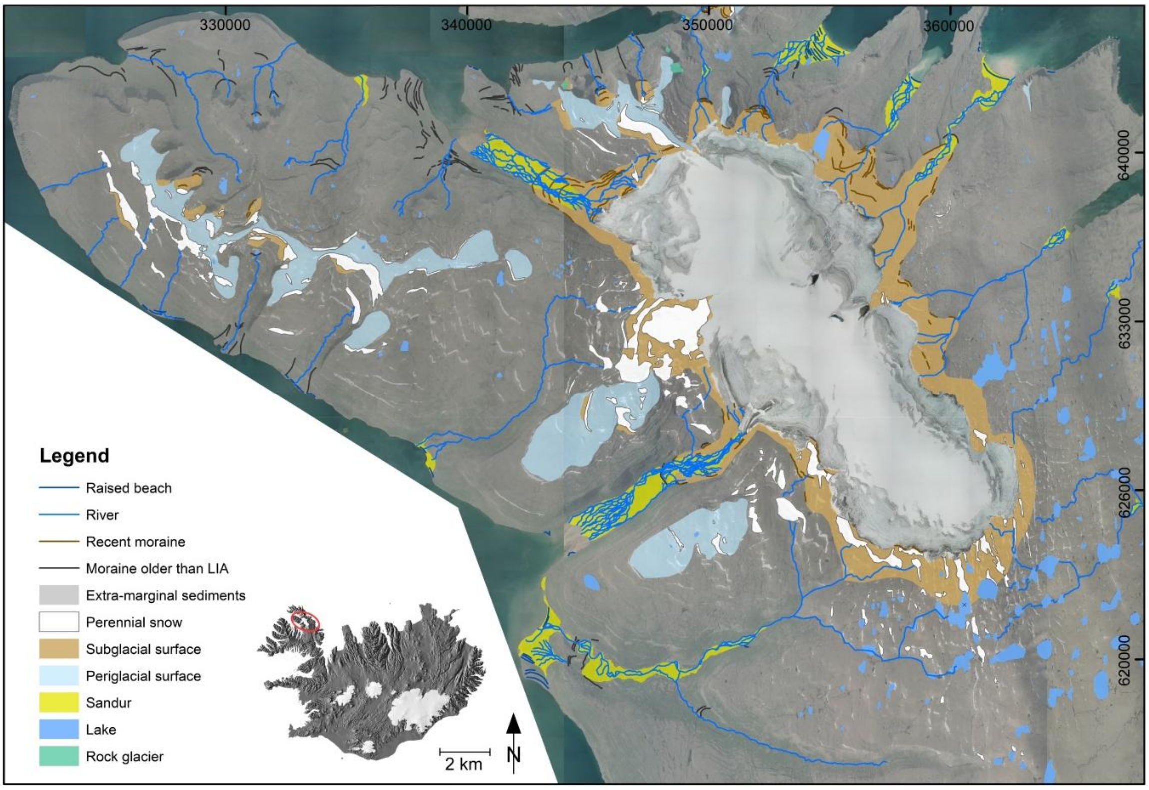

Figure 7. The main geomorphological and sedimentological characteristics in the Drangajökull area.

The geomorphology of the three surging glaciers is characterised by end moraines that correlate well with historical information on glacier advances since the LIA, and extensive sandar that cover the valley floors which are the main depocentres. The distal most moraines of the three surging outlets are 10-15 m high ridge shaped terminal moraines which consist of gravel, soil and peat pushed up by the glacier. The other moraines, constituting of coarse boulder-rich till, are usually not sharp crested but flat (2-5 m high) and 5-20 m wide and sometimes indistinct or eroded by the glacial rivers. Flutes are relatively common landforms proximal to the glaciers in Reykjarfjörður and Leirufjörður. Kames, eskers, pitted sandar and hummocky moraines are not frequently occurring landforms. In general, the sediment cover is thin around the Drangajökull ice cap and consists mainly of basal till or locally weathered bedrock. Ice sculpted bedrock sporadically protrudes the drift cover and large bedrock outcrops are often distinguishable below the thin till sheet. In general, the landform-sediment assemblage from the Drangajökull surging glaciers differs from other surging glaciers in Iceland, nor does it agree well with the landsystem model of Evans and Rea (1999, 2003). This could, among other things, be owing to low preservation potential of landforms and sediments due to intensive fluvial erosion at the valley bottoms.

The maximum extent of the ice cap during LIA was well constrained around the northern perimeter of the ice cap by ice marginal landforms and the historical data. Due to lack of moraines and other landforms, the maximum extent was challenging to reconstruct around the southern margin.

4.2 Paper II

This paper aimes at reviewing, re-interpreting and updating the historical surge descriptions of the Drangajökull ice cap since the LIA maximum, and to establish their climatic and glacial dynamic context. Furthermore, the maximum glacier conditions during the LIA were studied. The surge history records back to the beginning of the 18th century and is based on direct observations of marginal movements, recorded in church books and local annals. According to the 3-4 recorded surges, the glaciers reached their maximum extent asynchronously during surges in the 18th and 19th century, and have retreated 3000-4000 m since then.

Geomorphological data (Paper I) are used together with the historical data to improve the surge history of Drangajökull and to assess the accuracy of the historical data. End moraines were of special interest. Their position and the distance between them was measured with a hand held GPS and by GIS. Some of the moraines were of known age and related to the recorded surges. By using the historical data and correlating additional moraines of unknown age with the known moraines and a mean retreat rate in certain periods it was possible to establish a relative age of the additional moraines mapped in Paper I. DEMs were manually produced on the basis of moraines to represent maximum conditions during the LIA and to enable comparison with present conditions of the surging glaciers.

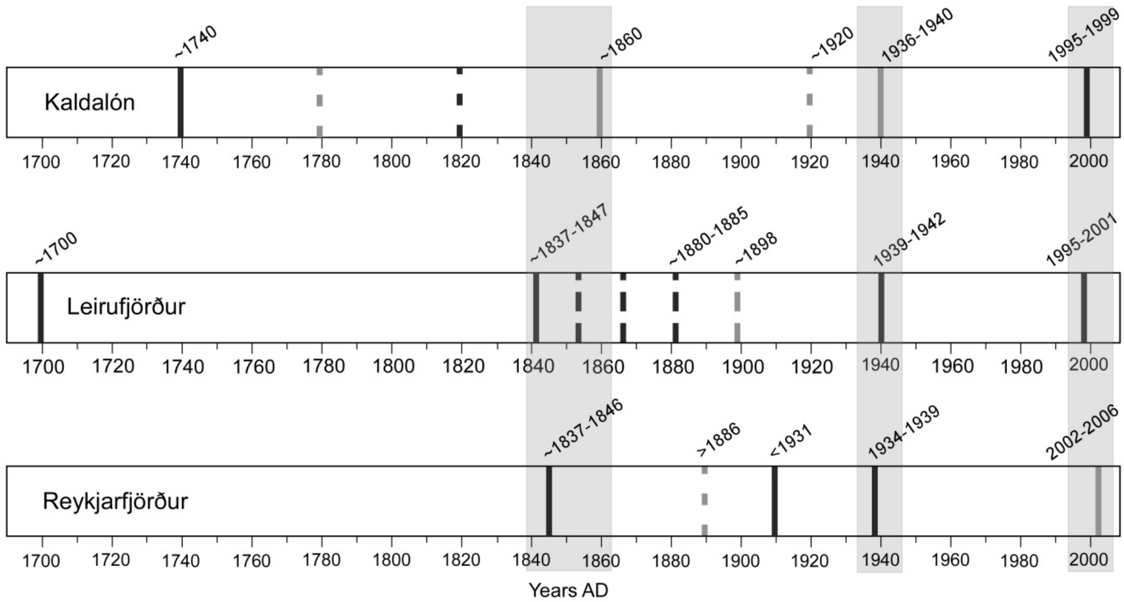

Figure 8. Overview of the surge history at the three surge-type outlet glaciers of Drangajökull. The grey vertical bars indicate moraine/marginal positions related to a surge that was not observed in the field, the dashed bars indicate new surges reconstructed during this study, bars without an yearare surges of unknown time, the shadowed areas (grey columns) demonstrate surges of all the outlets that occur almost synchronously.

The surge history of each glacier is plotted in detail on three individual figures in Paper II. The duration and timing of the surges is plotted with the longest temperature record from Iceland on an adjusted time and length scales (Figs. 2, 3 and 5 in Paper II). The main morphological features are shown on a shaded relief LiDAR map. For comparison, cross- and longitudinal sections of the ice surface in 2011 and the reconstructed maximum conditions during LIA are shown. Furthermore, the reconstructed thinning, volume changes and area changes since their LIA maximum until present are represented.

The results of this paper rely on the thorough geomorphological mapping presented in Paper I. In total, five surges were reconstructed from Reykjarfjörður, eight from Leirufjörður and seven from Kaldalón (Fig. 8). Earlier surges seem to have occurred asynchronously, but the last two surges of the three outlet glaciers occurred broadly synchronously, around 1940 AD and 2000 AD. The duration between surges is 10-140 years and varies between and within the glaciers, about 140 years without recorded surge between ~1700 and ~1840 in Leirufjarðarjökull perhaps indicating a lack of data rather than a long quiescent phase. Currently the three surge-type outlets are in their quiescent phase. Drangajökull reached a LIA maximum extent of ~216 km^2 compared to 143 km^2 in 2011. Each outlet reached their LIA maximum conditions asynchronously from ~1700-1846, when their volume and area, respectively, was 2.1-2.3 km^3 and 22-34 % greater than at present. Any clear relationship between climate and the surges was not established.

4.3 Paper III

The main aim of Paper III was to reconstruct the glacial history of the Drangajökull ice cap, assess the timing of terrestrial deglaciation and define terrestrial glacial conditions from the Late Weichselian to the early Holocene in the Vestfirðir peninsula. Cosmogenic isotope (^{36}Cl) surface exposure dating was applied to 24 samples from bedrock, erratic boulders and moraine boulders sampled from the perimeter of the ice cap and Hornstrandir (Fig. 9). Furthermore, river-cut sediment profiles in Reykjarfjörður were studied in order to reconstruct the environmental conditions there during the last deglaciation. The geomorphological mapping, presented in Paper I, was fundamental for the sampling process and to identify moraines, ice sculpted bedrock or other landforms that could be relevant to date for reconstruction of the glacial history.

Data from Paper I, II, and III together with results of earlier studies, based on marine sediment core data, sea level changes and terrestrial mapping and chronological data from the area, are compiled and represented in an overview glaciation model.

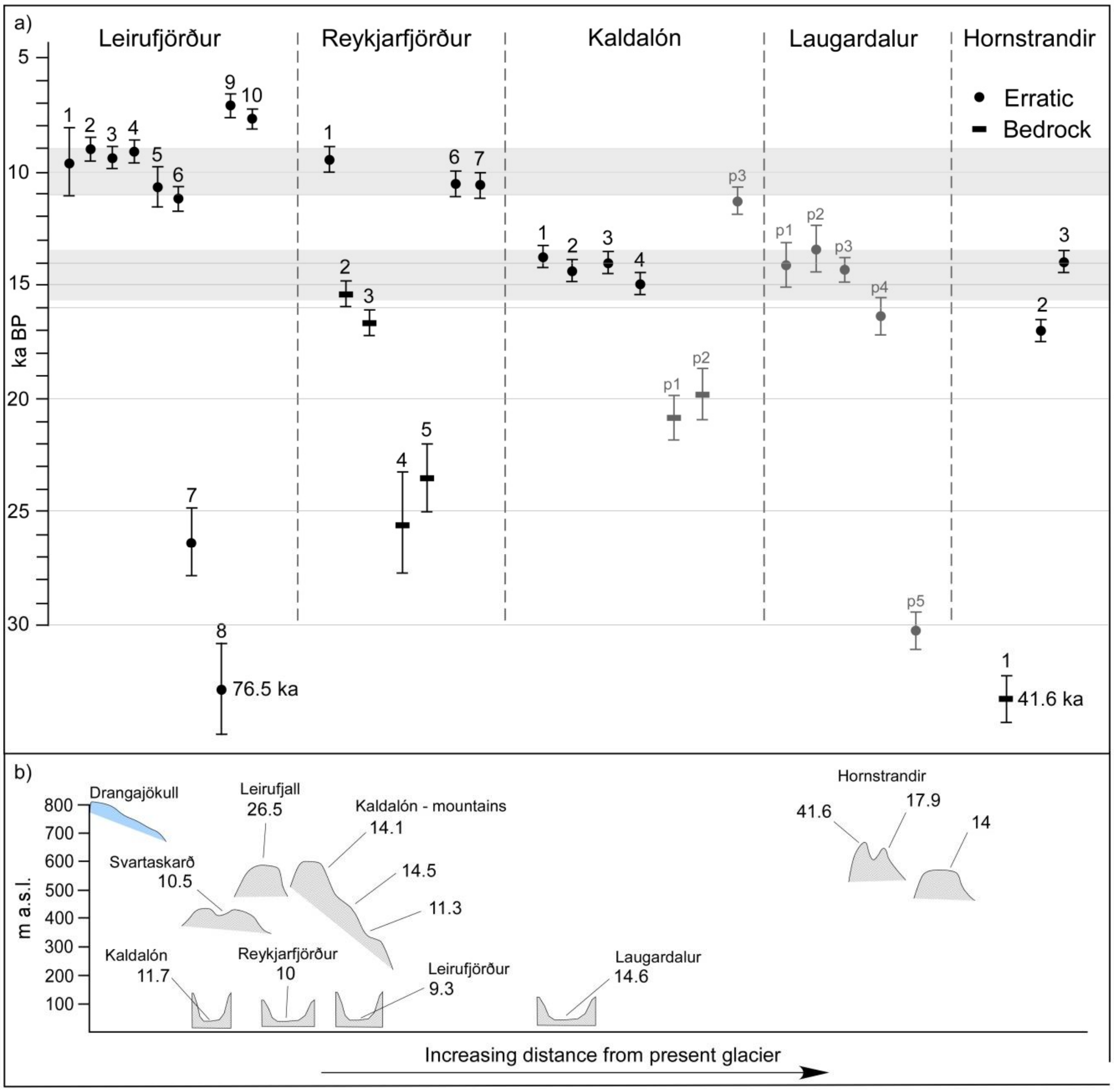

Figure 9. Overview of the cosmogenic exposure ages from the Drangajökull area. a) Bedrock samples are indicated with a bar and erratic and boulder samples with a circle. Previous exposure ages of Principato et al. (2006) are shown in grey colour for comparison. b) Exposure ages (ka BP) plotted against elevation and in relative distance from the present ice cap margin. For location of place names, see Figure 3b.

The resulting ^{36}Cl exposure ages were logical according to their topographical setting. The oldest samples (76.5 ±2 ka and 41.6 ±1.1 ka) were located on considerably weathered uplands and mountain plateaux in combination with younger erratic boulders (26.2-13.8 ka). This suggests a presence of non-erosive ice which left the oldest sample sites and the weathered surfaces relatively intact. In the lowland, glacially scoured bedrock and erratic boulders yielding younger ages (9.4-11.3 ka BP) indicate a warm-based, erosive, dynamic ice that occupied valleys and fjords during the last glaciation. The few higher ages obtained from bedrock in the valley of Reykjarfjörður were considered to result from nuclide inheritance. Notably, this demonstrate erosion <2-3 m during last glaciation. That observation of limited subglacial erosion corresponds with descriptions in Paper I of thin and patchy till cover in the forefield of Drangajökull.

The early deglaciation time of the plateau mountain Leirufjall (26.2 ±1.5 ka BP) indicates ice thinning and deglaciation over some mountains in the Vestfirðir peninsula that preceded any considerable lateral retreat of the ice sheet. The new exposure ages suggest asynchronous and stepwise deglaciation. Most of the uplands and some valleys were deglaciated 14-15 ka BP while other fjords and valleys drained the major outlets of Drangajökull at least until ~ 9 ka BP.

The outermost moraine in a series of concentric moraines in the mouth of Leirufjörður was dated to 9.3 ka. Those moraines are interpreted to have formed during glacier advances in response to climatic deterioration forced by reduced Atlantic Meridional Overturning Circulation (AMOC). The sediment profile data demonstrate an outlet glacier advancing and calving into a glacier lagoon in the valley of Reykjarfjörður after 9.4 ka BP.

The results suggest a strongly topographically controlled ice sheet, with slow non-erosive cold-based sectors over plateaux and uplands while warm-based erosive ice occupied fjords and valleys of the Vestfirðir peninsula during last glaciation. Furthermore, the new exposure data reveal an extensive Drangajökull ice cap which survived at least until c. 9 ka BP.

4.4 Paper IV

Paper IV focuses on surface elevation, volume changes and ice dynamics related to the recent most surges of Reykjarfjarðarjökull and Leirufjarðarjökull, aiming to clarify the main characteristics of the Drangajökull surging glaciers. Generally, surface lowering of the glacier reservoir areas is in the order of 20-100 m during surges, and surface thickening of a similar amount in the receiving area (Björnsson et al., 2003; Sund et al., 2009). Three DEMs from 1994, 2005 and 2011 were handled with a 3D extension of ArcGIS to quantify the surface and volume changes of Drangajökull in the period 1994-2011.

The main conclusion is that the surge duration, characteristics of ice discharge and dynamics of the Drangajökull surges resemble characteristics of Svalbard surges rather than the surges of the larger ice caps in central and south Iceland. At Drangajökull, the main surface thinning, of c. 10-30 m, during surges, occurs above ELA in the upper reservoir areas. Consequently, very variable thickening in order of 10-120 m occur in the receiving area, reflecting a high-relief topography that the surges have overridden. Ice discharge from the reservoir area to the receiving area was 0.054 km^3 for Reykjarfjarðarjökull and 0.152 km^3 for Leirufjarðarjökull. This contributes to at least 30-40% of the volume loss of the reservoir areas in the period 1994-2005. These values are minimum estimates because the DEMs pre- and post-date the surges. Based on annual glacier-frontal measurements of the surging outlet glaciers, the surges usually have a distinct maximum average flow rate in about one year, 2.1 m d^{-1} in 1995-1996 during the surge of Leirufjarðarjökull, 1.8 m d^{-1} in 1996-1997 during the surge of Kaldalónsjökull and only 0.2 m d^{-1} during the period of fastest flow of the Reykjarfjarðarjökull margin in 2003-2004. However, the accelerating and decelerating phases tend to take several years, resulting in a 4-10 long active phase of the surges. The average marginal flow rate was 0.46 m d^{-1} during the last surge of Leirufjarðarjökull and 0.12 m d^{-1} of Reykjarfjarðarjökull.

By comparison of the 2005 and 2011 DEMs, the condition of the glaciers during their present quiescent phase was evaluated. Despite negative mass balance of c. 0.15-0.22 m w.e. a^{-1}, depending on study areas, the surging glaciers gained mass and thickened by 0.5-0.7 m a^{-1} in the reservoir areas and thinned by 1 m a^{-1} on average in the receiving areas during the period 2005-2011.

Compared to a relative abrupt accelerating and decelerating phase, short surge duration and a main surface thinning in the upper ablation areas of most Icelandic surge-type glaciers (Björnsson et al., 2003), the expression of Drangajökull surges is unique; their surge duration is several years longer and characterised by a few years long accelerating and decelerating phase, and the main ice discharge coincide with the main accumulation zone of the quiescent phase in the upper reservoir areas.

4.5 Paper V

Paper V is a review article which focuses on past and ongoing research of geological fingerprints of Icelandic surge-type glaciers. The main aim is to give an overview of the processes and products of the Icelandic surging glaciers and recent progress of the understanding of surging glacier landsystems, and to outline future research challenges.

The main motivations for glacial geological studies of surging glaciers in Iceland has been that, form, flow, discharge and stability of present and past ice sheets and ice caps are controlled by ice streams and surging glaciers (Evans and Rea, 2003; Boulton, 2010). Thus, studying the geomorphological and sedimentological products and processes of surge-type glaciers enable better reconstructions of past glacier dynamics (Dowdeswell, 1995; Evans and Rea, 2003).

Studies of Brúarjökull, northern Vatnajökull, resulted in a new model suggesting rapid ice flow to be supported by overpressurized water which caused a decoupling of a thick sediment sequence from its bedrock and enabled the extreme fast flow of Brúarjökull surges (Kjær et al., 2006). Some occurrence of fast flow results from complex interplay of topography and the nature of the substrate and subglacial thermal and hydrological conditions. Therefore, the new model is not considered to explain all observed fast flow.

Benetiksson (2012) and Benediktsson et al. (2008, 2009, 2010) demonstrated the importance of glaciotectonic end moraines to understand the dynamic nature of surging and fast-flowing glacier. Different structural style and lateral variability in end moraine morphology and architecture at Brúarjökull and Eyjabakkajökull was explained by spatial variations in subglacial sediment conditions, thickness and rheology of foreland sediments, ice flow mechanism and subglacial hydrology. This extensive work on ice-marginal and sub-marginal landforms may support future reconstructions of ice-flow rates from geomorphological situations in the glacier forefields.

Studies of lake sediment cores suggest a climatic forced mass balance control on surge initiations and surge interval of Eyjabakkajökull, and the glacier switched to surge behaviour about 2200 BP during cooling Neoglacial conditions and surges of the glacier became even more frequent during the LIA (Striberger et al., 2011).

A landsystem model for surging glaciers, including three zones which each contained certain assemblage of landforms and sediments, was established by Evans and Rea (1999, 2003). A moderately refined version of that landsystem model was developed during studies of Brúarjökull and Eyjabakkajökull (Schomacker et al., 2014) and a new landsystem model was developed for surge-type cirque glaciers (Brynjólfsson et al., 2012). The main aim of developing and improve the landsystem models is to provide a method and tool to identify fingerprints of fast-flowing ice and obtain information on the conditions and processes of those glaciers on local, regional and temporal scale. Therefore well-developed landsystem models might enable us to identify surge-type glaciers in areas lacking any historical or observational record and reveal previously unknown surges at known surge-type glaciers. Thus, it may serve as analogue when identifying and describing the dynamics of glaciers and enable more thorough understanding of the potential relationship with climate.

5. *Discussion*

5.1 Geomorphological impacts of surging glaciers

Glacier surges are often considered to produce diagnostic collections of landforms and sediments. Studies of landform-sediment assemblages from the forefields of surging glaciers in Iceland, Alaska and Svalbard were used to establish a landsystem model, which aims to identify and describe characteristic landforms and sediments formed by surging glaciers (Evans and Rea, 1999, 2003). This landsystem model divides the surging glacier forefield into outer zone, middle zone and inner zone. Such an approach has been considered favourable to recognise imprints of former surging glaciers on the landscape and to identify unrecorded surge events of present surge-type glaciers (Sharp, 1985, 1988; Evans and Rea, 1999, 2003). However, more recent studies have demonstrated that the landform-sediment assemblages at some surging glaciers do not agree completely with the landsystem model of Evans and Rea (Brynjólfsson et al., 2012; Schomacker et al., 2014; Paper I). Brynjólfsson et al. (2012) emphasised the effect of the landscape settings, on the sediment-landform assemblage, which surge-type cirque glaciers experience on the Tröllaskagi peninsula in north Iceland.

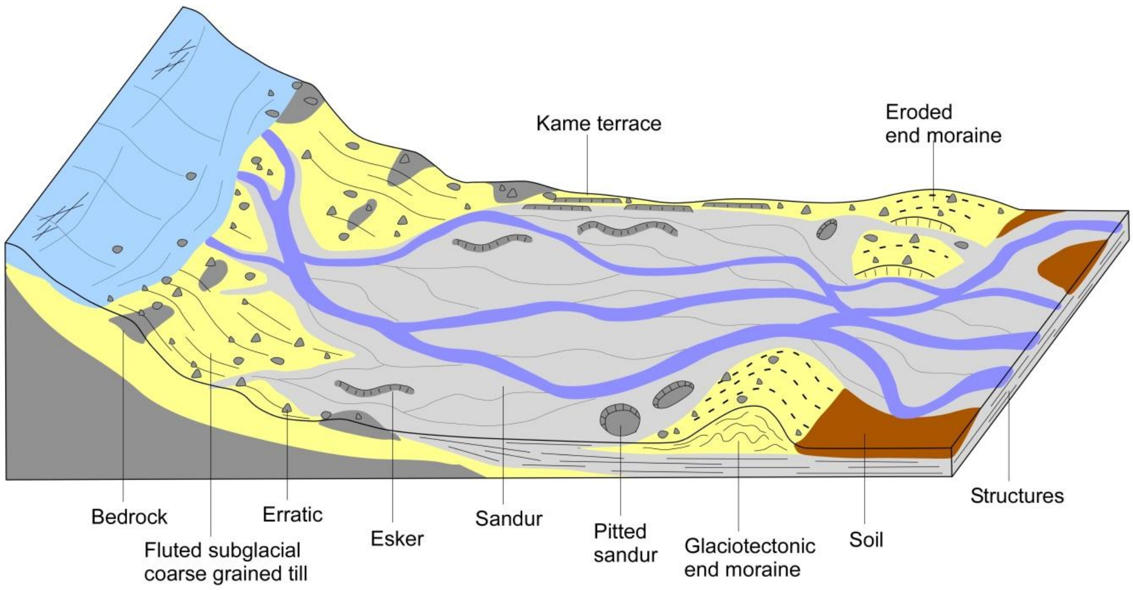

Figure 10. A landsystem model developed to clarify the main landform-sediment assemblages in forefields of strongly topographically confined surge-type outlet glaciers, based on the observations from the Drangajökull outlets.

One of the key aims of this study was to describe and analyse the landform-sediment assemblages of the three surging outlet glaciers of Drangajökull, develop a landsystem model for these glaciers for better understanding the relationship between surge dynamics and sediment/landform processes and products, and to compare the results to the landsystem model for surging glaciers in general (Evans and Rea, 1999, 2003). The ultimate aim was to understand better what controls the landform and sediment occurrence and distribution in the Drangajökull surging outlet glacier forefields.

It is concluded that the landform-sediment assemblages that define the landsystem model for the Drangajökull surge-type outlets (Fig. 10) differ considerably from those signifying surging glaciers elsewhere, and the landsystem model does not agree well with the forefields of other surge-type glaciers described in Iceland (Sharp, 1985, 1988; Evans et al., 2002, 2007 2009; Kjær et al., 2006, 2008; Brynjólfsson et al., 2012; Jónsson et al., 2014; Schomacker et al., 2014). The forefields of the Drangajökull outlets are characterised by extensive sandurs that cover the valleys floors and end-moraine ridges that mark the extent of former advances. Flutes are relatively common proximal to the glaciers while eskers, kame terraces, pitted sandur and hummocky moraine are in-frequent landforms. None of the landforms at Drangajökull are unique for surging glaciers. The general surging glacier landsystem model of Evans and Rea (1999, 2003) and the modified version of Schomacker et al. (2014) emphasise hummocky moraines, pitted sandur, crevasse-squeeze ridges and concertina eskers as important landforms of the surging glacier landsystem. The crevasse-squeeze ridges and the concertina ridges are absent, while hummocky moraine and pitted sandur only rarely occur in front of the Drangajökull outlets.

The main reasons for this difference could be (i) the generally thin and coarse-grained till produced by Drangajökull surging outlets cannot be squeezed from the substratum into crevasses during a surge; (ii) in the current quiescent phase, the main depocentres at the valley bottoms are characterised by sandur and glacial rivers that meander between the valley slopes indicating extensive fluvial erosion, which most likely erodes most surge imprints quickly; (iii) furthermore, the presently stagnant surge-type margins consist of relatively clean ice and are therefore not in the process of forming extensive dead-ice and hummocky moraine.

5.2 Are Drangajökull surges unique in terms of Iceland?

Another main objective of this study was the history and dynamics of the Drangajökull surge-type outlets, whether or not their surge behaviour was similar to other surge-type glaciers in Iceland, and if a distinct relationship between the surge behaviour and climate or other environmental factors could be detected.

Thorough geomorphological mapping of the forefields of the three surge-type outlet glaciers was fundamental to decipher their surge history. The geomorphological mapping in Paper I revealed several moraines in addition to those deposited by the recorded surges.

The ‘new’ moraines were interpreted to have formed during surges. The locations of end moraines were easily identified with direct observations and the historical review (Thoroddsen, 1933; Eyþórsson, 1935; Þórarinsson, 1943, 1969). Nevertheless, several of the moraines lack exact dating. Their relative ages are constrained by the neighbouring moraines. Thus, a combination of the geomorphological mapping, sedimentological studies and review of historical records, enabled reconstruction of a 300 year long surge history of the Drangajökull ice cap (Figs 8 and 11; Paper II).

Short series of measured precipitation and temperature in the vicinity of Drangajökull (Fig. 11) and almost no direct mass balance measurements of the ice cap make it challenging to identify any potential relationship of climate or mass balance with the surge characteristics, e.g. surge initiation, duration, periodicity or advanced distances (Fig. 8). Climatically forced mass balance may alter the surge periodicity of the Drangajökull outlets (Dowdeswell et al., 1995; Eisen et al., 2001; Hewit et al., 2007). The duration of the quiescent phases of the two recentmost surges was 50-60 years during warmer climate conditions after 1920 AD, compared to quiescent phases of only 10-50 years during the cooler 19th century. This agrees with observations of longer periods between surges of Svalbard, Vatnajökull and Langjökull surge-type glaciers during the warmer climate conditions after the termination of the LIA (Dowdeswell et al., 1995; Striberger et al., 2011; Larsen et al., 2015). The exceptionally long quiescent phase, about 140 years at Leirufjarðarjökull, in the 18th and 19th centuries was considered to indicate lack of data rather than a truly long quiescent phase, accounts for temporal uncertainties in the surge history (Fig. 9).

The glacier margins were observed to advance for 4-7 years during the last two surges of the Drangajökull outlets. The total duration of the active phase of Reykjarfjarðarjökull might be up to 8-10 years, because unusual crevasse formation in the upper reservoir area, indicating accelerated ice velocity were noticed several years before the marginal advance started (Þröstur Jóhannesson pers. comm.). Furthermore, zero retreat of the glacier margin the last three years before the marginal advance may also indicate increased ice velocity that overcome marginal retreat due to ablation. At Leirufjarðarjökull, the three DoDs for the period 1994-2011 show distinct surface thickening proximal to the southwest margin, and some advance of the margin at least until 2005. However, the glacier-frontal measurements at the westernmost tip of the Leirufjarðarjökull margin had indicated surge termination in 2001. This reveals an area proximal to the southwest margin which was still surging in 2005. Thus, the active phase of Leirufjarðarjökull lasted at least for 10 years since the first marginal advance was observed in 1995.

This is in pronounced contrast to the marginal advance during the active phases of most other Icelandic surge-type glaciers which commonly lasts 2-3 months (Björnsson et al., 2003). Therefore, the duration of the Drangajökull surges resemble better the 3-10 years

active phases of polythermal Svalbard surge-type glaciers (Dowdeswell, 1991; Hamilton and Dowdeswell, 1996, Jiskoot et al., 1998) and surge-type cirque glaciers in north Iceland (Brynjólfsson et al., 2012).
Similar to the surge periodicity, the surge distances are irregular and variable within periods and between glaciers. Reykjarfjörður surged 224 m in 2002-2006 but 839 m in 1933-1939. Similarly, Kaldalónsjökull surged 191 m in 1936-1940 and 1015 m in 1995-1999. However, Leirufjarðarjökull surged about 1000 m during the two recent most surges at similar time as Kaldalónsjökull (Fig. 11). Explaining those contrasts is challenging; shorter advances could reflect that a smaller portion of the reservoir area participated in a surge. Large local contrasts in climate and climatically forced mass balance are considered unlikely to explain those contrasting advanced distances.

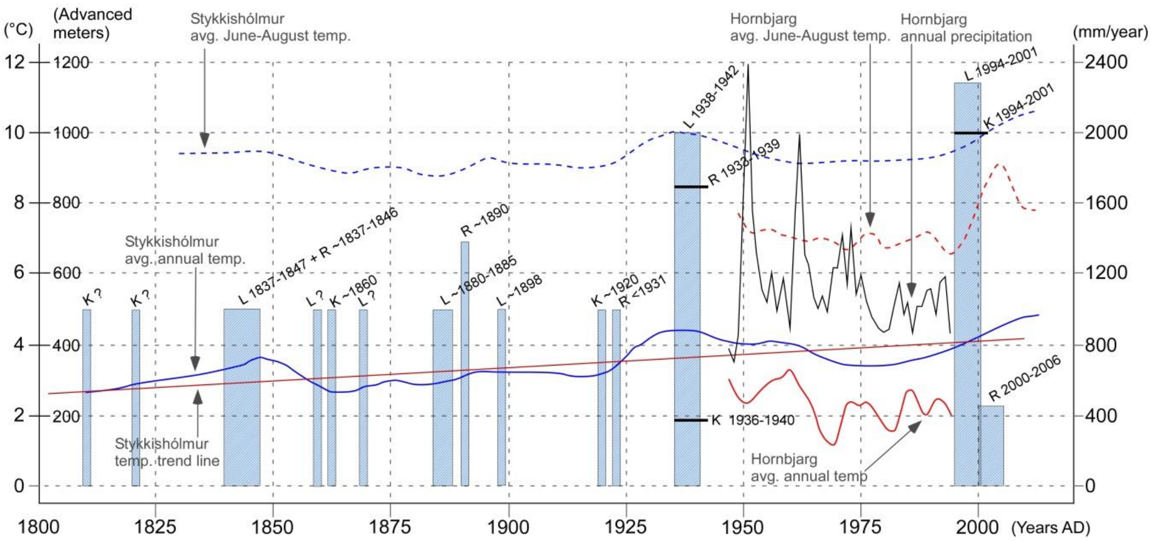

Figure 11. Surges of Drangajökull plotted against meteorological data from the Icelandic Met Office. The vertical columns represent surges, their width indicate the surge duration, except the thinnest columns which indicate surges of unknown duration. The column height represents the advanced distance, the columns are set to 500 m for surges of unknown distance (K = Kaldalónsjökull, L = Leirufjarðarjökull, R = Reykjarfjarðarjökull). See also Paper II.

The pattern of ice discharge during surges of Vatnajökull and other ice caps in Iceland is well established. The acceleration phase is characterised by an abrupt increase of the surface velocity in the upper ablation area (Björnsson et al., 2003). The main surface thinning coincides with the highest velocity in the upper ablation area and around the ELA. The total surge duration, from the first indication of high velocity and ice discharge until termination, can take up to 2-3 years (Björnsson, et al., 2003; Aðalgeirsdóttir et al., 2005). The average flow rate 2-22 m d^{-1} has been observed for some Vatnajökull surges

(Þórarinsson, 1969; Fischer et al., 2003; Aðalgeirsdóttir et al., 2005) and 0.1-7 m d^{-1} for some Svalbard surges (Dowdeswell et al., 1991; Murray et al., 2003; Kristensen and Benn, 2012).

The three DEMs from Drangajökull pre- and post-date the surges and can therefore not been used to make any direct measurements of surface velocities or propagation of a potential surge bulge during the Drangajökull surges. However, according to the ice-marginal measurements the average marginal flow rate during the recent most surges was 0.12 m d^{-1} for Reykjarfjarðarjökull and 0.46 m d^{-1} for Leirufjarðarjökull and the highest average flow rate within a year was c. 0.2-2.1 m d^{-1}. This is relatively low compared to average flow rate of the Vatnajökull and Svalbard surges (Þórarinsson, 1969; Dowdeswell et al., 1991; Fischer et al., 2003; Murray et al., 2003; Aðalgeirsdóttir et al., 2005). The relatively slow accelerating phases and especially the decelerating phases, which tends to take few years, accounts for those low average marginal flow rates of Drangajökull, resulting in active phases up to ten times longer than the active phase of the other ice caps in Iceland.

According to coinciding areas of the highest surface velocity and main surface draw-down in the upper ablation areas during the Vatnajökull surges (Björnsson et al., 2003), it might be assumed that the highest surface velocities of the Drangajökull surges would coincide with the areas of main surface thinning in the upper reservoir areas. Svalbard surges begin in the upper reservoir areas where the highest surface velocity and the main surface thinning is also observed, and coincides with the area of most accumulation during the quiescent phase (Sund et al., 2009, 2014).

The flow pattern and ice discharges of Drangajökull resembles Svalbard surging glaciers (Hamilton and Dowdeswell, 1996; Jiskoot et al., 1998; Murray et al., 2003; Sund et al., 2009) rather than other surge-type glaciers in Iceland which usually accelerate and decelerate abruptly (Björnsson et al., 2003). Furthermore, the observations described in Paper IV indicate the main thinning during Drangajökull surges, in the order of 10-30 m, to occur above the ELA in the upper reservoir area, coinciding with the area of most accumulation during the quiescent phase of the glaciers.

The ice discharge down to the reservoir area during some Vatnajökull surges contributes to about 75% of the lost volume of the reservoir area. The remaining 25% are explained by increased ablation enhanced by increased surface area and ice transportation to lower altitudes (Aðalgeirdóttir et al., 2005). The observations in Paper IV indicate that at least 30-40% of the reservoir area volume loss at Drangajökull surging outlets is due to ice discharge during the surges. The actual contribution of the ice discharge to the volume loss of the reservoir area must be considerably higher. Because the DEMs pre- and post-date the surges it allows some additional ablation of unknown amount before and after the surges.

Despite a negative gedetic mass balance of Drangajökull in the period 2005-2011, the reservoir areas of the surging glaciers have been thickening on average in the order of 0.5-0.7 m a^{-1} since the last surge termination (Fig. 12.). This coincides with positive mass balance observed by ablation stake measurements 2005-2007 (Shuman, et al., 2009). A distinct melting of the ablation areas and the thickening of the reservoir areas contributes to a steeper surface profile which gradually brings the glacier back to the pre-surge stage, which might lead to new surges in 45-65 years.

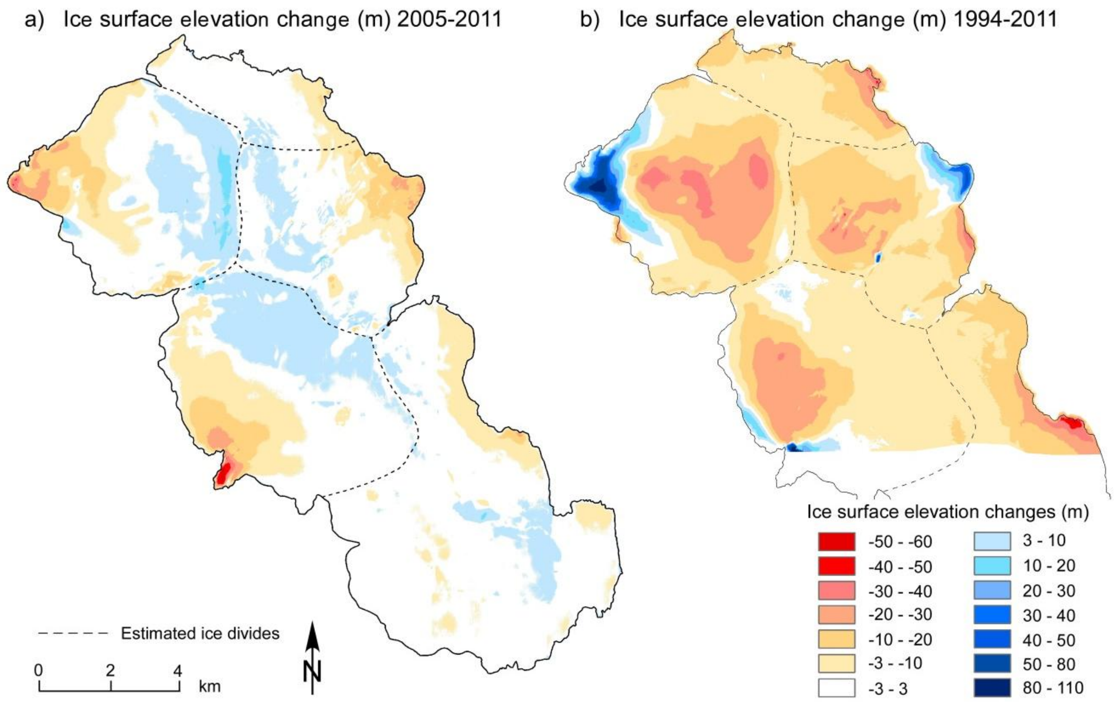

Figure 12. Surface elevation change of Drangajökull from 1994-2011. The surface elevation changes obtained from the DoDs clearly demonstrate the main ice discharge during the recent most surges of the outlet glaciers in a), and the surface thickening of the reservoir areas after the surge termination in b).

What control the characteristics of Drangajökull surges remains unknown.
The difference between the Drangajökull surges and surges of the larger Icelandic ice caps (Paper II and Paper IV), clearly shows that Drangajökull surges resemble surges of Svalbard glaciers (Dowdeswell et al., 1991; Jiskoot et al., 1998; Murray et al., 2003) and surge-type cirque glaciers in north Iceland (Brynjólfsson et al., 2012) rather than the surges of the larger ice caps in Iceland. The contrasting pattern of surface thinning, ice discharge, accelerating and decelerating phases and the surge propagation in general of Drangajökull surges compared to the Vatnajökull surges might be controlled by several parameters. The

high relief alpine setting of northern Drangajökull compared to the flat highland plateau which the Vatnajökull surging glaciers are situated on, and the Neogene plateau basalt substratum at Drangajökull, compared to palagonite dominated substratum of the larger ice caps in Iceland, might favour contrasting subglacial sedimentary and hydrological conditions of the glaciers and thereby contribute to the different dynamics and surge propagation. Furthermore, along with topographical conditions and geometry of the polythermal Svalbard surge-type glaciers, variable thermal regimes and margins frozen to their substratum have been considered to contribute to their long surge durations and long accelerating and decelerating phases (Hamilton and Dowdeswell, 1996; Jiskoot et al., 1998; Sund et al. 2009, 2014). This study suggests that the main characteristics of surge-type glaciers are similar in nature, but the pattern and expression of those characteristics depend on complex combination of environmental factors.

5.3 Exposure dating: problems and potentials

Despite accumulating data and improved knowledge on the Late Weichselian-Early Holocene glacial history of Iceland, configuration and thermal conditions of the last IIS still suffer from a lack of chronological data, especially direct dating of terrestrial landforms (Norðdahl et al., 2008; Geirsdóttir et al., 2009, 2013; Ingólfsson et al., 2010). Whether or not certain areas, like the Vestfirðir peninsula, were completely covered by a coalesced ice sheet or if the peninsula was partly occupied by a regional ice cap which drained through valleys and fjords leaving upland areas and some coastal proximal areas ice free during the last glaciation has been debated for several decades (Þórarinsson, 1937; Sigurvinsson, 1983; Hjort et al., 1985; Norðdahl, 1991; Rundgren and Ingólfsson, 1999; Principato et al., 2006; Hubbard et al., 2008; Norðdahl et al., 2008; Geirsdóttir et al., 2009).

The results of Principato et al. (2006) demonstrated the potentials of the ^{36}Cl exposure methodology, and indicated early ice free conditions of some uplands and valleys around Drangajökull. However, the age calculation suffered from uncertainties about the production rate of ^{36}Cl in Iceland (Principato et al., 2006).

This uncertainty of the ^{36}Cl method originate from discrepancy between different production rates which have been calculated from surfaces of very dispersed locations and ages (Stone et al., 1996; Philips et al., 2001; Swanson and Caffee, 2001; Licciardi et al., 2008). The uncertainty for Iceland was reduced significantly when a local production rate of ^{36}Cl was established in 2008, aiming to enable accurate exposure dating of postglacial landforms in Iceland (Licciardi et al., 2008). In addition, poorly defined weathering rates of bedrock in alpine landscapes have been considered to contribute to uncertainty in exposure ages (Balco, 2011). Correcting for several erosion scenarios in Paper III do not indicate a large effect of variable erosion on samples younger than 20 ka. Furthermore, more sophisticated recent calculators for scaling and shielding factors contribute to more accurate exposure ages (Vermech, 2007). Thus, the accuracy of ages from ^{36}Cl exposure dating in Iceland is considered well acceptable.

The results in Paper III confirm the potentials of the exposure dating, and have improved the knowledge about glacier conditions and the glacial history of Drangajökull since the Late Weichselian. The main results, based on the exposure ages from Paper III and geomorphic data from Paper I, suggest an early onset of a stepwise, asynchronous terrestrial deglaciation. A long lasting deglaciation, initiated about 26 ka BP over high mountains and terminated in valleys draining the main outlets of Drangajökull about 9 ka BP, compared to abrupt retreat of the shelf based ice sheet to the fjords and valleys about 14-15 ka BP (Andrews et al., 2000, 2002; Geirsdóttir et al., 2002; Principato et al., 2006; Ingólfsson et al., 2010).

Deglaciation of some valleys already c. 14-15 ka BP does not agree with the distinct IRD signals. However, this might rather indicate a complicated deglaciation pattern which makes precise reconstruction of the glacial history difficult without more direct datings of glacial landforms that identify the extent of a glacier in each specific area.

In Paper I, series of moraines were mapped and described at the mouth of Leirufjörður. According to marine data from the fjords distal to Leirufjörður, the ice retreated on land by 10.2 ka BP (Andrews et al., 2002; Geirsdóttir, et al., 2002). Therefore, the age of the moraines could be assumed to be younger than 10.2 ka. Subsequently, the exposure ages in Paper III showed the outermost moraine to be formed 9.3 ±0.8 ka BP. This allowed us to correlate the formation of the moraines with the climatic deterioration which has been interpreted to have resulted from freshwater pulses from Lake Agassiz and has been described from many localities around the North Atlantic (Alley et al., 1997, Clarke et al., 2004; Kleiven et al., 2008; Solomina et al., 2015). Therefore, the ^{36}Cl exposure dating in combination with the geomorphological mapping, for the first time in Iceland, revealed moraines formed by glacier re-advance in response to a cooler climate forced by meltwater pulses into the north Atlantic 9.5-8.2 ka BP.

In many presently glaciated environments, fjords or valleys occupied with glaciers have neighbouring fjords or valleys with ice free conditions. Hence, this modern analogue can be representative for the conditions on the larger peninsulas in Iceland, i.e. Vestfirðir, and Tröllaskagi during the last deglaciation and early Holocene. Similar to this study, such a complicated deglaciation pattern, based on cosmogenic exposure dating, have been suggested for Northwest Svalbard (Gjermundsen et al., 2013). Thus, marine sediment cores can convincingly indicate the timing of on-land retreat of glaciers and disappearance of calving glaciers. Direct dating of landforms, like exposure dating, is fundamental to reconstruct a detailed terrestrial glacial history and dynamics.

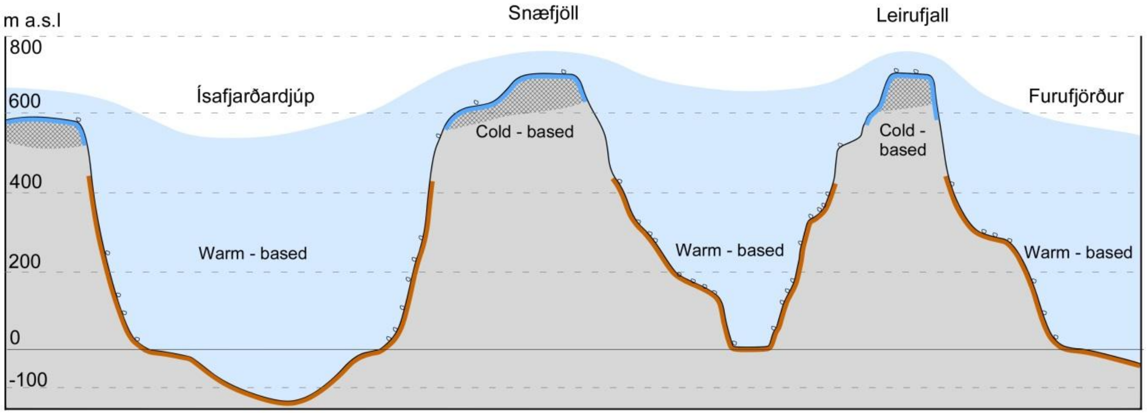

Figure 13. A conceptual overview of the glaciation of the Vestfirðir peninsula during the LGM.

The exposure ages from Paper III provided chronological control on the geomorphological and sedimentological observations in Paper I. Compiled results of those papers enabled us to reconstruct and establish a conceptual model of the glaciation of the eastern Vestfirðir peninsula during the last glaciation (Figure 13). The glaciation was characterized by topographically confined fjord/valley-type ice streams or glaciers. Warm-based ice in fjords and valleys flowed faster than the grounded, cold-based, ice domes over the mountains. A thin, non-erosive ice that flowed by internal deformation and left the substratum almost intact, covered uplands and plateaux, while the fjords and valleys were occupied with erosive, warm-based ice.

The exposure ages of erratic boulders and moraines show that an extensive ice cap was preserved over the eastern Vestfirðir peninsula at least until c. 9 ka BP. Contrastingly, at that time the other ice caps in Iceland have been reconstructed to similar size or even considerably smaller than at present (Geirsdóttir et al., 2009; Larsen et al., 2012). However, lots of un-dated moraines have been mapped in the central highlands of Iceland (Kaldal and Víkingsson, 1990). Because of an almost complete lack of organic material and tephra, those moraines have not been dated directly. By applying the cosmogenic nuclide exposure dating, absolute ages might be obtained.

Conversely, the exceptional extent of Drangajökull about 9 ka BP might indicate a strong contrast in climatically forced mass balance over the country enabling the size of the Drangajökull ice cap to be preserved longer. Perhaps those conditions of Drangajökull were promoted by enhanced strength of cold ocean currents just of the northwest coast, mainly affecting the climate conditions and glaciers on a local scale.

Despite ice sculpted bedrock and distinct glacial striations on the lowland of Reykjarfjörður valley, exposure ages obtained from bedrock samples were 15-25 ka, interpreted to indicate ^{36}Cl inheritance at least in some of the bedrock. Bedrock clearly indicating subglacial

erosion by striations and ice sculpted shape does not necessarily confirm enough erosion to reset pre-glaciation nuclide inheritance within the bedrock (Ivy-Ochs and Kober, 2008.
To remove any inheritance, the bedrock has to be eroded by 2-3 m (Ivy-Ochs and Kober, 2008; Balco, 2011). Thus, the old bedrock samples from Reykjarfjörður most likely indicate limited subglacial erosion, less than 2 m, during last glaciation. This supports the observations described in Paper I of a generally thin and non-coherent cover of till around the Drangajökull ice cap.

5.4 Late Weichselian-Holocene terrestrial glacial history of the Drangajökull ice cap

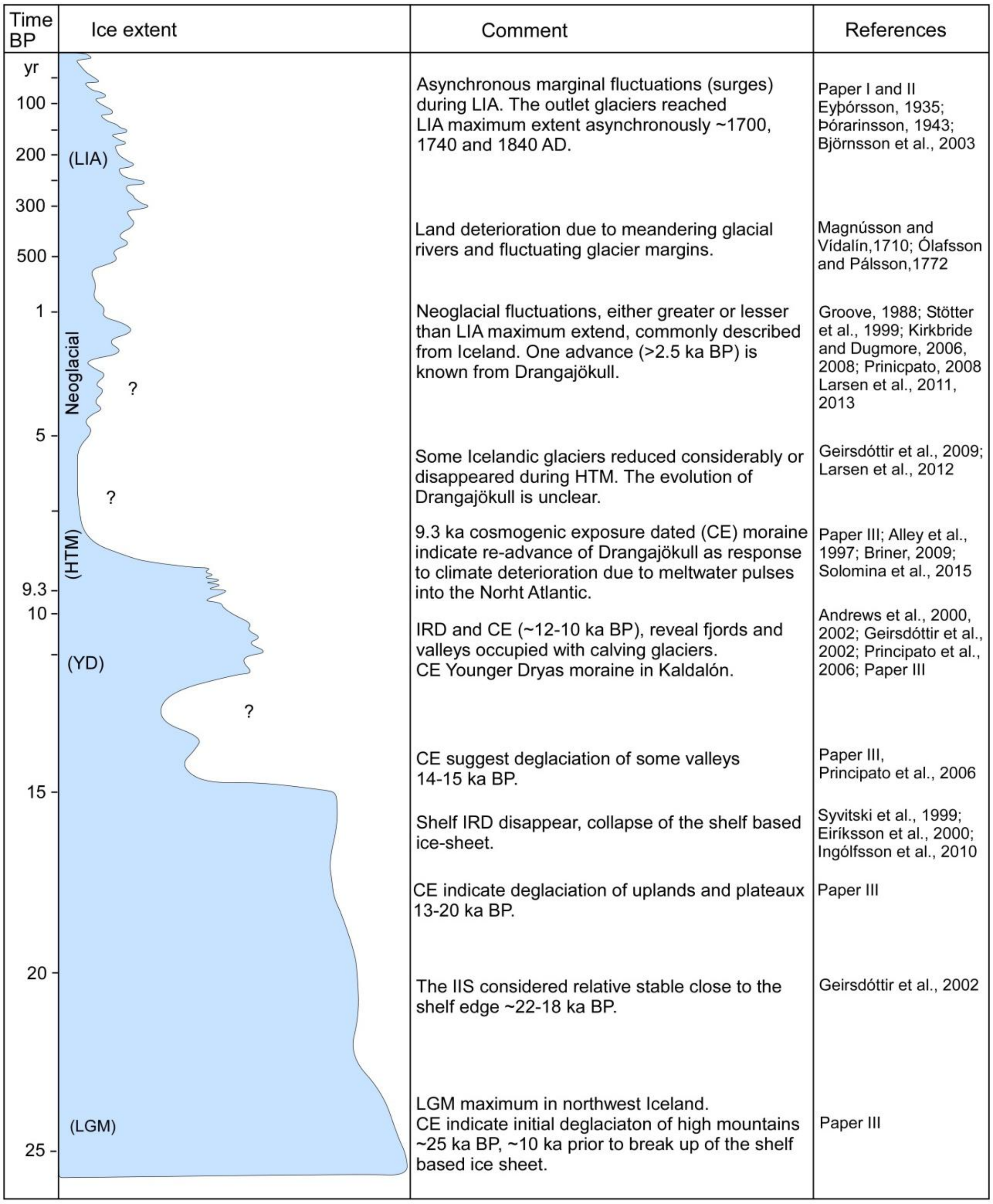

Figure 14. An overview time-distance model representing the glacier changes of the glacier over Vestfirðir and later the Leirufjarðarjökull outlet glacier. The model is based on data from Paper I,

II and III and several earlier studies in the Vestfirðir area, see references on the figure. Note that the time scale is not linear.

The main purpose of this study has been to improve the present understanding of the terrestrial glacial history and dynamics of the Drangajökull ice cap. Paper I, II and III are fundamental contributions aiming to serve that goal. Descriptions and interpretations of landforms, sediments, old literature and chronological control of certain landforms and localities accounted for improved glacial history of Vestfirðir (Fig. 14).

Previous research outlines the deglaciation history of the marine based part of the IIS. Interpretations based on marine sediment cores, geophysical data and ^{14}C dated shorelines indicate a sudden retreat about 15 ka BP of the IIS from an extended shelf based LGM position (Andrews et al., 2000, 2002; Geirsdóttir et al., 2002; Norðdahl and Pétursson, 2005). Calving outlet glaciers were active in the Vestfirðir until the IRD signal disappeared by 10.2 ka BP. Observations of the 10.2 ka BP Saksunarvatn tephra do also suggest at least some ice free localities in the Vestfirðir peninsula by that time (Hjort et al 1985; Principato, 2008).

The exposure ages represented in paper III reveal ice thinning and deglaciation over the plateau mountains about 26 ka BP, suggesting an early onset of terrestrial deglaciation of high mountains, compared to the break-up of the shelf based ice sheet about 15 ka BP and the glaciers retreat to the inner fjords and valleys about 12-10 ka BP. (Andrews et al., 2000, 2002; Geirsdóttir et al., 2002; Principato et al., 2006; Ingólfsson et al., 2010). The exposure ages from uplands and lower plateaux reveal deglaciation c. 14-15 ka BP, together with some valleys which were at least partly deglaciated at that time (Principato et al., 2006). Exposure ages from lateral moraines and uplands in Leirufjörður, Reykjarfjörður and Þaralátursfjörður argue for glaciers that approximately half-filled the valleys and extended to the sea 10.5-11.3 ka BP. This agrees with distinct IRD signals in marine sediment cores, interpreted to originate from calving glaciers in fjords and valleys which drained into the Ísafjarðardjúp fjord (Andrews et al., 2000, 2002; Geirsdóttir et al., 2002).

Exposure ages from Reykjarfjörður and Leirufjörður indicate that fjords and valleys draining the main outlets of Drangajökull were occupied with glaciers at least until about 9 ka BP.

Little is known about the postglacial evolution of the Drangajökull ice cap, which anyhow was considered to have retreated rapidly to the highland of the eastern Vestfirðir after the IRD disappearance (Geirsdóttir et al., 2002). The series of 10-12 undated moraines proximal to the 9.3 ka moraine at the mouth of Leirufjörður suggest glacier advances considered related to climatic deterioration events 9.5-8.2 ka BP (Alley et al., 1997; Solomina et al., 2015). Undated moraines formed by more extensive glaciers than during the LIA were mapped to the north and northwest of the present ice cap (Paper I). Some of those moraines are considered to have formed 9-11 ka BP as the recently dated moraines in Reykjarfjörður and Leirufjörður.

Neoglacial fluctuations are well known from many glaciers in Iceland (Grove, 1988; Guðmundsson, 1997; Stötter et al., 1999; Schomacker et al., 2003; Kirkbride and Dugmore, 2006, 2008; Solomina, et al., 2015). However, the mid-Holocene and Neoglacial history of Drangajökull is mostly unknown. Two degraded moraines in Reykjarfjörður and one in Kaldalón dated to Neoglacial ages indicate slightly more extensive glacial advances of Drangajökull some time during the Neoglaciation than during the LIA (Principato, 2008).

The historical data indicate land deterioration due to continual meandering of the glacial rivers and glacier fluctuations of the Drangajökull outlet glaciers in the centuries prior to the LIA maximum. However, potential geomorphological imprints of pre-LIA maximum advances have been eroded by advancing glaciers and the glacial rivers during the LIA. The most recent dynamics and fluctuations of the Drangajökull outlet glaciers are well constrained, and interpreted as surges (Papers I and IV).

A combination of geomorphological studies, historical studies and absolute dating of glacial features have successfully accounted for a reconstruction of the terrestrial glacial history, significantly improved the recorded surge history and clarified the configuration, thermal conditions and dynamics of the Late Weichselian Vestfirðir ice sheet and the Drangajökull ice cap (Fig. 14).

6. *Implications and future studies*

Until recently, direct dating of glacial landforms in Iceland has been very difficult due to lack of organic material and tephra in association with the landforms. The new Icelandic production rate for ^{36}Cl nuclides in Icelandic rocks (Licciardi et al., 2008) opens new opportunities to significantly improve the glacial history and properties of the IIS. The exposure ages presented here demonstrate complex ice conditions, dynamics and glacial history of Drangajökull and the IIS, which elucidate a long lasting debate about the glacial history and the glacial conditions over the Vestfirðir peninsula during the last glaciation. Future studies on the glacial history of Iceland can clearly benefit significantly from ^{36}Cl exposure dating, and ideally future studies would apply a combination of different chronological methods aiming to confirm and add further details to the results presented here, and the Icelandic glacial history in general.

The landform-sediment assemblage at Drangajökull surge-type glaciers is different from and lacking typical features compared to the general surging glacier landsystem model of Evans and Rea (1999, 2003) which is partly based on studies of Vatnajökull outlets. One of the potential explainations is considered to be the glacial rivers which meander around the valley floors, erode landforms and re-deposit sediments. This could be investigated by studying time series of aerial photographs. At least four sets of aerial photographs, from from the 1940s to 2005, exists from each glacier forefield. The older aerial photographs could reveal landforms produced by the surges in the 1930-1940s. Landforms that we do not see at present or on the aerial photographs from 2005, those potential landforms might have been eroded by the glacial rivers and the glaciers during their recent most surge. Thus, the aerial photographs could clarify this and enable remote studies of the forefield evolution after the 1930-1940s surges until the surges around 2000. Furthermore, the glaciers can be expected to retreat considerably during their present quiescent phase, perhaps the landform-sediment assemblage that will be revealed during the coming decades will be more similar to the landsystem model of Evans and Rea (1999, 2003).

Accumulating evidence during the last decades has revealed a variable nature and characteristics of surging glaciers (e.g. Hamilton and Dowdeswell, 1996; Harrison and Post, 2003; Ingólfsson et al., 2015). This study suggests that a combination of environmental factors affects the expression and pattern of the main surge characteristics of the Drangajökull ice cap, and makes it different from the surge-type glaciers of the larger ice caps in central and south Iceland. However, fundamental questions remains unresolved, like why do some glaciers surge, is there any trigger factor for surges afte all, or are surges rather result of gradual development of the glaciers surface profiles, why then is the balance velocity of those glaciers not correspondent to their mass balance and what enhances their fast flow during surges?

Though this study reveals the main characteristics and the surge history of the Drangajökull surge-type glaciers, it does not resolve all fundamental questions about surge dynamics. For doing that, a thorough real-time study and monitoring of surging glaciers, from before surge initiation until after surge termination, is necessary. Preferably, a whole surge cycle should be monitored and studied.

However, a unifying theory might not be expected to answer those questions and explain the nature of all surge-type glaciers. There are most likely different reasons and environmental conditions that can explain similar main characteristics of different glaciers.

This study suggests that the main characteristics of Drangajökull surges depend on a complex combination of their environmental conditions. Further investigation of the Drangajökull surge-type glaciers, e.g. extensive monitoring of climate and the glacier conditions and geophysical surveys both of the glaciers and their forefields might contribute to improved understanding of the properties of the Drangajökull surge-type glaciers.

7. Conclusions

This thesis presents geomorphological, sedimentological and chronological data gathered around the perimeter of the Drangajökull ice cap and Hornstrandir during fieldwork in the years 2011, 2012, 2013, and 2014. Those data along with a review of historical data from the area have enabled reconstruction and improvements of the terrestrial glacial history, and configuration and dynamics of the Drangajökull ice cap. The main conclusions are as follows:

- The sediment cover around Drangajökull is generally thin, coarse grained subglacial till and locally weathered bedrock, except the main valley depocentres characterised by sandur covered valley floors. Almost no geomorphological features were observed around the southern perimeter of the ice cap.
- The mapped landforms of the Drangajökull surge-type glaciers are not unique for surging glaciers, and furthermore, the mapped landform assemblage is not diagnostic for surging glaciers, nor does it agree well with other landsystem models developed for surge-type glaciers. This is most likely due to the generally thin, coarse grained basal till and the extensive fluvial erosion. The Neogene plateau basalt substratum of Drangajökull, and sub-glacial hydrological and thermal conditions during surges might also account for this pattern of erosion and sedimentation.
- A new landsystem model for strongly topographical confined surge-type outlet glaciers emphasises an extensive sandur covering the valley floor, distinct terminal moraines, and a coarse grained fluted till plain proximal to the glaciers. Kame terraces, eskers, pitted outwash and hummocky moraines are rarely occurring landforms.
- The surge-type outlets of Drangajökull reached their LIA maximum extents asynchronously during surges ~1700-1846 AD, reaching 3-4 km further down-valley and having ice volumes 2-2.5 km^3 greater than at present. Twice as many end moraines as previously recorded were mapped. The surge interval varies in order of 10-140 years between and within the outlets. Surges seem to have been more frequent during the 19^{th} century and the earliest 20^{th} century compared to about 50-60 year periodicity after 1920 AD. No clear relationship between surge initiation or periodicity and climate could be established.
- During surges, a maximum surface thinning in the order of 10-30 m occurs in the upper reservoir areas, which coincides with areas of maximum accumulation during quiescent phase. The surges contribute at least to 30-40% of the volume loss in the reservoir area during the period 1994-2005. During the present quiescent phase, the reservoir areas thicken by c. 0.5-0.7 m a^{-1} and the receiving areas thins by c. 1 m a^{-1},

which will bring the glacier surface profiles towards a pre-surge stage in 45-65 years.

- The ice sheet over Vestfirðir during the last glaciation was characterised by topographically controlled fjord/valley-type ice streams or glaciers which flowed faster than grounded ice domes over plateaux and some uplands. A thin, non-erosive and cold-based ice covered the plateaux and some uplands, while warm-based erosive ice filled valleys and fjords during LGM.
- Ice thinning and deglaciation of some mountains at least c. 26 ka BP, preceded any considerable lateral retreat of the ice sheet. The terrestrial deglaciation was stepwise and asynchronous. Uplands and some valleys were deglaciated 14-15 ka BP while some valleys were occupied with ice at least until c. 9 ka BP.

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Appendix I

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Geomorphology and the Little Ice Age extent of the Drangajökull ice cap, NW Iceland, with focus on its three surge-type outlets

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ABSTRACT

Detailed geomorphological maps from the forefields of three surging outlets of the Drangajökull ice cap, northwest Iceland, are presented. The maps are based on field studies in 2011–2013, high resolution orthorectified aerial photographs recorded in 2005–2006, and airborne LiDAR data from 2011. The maps cover an area of about 40–60 km^2 each. Furthermore, we present an overview map that covers the area surrounding the Drangajökull ice cap. Landforms and sediments were manually registered in a geographic information system (ESRI ArcGIS 10). We mapped glacial landforms such as flutes, ice-sculpted bedrock, hummocky moraine, kame terraces, and moraines. Fluvial landforms include outwash plains/sandur, pitted sandur, and eskers. In addition raised beaches were mapped. The Little Ice Age (LIA) maximum extent of Drangajökull and its outlet glaciers are fingerprinted by surficial till deposits and freshly glacially scoured bedrock. Sediments distal to the LIA deposits were recorded and consist mainly of late Weichselian and early Holocene sediments and locally weathered bedrock. Periglacial activity is demonstrated by patterned ground, mainly occurring on the 500–700 m high plateaux, and three rock glaciers. At least 3–4 surge events are described from each of the outlet glaciers, occurring over the last three centuries. In contrast to most other surge-type outlets from Icelandic ice caps, the Drangajökull outlets are confined within valleys, which affect the forefield geomorphology. Glaciofluvial landforms, moraines, and a thin sheet of till with numerous boulders are characteristic for the forefields of the Drangajökull outlets.

1. Introduction

The three main outlets of the Drangajökull ice cap (Reykjarfjarðarjökull, Leirufjarðarjökull and Kaldalónsjökull) are all surge-type glaciers (Þórarinsson, 1969; Björnsson et al., 2003; Sigurðsson, 2005; Sigurðsson and Williams, 2008). Eythorsson (1935) and Þórarinsson (1943) summarized the historical information regarding the fluctuations of the glacier margins, as chronicled by local farmers and previous explorers in the area. Leirufjarðarjökull and Kaldalónsjökull surged around A.D. 1700 and 1740. All three glaciers surged in the mid-nineteenth century and again in the 1940s. Finally, Leirufjarðarjökull and Kaldalónsjökull surged during the period 1995 to 2000 and Reykjarfjarðarjökull between 2002 and 2006 (Sigurðsson, 1998; Sigurðsson and Jóhannesson, 1998; Sigurðsson, 2003; Þrastarson, 2006). Björnsson et al. (2003) described the nature of surging glaciers in Iceland, concluding that the advance of the surging terminus usually lasts a few months. However, during the last two surges of the Drangajökull outlets, the advance of the terminus lasted about five years (Sigurðsson, 1998; Sigurðsson and Jóhannesson, 1998; Sigurðsson, 2003). Such long lasting advances of other surging glacier termini in Iceland have only been reported from Búrfellsjökull, a small surging cirque glacier in Tröllaskagi, north Iceland (Brynjólfsson et al., 2012). This makes the surge activity of Drangajökull unique, compared to the surge-type outlets of the other ice caps in Iceland. The duration of the Drangajökull surges resemble the surging of glaciers of Svalbard where the active phase typically lasts 3–10 years (Dowdeswell and Hamilton, 1991; Jiskoot et al., 1998; Murray et al., 2003).

Glacier surging is a cyclic flow instability generally thought to be triggered within the glacier system rather than by external climate forcing (i.e., Benn and Evans, 2010). However, notably, Striberger et al. (2011) suggested that a mass-balance control on surge frequencies was found on Eyjabakkajökull, a northeastern outlet of Vatnajökull ice cap, Iceland. During the active phase of surge, ice is transferred from the reservoir area to the snout of the glacier. Ice flow velocity of the active phase can be up to a thousand times faster than the quiescent phase, whereas the quiescent phase is characterised by snout stagnation and ice build-up in the reservoir area (Meier and Post, 1969; Raymond, 1987; Sharp, 1988; Harrison and Post, 2003; Benn and Evans, 2010). Landform-sediment assemblages on the foreland of such glaciers have been used to define a surging glacier land system model (Evans and

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Rea, 1999, 2003). The forefield of the nontopographically confined surging glaciers, Brúarjökull, an outlet of northern Vatnajökull, was one of the main study areas for the land system model (Evans and Rea, 1999, 2003; Kjær et al., 2008). Such an approach can be used to recognise the imprint of former surging on the landscape (Evans and Rea, 2003; Benn and Evans, 2010). However, sometimes this land system model has to be modified as each surging glacier is unique. A recent study of small surging cirque glaciers in Tröllaskagi, northern Iceland, shows that a modified version of the surging glacier land system model best described the fingerprints of surging cirque glaciers (Brynjólfsson et al., 2012), and Schomacker et al. (2014) described a modified surging glacier landsystem model for Eyjabakkajökull, eastern Iceland.

Geomorphological maps exist from the Drangajökull region (Eythorsson, 1935; Þórarinsson, 1943; John and Sugden, 1962; Groove, 1988; Principato, 2008). However, no detailed geomorphological maps exist of the forefield of the Drangajökull surging outlets, except the map of Kaldalón outlet by John and Sugden (1962). The aim of this paper is to present detailed geomorphological maps and describe the sediments and landforms that occur in the forefield of the three surging outlets of the Drangajökull ice cap. This is in order to better interpret the scarcely known surge dynamics before the observed surges in the 1930s and to identify characteristic landforms and sediments formed by the surges. Furthermore, based on the geomorphological mapping and historical information, we aim to reconstruct the Little Ice Age (LIA) maximum extent of the ice cap.

2. Regional setting

The Drangajökull ice cap reaches 915 m above sea level (asl) and is located on the eastern highland plateau of the Vestfirðir peninsula in northwest Iceland (Fig. 1). This dome shaped, fifth largest ice cap in Iceland, rests on Neogene flood basalts interbedded with thin sedimentary layers (Einarsson, 1991; Kristjánsson and Jóhannesson, 1994). The equilibrium line altitude (ELA) at 550–600 m asl is about half the altitude of ELA on the other ice caps in Iceland, reflecting low summer temperature, short melting season, and high precipitation over the eastern Vestfirðir peninsula (Eythorsson, 1935; Crochet et al., 2007; Björnsson and Pálsson, 2008). This unique glacial condition, in terms of Iceland, is considered to be amplified by the proximity of Greenland and the cold polar East Greenland Current (EGC) (Bergþórsson, 1969; Björnsson, 1979; Ingólfsson et al., 1997; Eiríksson et al., 2000). Short distance to the open ocean to the west, north, and east of the eastern Vestfirðir peninsula favours the access of moist air to the glacier (Eythorsson, 1935). The EGC converges with relatively warm Atlantic water of the Irminger Current off the Vestfirðir peninsula (Stefánsson, 1969; Eiríksson et al., 2000). Enhanced advection of either of the currents strongly affects sea ice extent north of Iceland and regional changes in temperature (Bergþórsson, 1969; Stötter et al., 1999; Hanna et al., 2004; Geirsdóttir et al., 2009).

The present climate of Iceland is classified as cool, temperate maritime (Einarsson, 1976). Regional climate conditions at Drangajökull are characterised by steep precipitation and temperature gradients from the northeast to the southwest. The mean summer temperature (June–September) is 6–7 °C on the northeast coast of the ice cap and 8–9 °C on the west coast of the ice cap (Eythorsson, 1935; Hanna et al., 2004; http://www.vedur.is/Medaltalstoflur-txt/Manadargildi.html accessed 15 August 2013). The precipitation gradient reflects prevailing wind direction in the area from the northeast (Einarsson, 1976), with about 1100 mm annual average precipitation close to sea level on the northeast coast and about 580 mm on the west coast on the lee side of the Drangajökull ice cap (Crochet et al., 2007; http://www.vedur.is/Medaltalstoflur-txt/Arsgildi.html accessed 15 August 2013).

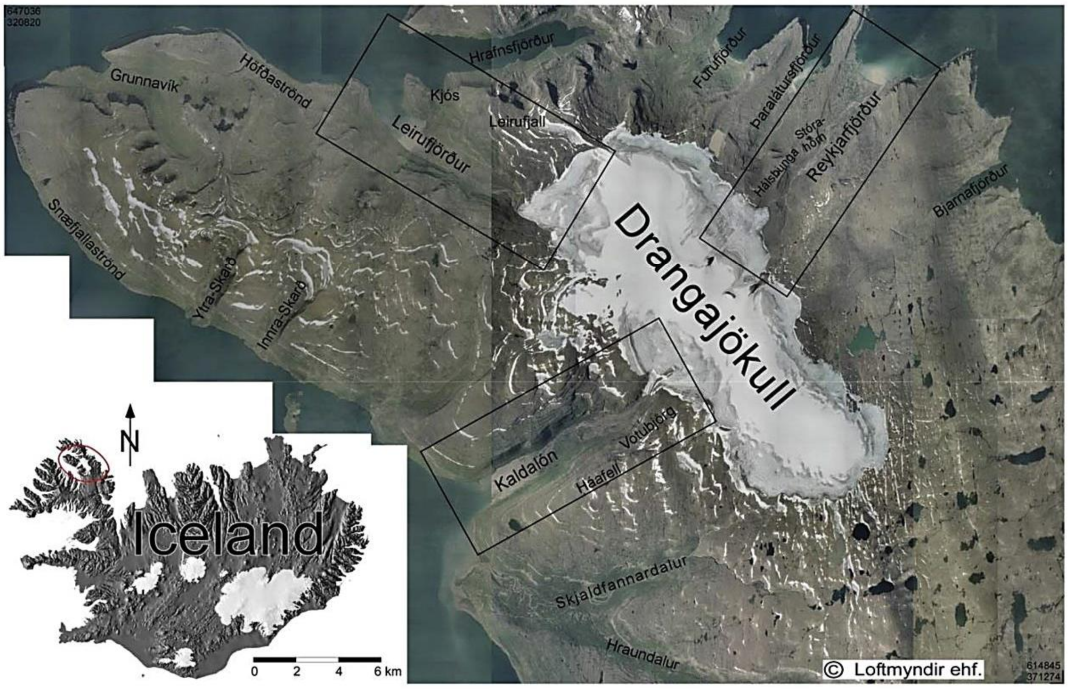

Fig. 1. The Drangajökull ice cap, located on the eastern Vestfirðir peninsula, north-western Iceland. The main geographical names are presented on aerial orthophotos from 2005 to 2006 taken by Loftmyndir ehf.

Three surging outlet glaciers have been recognised draining the Drangajökull ice cap, Reykjarfjarðarjökull to the northeast, Leirufjarðajökull to the northwest, and Kaldalónsjökull to the west. Records show that these glaciers are known to have surged at least three or four times since the settlement of Iceland (Sigurðsson, 1998; Björnsson et al., 2003). Timing of the last two surge events of each of the three surging outlets is well documented (Sigurðsson, 1998; Björnsson et al., 2003; Sigurðsson, 2003, 2005; Björnsson and Pálsson, 2008). The outermost end moraines, except the moraine in Kaldalón, were formed in the eighteenth and the mid-nineteenth centuries (Eythorsson, 1935; Þórarinsson, 1943; Sigurðsson, 1998; Björnsson et al., 2003; Sigurðsson, 2005; Principato, 2008). In Kaldalón the two outermost end moraines have been dated to Younger Dryas and Neoglacial ages (Principato et al., 2006). Since the LIA maximum, the Drangajökull outlet glaciers have retreated 3000–3500 m from their outermost terminal moraines, and extensive surfaces that formed during surge activity have been revealed. Surge activity has not been reported from other outlets or parts of Drangajökull. From SPOT 5 satellite imagery, the present size of the Drangajökull ice cap has been estimated to 150 km^2 (Þrastarson, 2006). Eythorsson (1935) estimated its size about 160 km^2, while the map survey of the Danish army in 1913–1914 shows its size to be about 200 km^2 (Þórarinsson, 1958; Björnsson, 1979). Two further estimates of the size of the ice cap were published by Björnsson (1979) who estimated its size to be 160 km^2 from air photos taken in 1960, and finally Jóhannesson et al. (2013) calculated its size in 2011 from LiDAR data to be 142 km^2. The area of Drangajökull, except the three surging outlets, thus appears to have been relatively stable during the last 50 to 80 years.

3. Methods

3.1. Data

The base data used for mapping the geomorphology of Drangajökull are aerial photographs with 0.5-m pixel size dating from 2005 and 2006 supplied by Loftmyndir ehf. and a LiDAR (light detection and ranging) derived digital elevation model (DEM) with 5-m ground resolution from 2011, supplied by the Icelandic Meteorological Office (IMO).

Extensive field work was carried out during the summers of 2011, 2012, and 2013. Valleys, fjords, mountains, and ice margins of the northern half of the ice cap were surveyed by walking around the perimeter from Reykjarfjörður on the east side to Kaldalón on the west side of the ice cap (Fig. 1). Landforms and sediments were described, interpreted, and classified according to genesis. Moraines representing former glacier extension, raised beaches, and other landforms and sediments relevant for the glacial history of Drangajökull were localized by Garmin GPSMAP 62sc, with horizontal accuracy close to ±3 m. The ISN93/WGS84 reference system was used for all data handling and the final maps.

3.2. Geomorphological mapping

The geographical information system (GIS), ESRI ArcMap 10, was used to manually digitize all the mapped features. Final layout of the maps was performed in the drawing software Canvas, version 11. On the overview map (Fig. 2), polygons are drawn to represent different sediments and surfaces, and polylines are used to delineate line-shaped features, e.g., moraines, raised beaches, and rivers.

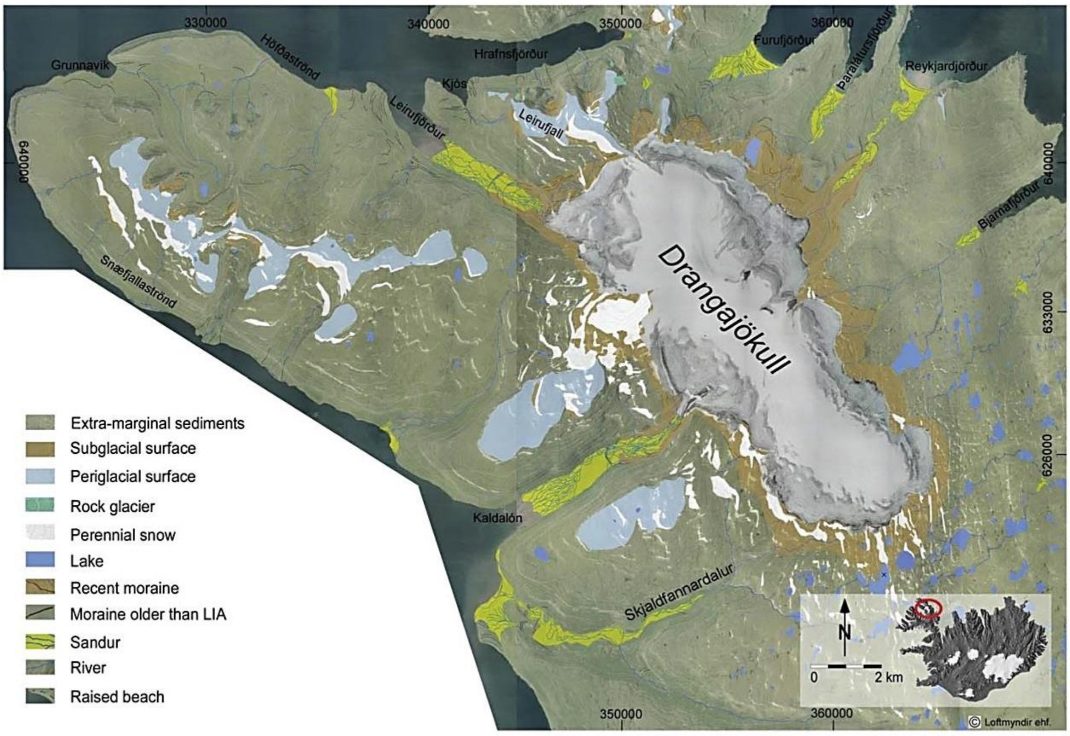

Fig. 2. An overview map of the geomorphology and sediments around Drangajökull, an x-mark south of the ice cap locates a lake sediment core. See online Supplementary data for full version of the map.

The detailed maps (Figs. 3, 4, 5) are processed in much more detail. Smaller landforms and indistinct sediments were registered in a scale of 1:1000 or better resolution. The LiDAR DEM, in addition to careful field exploration, results in accurate representation of landforms and sediments. Features such as *subglacial surface*, *sandur*, *pitted sandur*, *extra-marginal sediments*, *perennial snow*, *periglacial surface*, *glacial river*, *moraine*, *hummocky*, *esker*, *kame terrace*, and *ice sculpted bedrock* were delineated by polygons. Linear-shaped landforms such as *channels*, *flutes*, and *raised beaches* are represented by polylines. The maps are designed to be viewed digitally or printed in A0 format. Some landforms may appear as cluster or indistinct if the maps are printed on A3 or A4 format.

4. Mapped landforms

Below, we describe the geomorphology of Reykjarfjarðarjökull, Leirufjarðarjökull, and Kaldalónsjökull, which are unlike other described surge-type glaciers in Iceland. Their landform assemblage and distribution do not fit well into the land system model of Evans and Rea (1999, 2003). First we describe the general settings and the LIA maximum glacier extent of the Drangajökull region. Fig. 2 provides an overview of the distribution of the sedimentary units and landforms in the forefield of the Drangajökull ice cap and its surroundings. Then we describe the geomorphology related to the glacier surges. Detailed maps of the valleys of the three surging outlet glaciers are presented on separate maps shown in Figs. 3, 4, and 5. Finally, we describe other distinct geomorphological features in the region, not related to the surge activity.

4.1. The Drangajökull region

A former subglacial surface extends from the present glacier margin to the LIA limit (Fig. 2). In our mapping, this class (subglacial surface) includes basal till, locally weathered bedrock, and ice-sculpted bedrock dissected by meltwater streams (Fig. 6A). The class represents the LIA

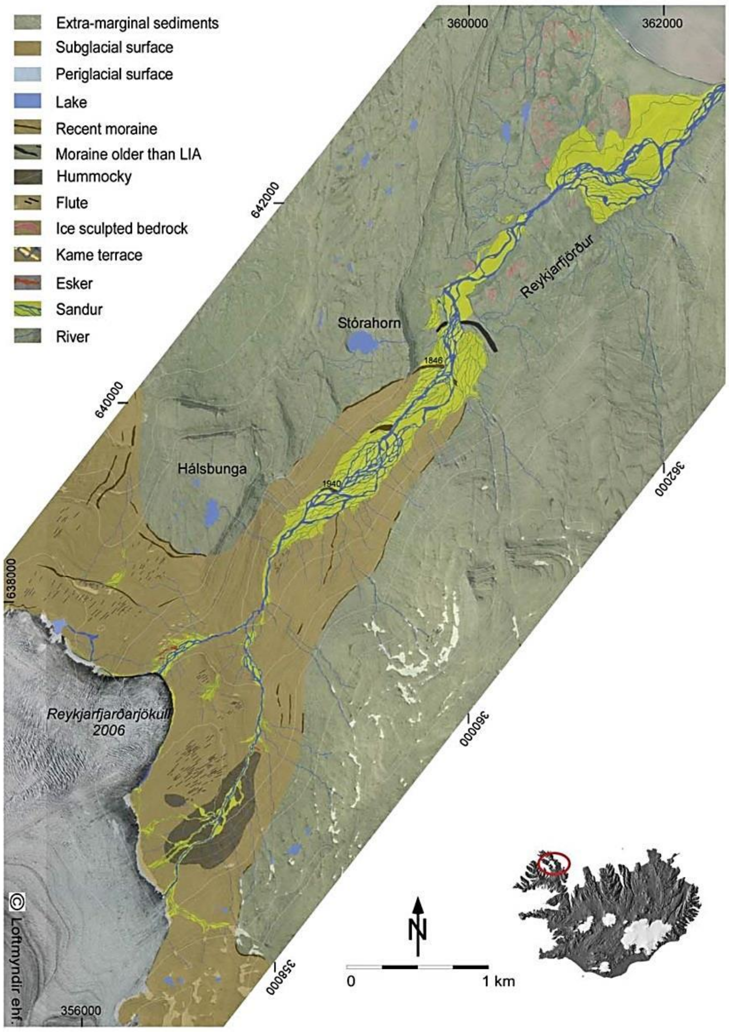

Fig. 3. A glacial geomorphology map of Reykjarfjarðarjökull. See online Supplementary data for full version of the map.

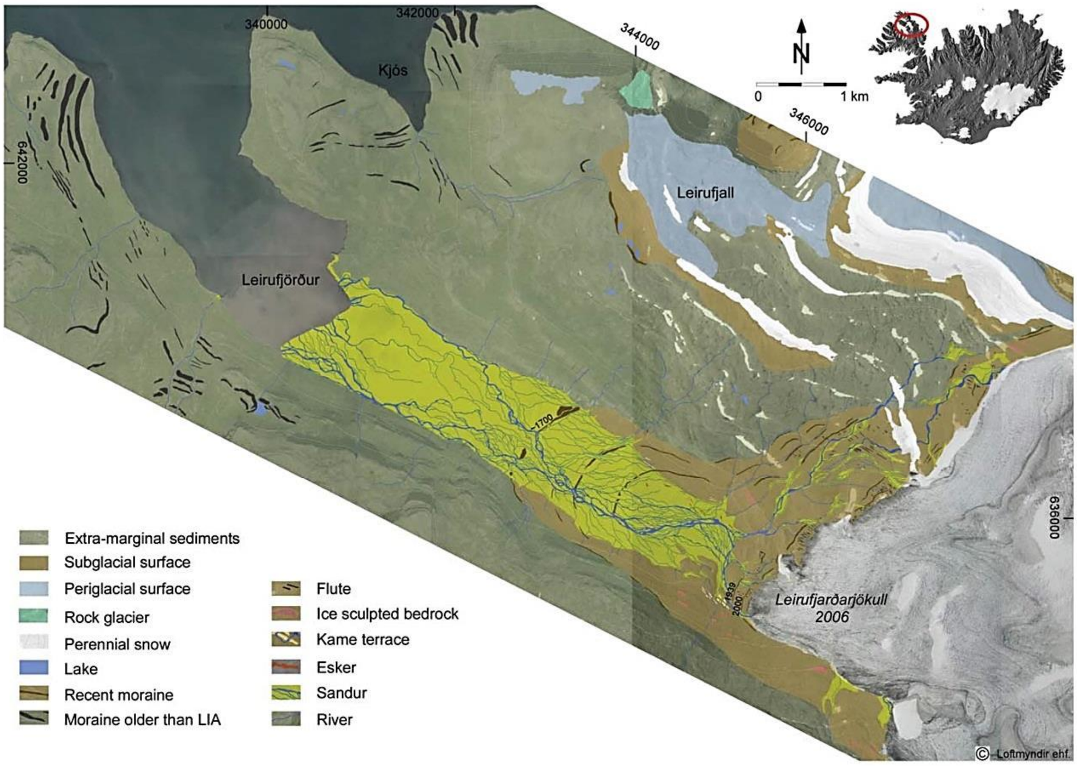

Fig. 4. A glacial geomorphology map of Leirufjarðarjökull. See online Supplementary data for full version of the map.

maximum extent of the glacier. Where possible it is defined by the LIA maximum moraines, otherwise with other landforms and sediments related to the LIA maximum position. Mapping of the subglacial surface is mainly based on field observations for the northern perimeter of the ice cap and remote sensing for the southern perimeter. Generally, the till is coarse grained, rich in subangular to subrounded basaltic pebbles and boulders (Fig. 6B). Glacially scoured bedrock sporadically protrudes the thin till cover. It is often streamlined and glacially striated heading approximately parallel to the valleys draining the glacier.

All superficial sediments beyond the mapped LIA extent of Drangajökull, except the sandur and the periglacial sediments, were mapped as extra-marginal sediments. This term includes remains of late Weichselian and early Holocene tills (Hjort et al., 1985; Geirsdóttir et al., 2002; Principato et al., 2006; Principato, 2008), bedrock, and locally weathered bedrock (Fig. 6C). It also includes scree material, soil, wetlands, and vegetation.

Erratics (Fig. 6D) are more densely scattered within known LIA terminal moraines than outside of them. The exact position of the LIA maximum ice limit is difficult to reconstruct all around the ice cap, in particular around the southern perimeter, where the geomorphological imprints are negligible. Lacustrine sediment cores retrieved about 1 km beyond the present southern ice cap margin (Fig. 2) indicate that the glacier extended at least to that location. At present this lake receives glacial meltwater. The core consists of about 30 cm of silt and clay on top of a diamict that is interpreted as till. This and the absence of organic material in the core are the main indication of glacier covering that area for a period during the LIA. The lack of geomorphological imprints indicates a thin and not very dynamic (perhaps cold-based) LIA ice margin in this area, where the elevation is about 500–600 m asl. At the south margin, the ice cap extended to the head of the Skjaldfannardalur valley (Fig. 2) in the mid-nineteenth Century (Shepard, 1867). Because of well preserved terminal moraines and fluted surfaces, as well as historical evidence, the LIA extent is well constrained from Bjarnafjörður on the northeast side to Leirufjörður on the northwest side of the glacier. Relatively small marginal variations since the LIA seem to be a general tendency on the plateaux around the ice cap. The main variations of the ice cap margins occur where the outlet glaciers descend into topographical lows (Fig. 2).

Moraines are presented in two different colour classes, i.e., recent moraines known to have formed from the LIA until present and other moraines older than LIA or of unknown age.

Moraines mainly occur around the northern half of the Drangajökull ice cap (Fig. 2). End moraines occurring on higher slopes and in the small hanging valleys north of the Drangajökull ice cap are generally inconspicuous low crest ridges 5 m in height and width, consisting of coarse diamict and occasionally predominantly of boulders (Fig. 6E). Lateral moraines are generally characterised by indistinct low relief moraine ridges, 1–5 m in height and width, of diamict and boulders. These features are particularly well preserved in Leirufjörður and Kjós (Fig. 4), where they are represented by a series of distinct lateral moraines, previously mapped by Principato (2008), up to 40 m wide and 15 m high (Fig. 7A), and can be traced along the valley sides. Furthermore, distinct moraines of unknown age occur far beyond the LIA maximum extent in Grunnavík and Skjaldfannardalur (Fig. 2).

Ice-sculpted bedrock (roche moutonneés and whalebacks) was spotwise recorded in front of the three surging outlets (Figs. 3, 4, and 5). The ice-sculpted bedrock outcrops are from 2 to 3 m long and wide, up to 100 m in diameter, and usually elongated in the former ice

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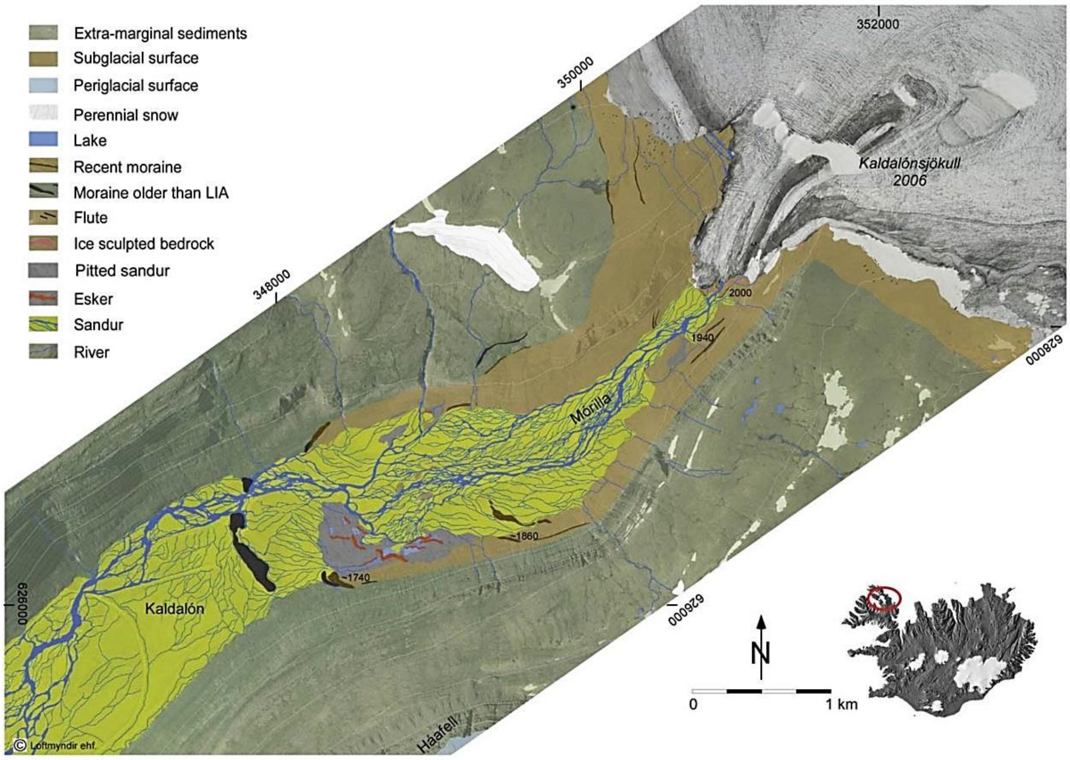

Fig. 5. A glacial geomorphology map of Kaldalónsjökull. See online Supplementary data for full version of the map.

flow direction. Smaller erosional forms like glacial striae also have a direction parallel to the valleys, and P-forms occasionally occur on the bedrock surface.

Flutes are elongated landforms aligned parallel to the former glacier flow (Fig. 7B). In addition to the flutes mapped in the forefields of the surging outlets, distinct flutes, up to 100 m long and 1 m in relief, were observed at the head of the Furufjörður, Þaralátursfjörður, and the hanging valleys north of the glacier within the LIA moraines.

Erratics, up to 1 m in diameter, are common around Drangajökull (Fig. 6D), particularly within moraines known to have formed during the LIA. However, they also occur outside the LIA limit, predominantly on mountain plateaux and along the coastlines fringing the ice cap. Because of their great number and wide dispersal, the erratics were not mapped.

Outwash plains and sandurs are gently down glacier sloping surfaces of glaciofluvial deposits, from braided glacial rivers, and are characterised by networks of abandoned channels and active river channels (Fig. 7C). Lowland valley floors around Drangjökull are partially filled with glaciofluvial sediments; eight sandurs related to present drainage of the glacier were mapped. The sandurs covering the largest areas are located in Kaldalón, Leirufjörður, and Reykjarfjörður. Sandurs covering smaller areas occur in Þaralátursfjörður, Bjarnafjörður, Furufjörður, and Skjaldfannardalur, which all drained outlet glaciers during the LIA (Shepard, 1867; Eythorsson, 1935). Sandurs were also mapped at Höfðaströnd, Unaðsdalur, and occasionally along the coast on the southwest side the ice cap. That is not known if the latter were active during the LIA or later. Braided meltwater channels commonly occur on the sandurs. Some are recurrently occupied by water, others fill only during flood events. Owing to their chaotic and dense distribution, the major stream channels were mapped as polygons while smaller streams and dry channels were mapped as polylines.

The periglacial surfaces (Fig. 8A) were primarily documented on flat, wind-exposed plateaux at 500–750 m asl (Fig. 2), where they are characterised by block fields, polygons, and stripes. The polygons at this elevation are usually 2–5 m in diameter and up to 1 m in relief.

4.2. Geomorphology of the surging outlets in Reykjafjörður, Leirufjörður and Kaldalón

4.2.1. Reykjarfjörður

Four separate end moraines were mapped on the valley floor in Reykjarfjörður (Fig. 3). This is one more than mapped by Eythorsson (1935), Þórarinsson (1943), and Principato (2008). The outermost moraine, about 4500 m distal to the present ice front, is a poorly defined, vegetated, horseshoe-shaped moraine. It is 5 m high and 10–30 m wide, consisting of gravelly and boulder-rich diamicton. Its age is unknown, and since no sections exist where sediments and structures can be observed, and therefore not known if it is primarily formed by dumping or active thrusting. About 250–350 m proximal to the glacier from the outermost moraine is a 10–15 m high moraine formed by the surge in AD 1846, consisting mainly of gravels pushed/thrusted up from the pre-surge outwash plain (Eythorsson, 1935; Principato, 2008). A lateral moraine continuing from the 1846 end moraine can be followed for about 900 m along the northern valley side. Located about 1600 m proximal to the glacier from the 1846 moraine is a poorly developed end moraine formed by a surge in AD 1936–1940

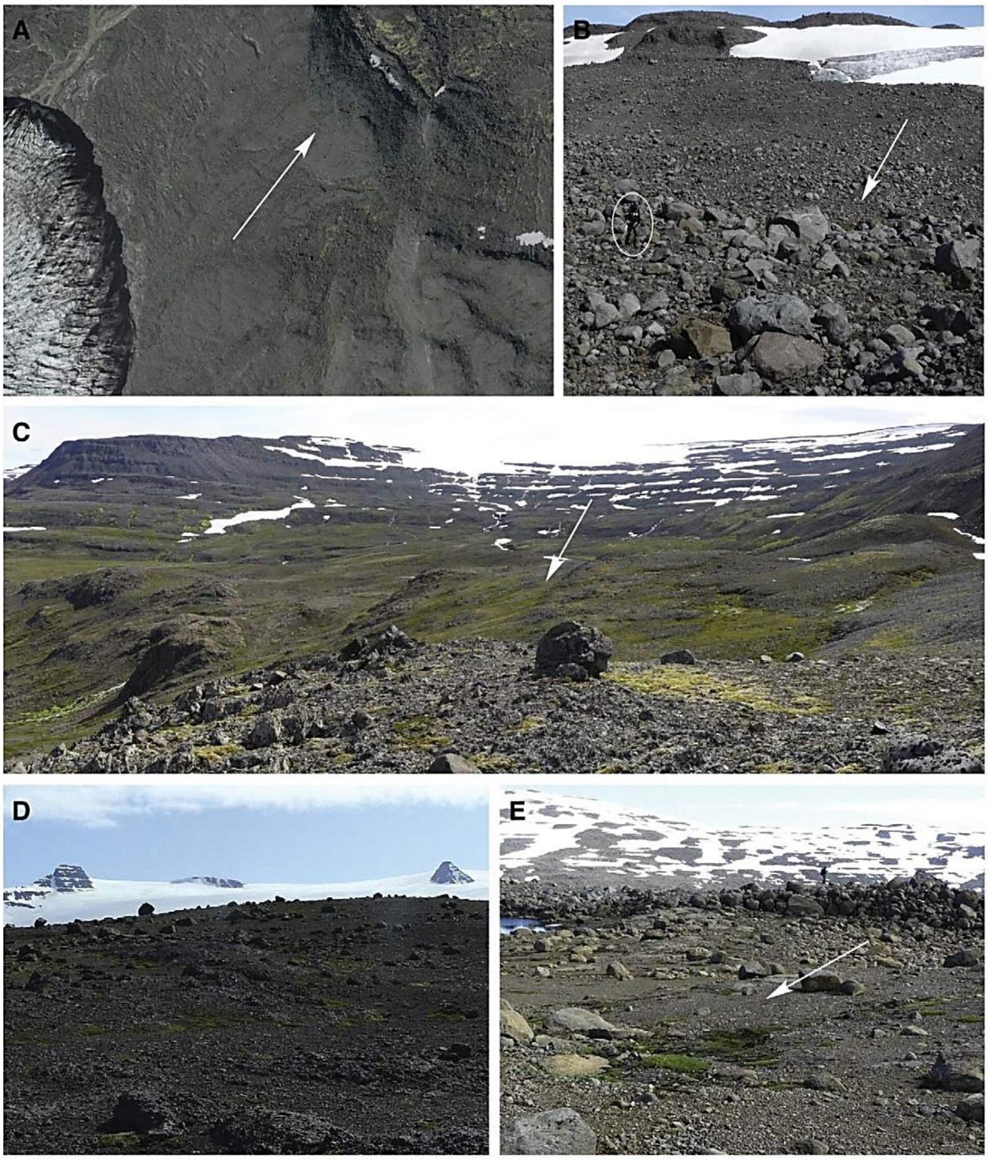

Fig. 6. Examples of landforms in the forefield of Drangajökull. The arrows indicate ice flow direction. (A) Overview of 'subglacial surface' in front of Reykjarfjarðarjökull. (B) Typical 'subglacial surface' proximal to present margin of Leirufjarðarjökull. (C) Extra-marginal sediments at the valley floor of Furufjörður. Today, the glacier barely reaches into the head of the valley. (D) Erratics on the mount Hálsbunga in Reykjarfjörður, proximal to present glacier margin. (E) The distal most terminal moraine of Reykjarfjarðarjökull formed during a surge around AD 1840. Located on top of the mountain Hálsbunga where the sediment supply is limited; as a consequence, the end moraine mainly consists of boulders at this place.

(Sigurðsson, 1998, 2003; Björnsson et al., 2003). Between the moraines formed by the 1846 and 1940 surges the third moraine is located, which age and process of formation are unknown. The 1940 moraine and the third moraine have been extensively eroded by the meltwater streams; the moraines today are primarily recognised by trains of remnant boulders on the outwash plains. The glacier last surged in 2001–2006 (Þrastarson, 2006), but because of snow cover along the margin during fieldwork in summer of 2011, we were not able to recognise or map a corresponding moraine. Dissected lateral moraines occur along the southern slope of the valley and discontinuous end moraines on Hálsbunga, which most likely correlate to the glacial events that caused the formation of the three end moraines since AD 1846. However, the lateral moraines remain undated and their correlation is therefore not clear.

Inconspicuous flutes, up to a few tens of metres long, were mapped within the area influenced by the 1940 surge and proximal to the end moraines at Hálsbunga. Hummocky moraine consisting of diamict—characterised by small depressions, hills, and hillocks—covers a

Fig. 7. Examples of landforms in the forefield of Drangajökull. The arrows indicate ice flow direction. (A) Lateral moraines along the lower part of the 350-m high mountain, Kjósarnúpur. (B) Distinct flutes within the Little Ice Age limits of the ice cap, along its northern margin. (C) Sandur in the valley Reykjarfjörður, farther out on the plain is the Little Ice Age limit of Reykjarfjarðarjökull.

topographic depression proximal to the present glacier margin in the southern part of the valley Reykjarfjörður (Fig. 8B). It is interpreted to represent melting of stagnant glacier ice after the 1940 surge termination. Fragmented kame terraces composed of gravel, cut by meltwater channels, and slumped as debris flows occur within the AD 1940 moraine. They could indicate stepwise shrinkage of the glacier after surge stagnation. An ~5-m-high, 10-m-wide, and 200-m-long esker, consisting of coarse-grained gravel, occurs within the area of the AD 1940 surge (Fig. 8C). Two smaller eskers are located nearby and a small esker in the southern part of the valley.

4.2.2. Leirufjörður

Seven separate moraines were mapped in the forefield of the Leirufjörður surge-type outlet (Fig. 4). The two outermost end moraines are ~3000 m distal to the present glacier margin and located together. The terminal moraine is heavily eroded, mostly vegetated, about 10 m high and 15 m wide, and extending about 200 m. The end moraine located at its proximal side is 5–15 m high, 5–20 m wide, and partly vegetated (Fig. 8D). Its composition (mainly silt, sand, soil, and gravel) is revealed in a 10-m-wide and 3-m-high river-cut section. Thrust planes and folds show that the moraine contains pushed and deformed sediments. On the west side of the valley floor the moraine has been eroded by the glacier river and consists mainly of boulders and diamict. Between 300 and 1000 m within the two outermost moraines, we mapped tree indistinct ridges that we interpret to be end moraines, each 1–3 m high and 3–10 m wide. The ridges are fluvial eroded, consisting of remnants of coarse-grained diamicton and boulders. Their process of formation is unknown. Most proximal to the glacier, located at an elevation of 80–100 m asl are two moraines formed by the surges in AD 1938–1942 and AD 1995–2001, respectively (Björnsson et al., 2003; Sigurðsson, 2003, 2005). They are 1–3 m high and 2–6 m wide, consisting of the coarse-grained diamicton. Six separate lateral moraines, of similar appearance as the AD 1942 and AD 2001 moraines, occur on the eastern valley side. They most likely correlate with the end moraines described above.

Clusters of poorly preserved flutes, up to 100 m long and perpendicular to the end moraines, are located at the eastern side of the valley. Distinct kame terraces of gravel are situated in a river gully at the head of the valley about 300 m outside the present ice margin (Fig. 9A). Fragmented kame terraces of gravel also occur about 1 km downvalley. Few eskers, about 1–3 m high and 5 m broad and segmented by meltwater channels, are present on the sandur distal to the AD 1942 moraine.

4.2.3. Kaldalón

Eight end moraines were mapped in the forefield of the Kaldalón surging outlet (Fig. 5). The two outermost moraines have been dated to Younger Dryas and Neoglacial ages (Principato et al., 2006; Principato, 2008). The younger of those is not considered to be related to the LIA surge history of the Kaldalón glacier.

We mapped six moraines that were formed by ice marginal fluctuations and surge events since the LIA maximum. On the proximal side of the two old moraines is a steep crested moraine ridge, composited of gravel and diamicton, scarcely vegetated, and 10 m high. It is considered to have formed as a result of a surge around AD 1740 (Sigurðsson, 1998; Björnsson et al., 2003). The next five moraines have been

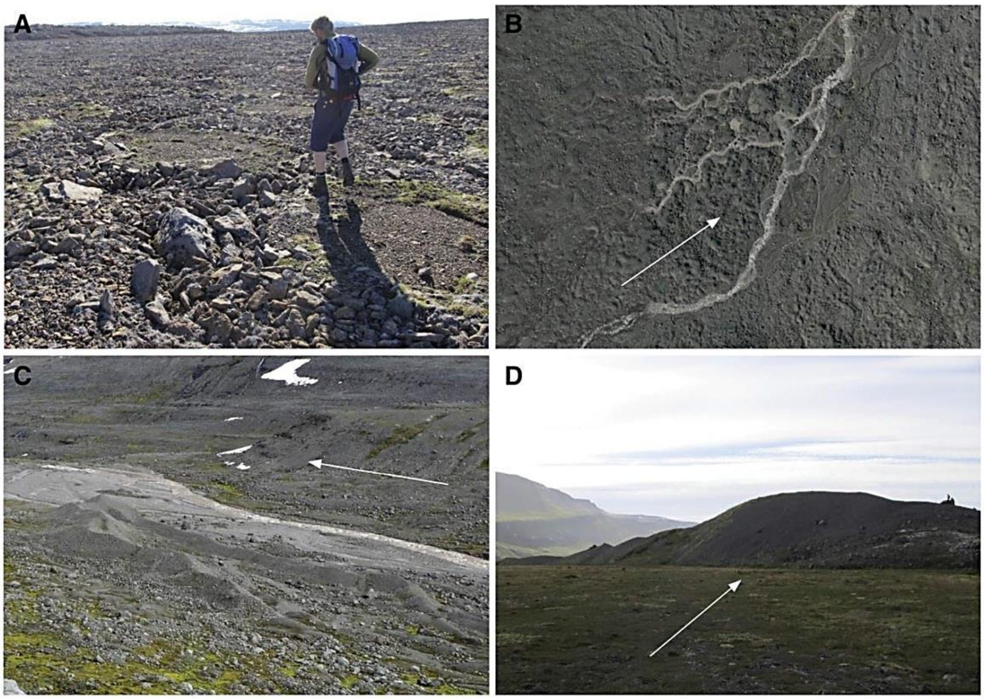

Fig. 8. Examples of landforms in the forefield of Drangajökull. The arrows indicate ice flow direction. (A) Polygons, 2–5 m in diameter, located about 550 m asl on mount Leirufjall. (B) Hummocky moraine about 500 m beyond the present margin of the Reykjarfjörður outlet. (C) A 5-m-high and 200-m-long esker located 50 m in front of the present margin of Reykjarfjarðarjökull. (D) The distal most end moraine in Leirufjörður formed during a surge in the early eighteenth century, about 15 m high and about 3000 m distal to the present glacier margin.

extensively eroded by the glacial rivers and can be identified as low relief latero-frontal ridges, 1–5 m high, consisting predominantly of coarse-grained diamicton and trains of boulders. The fourth moraine in this sequence of end moraines is considered to have formed by surge around AD 1860 (Björnsson et al., 2003). Its 1–4 m high ridge on the southern side of the valley floor consists of coarse-grained diamicton and boulders. Finally, two indistinct moraines proximal to the glacier correlate to the surges that terminated in AD 1940 and AD 1999.

Poorly preserved flutes, 50 m long, are visible in the glacier forefield on the plateau on each side of the valley. Pitted sandur occurs at the foot of the southern valley slope; its surface is uneven with depressions and mounds. Though its appearance resembles the hummocky moraine, the deposit is different; composed of esker segments and glaciofluvial sediments, coarse and sandy gravel with subrounded to rounded pebbles and boulders. The pitted sandur, indicative of dead-ice downwasting, was probably mostly deposited supraglacially. Several eskers occur in the middle of the valley; they range from a few metres in width and height and up to a few hundred metres in length. The most distinct eskers were mapped, but some esker fragments have a chaotic appearance owing to melting of dead-ice, and a hummocky surface that indicates at least some were formed englacially.

4.3. Other geomorphological features

Three distinct geomorphological forms considered to be rock glaciers occur on the northern side of the ice cap: one is situated in a narrow north-facing gully between Kjós and Bæjardalur, and two occur below a northeast-facing sharp mountain edge at the head of Furufjörður (Fig. 9B). They are a few tens of metres thick, and their surface is bare of vegetation and with multiple lobed-tongue—shaped piles of coarse and angular rocks. Their structure and geomorphology closely resemble rock glaciers described by Clark et al. (1998) and Humlum (2000). The rock glaciers have a common location and aspect with actively frost shattered backwalls and restricted exposure toward solar radiation.

Lakes are particularly widespread on the plateaux south and southeast of the ice cap. They seem to occupy structural depressions in the bedrock rather than being formed as moraine-dammed ice-marginal lakes. Because of restricted glacial drift and no moraine dammed lakes, Principato and Johnson (2009) suggested that the lake basins are formed by ice scour, indicating intensively eroded bedrock. Some lakes serve as threshold lakes and have received glacial meltwater sediments during more extended stages of the glaciers. Few of them receive glacial meltwater at present.

Distinct perennial snow fields were mapped where ice and old dark snow were exposed. The aerial photos used for the mapping were captured during late summer when perennial snow fields can be expected to remain after the summer melt.

Raised beaches at about 5 m asl are common and were mapped in the fjords east and north of the glacier (Fig. 9C). Distinct raised beaches 5–8 m asl are common on Snæfjallaströnd and occur sporadically along the whole coast southwest of the ice cap. Additional raised beaches and marine limits have previously been mapped at altitudes up to 30 m asl (John and Sugden, 1962; Principato, 2008), but were not recognised during our mapping campaign. However, ~20 m asl we mapped sandur terraces proximal to the present beach south of Kaldalón. Topset and foreset are revealed in a 2–4 m deep gully intersecting the sandur. We

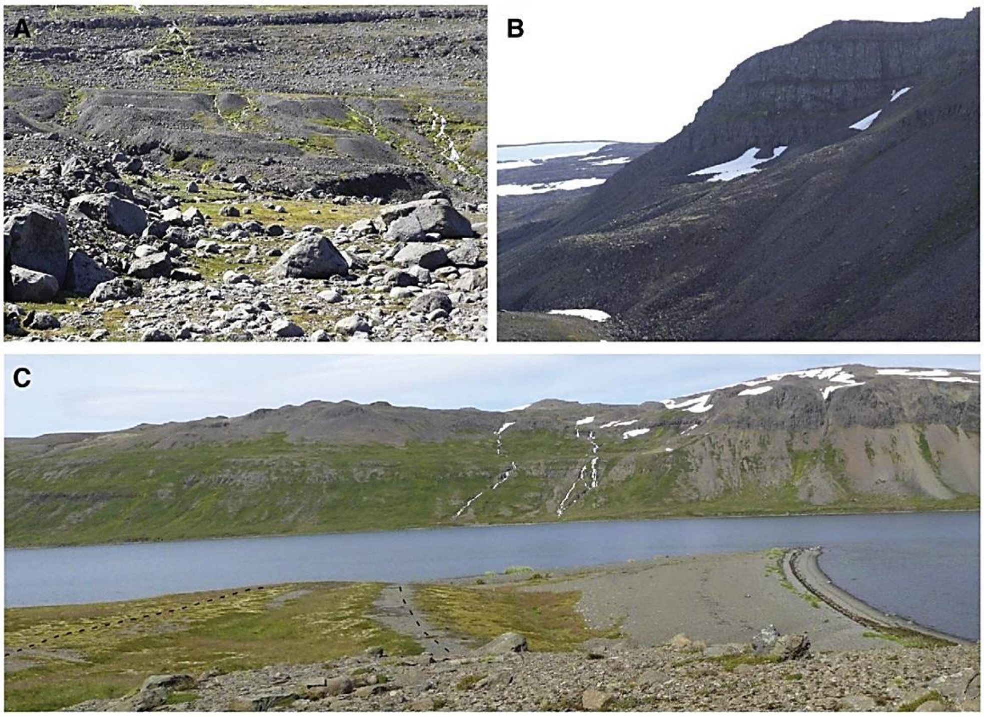

Fig. 9. Examples of landforms in the forefield of Drangajökull. (A) Fragmented kame terraces proximal to the present glacier margin of Leirufjörður outlet. (B) A rock glacier located at the head of Furufjörður, facing northeast. (C) About 5 m high raised beach ridges in Hrafnsfjörður north of the ice cap.

conclude that this landform was either deposited in the sea during higher sea level or into a lagoon, lateral to a glacier extending to the fjord, during deglaciation.

Numerous distinct sharp edged-boulders up to 1.5 m in diameter are present on the sandur on the valley floor of Kaldalón around tidemark at the head of the valley. They appear in clusters and extend from the valley sides to the middle of the valley bottom. Farmers that used to live on now-abandoned farms in the valley have described deposition of some of these boulders by snow avalanches coming down the steep valley slopes (John and Sugden, 1962; Indriði Aðalsteinsson, affiliation, personal communication, 2013). Big snow fields and cornices build up on the lee sides of the prevailing snow drift direction from the north and northeast. Under certain conditions the cornice and the snow fields can fail resulting in a release of an avalanche that plucks boulders from the slope. Some of the boulders are transported 200–300 m across the flat valley bottom.

5. Discussion

We mapped landforms in front of the three surging outlets draining the Drangajökull ice cap. None of the landforms mapped are diagnostic for surging glaciers, and furthermore, the landform assemblages we have mapped are not diagnostic for surging glaciers, according to the surging glacier landsystem model of Evans and Rea (1999, 2003). Moraines, flutes, and sandurs are predominant at Drangajökull surging glaciers; they are also prominent in the land system model of Evans and Rea (1999, 2003). Hummocky moraine, pitted sandur, crevasse fill ridges, and concertina eskers all serve as important components of their land system model. However, they are restricted, obscure, or absent in our study. The conceptual model for a surging glacier landsystem was developed by Evans and Rea (1999, 2003) based on observations from surging glacier forefields in Iceland, Svalbard, U.S.A., and Canada. The Icelandic examples from Vatnajökull have well-preserved landform assemblages conforming well with the land system model and are considered to be good analogues for palaeo surge ice sheet lobes (Evans and Rea, 1999, 2003; Kjær et al., 2008; Schomacker et al., 2014).

5.1. Geomorphology related to surges of Drangajökull

The geomorphology of Reykjarfjarðarjökull, Leirufjarðarjökull, and Kaldalónsjökull are unlike other described surge-type glaciers in Iceland, and their landform assemblage and distribution does not agree well with the landsystem model of Evans and Rea (1999, 2003). In contrast to most other surging outlets in Iceland, the surging outlets of Drangajökull are strongly topographically confined. They could possibly serve as analogues for terrestrial palaeoglaciers and contemporary surging outlets that terminate in strongly topographically confined fjords and valleys.

The absence of diagnostic landforms such as crevasse-fill ridges (Sharp, 1985; Evans and Rea, 1999, 2003; Kjær et al., 2008) and very limited occurrence of hummocky moraine, pitted sandur, and flutes could be a result of the relatively steep and confined topography. Crevasse ridges form when water-saturated basal till is squeezed into fractures extending upwards from the glacier bed; the process appears to be uniquely associated with glacier surges (Sharp, 1985; Benn and Evans, 2010). The valley floors in front of Drangajökull are covered by glaciofluvial sediments; and because of continuous erosion by the meandering glacial rivers, the preservation potential of glacial landforms and sediments is poor. The erosion has degraded many of the foreland moraines, washing away the fines and leaving boulders that mark the end moraine positions. The outermost end moraines of the surging outlets (Figs. 3–5) resemble best moraines described from the not topographically confined surging outlets of Vatnajökull (Benediktsson et al., 2008, 2009). Glaciotectonic structures that indicate that sediment in an end moraine had been pushed and thrusted by the glacier occur in a section in the A.D. 1740 moraine in Leirufjörður. Apart from that site, available sections for careful studies of internal structures and sediments are very limited. We think the indistinct appearance of the moraines closer to the glacier is best explained by intensive river erosion and resedimentation on the topographically confined valley bottoms. Furthermore, the steep valley slopes make the preservation potential of landforms low, which is well illustrated by the fragmented appearance of moraine ridges and kame terraces on the valley sides. However, relatively flat topographically high areas occupied by flutes exist proximal to the present margins of Kaldalónsjökull, Leirufjarðarjökull, and Reykjarfjarðarjökull. The absence of crevasse-fill ridges in those areas proximal to the present margins may be explained by the coarse-grained and thin till. The subglacial till is predominantly gravelly, whereas finer grained sediments are inconspicuous (Fig. 6B). This could also suggest high permeability and low water saturation of the till. The till cover is usually thin on topographic high areas, bedrock structures are often easily discernible below the thin till sheet, and intermittently glacially scoured bedrock protrudes (Figs. 3–5). Where surging glaciers are underlain by thin till, characteristic components such as crevasse ridges may be absent (Benn and Evans, 2010). We consider the till resulting from the Drangajökull outlet advances to be unfavourable for the process of crevasse-squeeze ridge formation. This might also explain the limited and often indistinctive occurrence of flutes, which is common in front of surging and nonsurging outlets from the other major Icelandic ice caps. The characteristics of the subglacial till are generally valid for the Drangajökull forefields and are not limited to the surging outlets. The surficial sediment could also be partly supra- and englacial sediment deposited as the ice retreated. This is also reflected by size and architecture of the end moraines. They are most prominent in basins that have contained sediments for the glacier to push or thrust. They are usually indistinct on topographic rises, constituting of coarse-grained diamict with very little fine material and in some cases only consisting of boulders (Fig. 6E). Colin (1964) described several small ridges of diamicton in front of the margin of Kaldalónsjökull, which he interpreted as annual moraines. However, during our field campaign no annual moraines were identified.

5.2. Reconstruction of Little Ice Age extent

The subglacial surface represents the reconstructed maximum extent of the LIA glaciation. The most important data for reconstruction of the LIA glacier extent are terminal moraines, which together with historical information provide good control on the LIA glacier extent of the northern half of the ice cap.

Except for the moraines in Kaldalón, only few and indistinct moraines occur along the southern and western margins of the ice cap. The density of erratics appears to be higher within moraines known to be formed since LIA and can therefore aid in reconstructing the LIA extent. Flutes are easily eroded, and their occurrence is considered an indication of glacial activity since LIA maximum and reveals one measure of the LIA extent. A complete lack of glacial geomorphological imprints on the plateau between Kaldalón and Leirufjörður made the LIA reconstruction there difficult. Lateral moraines, correlating with the distalmost LIA moraines in Leirufjörður and Kaldalón, are located adjacent to the plateau and immediately proximal to the present ice cap margin. This indicates little variations of the ice cap margin on the plateau. However, the plateau is presently partly covered with perennial snow fields up to 3.5 km^2 in size. Some of the snow fields have been suggested to be remnants of thin glacier ice that covered the plateau until recently (Þrastarson, 2006; Jóhannesson et al., 2013).

Lacustrine sediment cores taken about 1 km beyond the present southern ice cap margin (Fig. 2), in connection with a separate study, indicate that the glacier extended at least to this position in the LIA. At present this lake receives glacial meltwater drained from an elongated perennial snow field, with exposed glacier ice, extending about 1 km south from the present southern margin, approximately to the lake. The core consists of about 30 cm of silt and clay on top of a diamict that is interpreted as till. We consider these circumstances and the absence of organic material in the core to indicate that the area around the southern perimeter of the glacier was ice covered for some period during LIA. At the southwestern margin, the glacier extended to the head of the Skjaldfannardalur valley in the year AD 1862 (Shepard, 1867). Negligible geomorphological imprints suggest a thin and not very dynamic glacier ice in those specific areas at elevations of 500–600 m asl, perhaps it infers that the ice was polythermal and frozen to its base at the margins there. Furthermore, many of the topographical depressions that could serve as depocentres are occupied by lakes. Relatively small marginal variations since LIA seem to be the general tendency on the plateaux, the main variations of the ice cap occur where the outlet glaciers drain into topographical lows (Fig. 2).

Three landforms considered to be rock glaciers occur below 500 m asl on steep north-facing slopes north of the glacier. They could indicate sporadic permafrost conditions and favourable conditions for rock glacier formation. As suggested from Tröllaskagi peninsula, north Iceland (by Whalley and Martin, 1994; Whalley et al., 1995) they might as well indicate changed climate affecting the input rate of weathered rock debris rather than being related to permafrost. Whalley (2009) used the term 'discrete debris accumulation' for such landforms related to debris accumulation of unknown origin.

However, it remains unknown if the rock glaciers are active at present. Hjort et al. (1985) interpreted entirely frozen ground below a depth of 15–30 cm, on 450–500 m high plateaux about 20 km north of the Drangajökull, as permafrost. Solid frozen ground was also observed at 15–30 cm soil depth on the 550–700 m high plateaux around the Kaldalón outlet during fieldwork in late July 2013. Recent studies from the coastal areas in north and east Iceland indicate sporadic permafrost above 800 m asl (Etzelmüller et al., 2007; Farbrot et al., 2007), but the elevation for permafrost could be significantly lower around Drangajökull ice cap, as a consequence of lower summer temperatures and shorter melting season. Patterned ground and 2–5 m wide polygons that occur on the plateaux around Drangajökull above 500 m asl could be products of permafrost. Polygons of that size have been considered as indicators of present or recent permafrost in the central highland of Iceland (Schunke and Priesnitz, 1983). Further studies are needed to conclude if the polygons and the rock glaciers are related to occurrences of permafrost, or frequent and numerous frost and thaw cycles.

6. Conclusions

- We present detailed geomorphological maps from the forefields of the Drangajökull surging outlet glaciers (Reykjarfjarðarjökull, Leirufjarðarjökull, and Kaldalónsjökull). Except for Kaldalón, those are the first detailed geomorphological maps published from the forfields of the surging outlets. An overview map presents the geomorphology around the entire Drangajökull ice cap.
- In general the sediment cover is thin around Drangajökull and mainly consists of basal till and locally weathered bedrock. Large bedrock outcrops are often distinguishable below the thin sediment sheet, intermittently glacially scoured bedrock protrudes the drift cover. The valleys are the main depocentres for sediments.
- The Little Ice Age limits of the ice cap are constrained by terminal moraines and confirm with historical evidence on glacier extent around the northern and eastern part of the ice cap. Glacial geomorphological imprints are mostly lacking along the southern and part of the western margin of the ice cap, which makes the LIA reconstruction challenging. Fragmentary historical information, distribution of erratics

and flutes, correlation to adjacent known LIA landforms, and lake sediment cores were used to estimate the LIA extent in such areas.

- The geomorphology of the surging outlets is characterised by extensive sandurs, covering the valley floors, and end moraines. The end moraines correlate well with reported surges since the Little Ice Age, and together they reveal the surge history over the last two centuries.
- Flutes, eskers, kame terraces, pitted sandur, and hummocky moraines were mapped in front of the surging outlets but are not frequently occurring landforms. The till is generally coarse grained and rich in boulders. The distalmost end moraines of the three surging outlets are 10–15 m in relief consisting mainly of gravel pushed up from the outwash plain and (to a lesser extent) of peat, fines, and diamict. Other moraines are usually low in relief and often indistinct or eroded by the glacial rivers, constituted of coarse-grained till and boulders.
- The landform distribution in the forefields of Reykjarfjarðarjökull, Leirufjarðarjökull, and Kaldalónsjökull does not agree well with the surging glacier land systems model of Evans and Rea (1999, 2003) and differs from descriptions of other surge-type glaciers in Iceland. This could be owing to the thin basal till, generally coarse-grained till matrix, subglacial hydrological conditions, or basal temperatures and intensive fluvial erosion on the valley bottoms.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.geomorph.2014.01.019.

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Appendix II

Research paper

The Holocene
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A 300-year surge history of the Drangajökull ice cap, northwest Iceland, and its maximum during the 'Little Ice Age'

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Abstract

Over the last 300 years, each of the three surge-type outlet glaciers of the Drangajökull ice cap in northwest Iceland has surged 2–4 times. There is valuable historical information available on the surge frequencies since the 'Little Ice Age' (LIA) maximum because of the proximity of the surging outlets, Reykjarfjarðarjökull, Leirufjarðarjökull and Kaldalónsjökull, to farms and pastures and monitoring of these glaciers since 1931. We have reconstructed the surge history of the Drangajökull ice cap, based on geomorphological mapping, sedimentological studies and review of historical records. Geomorphological mapping of the glacier forefields reveals twice as many end moraines as previously recognized. This indicates a higher surge interval than earlier perceived. A clear relationship between the surge interval and climate cannot be established. Surges were observed more frequently during the 19th century and the earliest 20th century compared with the relatively cool 18th century and the late 20th century, possibly reflecting a lack of information rather than a long quiescent phase of the glaciers. We have estimated the magnitude of the maximum surge events during the LIA by reconstruction of Digital Elevation Models (DEMs) that can be compared with modern DEMs. As reference points for the digital elevation modelling, we used the recently mapped lateral moraines and historical information on the exposure timing of *nunataks*. During the LIA maximum surge events, the outlet glaciers extended 3–4 km further down-valley than at present. Their ice volumes were at least 2–2.5 km^3 greater than in the beginning of the 21st century.

Keywords

digital elevation model, Drangajökull ice cap, glacier surge, Little Ice Age, moraine

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Introduction

The classical definition of surge-type glaciers emphasizes a cyclic flow instability in which glaciers undergo quasi-periodic fluctuations (Benn and Evans, 2010; Harrison and Post, 2003; Meier and Post, 1969; Þórarinsson, 1964). The active period of fast flow, the surge, lasts from a few months to several years, and the quiescent phase lasts from decades to several hundred years. A glacier surge is a dramatic event where ice velocity can be up to two or three orders of magnitude higher than the velocity of the quiescent phase (Kamb et al., 1985; Meier and Post, 1969; Sharp, 1988a, 1988b; Þórarinsson, 1964). Non-surging glaciers experience much more stable ice flow. The flow rate can be both slow, less than 1 m/day centrally in ice sheets, and fast, tens of metres on a day in ice-streams. In contrast to surging glaciers, the dynamics of non-surging glaciers is climatically controlled, thus they respond directly to changes in their mass balance. Because non-surging glaciers respond virtually directly to mass balance (climate), their end moraines are much better proxies for climatic fluctuations than end moraines at surging glaciers (Benn and Evans, 2010).

During a surge, ice is transported down to the glacier front from the reservoir area, sometimes resulting in a frontal advance of several kilometres (Harrison and Post, 2003; Meier and Post, 1969; Raymond, 1987). Those drastic changes of the glacier dynamics are believed to be mainly driven by thermal and hydrological changes at the glacier bed. Initially a surge is triggered by a complex combination of internal dynamic processes and external environmental factors (Fowler et al., 2001; Kristensen and Benn, 2012; Murray et al., 2003). Two main mechanisms have been proposed; either a threshold behaviour of underlying sediments appearing in water saturation which enhance the fast flow (Clarke et al., 1984; Kjær et al., 2006), or an abrupt shift from a subglacial tunnel drainage system to a linked-cavity system (Björnsson, 1998; Kamb, 1987; Kamb et al., 1985). Surges can be fast (velocity of several kilometres per year) or slow (velocity of a few tens of metres per year), and they occur in both temperate and polythermal glaciers. Polythermal glaciers typically experience an active phase lasting 3–10 years, while the active phase of temperate glaciers is often 1–2 years in Alaska and Iceland (Dowdeswell

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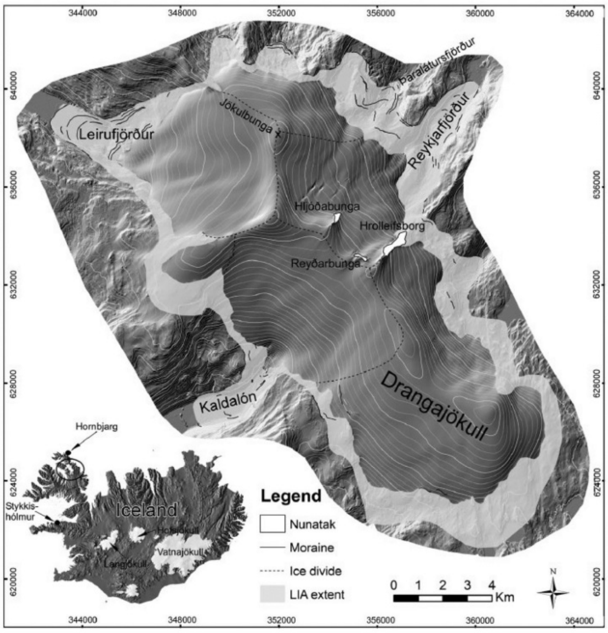

Figure 1. Overview map of the Drangajökull ice cap in northwest Iceland. The approximate ice divides of the three surging outlets are drawn in the figure. The 'LIA extent' presents the glacier maximum extent over the LIA period and its reconstruction is based on geomorphological mapping and aerial photograph analysis.

Annual marginal positions of the three surging outlets have been measured since 1931 (Eyþórsson, 1963; Sigurðsson, 1998, 2011). Their two last surges are therefore temporally and spatially well documented.

Data and methods

Fieldwork, aerial photographs and DEM

Fieldwork was carried out during the summers of 2011, 2012 and 2013 with focus on Reykjarfjarðarjökull, Leirufjarðarjökull and Kaldalónsjökull (Figure 1). Landforms and sediments in front of the glaciers were mapped, described and interpreted (Brynjólfsson et al., 2014). Samples of peat and tephra for dating of glacial landforms and sediments were also collected from a moraine section in Leirufjörður. The ^{14}C ages of the peat samples were calibrated with IntCal09 according to Reimer et al. (2009); the ^{14}C samples were processed at the Ångström Laboratory, Uppsala University, Sweden. The geochemistry of the tephra layer was analysed at the University of Copenhagen, Denmark, using a JEOL Super probe JSL 8200 with an acceleration voltage of 15 kV, a 10-nA beam current and a beam diameter of 7 µm. In addition to natural and synthetic minerals, glass standards (K22_ATHO and K15_KL2) were used as standards. The geomorphological overview of Brynjólfsson et al. (2014; Figures 2, 3 and 5) is based on aerial photographs with 0.5-m pixel size dating from 2005 and 2006 supplied by Loftmyndir ehf, and a LiDAR (Light Detection and Ranging) derived DEM with 5-m ground resolution from 2011 supplied by the Icelandic Meteorological Office (IMO).

Historical records of glacier fluctuations

Some of the fluctuations of Drangajökull in the last four to five centuries are known from written sources (Eyþórsson, 1935; Jóhannesson and Jónsson, 2012; Magnússon and Vídalín, 1710; Ólafsson and Pálsson, 1772; Thoroddsen, 1933, 1958). Farmers that lived in the vicinity of the glaciers were conscious about their fluctuations, and information was recorded in local annals and church books. Two particularly valuable surveys on glacier fluctuations of Drangajökull in the last three centuries exist. Thoroddsen (1933, 1958) and Jóhannesson and Jónsson (2012) recorded information on glacier fluctuations given by local residents during Thoroddsen's exploration in 1886–1887. Eyþórsson (1935) explored fluctuations of Drangajökull and installed markers for measurements of glacier margin positions on some of the outlets in the year 1931. Furthermore, some ambiguous information on glacier fluctuations extending back to the late Middle Ages was summarized by Þórarinsson (1943). Since Eyþórsson's (1935) initiative, the Icelandic Glaciological Society (IGS) has measured frontal fluctuations of glaciers in Iceland since the 1930s, published in semi-annual reports in the journal *Jökull* (e.g. Sigurðsson, 2000, 2011).

Reconstruction of the surge history and interval

The reconstruction of the surge history of the Drangajökull ice cap is based mainly on the recently produced geomorphological maps (Brynjólfsson et al., 2014) and a review of historical data. End moraines and lateral moraines were of special interest. They were surveyed on the aerial photographs and localized in the field with GPS (Garmin GPSMAP 62sc) with a horizontal accuracy of ±3 m. The distances between each end moraine were measured on the aerial orthophotographs using the Geographical Information System (GIS), Esri ArcMap 10.

Some moraines are of known age and relate to the recorded surges (Björnsson et al., 2003; Eyþórsson, 1935; Þórarinsson, 1969; Thoroddsen, 1933, 1958), while the exact ages of other moraines are not known, but they are assumed to have formed in colder periods (Eyþórsson, 1935). Organic material or tephra for absolute dating of the moraines is very limited, but we were able to constrain the age of two moraines by absolute dating.

Furthermore, we constrained a relative age of the undated additional moraines that were mapped by Brynjólfsson et al. (2014). This was done by using the historical data and correlating moraines of unknown age with moraines of known age and a mean retreat rate in certain periods.

Reconstruction of the maximum surge conditions during LIA

The lateral moraines of the Drangajökull surging outlet glaciers formed along the valley slopes during the LIA maximum extent provide a reference height for the ice surface and maximum ice thickness at that time. Lateral moraines are characteristic along the margins of outlet glaciers confined by valleys and considered to demonstrate the ice thickness during advance or steady state (Benn and Evans, 2010; Hannesdóttir et al., 2015). In case of absence on one of the sides, the elevation of a lateral moraine on the other side of the valley was used for reconstruction of the absent moraine. The area of each of the three surging outlets was also estimated, based on their approximate ice divide location, identified on shaded relief LiDAR derived DEM, and the LIA maximum moraines.

The *nunatak*, Reyðarbunga, located about 800 m a.s.l. in the accumulation area of Reykjarfjarðarjökull proximal to the ice divide (Figure 1), provides valuable information on minimum ice thinning in the reservoir areas of the surging outlet glaciers, since its exposure in the first decade of the 20th century (Eyþórsson, 1963). When producing the LIA maximum DEMs, we used the approximate 50-m ice thinning around Reyðarbunga since ~1910 as fixed thinning for the reservoir areas since LIA maximum. The outlet tongues are better constrained using the lateral moraines extending adjacent to their margins as reference height.

The DEMs representing the LIA maximum ice conditions were produced in GIS. First, a grid of points with given coordinates and elevation was produced, with emphasis on the lateral moraines. The ice surface topography was assumed flat between the lateral moraines of the outlet tongues. The DEMs were processed using an ArcGIS 3D extension, where the point grids were interpolated using *natural neighbour* as interpolation method. Finally, the DEMs were compared with the LiDAR derived DEM enabling estimations of ice thickness and volume changes of the surging outlets between the LIA maximum extent and 2011.

Historical record of glacier variations and surges

End moraines formed since the LIA maximum were recently mapped by Principato (2008) and Brynjólfsson et al. (2014). Four moraines occur in Reykjarfjörður, six in Leirufjörður, and six in Kaldalón. The number of end moraines is not in agreement with the number of recorded surges of the outlet glaciers as read from historical records. They document two surges in Reykjarfjörður, four in Leirufjörður and four in Kaldalón (Björnsson et al., 2003; Sigurðsson, 1998; Þórarinsson, 1969). By mapping the geomorphology, more advances have been identified compared with what is known from the surge record during the last c. 300 years.

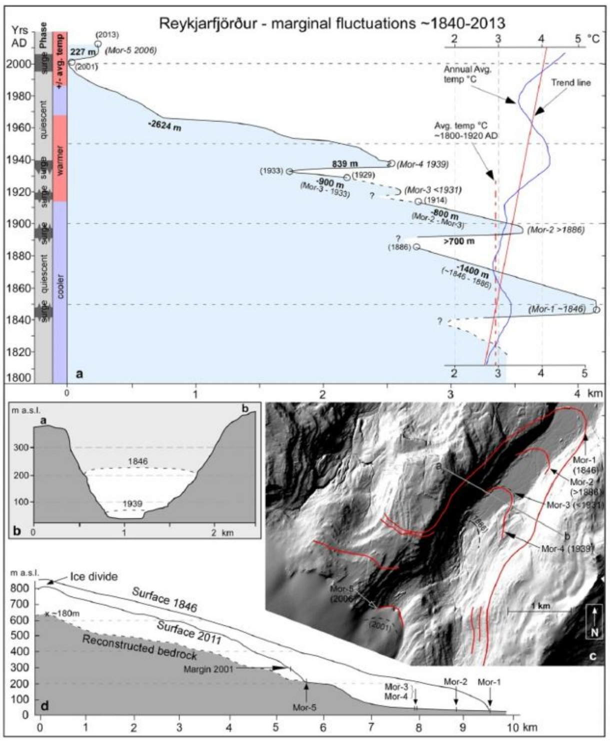

Figure 2. Fluctuations of Reykjarfjarðarjökull since the LIA maximum extent. (a) Open circles mark marginal position at known times; recorded marginal advances and retreats are indicated with bold numbers. A solid line in the diagram indicates known pattern of the glacier, while dashed line indicates unknown. The temperature plot is the longest measured record in Iceland, from Stykkishólmur in west Iceland. Cooler and warmer periods are shown on the bar to left. (b) Cross section indicating the different ice conditions during LIA maximum extent and approximately the mid-20th century. (c) LiDAR derived shaded relief surface model; solid lines show moraine positions in the glacier forefield, and dashed lines ice marginal positions or described moraines not observable at present. (d) Longitudinal section of the present glacier surface, reconstructed LIA maximum surface, moraine positions, and a reconstructed bedrock based on few ice-radar point measurements (Magnússon et al., 2004).

Reykjarfjarðarjökull

The Reykjarfjarðarjökull outlet advanced abruptly in 1837 (Jóhannesson and Jónsson, 2012; Thoroddsen, 1933) and reached its LIA maximum position about AD 1846 (Eyþórsson, 1935; Þórarinsson, 1943; Thoroddsen, 1933). Based on Thoroddsen's (1933) descriptions of this sudden advance and of the subsequent geomorphological characteristics, this event is concluded to have been a surge (Björnsson et al., 2003; Þórarinsson, 1969). The glacier front was described as steep and high, pushing and overriding older moraines and fluvial material, and finally forming a distinct end moraine, which in this study is referred to as Moraine 1 (Figure 2; Thoroddsen, 1933). An associated lateral moraine is partly preserved along the valley slopes (Brynjólfsson et al., 2014; Eyþórsson, 1935).

Initially after the surge termination, the glacier melt was characterized by surficial thinning and no distinct frontal retreat. A total of 8–9 years later, the glacier had thinned considerably and separated by 10–20 m from Moraine 1. The glacier retreated quickly about 1400 m in the period until 1886 (Thoroddsen, 1933). At that time, the valley floor was covered by sandur deposits, and there were deep ponds located between gravelly hillocks indicating dead-ice melting (Jóhannesson and Jónsson, 2012; Thoroddsen, 1933).

Moraine 2 is situated about 800 m proximal to Moraine 1, formed after at least 600-m advance of the ice front (Figure 2). An exact timing of its formation is unknown. Eyþórsson (1935) assumed it formed about 1860–1870 as a response of the glacier to colder climate in that period. As Thoroddsen only recorded Moraine 1 in the year 1886 and according to known surge activity of the glacier, we assume that Moraine 2 was formed by surge rather than as direct response to a period of cooler climate. The third moraine, located about 800 m inside Moraine 2, was formed by a surge in 1934–1939 (Björnsson et al., 2003; Þórarinsson, 1969). Notably, Eyþórsson (1935) described a moraine at a similar location during his exploration in 1931 and measured the glacier margin being located about 870 m up-valley from that moraine. Therefore, we assume the surge in 1934–1939 was at least the fourth advance since the glacier reached its LIA maximum extent in 1846; the associated moraine is hereafter referred to as Moraine 4, and the moraine that Eyþórsson (1935) described referred to as Moraine 3 (Figure 2). The annual glacier-front measurements indicate an advance of 830 m during the surge in 1934–1939 (Eyþórsson, 1963).

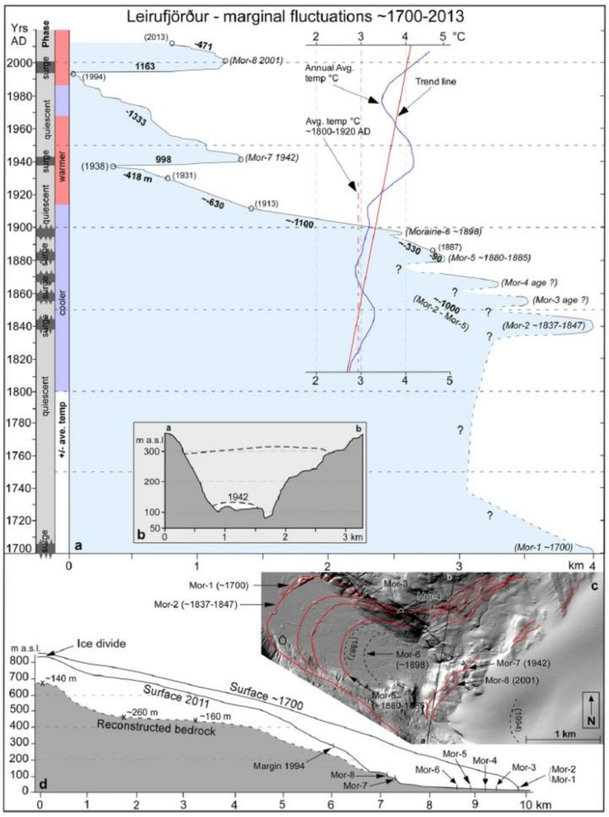

Figure 3. Fluctuations of Leirufjarðarjökull since the LIA maximum extent. (a) Open circles mark marginal position at known time; recorded marginal advances and retreats are indicated with bold numbers. A solid line in the diagram indicates known pattern of the glacier, while dashed line indicates unknown. The temperature plot is the longest measured record in Iceland, from Stykkishólmur in west Iceland. Cooler and warmer periods are shown on the bar to left. (b) Cross section indicating the different ice conditions during the LIA maximum extent and approximately the mid-20th century. (c) LiDAR derived shaded relief surface model; solid lines show moraine positions in the glacier forefield, and dashed lines ice marginal positions or described moraines not observable at present. An approximate position of the farm Öldugil is indicated by Ö. (d) Longitudinal section of the present glacier surface, reconstructed LIA maximum surface, moraine positions, and a reconstructed bedrock based on few ice-radar point measurements (Magnússon et al., 2004).

During the quiescent phase 1940–2002, the glacier retreated about 2300m and in the course of retreat, the location of the margin changed from about 30 to about 250 m a.s.l. The glacier advanced about 227m during the most recent surge in 2002–2006 and has only retreated about 15m since then. Because of snow cover at the glacier margin, no moraine was visible in front of the glacier in late July 2011. Sigurðsson (2011) mentions that the glacier margin was about to become separated from an indistinct gravelly moraine in the autumn of 2010. Reykjarfjarðarjökull has surged at least five times and retreated about 4100m from its LIA maximum position in 1846 until 2011.

Leirufjarðarjökull

Leirufjörður is an about 6-km-long valley orientated NW-SE. The flat valley floor is mostly covered with fluvial sediments. The 600m closest to the present glacier margin are gently inclined slopes reaching elevations of about 200 m a.s.l. at the glacier margin, characterized by ice sculpted bedrock protruding through a coarse-grained subglacial till. The farm Öldugil (Figure 3c) was abandoned in the 15th–16th century because of deteriorating farmland related to expansion of the glacial rivers and the advancing glacier (Magnússon and Vídalín, 1710). No landforms can be correlated with the historical records earlier than approximately AD 1700.

In the late 17th century or at about AD 1700, the farm Öldugil was overridden during a rapid advance of the glacier (Magnússon and Vídalín, 1710; Thoroddsen, 1933, 1958). The associated glacier marginal position, Moraine 1 (Figure 3), is represented by one moraine segment about 3100m distal to the glacier-front position in 2011. Moraine 1 is not dated, but by correlating the historical information and the geomorphological imprints, we consider it to have formed in the glacier surge about AD 1700.

Moraine 2 formed approximately AD 1840, just proximal to Moraine 1 (Eyþórsson, 1935; Thoroddsen, 1933, 1958). The local farmers informed Thoroddsen (1933) that the glacier margin was at Moraine 2 approximately AD 1837–1847. This correlates with tephra and organic material that were dated from glaciotectonized sediments within Moraine 2, indicating formation no earlier than AD 1693 and at latest about AD ~1840 (Figure 4). The geochemical analyses show that the tephra layer originates from the Hekla volcano (Gudmundsdóttir et al., 2011; Larsen and Eiríksson, 2008; Larsen et al., 1999; Figure 4 and Appendix 1). Based on chemistry, the ^{14}C dates and distribution of known tephra layers from Hekla, this tephra layer is correlated to the Hekla eruption in AD 1693 (Dugmore et al., 2007; Larsen, unpublished data; Þórarinsson, 1968). The ^{14}C samples were dated to: Ua-46081 199 ± 30 cal. BP, Ua-46082 111 ± 30 cal. BP and Ua-46083 1211 ± 30 cal. BP.

We consider sample Ua-46083 an outlier. It could possibly be due to old, organic material that has been displaced and reworked by the glaciotectonic deformation.

Heavily deformed sediments within Moraine 2 are exposed in a 10-m-wide and a 3-m-high river cut section. The unit FPT (Fine-Peat-Tephra) is deformed fine sediments, mostly silt to sand, with folded layers of tephra and peat (Figure 4). The units Fine 1 and Fine 2 both consist mainly of silt and sand with thin lenses of clay and fine gravel; the main difference is more deformation and no preservation of sedimentary structures in Fine 1. Pockets of massive gravel are found in the upper parts of the section and a massive diamict, interpreted as till, occurs on the proximal side of the moraine.

The glaciotectonically deformed sediments resemble descriptions of end moraines formed during surges of large Vatnajökull outlet glaciers (Benediktsson et al., 2008, 2009, 2010) which suggest that this moraine was formed during a surge. Furthermore, all the observed advances of Leirufjarðarjökull were abrupt (Eyþórsson, 1935; Thoroddsen, 1933, 1958) and subsequently interpreted as surges (Björnsson et al., 2003). The appearance of Moraine 2 corresponds well to Thoroddsen's (1933, 1958) descriptions, and an associated lateral moraine is partly preserved along the valley slope (Brynjólfsson et al., 2014; Eyþórsson, 1935).

Thoroddsen (1933, 1958) described four distinct horseshoe shaped moraines, here referred to as Moraine 2–Moraine 5, located within about 1000-m distance from the glacier in 1887. The glacier snout was steep and heavily crevassed and icebergs had been falling off the margin few years earlier. In 1887, the glacier had just separated by a few tens of metres from Moraine 5. Judging from those descriptions, we conclude that the glacier surged a few years before, probably around AD 1880–1885. Thoroddsen (1933, 1958) did not describe the location or age of Moraines 3 and 4, but their locations have been mapped and described recently (Brynjólfsson et al., 2014; Principato, 2008). The exact ages are unknown, but their relative ages are confined by Moraine 2 and Moraine 5. Eyþórsson (1935) described a moraine about 330m proximal to Moraine 5. His observation was supported by his local guide who confirmed the glacier position at the moraine approximately in the year 1898. This moraine, Moraine 6, is not recognizable at present but its location can be approximated from Eyþórsson's (1935) data. The location of Moraine 6 is marked as an ice marginal position in Figure 3. This suggests an average surge interval of 10 years in the period ~1840 to ~1898. Between 1898 and 1938, the glacier retreated about 1800m.

The glacier surged approximately 998m in the years 1939–1942, and formed Moraine 7 about 1200m up-valley from Moraine 6. During the next 53 years, the glacier retreated about 1300m. A new surge took place in the years 1995–2001, where the glacier advanced about 1150m and formed the most recent moraine, Moraine 8. In 2012, the margin had retreated about 430m from Moraine 8, experienced at least seven surges and retreated about 3100m since it reached a maximum extent during LIA, about AD 1700.

Kaldalónsjökull

Kaldalón, orientated approximately NE-SW, is an about 8-km-long glacially carved valley. Fluvial sediments dominate on the flat valley floor which is confined by steep 300- to 500-m-high mountain slopes. Advances and retreats of the outlet glacier Kaldalónsjökull in earlier centuries are recorded in historical documents and considered to have caused abandonment of at least two farms before AD 1700 (Magnússon and Vídalín, 1710; Thoroddsen, 1933, 1958). These fluctuations are most likely reflected in a pit section on the distal side of Moraine 1 (Figure 5). There, 15- to 35-cm-thick layers of fine, massive gravel and sand are interbedded with a sequence of peat and soil (Figure 6). The clast supported sand and gravel layers are interpreted as glaciofluvial sediments. The soil layers are massive and consist of organic material and small amounts of minerogenic sediments of grain size from clay to fine sand.

A natural meadow, of a size which 'took 12 men to harvest over a summer', was abruptly overrun by an advancing glacier about 20 years before an official survey of the area was conducted in 1754 (Eyþórsson, 1935; Ólafsson and Pálsson, 1772; Þórarinsson, 1943; Thoroddsen, 1933, 1958). A corresponding distinct moraine occurs about 3400m distal to the marginal position in 2011, referred to as Moraine 1 in this study. It has been considered as the first historically confirmed surge of Kaldalónsjökull, occurring approximately AD ~1740 (Björnsson et al., 2003; Sigurðsson, 2005; Þórarinsson, 1969).

Thoroddsen (1933, 1958) described three arc-shaped moraines in front of the glacier, observed during his travels in 1887. Similar to Leirufjörður, he described a steep, 120- to 150-m high and heavily crevassed glacier snout. Distal to Moraine 1, he described an old vegetated moraine, dated to Younger Dryas age by Principato et al. (2006). A poorly conserved lateral moraine occurs on the

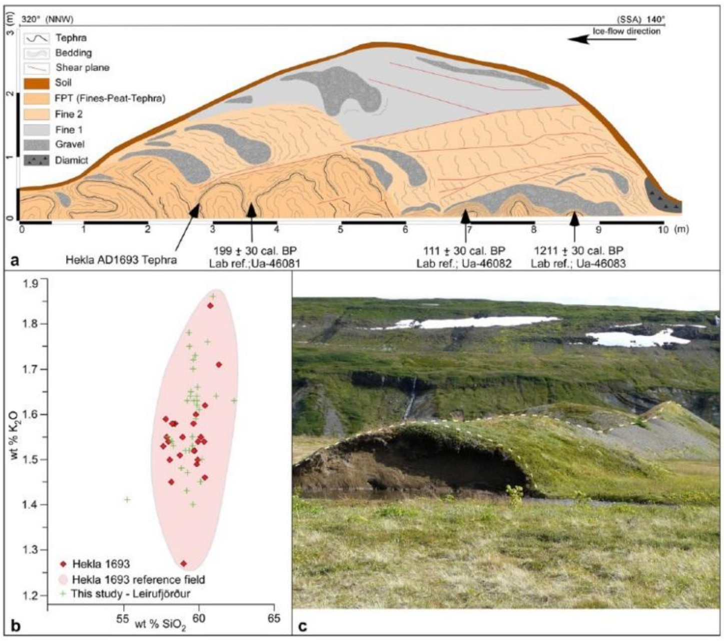

Figure 4. Glaciotectonically deformed sediments in a section in Moraine 2 in Leirufjörður. (a) Diagram of the section. The ^{14}C ages were calibrated according to Reimer et al. (2009). (b) Geochemistry of the tephra in Moraine 2 compared with analyses of Hekla AD 1693 tephra (Dugmore et al., 2007; Larsen, unpublished data; Þórarinsson, 1968). (c) Overview of Moraine 2; white dashed lines define the moraine crest and the cleaned section wall in Figure 4a.

southern valley slope, between Moraine 1 and Moraine 3. However, its terminal position is impossible to locate on the valley floor, and is therefore only marked with a dashed line in Figure 5. Moraine 3 is located about 1200 m up-valley from Moraine 1. At present, it is recognizable as a 1- to 4-m high, discontinuous, gravel ridge on the valley floor near the southern valley slope, and as a lateral moraine along the southern valley slope. The age of Moraines 2 and 3 is unknown, but a relative age is provided by Moraine 1 and Moraine 4, that is, between AD ~1740 and 1860.

Moraines 4, 5 and 6, described below, are not recognizable anymore, because of a 1998 outburst flood in the glacial river Mórilla, which drains Kaldalónsjökull (Sigurðsson, 2000). The river transported icebergs of many metres in diameter 2–3 km down-valley and deposited an 8- to 10-m thick sheet of coarse-grained gravel on the valley floor distal to the glacier margin. A thinner sheet of gravel was deposited to about 1 km down-valley.

Thoroddsen (1958) described a moraine, referred to as Moraine 4 in this study, about 700 m up-valley from Moraine 3 (Figure 5). At that time, the glacier margin was located 400–500 m up-valley from Moraine 4. Local residents informed Thoroddsen that the glacier was positioned at Moraine 4 about 20–30 years before his visit, that is, approximately in the year 1860 (Eyþórsson, 1935; Thoroddsen, 1933, 1958). Eyþórsson (1935) described this moraine as a distinct ridge on the southern side of the valley floor in 1931. It is not recognizable at present, but its location can be approximated from correlating the Iceland Glaciological Society (IGS) ice marginal measurements and measurements from Thoroddsen (1933, 1958) and Eyþórsson (1935).

Moraine 5 is located about 750 m up-valley from Moraine 4, but is not recognizable at present. Eyþórsson (1935) described it as conical mounds of boulders about 5 m in height in 1931, marking a still stand of the glacier margin because of relatively cold summers in 1920–1925. The margin had retreated about 100 m from Moraine 5 in 1931. Because of this and the fact that it generally takes at least few years before the margin and a moraine separate at Drangajökull, we consider that it was formed during a surge 10–15 years before Eyþórsson's survey in 1931.

The glacier surged 191 m in the years 1936–1940 (Eyþórsson, 1963). Only the associated lateral moraines are recognizable. However, the terminal position, referred to as ice marginal position (Moraine 6), was easily established about 120 m up-valley from Moraine 5 by using the IGS glacier-front measurements. The glacier retreated about 1500 m in the following 55 years until 1995.

Moraine 7 was formed during the surge in 1995–1999 when the glacier advanced about 1015 m. A total of 4 years after the surge termination, in 2003, the margin started to retreat and has retreated about 290 m until 2013.

Two additional lateral moraines have recently been mapped between Moraine 6 and Moraine 7 (Brynjólfsson et al., 2014). No corresponding terminal moraines exist for exact marginal delimitation. The IGS data report minor advances or still stands in 1948–1951 and 1967–1973; the snout was described relatively flat and smooth and not surging at those times. The lateral moraines might have formed during those periods. Since Kaldalónsjökull reached a maximum extent during LIA, it has retreated about 3400 m and experienced at least six surges.

Surge interval and the maximum conditions during LIA

Surge interval

The surge intervals are summarized in Table 1. It appears that the surges are non-synchronous between the three outlets Reykjarfjarðarjökull, Leirufjarðarjökull and Kaldalónsjökull. The surge interval varies from 10 to 140 years.

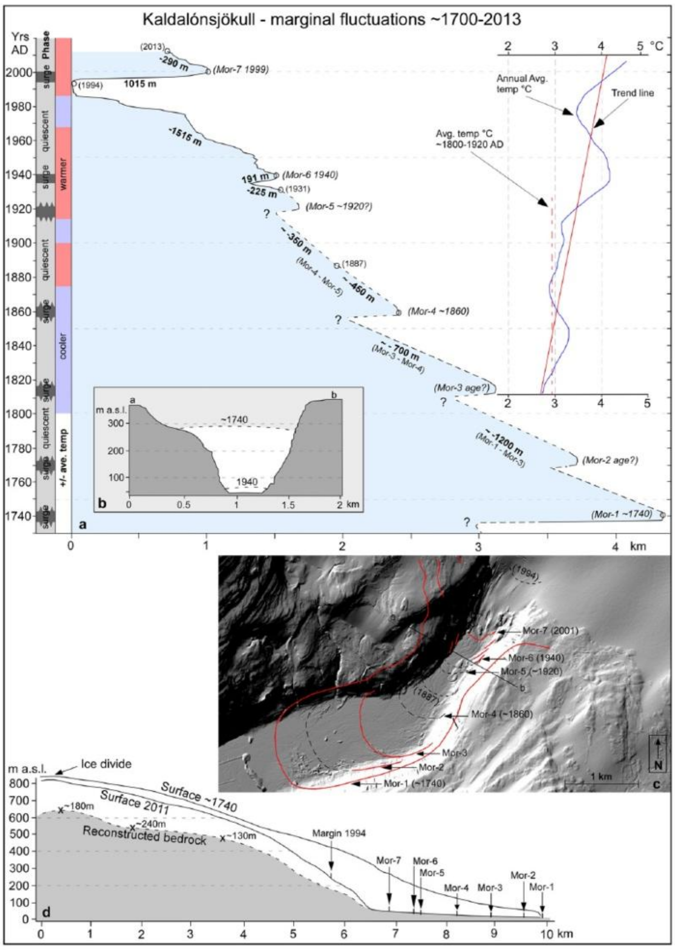

Figure 5. Fluctuations of Kaldalónsjökull since the LIA maximum extent. (a) Open circles mark marginal position at known time; recorded marginal advances and retreats are indicated with bold numbers. A solid line in the diagram indicates known pattern of the glacier, while dashed line indicates unknown. The temperature plot is the longest measured record in Iceland, from Stykkishólmur in west Iceland. Cooler and warmer periods are shown on the bar to left. (b) Cross section indicating the different ice conditions during LIA maximum extent and approximately the mid-20th century. (c) LiDAR derived shaded relief surface model; solid lines show moraine positions in the glacier forefield, and dashed lines show ice marginal positions or earlier described moraines not observable at present. (d) Longitudinal section of the present glacier surface, reconstructed LIA maximum surface, moraine positions, and a reconstructed bedrock based on few ice-radar point measurements (Magnússon et al., 2004).

The 140-year long quiescent phase in Leirufjörður, from AD 1700 to 1840 is in contrast to the shorter surge interval in other periods for all the three glaciers. No information exists from this period between the formation of Moraine 1 and Moraine 2. All older geomorphological features appear to have been removed by the ~1840 advance.

The quiescent phase seems to be generally shorter in the 19th century and the beginning of the 20th century. This could relate to lower average temperatures (Figures 2, 3 and 5) in the period and thus potentially a shorter recovery time of the reservoir areas to build up for a new surge, either related to less melting during the ablation season or more accumulation during winters, or a combination thereof.

The quiescent phase is about 50–60 years between the most recent surges. This is longer than before the 1940s surges in Reykjarfjörður and Leirufjörður. The longer quiescent phase coincides with increased mean annual air temperature by 1°C at the Stykkishólmur weather station after ~1920 compared with the period ~1850 to ~1920.

Maximum extent during LIA

We have mapped the maximum area of the Drangajökull ice cap to about 216 km² during LIA, compared with about 142 km² at present (Jóhannesson et al., 2013). The whole area for each of the three surging outlets was also mapped and compared (Table 2).

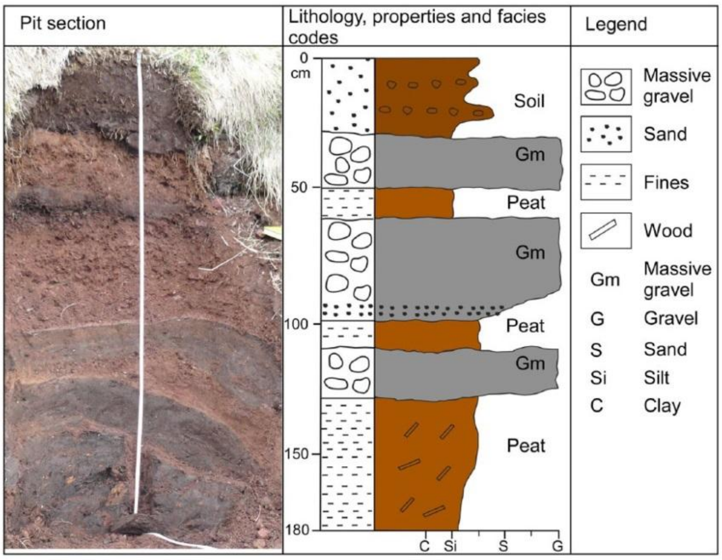

Figure 6. Section in a peat bog below Moraine 1 in Kaldalón.

Lateral moraines occur along the valley slopes of the three glaciers. They are located up to 240 m above the valley floor, revealing the minimum ice thickness at the time of formation. Beyond presently glaciated areas, the ice has been thickest just where the valley floors start to rise in elevation, generally half way between the present margins and the moraines formed during maximum extent in LIA. The ice has thinned towards the LIA maximum end moraines (Figure 7). Based on the lateral moraines, we produced a DEM representing maximum glacier conditions for each outlet glacier during LIA. Reconstructing any topographical details of the maximum ice surface is difficult, as it is only based on the lateral moraines and the present surface topography. Therefore, we assumed a flat ice surface between the lateral moraines for the DEMs. This gives an approximation of the ice thickness during the LIA and, hence, approximate volume changes since the maximum conditions during LIA.

The only reference we have on surface elevation change for the presently glaciated areas, prior to the first aerial photos in 1945, is the exposure time of the *nunatak* Reyðarbunga (Figure 1), in the first decades of the 20th century. The *nunatak* is said to have protruded 1–2 m through the ice surface until at least 1931, and subsequently towered about 100 m above the ice surface approximately in 1960 (Eyþórsson, 1963). This great change in the period 1931–1960 coincides with the surge of Reykjarfjarðarjökull 1936–1940 and a relatively warm period (Figure 2). We conclude that this dramatic deglaciation of the *nunatak* is mainly a result of the surge in 1936–1940, when ice was transported from the reservoir area down to the ablation area. At present, Reyðarbunga reaches at maximum 50–65 m above the glacier surface at its down-glacier side. The up-glacier side is covered by snow and ice of similar height as the summit of the *nunatak*. Thus, the surface elevation has increased by 35–50 m in this place since 1960. According to this, we used ~50 m as a minimum ice thinning since the LIA maximum of the reservoir areas in the presently glaciated areas of the surging outlets.

Comparison of the DEMs, representing maximum conditions during LIA, with the LiDAR derived DEMs from 2011 shows a significant ice volume loss. Reykjarfjarðarjökull has lost about 2.5 km^3 between 1846 and 2011, Leirufjarðarjökull has lost about 2.2 km^3 and Kaldalónsjökull about 2.1 km^3 (Table 2). Both the decrease in area and thickness of the surging outlets are considered strongly related to the general warming trend since the glaciers experienced their maximum extent during LIA.

Discussion

Surge activity prior to the maximum extent during LIA is challenging to reconstruct. No geomorphological imprints or precise descriptions in the historical information can confirm surges prior to the LIA. However, the historical data provide information on advances and retreats of the outlets prior to the LIA maximum extent. In general, glaciers in Iceland are considered to have been growing during the 17th and 18th centuries (Magnússon and Vídalín, 1710; Sigurðsson, 2005; Þórarinsson, 1943). Around Drangajökull, farmlands were deteriorating in the forefields of the three surging outlets, Reykjarfjarðarjökull, Leirufjarðarjökull and Kaldalónsjökull, which were destroying pasture areas. Farms were abandoned in all of the valleys because of the approaching glaciers and the braided glacial rivers which were constantly changing courses (Magnússon and Vídalín, 1710; Ólafsson and Pálsson, 1772; Þórarinsson, 1943). According to the directly observed surge history of Drangajökull since the maximum extent in LIA, we consider it most likely that the potential advances which the three surging outlets experienced, during late Holocene, prior to the LIA maximum extents were also surges. Proglacial lake sediment records from Eyjabakkajökull indicate that this Vatnajökull outlet glacier has been experiencing surges prior to the LIA, back to c. 2000 yr BP (Striberger et al., 2011). Hence, surge behaviour of Icelandic glaciers is not necessarily confined to the LIA.

Principato (2008) concluded that the Drangajökull outlets reached their Holocene maximum size in Neoglacial time. This is most likely true for Reykjarfjörður and Kaldalón, but we have not been able to confirm this in Leirufjörður. Principato et al. (2006) used ^{36}Cl cosmogenic exposure dating to date the furthest end moraine in Leirufjörður, Moraine 2, which marks the glacier maximum extent during the LIA. Only one rock sample provided a

Table 1. Ages of moraines and ice marginal positions from the three surging outlets. The surge interval, the advanced distance during surges and the retreat between surges are shown in metres. The ages of Moraines 2 and 3 in Reykjarfjörður, Moraines 3 and 4 in Leirufjörður and Moraines 2 and 3 in Kaldalón are estimated. However, their relative age and interval are constrained by prior and subsequent moraines.

Glacier	Moraine	Year (AD)	Interval (yr)	Advance (m)	Retreat (m)
Reykjarfjarðarjökull	Mor-1	1846		?	>1400
	Mor-2	~1890	~45	~750	>800
	Mor-3	~1910	~20	~800	~900
	Mor-4	1934–1939	~25	839	2624
	Mor-5	2001–2006	62	221	19
Leirufjarðarjökull	Mor-1	~1700		Substantial	?
	Mor-2	~1837–1847	~140	?	>300
	Mor-3	~1860	~15		>250
	Mor-4	~1870	~10		>450
	Mor-5	~1880–1885	~15		>330
	Mor-6	~1898	~13		~1800
	Mor-7	1938–1942	53	998	1333
	Mor-8	1996–2001	54	1163	471
Kaldalónsjökull	Mor-1	~1740		Substantial	>400
	Mor-2	1780??	40?		>600
	Mor-3	1820??	40?		>700
	Mor-4	1860	~60?		>800
	Mor-5	~1920	~60		~330
	Mor-6	1936–1940	~16	191	1515
	Mor-7	1996–2001	56	1015	290

Table 2. The measured area (km^2) of each surging outlet, and the whole ice cap, in 2011 and during the LIA maximum extent. Proportional areal changes, and estimated volume changes, between the LIA maximum extent and 2011 are shown. The decrease of the whole ice cap is most likely a bit overestimated; thus the three surging outlets did not reach their maximum extent at absolutely the same time.

Glacier	Extent (km^2) in 2011	Maximum extent (km^2) during LIA	Change (%)	Change (km^3)
Reykjarfjörður outlet	22	33	−34	−2.5
Leirufjörður outlet	28	40	−30	−2.2
Kaldalón outlet	33	42	−22	−2.1
Drangajökull	142	216	−34	

measureable amount of ^{36}Cl, dating to about 4500 yr BP. Our results do not agree with this conclusion. We dated tephra and organic material from within the moraine (Figure 4), indicating formation of Moraine 2 no earlier than AD 1693 and no later than AD 1850. Thus, the Holocene maximum extent seems to have been reached twice, and is represented by Moraine 1 and Moraine 2 in Leirufjörður, formed AD ~1700 and ~1840.

However, in Reykjarfjörður we have mapped two diffuse terminal moraines 200–300 m distal to Moraine 1, the LIA maximum position, indicating a more advanced Holocene position of Reykjarfjarðarjökull as suggested by Principato (2008; Brynjólfsson et al., 2014).

Furthermore, two moraines distal to Moraine 1 (LIA maximum extent) in Kaldalón confirm maximum glacier extent during the Holocene prior to the LIA maximum extent (Brynjólfsson, 2014; Eyþórsson, 1935; Principato, 2008; Thoroddsen, 1933). The furthest is dated to Younger Dryas age, by ^{36}Cl cosmogenic exposure dating. The other moraine has a minimum age of 2600 yr BP, indicated by three radiocarbon datings from peat on top of the moraine (Principato, 2008). We explored a 2- to 3-m-high river cut section in a peat bog, just in front of and below the second moraine (Figure 6). The peat and soil are interbedded with sand and gravel layers which we interpret as deposited by a glacial river braiding on the valley floor.

Surge history

Reykjarfjörður surge history. Thoroddsen (1933) described deep ponds sitting between gravelly hills and hillocks interrupting the 1400-m-long, flat sediment plain between Moraine 1 and the glacier margin in the year 1886. His description of this hummocky surface indicates downwasting of stagnant ice after the surge termination approximately AD 1846. Such dead-ice environment is characteristic for surging glaciers during the quiescent phase, in line with the landsystems model of Evans and Rea (1999, 2003). We only observed a small patch of hummocky moraine close to the present margin of Reykjarfjarðarjökull (Brynjólfsson et al., 2014). Apparently, the last surge in 2002–2006 did not produce enough debris-rich dead-ice for the formation of any major hummocky moraine in the area. We concur with Þórarinsson (1969) and Björnsson et al. (2003) that Moraine 1 was formed by a surge and consider that Thoroddsen's (1933) description of the hummocky terrain supports that conclusion.

The exact ages of Moraine 2 and Moraine 3 remain unknown (Figure 2). As Thoroddsen (1933) only observed Moraine 1 in the year 1886 and Eyþórsson (1935) described Moraine 1, Moraine 2 and Moraine 3 in the year 1931, the relative ages of Moraine 2 and Moraine 3 are confined by Moraine 1 and Eyþórsson's exploration in 1931. Exactly when they formed, in this period between 1886 and 1931, remains unknown. Thoroddsen (1933) described gravelly hills, heap of rocks and ponds in between, and we do not exclude the possibility that he did not observe all moraines. However, Eyþórsson (1935) considered Moraine 2 to be formed prior to Thoroddsen's exploration, during a cold period in 1860–1870. We consider this unlikely because all the observed advances of the outlets were surges, and furthermore, we can confirm that Thoroddsen's descriptions of the moraines are reliable. Moraine 3 must have formed after 1914. The Ordinance Survey map from 1914 located the glacier

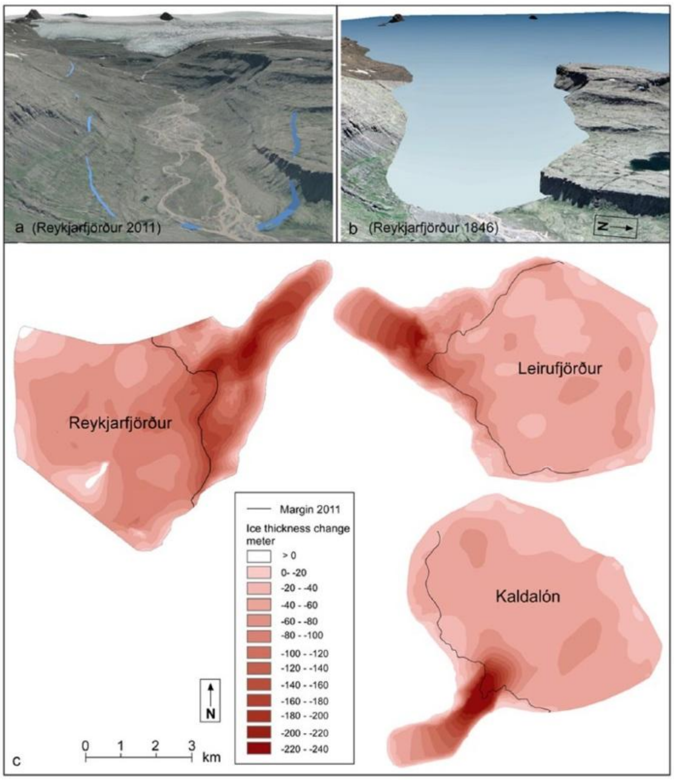

Figure 7. The maximum surge conditions during LIA compared with the present conditions on the 2011 LiDAR derived DEM, (a) the 1846 lateral moraine shown on aerial photos from 2006 draped over the LiDAR derived DEM, (b) reconstruction of Reykjarfjarðarjökull in 1846, and (c) approximate ice thickness change of Reykjarfjarðarjökull from 1846 to 2011, approximate ice thickness change of Leirufjarðarjökull from 1700/1847 to 2011, approximate ice thickness change of Kaldalónsjökull ~1740 until 2011 and scale for the ice thickness change.

margin about 100 m further down-valley than Moraine 3. Eyþórsson (1935) considered Moraine 3 to have been covered by the glacier at that time, which we find unlikely as it would probably have been eroded away, or at least deformed considerably, by the glacier. Moraine 3 was described by Eyþórsson (1935) to constitute an about 9-m high, distinct ridge, formed prior to 1931; it has only been explored and described by Eyþórsson (1935). Consequently, we assume it was either buried by glaciofluvial sediments or eroded by the glacial river during the next surge in 1935–1939 which formed Moraine 4 at a similar location (Figure 2).

The 1935–1939 surge is of specific interest because of its halt in 1937 when the margin retreated 27 m, and advanced again 57 and 2 m in the next 2 years (Eyþórsson, 1963). Its pause and restart remain unexplained, but could suggest two-phased surge wave reaching the ice front.

Leirufjörður surge history. Thoroddsen (1933, 1958) described steep, high and crevassed glacier snouts both in Kaldalón and Leirufjörður during his exploration in 1887. In Leirufjörður, local residents reported 'icebergs' collapsing from the glacier margin a few years prior to Thoroddsen's exploration (Eyþórsson, 1935; Thoroddsen, 1958). This indicates a major advance or surge of the glacier few years prior to Thoroddsen's (1958) survey. An exact timing of this advance is unknown, but we suggest the glacier surged approximately in the years 1880–1885 and formed Moraine 5. Moraine 3 and Moraine 4 are indistinct fluvially eroded gravelly ridge segments not providing any tephra or organic material for dating; their age is confined by Moraine 2 and Moraine 5. Eyþórsson (1935) described a moraine, Moraine 6, about 330 m proximal to Moraine 5. The glacier frontal position was reported at Moraine 6 approximately in the year 1898. This suggests a period with average surge intervals of 10 years in the years ~1840 to ~1898.

One could consider that 10 years would be an interval too short for mass build-up within the reservoir area so as to enable five surges in about 58 years. However, such high surge recurrence is also reported from some outlets of the large ice caps in central and south Iceland (Björnsson et al., 2003; Johnson et al., 2010). We do not have information on the ice dynamics in the 140-year period between the formation of Moraine 1 and Moraine 2. All older geomorphological features appear to have been removed by the ~1840 advance. The surge interval could have been similar in the period AD 1700–1840.

Heavily deformed sediments have been described in end moraines formed during surges of Brúarjökull and Eyjabakkajökull, surging outlet glaciers of northern Vatnajökull. The characteristic glaciotectonic deformation structures in tephra and fine sediments are considered strongly related to surge behaviour, although they cannot be considered diagnostic for surges alone (Benediktsson et al., 2008, 2009, 2010). Moraine 2 in Leirufjörður consists of heavily glaciotectonized sediments (Figure 4) that strongly resemble descriptions of the moraines from Brúarjökull and Eyjabakkajökull. This and the observed surge behaviour of the glacier supports the conclusion of Þórarinsson (1958) and Björnsson et al. (2003) that Moraine 2 was formed by a surge.

Kaldalón surge history. An area with pitted sandur and hummocky moraine, indicating melting of stagnant ice, on the proximal side of Moraine 1 has been mapped in Kaldalón (Brynjólfsson et al., 2014; John and Sugden, 1962). We consider it to support earlier conclusions (Björnsson et al., 2003; Þórarinsson, 1969), that Moraine 1 in Kaldalón was formed by a surge. The exact ages of Moraines 2 and 3 are unknown, but confined by the surges in AD ~1740 and ~1860 (Figure 5).

Similar to Leirufjörður, the glacier snout was steep, high and crevassed during Thoroddsen's visit in 1887. However, no end moraine was recognizable proximal to the 1887 ice marginal position, during our fieldwork in 2013. Thoroddsen's (1933, 1958) description of the glacier snout could indicate recent or active surge about the time of his exploration in 1887. Eyþórsson (1935) discussed discernible lateral moraines on the southern valley slope which we correlate with surge activity. He concluded that the lateral moraines indicated fluctuations, related to colder periods, that could not be clarified exactly due to lack of terminal moraines.

Eyþórsson (1935) proposed that Moraine 5 was formed during a still stand of the margin because of relatively cold summers in 1920–1925. However, glaciers are generally not considered to shift from surge-type to a climatically forced non-surging glacier (Benn and Evans, 2010). Hence, according to the observed surge behaviour of Kaldalónsjökull we suggest that Moraine 5 was formed by a surge. The glacier margin was about 100 m proximal to Moraine 5 in 1931. According to the few years' time the outlets needs to separate from a moraine and a 45-m mean annual retreat during the last quiescent phase of Kaldalónsjökull, we conclude that Moraine 5 was formed by a surge approximately AD 1920.

Surge interval, duration and LIA maximum conditions

Surge interval and duration. The surge interval of Icelandic glaciers is from c. 10 years up to more than a century. Most surging glaciers in Iceland do not experience regular surge periodicity (Björnsson et al., 2003), demonstrating that the time span needed to recover from a surge, build up a steep profile and accumulate mass in the reservoir area before a new surge is initiated can be variable for a particular glacier.

On the other hand, surge events can be missing in the historical record. This is well demonstrated in the geomorphological record of the Drangajökull outlets, where, for example, the number of end moraines is higher than the number of historically recorded surges (Brynjólfsson et al., 2012, 2014). In Leirufjörður, five surges occurred in the 58-year period between ~1840 and ~1898, forming Moraine 2 to Moraine 6. About 10-year surge interval in this period is short for the glacier to recover between the surges, and contrasting to the 50-year quiescent phase of the two subsequent surges. It remains unknown if any surges occurred in the 140-year period between the surges in AD ~1700 and ~1840 of Leirufjarðarjökull. On Svalbard, more frequent surges have been related to colder conditions in the 19th and early 20th centuries, compared with the late part of the 20th century (Dowdeswell et al., 1995). Surge interval less than 10 years is known from a few glaciers in south and central Iceland (Björnsson et al., 2003; Johnson et al., 2010). A 40- to 50-year variation in the length of quiescent phase within the same ice caps is also known from other Icelandic surging glaciers (Björnsson et al., 2003; Þórarinsson, 1964, 1969).

Surges of the larger ice caps in central and south Iceland, often lasting 2–3 years, start with glacier surface velocity acceleration (Björnsson et al., 2003; Fischer et al., 2003; Pálsson et al., 1992). Subsequently, a surface bulge propagates down-glacier, and after reaching the glacier front, it results in a glacier advance of several hundred metres to some kilometres, which normally takes only 2–3 months before termination (Aðalgeirsdóttir et al., 2005; Björnsson et al., 2003). For Drangajökull, the advances of the glacier margins took 3–6 years in the last two surges of the three surge-type outlets. This greatly contrasts with the general few months' duration of the marginal advance during surges of the bigger ice caps in central and south Iceland (Björnsson et al., 2003; Þórarinsson, 1969). Additionally, 3 years with no or negligible retreat before the last surges of Reykjarfjarðarjökull and Kaldalónsjökull could indicate that the ice velocity in the receiving area had already accelerated and overcome the potential marginal retreat. Thus, these two surges have potentially been going on for at least 8 years. On the larger Icelandic ice caps, it generally takes 2–3 years from the first signs of increased sliding and the subsequent down-glacier propagation of a surge bulge (Björnsson et al., 2003; Sigurðsson, 1995). Surges of at least some Icelandic glaciers start at central locations on the glaciers, with ice flow rates then accelerating both upward and downward along their longitudinal profiles (Fischer et al., 2003; Pálsson et al., 1992). The active phase of surge turns out to be longer for smaller surging glaciers in north and northwest Iceland than for the bigger ice caps in Iceland. The advance of Búrfellsjökull, a small surge-type cirque glacier in northern Iceland, lasted for about 3 years after a surge bulge reached the margin in 2001 (Brynjólfsson et al., 2012). For Drangajökull, it is unknown when the ice velocity accelerated, and how long it took a potential surge bulge to travel down the glacier before the marginal advance started in the recent most surges.

Apparently, the long surge duration of Drangajökull outlets is more similar to that of polythermal surging glaciers in Svalbard, where the surge mechanism is often characterized by a thermal transition (Dowdeswell et al., 1991; Jiskoot et al., 1998; Murray et al., 2003). However, all Icelandic glaciers are warm based, and the long active phase of Drangajökull surges can therefore not be related to a surge mechanism that relies on thermal transition. The surge mechanism and propagation of a surge bulge down-glacier during surges are rather considered to be controlled by a switch between a channelized basal drainage system and a distributed drainage system (Björnsson et al., 2003). The different substratum or climatic settings of Drangajökull compared with the other ice caps of Iceland might explain the longer surge duration. It is interesting to note that Drangajökull outlet surge recurrence has generally decreased after the LIA, particularly since the 20th century warming trend set in. This agrees with the observations of Striberger et al. (2011), based on a more than 2000-year long surge record of Eyjabakkajökull, that it doubled its surge recurrence during the LIA. They suggested that surge periodicity of Eyjabakkajökull was forced by climatically driven mass balance changes which might be a common forcing factor for similar surge-type outlet glaciers.

Maximum conditions during LIA. The surface drawdown of the reservoir areas is often tens of metres during surges of Icelandic glaciers (Aðalgeirsdóttir et al., 2005; Björnsson et al., 2003; Jóhannesson et al., 2013). We consider that large parts of the dramatic ice thickness change of Reykjarfjarðarjökull can be explained by drawdown of the surface, related to the surge in 1936–1939, and not only because of warmer climate in the 20th century. This is supported by the ice level observations at the Reyðarbunga *nunatak* (Figure 1). Thus the ice thickness around Reyðarbunga has increased by tens of metres during the last decades despite similar climate as in the period 1930–1960 when the main ice thinning was observed (Eyþórsson, 1963).

When producing the DEMs for maximum conditions during LIA, we used about 50-m ice thinning around Reyðarbunga, from

~1910 until present, as an evenly distributed surface lowering for the reservoir areas since LIA maximum conditions. This is a simplification, but it gives a minimum estimate of the ice thinning and volume changes of the surging outlets since their maximum conditions in LIA. Unfortunately, calculating the ice volume loss as a proportion of the whole ice volume is not possible as the thickness of the ice cap is currently only known in a few points (Figures 2, 3 and 5; Magnússon et al., 2004).

The surface lowering of the Drangajökull outlets reached a maximum lowering of 160–240 m, near the present marginal locations (Figure 7). The surface area decrease of the surging glaciers was about 20–35% from maximum extent during LIA until present (Table 2). About 34% area decrease of the whole ice cap, in the same period, is most likely overestimated, because of asynchronous maximum extent of the outlets. Those values are comparable with about 150- to 270-m outlet surface lowering since the LIA maximum conditions of non-surging glaciers that drain the southern Vatnajökull ice cap. They decreased in area by 10–30%, and individual glaciers lost about 10–50% of their volume since ~1890 (Hannesdóttir et al., 2014). Based on a lacustrine sediment record from two non-surging outlets of the Langjökull ice cap, Larsen et al. (2013) propose two main growth phases of the ice cap during the LIA, c. AD 1400 to 1550 and c. AD 1680 to 1890.

It is difficult to reconstruct the exact timing for maximum conditions of the Drangajökull ice cap during LIA. Apparently, the three main outlets reached their maximum area, and probably thickness, asynchronously. The fine sediments and organic material that was pushed up to form Moraine 2 in Leirufjörður (Figure 4) indicate a sufficiently long period with favourable conditions for vegetation growth and low energy sedimentation without disturbance from glacier advances. Framed in by their largest surges, the individual outlets of Drangajökull reached their LIA maximum extent asynchronously in the period AD ~1700–1846.

Surges and climate

External forcing such as climate does not trigger a surge directly but can alter the surge periodicity by affecting the mass accumulation of a glacier, and prevent a new surge event in terms of insufficient mass build-up in the reservoir area (Dowdeswell et al., 1995; Hewitt, 2007; Striberger et al., 2011). However, for some glaciers in the Karakoram region, in Pakistan, the periodicity of surges seems to be consistent since the LIA, despite significant climate changes (Quincey and Luckman, 2014).

The surge interval varies between and within the three surging outlets of Drangajökull. A longer quiescent phase, 55–60 years, in the second half of the 20th century could be considered a result of generally warmer climate compared with the 19th century (Figure 8). Thus, longer time was needed for snow accumulation to refill the reservoir area (Eisen et al., 2001) because of more ablation in this period. Interestingly, all the three outlets surged during a period of relatively high average temperature in the 1930s–1940s and again around the 2000s. How or whether this could be related to increased melting and potentially more water in the englacial and subglacial meltwater systems remains unknown. Some surges are known to terminate in the melting season or during exceptionally high ablation events (Eisen et al., 2001, 2005). This is considered a result of increased meltwater entering from the surface and a shift from inefficient distributed subglacial drainage system to an efficient channelized system (Kamb, 1987; Kamb et al., 1985). A strong relationship between sufficient accumulated mass over a period and surge initiations has been described from Variegated glacier in Alaska (Eisen et al., 2001). The process leading to initiation of Drangajökull surges seems to be independent of average temperature since AD ~1800 (Figure 8). The shorter surge distances of Kaldalónsjökull in the 1930s and Reykjarfjarðarjökull in the 2000s compared with the other surges at the same time could indicate a smaller amount of accumulated mass in the reservoir areas for driving the surge bulge down-valley (Figure 8).

Both decrease of the area and thinning of the surging outlets are most likely related to a generally warmer climate and thus increased melting since the early 20th century. Short data sets for precipitation in the area do not allow concrete assumptions of changed pattern for accumulation (Figure 8). It could be assumed that an increased percentage of the precipitation falls as rain on the glacier as the mean average temperature in Iceland has increased by 1–2°C since the LIA maximum (Figure 8; Guðmundsson, 1997; Hanna et al., 2004). Increased annual air temperature of 1–2°C has been estimated to increase the ELA by 100–150 m on valley and cirque glaciers in north Iceland (Caseldine and Stötter, 1993; Stötter et al., 1999).

The three surging outlets show different patterns of retreat since their LIA maximum extent. The extent of Leirufjarðarjökull remained rather stable until the turn of the 19th and 20th centuries when the average temperature started to increase (Figure 3). Prior to Thoroddsen's exploration in 1887, the average retreat of the glacier was about 6 m per year, compared with about 17 m afterwards. The average retreat rate of Reykjarfjarðarjökull changed greatly around 1939 (Figure 2). In 1846–1939, its average retreat was about 17 m per year and about 32 m in 1939–2013. No specific change of the retreat rate of Reykjarfjarðarjökull is noted in the beginning of the 20th century. Unexpectedly, on average, Kaldalónsjökull retreated faster before the temperature rise in the early 20th century (Figure 5). From 1740 to 1887, it retreated about 16 m per year, and about 9 m per year in 1887–2013. This demonstrates the importance of also considering the relationship to non-climatic factors such as topography confining the glaciers, and that fluctuations of surge-type glaciers are not well suited as a proxy for short-term climate changes (Yde and Paasche, 2010).

Conclusion

- The historical data reveal information on advances and retreats of the surging outlets prior to their maximum extent during LIA. Descriptions of fast and sudden frontal advances during the LIA are most likely surges rather than glacier responses directly triggered by climate.
- Five surges are reconstructed from Reykjarfjarðarjökull, seven from Leirufjarðarjökull and six from Kaldalónsjökull.
- The surge interval varies between and within the outlets. About 50–60 years are between the last two surges of each outlet. The interval is generally shorter in the 19th century and in the beginning of 20th century. Surges of Leirufjarðarjökull occurred with about 10-year intervals during the period AD ~1840 to ~1898. About 140 years elapsed with no surges recorded from Leirufjarðarjökull from AD ~1700 to ~1840, possibly reflecting a lack of information rather than a long quiescent phase of the glacier.
- Extent, ice thickness, and ice volume changes of the maximum surges during LIA were quantified. The surging glaciers' volumes were estimated 2.1–2.5 km^3 higher than at present, and the area has diminished by 22–34% since the maximum extent in the LIA.

The ice cap decreased in area from ~216 km^2 during the LIA maximum extent to 143 km^2 in 2011. Individual outlets of Drangajökull reached their LIA maximum extent asynchronously during surges in the period AD ~1700–1846. Leirufjarðarjökull reached its Holocene maximum extent about AD 1700 and again about AD 1840. Reykjarfjarðarjökull reached its LIA maximum

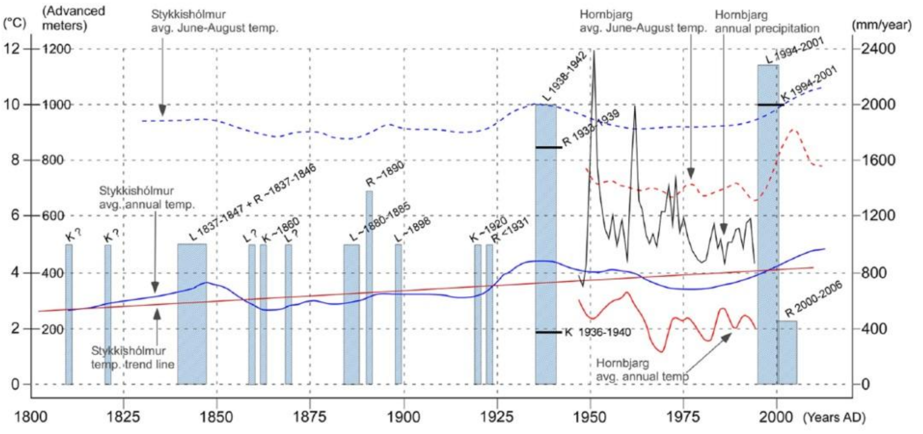

Figure 8. Surges of the three surge-type outlets of Drangajökull ice cap plotted against meteorological data; vertical columns represent surges (K = Kaldalónsjökull, L = Leirufjarðarjökull, R = Reykjarfjarðarjökull). Their widths indicate the active phase duration, except the thinnest columns which are active phases of unknown duration. The column heights indicate the surge distance in metres; for unknown surge distances, the columns are set to 500 m. The meteorological data are provided by the Icelandic Met Office.

extent about AD 1846. Two more distal indistinct moraines indicate greater Holocene advances of Reykjarfjarðarjökull. Kaldalónsjökull reached its LIA maximum extent approximately AD 1740. Two more distal moraines indicate greater Holocene advances of Kaldalónsjökull.

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Appendix 1. Chemical composition of the tephra; Hekla AD 1693 tephra, from within Moraine 2 in Leirufjörður (Figure 4).

Hekla 1693 AD	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	SUM
Leirufjörður	62.33	1.27	15.77	9.48	0.32	2.01	4.62	4.27	1.63	101.70
	61.14	1.08	15.21	8.96	0.28	1.77	5.06	3.81	1.64	98.95
	60.92	1.15	15.13	8.81	0.20	1.85	4.95	4.12	1.86	98.99
	60.57	1.22	15.28	9.05	0.25	1.84	5.00	3.89	1.76	98.87
	60.19	1.28	15.33	9.91	0.26	2.01	5.35	3.90	1.50	99.73
	60.11	1.21	14.79	9.46	0.28	1.92	5.20	4.55	1.45	98.97
	60.02	1.19	15.25	9.21	0.20	1.95	4.99	4.01	1.61	98.44
	59.92	1.22	15.10	9.01	0.16	2.08	5.24	4.02	1.66	98.41
	59.90	1.24	15.24	9.55	0.15	1.86	5.07	3.74	1.62	98.37
	59.87	1.19	15.42	9.75	0.11	1.82	5.27	3.92	1.64	98.99
	59.87	1.22	15.40	9.74	0.28	1.97	5.05	3.84	1.62	98.99
	59.87	1.27	15.18	9.35	0.28	2.03	5.13	3.84	1.63	98.58
	59.78	1.23	15.48	10.04	0.38	1.91	5.37	3.82	1.59	99.60
	59.75	1.28	14.86	9.43	0.11	1.79	5.37	4.32	1.73	98.64
	59.68	1.22	14.92	9.87	0.32	1.96	5.10	3.65	1.63	98.35
	59.63	1.31	15.27	9.33	0.16	1.86	5.36	3.97	1.72	98.61
	59.61	1.26	15.28	9.47	0.25	1.90	4.85	3.87	1.70	98.19
	59.60	1.21	15.51	9.22	0.26	1.85	5.42	4.28	1.40	98.75
	59.57	1.27	15.15	9.63	0.29	2.07	5.17	3.96	1.55	98.67
	59.57	1.19	15.00	9.59	0.26	1.98	5.12	3.83	1.53	98.08
	59.47	1.34	15.23	9.86	0.25	2.00	5.22	3.80	1.65	98.82
	59.43	1.15	14.78	9.99	0.16	1.92	5.27	3.87	1.64	98.21
	59.37	1.27	15.19	9.37	0.26	1.95	5.04	4.11	1.75	98.31
	59.36	1.24	14.94	10.47	0.35	2.10	5.35	3.67	1.52	99.00
	59.34	1.19	14.88	9.57	0.29	1.93	5.21	3.93	1.78	98.12
	59.24	1.27	15.05	9.85	0.22	2.06	5.42	3.78	1.47	98.36
	59.22	1.11	15.11	10.27	0.36	1.98	5.24	4.07	1.63	98.99
	59.19	1.15	15.32	10.34	0.29	3.29	6.14	3.89	1.43	101.03
	59.12	1.15	15.18	9.94	0.24	1.99	5.17	3.87	1.52	98.17
	58.83	1.30	15.10	10.42	0.19	2.00	4.97	3.76	1.48	98.05
	58.31	1.22	14.33	9.45	0.26	1.84	4.96	3.36	1.53	95.27
	58.21	1.25	15.25	10.31	0.27	2.22	5.33	4.38	1.54	98.76
	57.95	1.20	14.86	9.79	0.19	1.85	5.04	3.58	1.55	96.01
	55.25	1.22	14.80	9.39	0.18	2.04	5.61	4.69	1.41	94.59
Mean	**59.54**	**1.22**	**15.14**	**9.64**	**0.24**	**1.99**	**5.20**	**3.95**	**1.60**	**98.52**
Stdv	1.11	0.06	0.27	0.43	0.07	0.25	0.26	0.27	0.11	1.28

Appendix III

Quaternary Science Reviews (in review)

Late Weichselian-Early Holocene glacial history of northwest Iceland, constrained by ^{36}Cl cosmogenic exposure ages

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Highlights

- 24 new ^{36}Cl exposure ages constrain the glacial history of northwest Iceland.
- A strongly topographically confined ice cap covered northwest Iceland during LGM.
- Warm-based erosive ice occupied valleys, while cold-based ice covered uplands.
- Glaciers advanced 9.3 ka BP as a response to climate deterioration.
- Extensive glaciation persisted longer in NW Iceland than elsewhere in Iceland.

Abstract

We present 24 new cosmogenic isotope (^{36}Cl) surface exposure ages from erratic boulders, moraine boulders and glacially eroded bedrock that constrain the late Weichselian to Holocene glacial history of northwest Iceland. The results suggest a topographically controlled ice sheet over the Vestfirðir (Westfjords) peninsula during the last glaciation. Cold based non-erosive sectors of the ice sheet covered most of the mountains while fjords and valleys were occupied with erosive, warm-based ice.
High ^{36}Cl exposure ages from uplands and mountain plateaux (L8; 76.5 ka and H1; 41.6 ka) in combination with younger erratic boulders (L7; 26.2 and K1-K4; 15.0–13.8 ka) superimposed on such surfaces suggest the presence of non-erosive ice over uplands and

plateaux in the Vestfirðir peninsula during the last glaciation. Glacially scoured terrain and erratic boulders yielding younger exposure ages (L1-L6; 11.3–9.1 ka and R1, R6-R7; 10.6–9.4 ka) in the lowland areas indicate that the valleys and fjords of the Vestfirðir peninsula were occupied by warm-based, dynamic ice during the last glaciation.
The 26.2 ka deglaciation time of mountain Leirufjall indicate ice thinning and deglaciation of some mountains and plateaux that preceded any considerable lateral retreat of the ice sheet. Subsequently this initial ice thinning was followed by break-up of the shelf based ice sheet off Vestfirðir about 15 ka BP. Hence, the new exposure ages suggest a stepwise asynchronous deglaciation on land, following the shelf break-up with some valleys and most of the uplands, deglaciated about 14-15 ka BP.
The outermost moraine at the mouth of Leirufjörður was dated to 9.3 ka BP, and we suggest the moraine to be formed by a glacier re-advance in response to a cooler climate forced by the reduced Atlantic Meridional Overturning Circulation at around 9.3 ka BP. A system of moraines proximal to the 9.3 ka moraine in Leirufjörður and a 9.4 ka deglaciation age in the coastal area of Reykjarfjörður suggest that an extensive ice cap was preserved over the eastern Vestfirðir peninsula at least until c. 9 ka BP.

Keywords: Drangajökull ice cap, 9.3 ka event, deglaciation, moraine, erratic boulder, blockfield, cosmogenic exposure age

1. Introduction

Identifying temporal and spatial changes of the last Icelandic ice sheet (IIS) is essential to improve our understanding of its interactions with ocean-atmospheric systems in the North Atlantic during the late Pleistocene and early Holocene (Ingólfsson, 1991; Eiríksson et al., 2000; Hubbard et al., 2006; Norðdahl et al., 2008; Geirsdóttir et al., 2009; Ingólfsson et al., 2010). The location of Iceland at the boundary of relative warm Atlantic water of the Irminger Current and cold polar water in the East Greenland Current makes past and present Icelandic glaciers very sensitive to changes in oceanic and atmospheric circulation (Bergþórsson, 1969; Eiríksson et al., 2000; Flowers et al., 2007, 2008; Geirsdóttir et al., 2009). The Atlantic Meridional Overturning Circulation (AMOC) supplies the higher latitudes of the Atlantic with warmer surface ocean water which makes the North Atlantic climate relatively mild, considering its high latitude. However, large pulses of freshwater into the surface of the North Atlantic have been considered to reduce or temporarily shut down the northward heat transport by the AMOC, and therefore cause climatic deteriorations (Alley and Ágústsdóttir, 2005; Lewis et al., 2012). Climatic cooling events about 9.2-9.3 ka and 8.2 ka BP were probably caused by meltwater pulses from Lake Agassiz to the Atlantic Ocean, related to the deglaciation of the Laurentide ice sheet (Alley et al., 1997; Clarke et al., 2004; Teller et al., 2005; Kleiven et al., 2008; Murton et al., 2010). The signature of these events can be identified in various palaeo-archives (Alley et al., 1997; Kaufman et al., 2004; Alley and Ágústsdóttir, 2005; Fleitman et al., 2008; Solomina et al., 2015). Proxies for climatic deterioration during these periods in Iceland have been identified in marine and lacustrine sediment cores (Eiríksson et al., 2000; Geirsdóttir et al., 2009; Larsen et al., 2012). However, there are no observations of moraines or other landforms directly related to glacier advances in Iceland related to these events.

Our understanding of the Late Weichselian and Holocene environmental history of Iceland has improved significantly during the last two to three decades. Perhaps the most important progress is new models of a dynamic and rapidly changing ice sheet, compared to a slowly responding ice sheet in early reconstructions of the IIS (Norðdahl et al., 2008; Ingólfsson et al., 2010). Configuration and thermal conditions of the IIS during the Last Glacial Maximum (LGM) and the deglaciation history still suffer from a considerable lack of chronological data, especially direct dating of terrestrial landforms and sediments (Andresen et al., 2005; Caseldine et al., 2006; Hubbard et al., 2006; Norðdahl et al., 2008 Geirsdóttir et al., 2009, 2013).

Two main theories have been proposed for the LGM in NW Iceland; an independent, restricted ice cap that partly occupied the Vestfirðir peninsula without merging with the main IIS, i.e. leaving nunataks, coastal areas and mountains ice free in between ice-streams which drained the ice sheet through the main fjords and valleys (Thoroddsen, 1911; Þórarinsson, 1937; Sigurvinsson, 1983; Hjort et al., 1985; Ingólfsson, 1991; Norðdahl, 1991; Rundgren and Ingólfsson, 1999; Andrews et al., 2002; Principato et al., 2006). More recently it was suggested that the Vestfirðir ice cap and the main IIS coalesced into one large ice sheet probably covering most of the country except some coastal mountains (Syvitski et al., 1999; Norðdahl and Pétursson 2005; Hubbard et al., 2006; Norðdahl et al., 2008; Geirsdóttir et al., 2009). The latter case requires a single domed, up to 2000 m thick, ice sheet that occupied the whole island and extended far out on the Icelandic shelf (Norðdahl and Pétursson, 2005; Hubbard et al. 2006). Thus, the extent of glaciation in Vestfirðir is still not known in detail, and it remains to firmly establish whether ice free refugia and nunataks existed during the Weichselian (Rundgren and Ingólfsson, 1999; Norðdahl et al., 2008; Ingólfsson et al., 2010).

At the LGM, the outlet glacier occupying the main fjord on the Vestfirðir peninsula, Ísafjarðardjúp, is considered to have reached either of two potential moraine banks located 10 km and 30 km off the mouth of Ísafjarðardjúp, respectively (Fig. 1; Andrews et al., 2002; Geirsdóttir et al., 2002). Hjort et al. (1985) considered the LGM ice sheet to have extended about 6 km on to the shelf north of the Vestfirðir peninsula.

Marine sediment cores, geophysical data, and correlation of ^{14}C dated shorelines, indicate a sudden retreat of the IIS from the shelf area due to rapidly rising sea level about 15 ka BP (Syvitski et al., 1999; Andrews et al., 2000, 2002; Eiríksson et al., 2000; Ingólfsson and Norðdahl, 2001; Geirsdóttir et al., 2002; Norðdahl and Pétursson, 2005). Absence of ice rafted debris (IRD) by 15 ka BP in sediment cores from the shelf off Vestfirðir was interpreted as rapid deglaciation of the Vestfirðir shelf (Andrews et al., 2002; Geirsdóttir et al., 2002). IRD was commonly observed between 12 and 10 ka BP in sediment cores from the Jökulfirðir fjord system, indicating valleys and fjords occupied with calving glaciers in the eastern Vestfirðir peninsula at that time (Fig. 1; Geirsdóttir et al., 2002, 2009).
IRD in the Jökulfirðir and Húnaflói sediment cores ceases after 10.2 ka BP, suggesting rapid ice retreat on shore and only minor ice caps covering highland areas (Geirsdóttir et al. 2002; Castañeda et al., 2004). Furthemore, presence of the Saksunarvatn tephra in a few terrestrial localities around Drangajökull confirms that eastern Vestfirðir were at least partly deglaciated about 10.2 ka BP (Hjort et al., 1985; Principato et al., 2006; Hole, 2015). However, due to the lack of directly dated glacial landforms and sediments, the history of the Vestfirðir ice cap and Drangajökull is poorly known after glaciers retreated on shore c. 10.2 ka BP (Geirsdóttir et al., 2009).
Simultaneously, the IIS was retreating rapidly to the central highlands of Iceland (Kjartansson, 1955, 1964; Kaldal and Víkingsson, 1990; Tómasson, 1993; Caseldine et al., 2003; Geirsdóttir et al., 2009; Striberger et al., 2012). Valleys and cirques in the Tröllaskagi peninsula in central north Iceland were deglaciated or hosted glaciers of similar size as at present (Stötter et al., 1999; Caseldine et al., 2003, 2006; Wastl et al., 2005) about 10 ka BP. Furthermore, studies of lacustrine sediments indicate absence of glacial meltwater from the Langjökull ice cap into the proglacial lake Hvítárvatn in central Iceland and a smaller Langjökull ice cap than at present by 9.5 ka BP (Geirsdóttir et al., 2009; Larsen et al., 2012).
Cosmogenic exposure dating is widely applied to glacial landforms where other sources for chronological control and reconstruction of glacier fluctuations are restricted (Gosse and Phillips, 2001; Briner et al., 2005; Dunai, 2010; Balco, 2011; Young et al., 2011, 2013). Principato et al. (2003, 2006) applied the ^{36}Cl exposure dating method to reconstruct the deglaciation of NW Iceland. They demonstrated the potential to obtain absolute ages of glacial landforms that cannot be dated by other techniques. However, the age calculation suffered from uncertainty of the ^{36}Cl production rates (Principato et al., 2006). The exposure age of erratic boulders and bedrock indicated at least some ice free coastal mountains about 20 ka BP, and deglaciation of valleys already about 11.7 ka and 14.6 ka BP (Principato, 2003; Principato, et al., 2006).
Here, we present twenty-four ^{36}Cl exposure ages of rock samples from erratic boulders, moraine boulders, and glacially sculpted bedrock. They are used along with the exposure

ages from Principato et al. (2006) to reconstruct the glacial history of the Vestfirðir area. Additionally, we investigated six river-cut sediment profiles in the Reykjarfjörður valley, the aim is to reconstruct the environmental conditions there during the final deglaciation.
The aim of this study is to reconstruct the Late Weichselian to Holocene glacial history, obtain chronological control of the last deglaciation and constrain the terrestrial glacial conditions of the Drangajökull ice cap and Vestfirðir peninsula from LGM until the early Holocene.

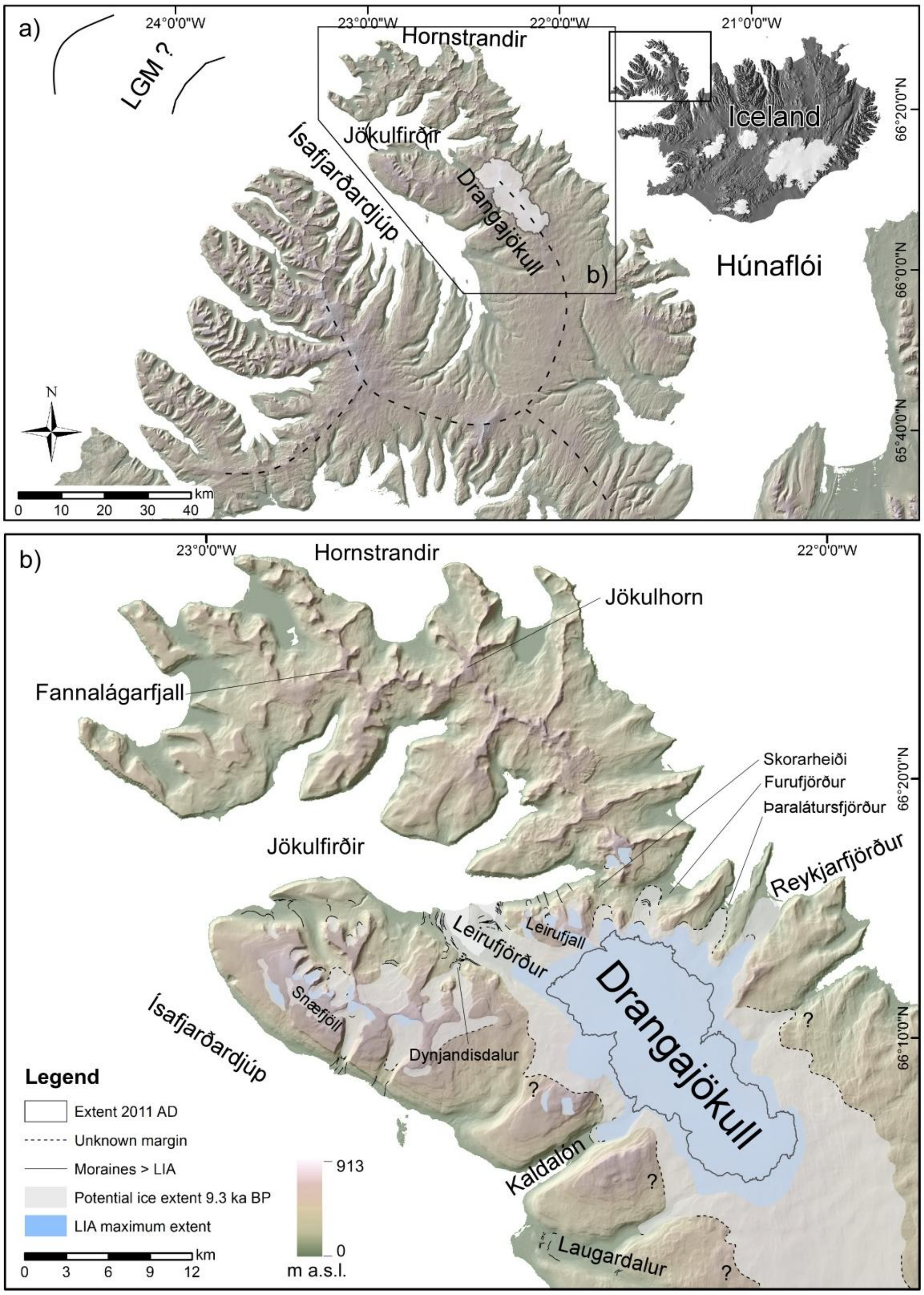

Fig. 1. The study area at Drangajökull ice cap and its surroundings in northwest Iceland. The topographical maps are based on a digital elevation model with 20 m resolution. a) Overview map of Vestfirðir, northwest Iceland. The dashed line represents the ice divide over the Vestfirðir peninsula during LGM (Norðdahl, 1991; Ingólfsson and Norðdahl,

2001). b) The Drangajökull area, the present extent is measured, the LIA extent is adapted from Brynjólfsson et al. (2015) and the potential extent 9.3 ka BP is reconstructed from geomorphology and exposure ages.

2. Setting

The Vestfirðir peninsula (Fig. 1), located approximately at 66°N and 23°W, consists mainly of Miocene sub-aerial tholeiitic and porphyritic basalts, interbedded with thin sediment layers, and some outcrops of olivine basalts and volcanoclastic sedimentary horizons (Sæmundsson, 1979; Einarsson, 1991; Kristjánsson and Jóhannesson, 1994; Guðmundsson et al., 1996; Harðarson et al., 1997). The landscape is characterized by steep glacially eroded fjords and valleys, often confined by 500-700 m high basaltic plateaux. However, the eastern part of Vestfirðir is a relatively large, 350-600 m high upland hosting the Drangajökull ice cap (Fig. 1a). Surfaces of the basaltic plateaux are commonly characterized by block fields with 2-4 m wide surface polygons or sorted stripes. Glacial sediments are mostly absent in the block fields, except for scattered, relatively fresh-looking erratic boulders. The erratic boulders are very common in lowland locations, but are rare on the uplands and plateaux 400-600 m a.s.l. on the eastern Vestfirðir peninsula (Brynjólfsson et al., 2014). However, a diamict, either locally weathered bedrock or till, was occasionally observed on the uplands.

An alpine landscape with cirques cut into the plateau basalts dominates the Hornstrandir area, about 25 km north of the Drangajökull ice cap (Fig. 1), indicating glaciation by valley or cirque glaciers. This topography suggests that mountain plateaux and coastal capes could have been ice free during the last glaciation (Sugden and John, 1976; Símonarson, 1979; Hjort et al., 1985; Principato et al., 2006).

The dome shaped Drangajökull ice cap, situated 100-915 m a.s.l., has a very low mean equilibrium line altitude, at about 550-600 m a.s.l., compared to 1000-1300 m a.s.l. on other ice caps in Iceland (Eyþórsson, 1935; Björnsson, 1979; Björnsson and Pálsson, 2008). Drangajökull extended over about 190-215 km^2 during the Little Ice Age (LIA) maximum (Sigurðsson et al., 2013; Brynjólfsson et al., 2014, 2015) and now covers 142 km^2 (Jóhannesson et al., 2013). Moraines formed during the LIA have been mapped all around the northern perimeter of the ice cap. Undated moraines located beyond the LIA moraines in valleys and cirques around the northern perimeter indicate advanced glacial positions in the period between the last deglaciation and the LIA maximum (Principato, 2008; Brynjólfsson et al., 2014). Three surge-type glaciers, Reykjarfjarðarjökull, Leirufjarðarjökull, and Kaldalónsjökull are the main outlets of the Drangajökull ice cap at present (Fig. 1b; Þórarinsson, 1969; Sigurðsson 1998; Björnsson et al., 2003; Brynjólfsson et al., 2015).

Geological sections in Quaternary sediments for stratigraphical and sedimentological studies are limited in the area. About 1-3 m high river-cut sections occur in the valleys Reykjarfjörður, Leirufjörður and Kaldalón. Sediments are often thin and protruded by ice sculpted bedrock except in the valleys, which are the main depocenters (Brynjólfsson et al., 2014, 2015). Raised beaches are located 5-48 m a.s.l. on the eastern Vestfirðir. They reach up to 30 m a.s.l. in the mouth of Kaldalón, 14 m a.s.l. in the mouth of Jökulfirðir and not higher than 5 m a.s.l. in Reykjarfjörður and Leirufjörður (Principato, 2008).

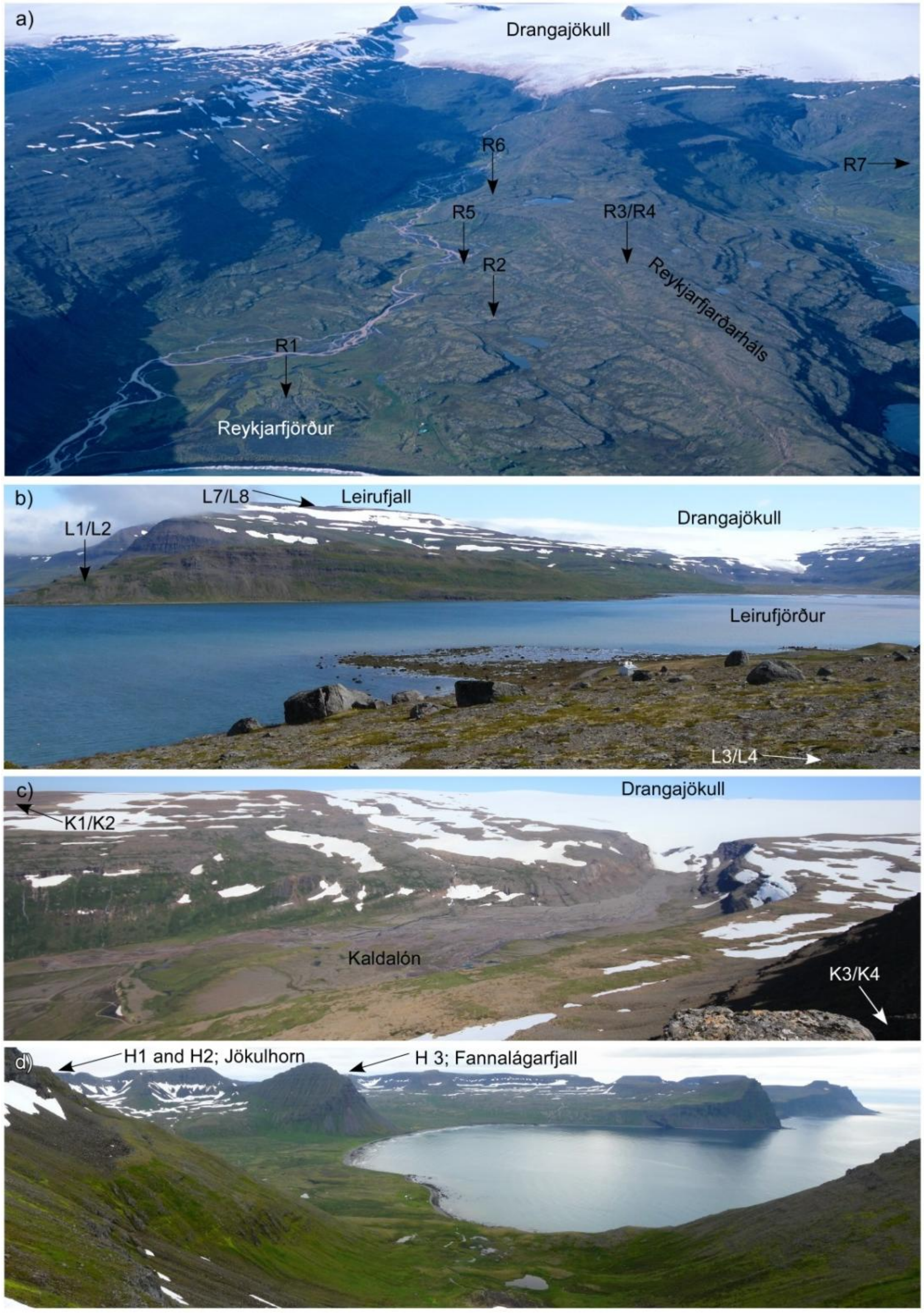

Fig. 2. Photographs showing the three main outlet glaciers of Drangajökull at present, and some of the rock sample localities. a) View into Reykjarfjörður towards Drangajökull in the southwest (Photograph: Oddur Sigurðsson, 2001). b) View towards Drangajökull and the east side of Leirufjörður, July 2012. c) The inner part of Kaldalón viewed towards Drangajökull in the east, July 2013. d) View towards Jökulhorn and Fannalágarfjall on Hornstrandir. July 2014.

This study focuses on the three valleys hosting the main outlet glaciers that drain Drangajökull the ice cap at present (Fig. 2). Ice sculpted bedrock and erratic boulders are common in the 7 km long NA-SW orientated valley Reykjarfjörður (Fig. 2a). The 10 km long NW-SA oriented Leirufjörður valley (Fig. 2b), hosts a group of about 10 undated moraines in the valley mouth (Brynjólfsson et al., 2014). The surface of the plateau mountain, Leirufjall, on the east side of the valley is characterized by a block field with only few erratic boulders. Distinct lateral and terminal moraines are also located in the tributary valley, Dynjandisdalur, on the west side of Leirufjörður (Brynjólfsson et al., 2014).

Kaldalón (Fig. 2c) is c. about 10 km long SW-NA orientated valley confined by plateau mountains on each side. Infrequent erratic boulders superimposed on the block fields covering the plateaux above 450-500 m a.s.l. are often degraded and difficult to sample (Fig. 3). Principato et al. (2006) mapped and dated a moraine ridge to 11.7 ka BP about 500 meters distal to the LIA maximum moraine.

3. Methods

Fieldwork with focus on documenting glacial sediments and landforms was carried out in the Drangajökull area during the summers 2011-2013. Initially, detailed geomorphological maps were produced of the Drangajökull forefields to identify moraines, ice-sculpted bedrock or other landforms that could be sampled for exposure dating (Brynjólfsson et al., 2014). All sample details and locations are indicated in Figs 4 and 5, Table 1 and the supplementary data.

Fig. 3. Overview of selected sampling localities for cosmogenic exposure dating. a) Sampling the outermost lateral moraine on the west side at the mouth of Leirufjörður (sample number L3), b) sampling an erratic boulder resting on bedrock in Reykjarfjörður (sample number R6), c) an erratic boulder resting on bedrock, surrounded by block field, on the plateau mountain north of Kaldalón (sample number K3), d) sampling from ice-sculpted bedrock in Reykjarfjörður (sample number R5).

3.1 Samples and sampling procedure

About 35 samples were chipped off erratic boulders, moraine boulders and bedrock in the recently mapped forefield of Drangajökull ice cap (Fig. 3; Brynjólfsson et al., 2014). Twenty-four samples with a good spatial coverage were selected for dating, six of them from moraine boulders, twelve from erratic boulders and six from ice sculpted bedrock (Table 1; Fig. 4, supplementary figure). The samples were collected from four sub-areas; three around Drangajökull (Reykjarfjörður, Leirufjörður, Kaldalón) and finally a set of boulders were sampled further north from the Hornstrandir peninsula (Figs 1 and 2d). In Reykjarfjörður, seven samples were sampled on a transect from the coast towards the glacier. Ten samples in three groups were sampled from the Leirufjörður area. Four paired samples were sampled from plateaux around Kaldalón and three from mountain summits in the Hornstrandir area.

Winter expeditions in the area confirmed that snow is generally blown off topographic highs, but icing and rime is common on the higher mountains. Fresh-looking rocks were sampled, preferably from topographic highs. Topographic shielding is well known error source for cosmogenic dating in mountainous areas (Ivy-Ochs and Kober, 2008; Stroeven et al., 2011). The topographic shielding was measured by compass and clinometer in the field and corrected for (Table 1). Potential weathering of sampled surfaces was estimated in the field for each sample, e.g. by looking for striations, pitting and other original or erosional surface textures.

Typical erosion rate for crystalline rock in most alpine environments is considered 0.5-2.5 mm/ka (Balco, 2011). Because this can be difficult to estimate, uncertainty of the erosion rate can contribute significantly to the exposure age of samples since the LGM or before (Balco, 2011). Problems with potential past burial (exhumation) or tilting of erratic boulders were minimized by choosing erratics sitting on bedrock or relatively stable block field, and large moraine boulders not showing evidence of any post-depositional movements (Ivy-Ochs and Kober, 2008; Balco, 2011). The positions of all samples were recorded with a handheld Garmin GPSmap 62sc, and the ISN93/WGS84 reference system was used for all data handling.

The boulders selected for comogenic age dating are all sub-alkaline basalts with 5-9 wt. % MgO (Supplementary Table) and are predominately aphyric, fine-grained vesicular lavas but porphyritic varieties, however, also occur. The phenocryst assemblage in the latter type is mainly plagioclase +/- clinopyroxene and olivine. Some of the lavas show flow banding and contain patches or small pockets of segregation veins. Generally, the analyzed samples, analytical methods described below, are fresh and unaltered as also evidenced by their relative low content of loss on ignition having an average value of 1.2 wt. % (Supplementary Table).

3.2 Sample preparation and exposure calculations

Because of chemical characteristics and lack of phenocrysts in the Icelandic basalt we used the cosmogenic ^{36}Cl nuclide to date whole-rock samples. A prerequisite for calculation of the exposure ages by the ^{36}Cl method is that the geochemistry and specific gravity of the sample is known. The uppermost 3-5 cm were cut of the samples, specific gravity was measured, each sample was crushed, and the 147-250 µm fraction was sieved at the lab at the Norwegian University of Science and Technology (NTNU) and the Activation Laboratories (ACTLABS), in Ancaster, Canada. Whole-rock major and trace elements concentration of the samples, data given in Supplementary Table, were analyzed by

ACTLABS by a combination of lithium metaborate/tetraborate fusion inductively coupled plasma mass spectrometry (ICPMS) for major elements and by traditional solution ICPMS for trace element analysis using a Perkin Elmer Sciex ELAN 6000 ICPMS. Chlorine concentrations were determined by instrumental neutron activation analysis (INAA). Loss on ignition (LOI) was determined by measuring weight loss upon heating to 1100° C over a three hour period. Several certified reference materials were used for data control. Totals of major elements are 100 ± 1.5 wt. %, with an analytical precision of 1–2 % for most major elements (Supplementary Table). The analytical precision for trace elements is better than 10 %.

Further work and sample preparation was carried out by the PRIME lab, Purdue University, Indiana, USA, and is briefly outlined below. During preparation and chemical processing of the samples, the procedure of Zreda et al. (1999) and Philips (2003) was followed. First the samples were chemically treated to remove any secondary materials. After cleaning, the crushed samples were dissolved in a mixture of hydrofluoric and nitric acids. To ensure there was sufficient concentration of Cl to be determined during the accelerator mass spectrometry (AMS), a spike of Cl was added during the step of dissolution. Following the dissolution, Cl was precipitated as AgCl. The ratio of ^{36}Cl/Cl and ^{37}Cl/^{35}Cl was determined by AMS analysis on AgCl targets.

The age of the samples was calculated using an Excel spreadsheet calculator developed for calculating ^{36}Cl ages by Schimmelpfennig et al. (2009). We used a production rate for Iceland of 57 ±5 atoms ^{36}Cl (g Ca)$^{-1}$ yr^{-1} (normalized to sea level with the standard atmosphere; Licciardi et al., 2008). However, for comparison, the ages were also calculated according to earlier production rates (Stone et al., 1996; Phillips et al., 2001; Table 1).

Scaling and shielding factors were calculated and corrected for each sample, i.e. latitude, elevation, and estimated shielding by snow cover, topography and erosion. Snow cover was assumed to be similar throughout the Holocene as at present. The scaling effect of latitude and elevation was calculated by CosmoCalc (Vermeesch, 2007) using the scaling model of Dunai (2001) and Desilets et al. (2006), which both yielded similar scaling factors and are considered more sophisticated than older scaling models by Lal (1991). CosmoCalc was also used to correct for potential snow cover. Uncertain weathering rates of bedrock in alpine landscapes and rapid weathering rates of Icelandic basalt can contribute to significant errors in the exposure age (Gíslason et al., 1996; Balco, 2011). For comparison, we corrected each sample for several scenarios of erosion and evaluated the results according to our field observations (Table 1).

3.3 Sedimentological studies

Six river-cut geological sections along the main river in Reykjarfjörður (Fig. 1) were cleaned in order to investigate the stratigraphy and sedimentology. Sediment structures and lithologies were logged using the data chart of Krüger and Kjær (1999). The sedimentological results along with the geomorphological maps of Brynjólfsson et al. (2014) enabled us to reconstruct the environmental setting during the final phase of deglaciation in Reykjarfjörður.

4. Results

The 24 ^{36}Cl exposure ages range from 76.5 ka to 7.2 ka (Table 1). The samples are located from valley floors about 10 m a.s.l. up to mountain plateaux and passes reaching 650 m a.s.l. (Fig. 4). All dated samples, except some in Reykjarfjörður and two at Skorarheiði, are in good agreement with their topographical position.

[Figure 4 here]

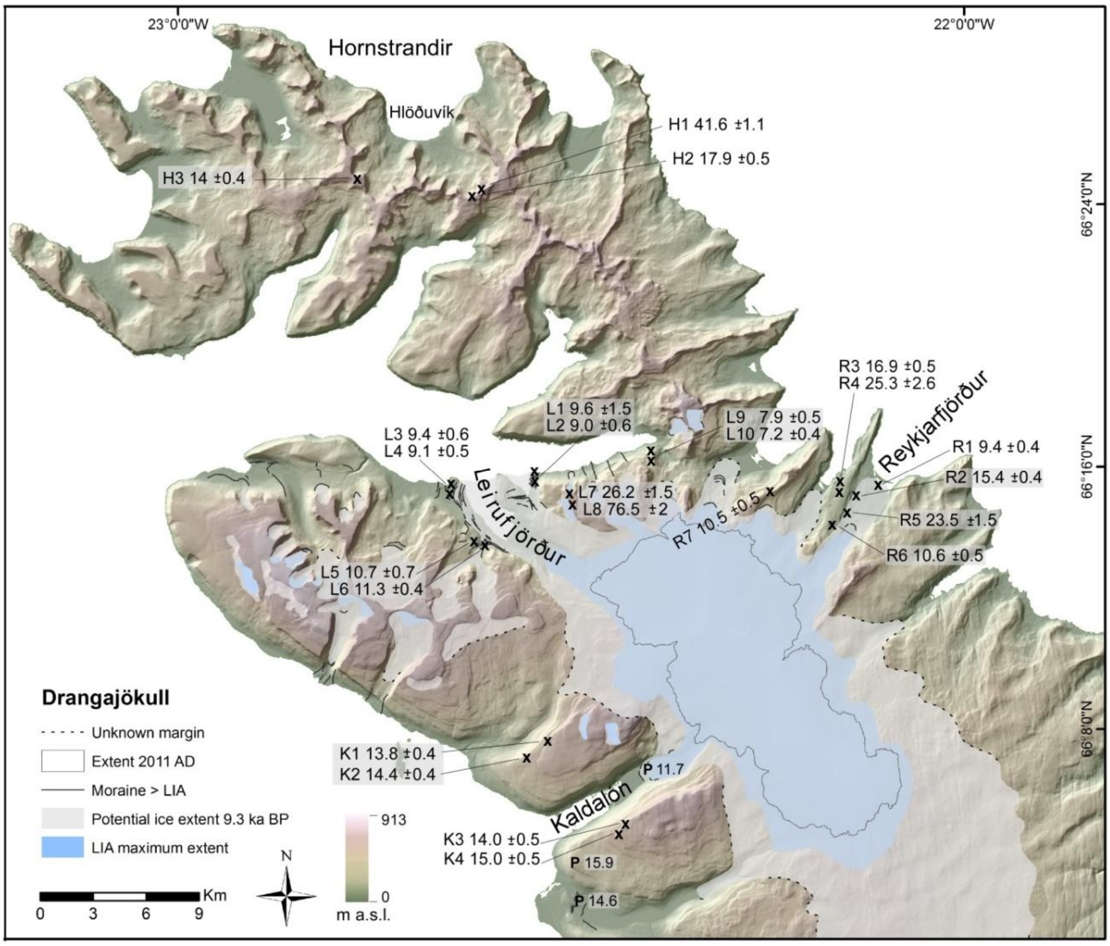

Fig. 4. Map overview of sample locations and ^{36}Cl ages with 1σ AMS errors. All ages are given in thousands years (ka). Samples from this study are marked with x, earlier average ages from Principato et al. (2006) are marked with P. Three size scenarios of the Drangajökull ice cap are shown.

4.1 ^{36}Cl exposure ages

4.1.1 Reykjarfjörður – northeast of Drangajökull

Seven samples were dated from the Reykjarfjörður area northeast of the ice cap. R1-R6 were sampled on a transect from the coast towards the LIA maximum end moraine which is located centrally in the valley about 3.5 km from the coast (Fig. 4). The individual ages are scattered and challenging to interpret (Fig. 5). The youngest sample, R1, yielding 9.4 ± 0.4 ka is from an ice sculpted bedrock hill 10 m a.s.l. located centrally in the valley about 500 m from the present beach. Another bedrock sample R2, 15.4 ± 0.6 ka old, situated about 500 m further into the valley. Next were two samples, R3 giving 16.9 ± 0.8 ka and R4 giving 25.3 ± 2.6 ka, both from ice-sculpted bedrock with striations located on a 140 m a.s.l. at the ridge between Reykjarfjörður and the neighbouring valley, Þaralátursfjörður (Fig.1). The fifth sample, R5, giving 23.5 ± 1.5 ka is from an about 30 m high glacially sculpted bedrock hill at the valley floor about 2.6 km from the coast. All these bedrock samples come from moderately weathered surfaces where striations can be recognized. Sample R6, giving 10.6 ± 0.5 ka, is from an erratic boulder located 130 m a.s.l. on the ridge between Reykjarfjörður and Þaralátursfjörður, about 3.3 km from the coast and approximately 200-300 m distal to the LIA maximum end moraine. Finally, an erratic boulder, sample R7, was dated to 10.5 ± 0.5 ka, on the 440 m high mountain pass, Svartaskarð, about 3 km to the northeast from the present glacier margin, between the valleys Furufjörður and Þaralátursfjörður (Fig. 4).

4.1.2 Leirufjörður – north of Drangajökull

Ten samples were dated from the Leirufjörður area; six moraine boulders from the lowland of the Leirufjörður valley, two erratic boulders from the plateau mountain Leirufjall, and finally two erratic boulders from the mountain pass Skorarheiði approximately 3.5 km north of the present glacier margin (Fig. 4). A system of 10-12 moraine ridges were recently mapped at the mouth of the fjord (Principato, 2008; Brynjólfsson et al., 2014). Two samples L1 and L2, giving 9.6 ± 1.6 ka and 9.0 ± 0.6 ka, were obtained from boulders on the outermost moraines on the east side of the fjord. Two samples L3 and L4, giving 9.4 ± 0.6 ka and 9.1 ± 0.5 ka were obtained from boulders on the moraines on the west side of the fjord. These four samples from the outermost moraine system yield an average age of 9.3 ± 0.8 ka. The moraines located proximal to the outermost moraine are younger and remain undated. These four dates from the moraine ridges suggest formation during a glacier advance or standstill at around 9.3 ka BP. The moraines inside the 9.3 ka moraine formed during subsequent advances or standstill of the glaciers.

Two samples L5 and L6, 10.7 ± 0.7 ka old and 11.25 ± 0.4 ka old, were obtained from a lateral moraine located about 130 m a.s.l. in the mouth of the tributary valley Dynjandisdalur (Fig. 4). This age correlates well with a more extensive and calving Leirufjörður outlet glacier before 10.2 ka BP (Andrews et al., 2002; Geirsdóttir et al., 2002).

Two erratic boulders, samples L7 and L8, were dated to 26.2 ± 1.5 ka and 76.5 ± 1.99 ka, respectively. They are located about 560 m a.s.l., approximately 5 km beyond the present ice cap on the plateau mountain Leirufjall (Fig. 4; Table 1). They suggest ice free conditions at Leirufjall by 26 ka BP, i.e. during the LGM. To the contrary, we obtained surprisingly young ages, 7.2 ± 0.6 ka and 7.9 ± 0.7 ka, from the two erratic boulders, sample L9 and L10, located about 330 m a.s.l. in the mountain pass Skorarheiði. Despite the consistency of those two samples, they are not considered to represent the deglaciation age of that locality. Because they do not agree with other samples in this study and their topographical and spatial setting would demand very large extensional anomaly of the glacier at this locality compared to the glacier extent suggested by the other exposure ages.

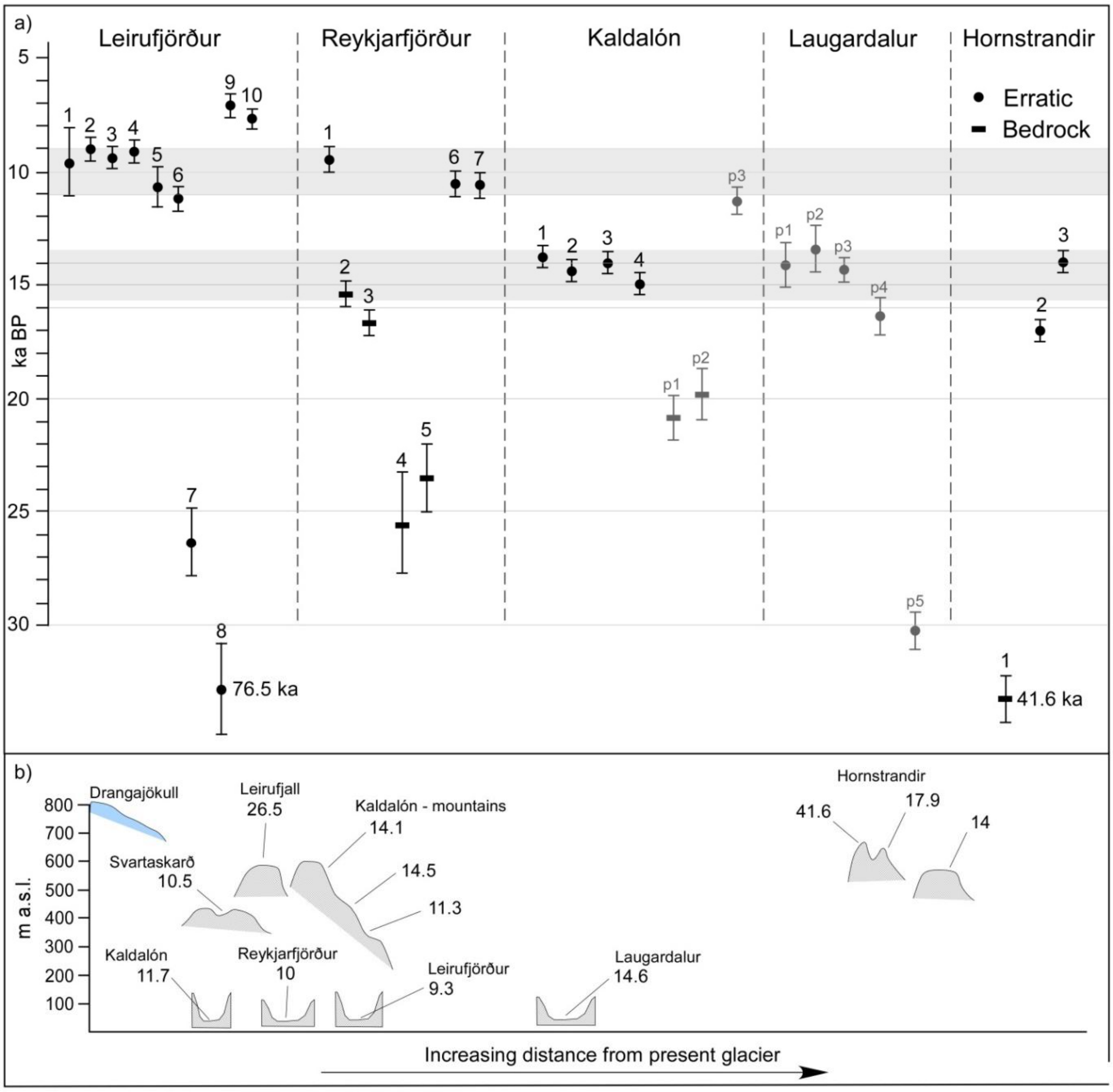

Fig. 5. a) Summary of the obtained cosmogenic exposure ^{36}Cl ages from the Drangajökull area. Erratic and moraine boulder samples are indicated by circles and bedrock samples with bars. For comparison previous exposure datings from the area (Principato et al., 2006) are shown in grey and marked with p. Error bars of 1σ represent analytical uncertainty. b) Exposure ages (ka BP) plotted against elevation and relative distance from the present ice cap margin.

4.1.3 Kaldalón – southwest of Drangajökull

Four erratic boulders located 6-8 km to the west of the present glacier margin on the plateaux on the north and south side of Kaldalón were dated (Fig. 4). Erratic boulders rarely occur above 400 m a.s.l. on the north side of the valley. The sample K1 located 650 m a.s.l. yielded 13.8 ± 0.4 ka, and sample K2 located 530 m a.s.l. yielded 14.4 ± 0.4 ka. On the south side, the erratic boulders were also rare and not observed at all above 500 m a.s.l. The sample K3 located 450 m a.s.l. was dated to 14.0 ± 0.5 ka and the sample K4 from about 400 m a.s.l. yielded 15.0 ± 0.5 ka (Fig. 5). The results suggest exposure age of the mountains on each side of Kaldalón about 14-15 ka BP. This is in good agreement with the collapse of the shelf based parts of the IIS about 15 ka BP and subsequent glacier calving in the fjords of Ísafjarðardjúp (Andrews et al. 2000, 2002; Geirsdóttir et al., 2002).

4.1.4 Hornstrandir

Two erratic boulders and one bedrock sample were dated from the Hornstrandir area (Figs 4 and 5). The bedrock sample H1 yielded 41.6 ± 1.1 ka BP and is located 675 m a.s.l. near the summit of the mountain Jökulhorn above Hlöðuvík (Figs 2d and 4). The bedrock is affected by local weathering, and scattered tors are located near the summit. The sample H2, 17.9 ± 0.5 ka is located 630 m a.s.l. on a small plateau near the summit of Jökulhorn along with several other erratic boulders. The sample H3 is a 14 ± 0.4 ka old erratic boulder located 585 m a.s.l. on the plateau mountain Fannalágarfjall (Fig. 4). Only very few erratic boulders were observed at Fannalágarfjall. All of them were almost without any fractures and very hard to sample by hammer and chisel.

The summit areas and the plateaux are characterised by local weathering and periglacial features, such as blockfields, polygons, sorted stripes and frost shattered bedrock. In these areas, we found no indications of recent glaciation except the few erratic boulders superimposed on the blockfield and the bedrock.

Table 1. Analytical data and input data for the Excel spread sheet age calculation (Schimmelpfenning et al., 2009). Bold, underscored numbers indicate the final age obtained after evaluation of all affecting factors, i.e. topographic shielding, snow shielding, scaling factors due to elevation and latitude, and surface erosion/weathering of the samples.

Ages according to different production rates, topography/snow shielding, scaling factors anderosion (e=erosion; mm/ka)

Sample	PRIME-lab	m a.s.l.	Type	Northing	Easting	Thick cm	Shielding topography	Shielding snow	Scaling factor	^{36}Cl ppm	^{36}Cl atoms/g	AMS ^{36}Cl error ± %	Licciardi et al. 2008 No correct.	Licciardi et al. 2008 e = 0	Licciardi et al. 2008 e = 2	Licciardi et al. 2008 e = 3	Licciardi et al. 2008 e = 5	Licciardi et al. 2008 e = 10	Stone et al. 1996; e = 2	Philips et al. 2001; e = 2
L1	201300697	33	erratic	7348189	432078	3,5	0,988	0,98	1,04	28,7	66979	16,3	9,69	9,57	**9,58**	9,6	9,63	9,75	10,8	8,4
L2	201300698	38	erratic	7348232	432067	3,5	0,988	0,98	1,03	25,3	59448	6,3	9,03	9,02	**9,04**	9,06	9,09	9,22	10,04	8,04
L3	201300699	31	erratic	7347385	427764	3,5	1,000	0,98	1,03	17,3	58001	6,3	9,42	9,29	**9,35**	9,38	9,45	9,34	10,43	8,27
L4	201300700	31	erratic	7347461	427776	3,5	1,000	0,98	1,03	21,6	74104	5,9	9,21	9,09	**9,12**	9,14	9,19	9,12	10,16	8,09
L5	201300704	135	erratic	7344384	429314	3,5	0,996	0,87	1,13	15	65541	6,9	10,67	10,62	**10,68**	10,71	10,79	11,05	11,88	9,48
L6	201402039	130	erratic	7344384	429314	3,5	0,996	0,87	1,13	**2,6**	62000	3,5	11,04	11,1	**11,25**	11,35	11,53	12,04	13,58	10,54
L7	201300703	565	erratic	7346698	434454	3,5	1,000	0,89	1,65	16,8	210641	5,6	37,92	25,03	25,35	25,6	**26,2**	28,4	29,12	23,23
L8	201402040	570	erratic	7346723	434377	3,5	1,000	0,89	1,67	5,6	451971	2,6	85,7	59,9	64,9	68,1	**76,5**	126,5	73,7	56,4
L9	201300701	310	erratic	7349334	438757	3,5	0,998	0,97	1,32	23,8	67148	6,4	10,01	7,81	7,83	7,84	**7,87**	7,97	8,69	6,96
L10	201300702	310	erratic	7349334	438757	3,5	0,998	0,97	1,32	17,3	56635	6,1	9,21	7,15	7,18	7,19	**7,23**	7,34	8,06	6,4
R1	201104285	10	bedrock	7348296	451558	3,5	0,999	0,98	1,01	10,3	58255	4,4	9,14	9,22	9,3	9,35	**9,44**	9,7	10,41	8,2
R2	201104280	35	bedrock	7348281	450744	3,5	1,000	0,97	1,03	14,8	94478	3,6	15,32	15,25	**15,41**	15,51	15,74	16,44	17,13	13,7
R3	201104283	140	bedrock	7348284	449665	3,5	1,000	0,97	1,13	10,4	110413	3,2	17,79	16,13	16,4	16,55	**16,85**	17,82	18,3	14,5
R4	201104284	140	bedrock	7348284	449665	3,5	1,000	0,97	1,13	12,3	157745	10,1	26,23	23,75	24,3	24,6	**25,34**	27,7	27,13	21,48
R5	201104281	30	bedrock	7346746	450069	3,5	0,999	0,97	1,02	13,4	145077	6,3	22,81	22,98	**23,48**	23,75	24,4	26,6	25,85	20,85
R6	201104282	130	erratic	7345670	449390	3,5	0,999	0,97	1,13	10,4	69700	3,8	9,25	10,5	**10,6**	10,7	10,8	11,2	12,02	9,35
R7	201104286	440	bedrock	7347672	445341	2,5	1,000	0,86	1,49	3,4	79550	4,6	13,36	10,18	10,32	10,38	**10,53**	10,92	11,63	9,04
K1	201403166	650	erratic	7335064	436163	3,5	1,000	0,89	1,78	4,1	118601	2,7	21,43	13,19	13,41	13,53	**13,75**	14,44	15,13	11,45
K2	201403167	530	erratic	7332893	433531	3,5	1,000	0,97	1,60	7,3	130381	2,7	22,14	14,1	14,3	**14,42**	14,7	15,45	16,1	12,25
K3	201403168	450	erratic	7327963	437013	3,5	1,000	0,89	1,50	3,6	106070	3,8	18,57	13,63	13,88	**14**	14,27	15,1	15,68	11,92
K4	201403169	400	erratic	7327909	436732	3,5	0,977	0,89	1,43	4,5	109888	3,6	18,55	14,6	14,88	**15,02**	15,32	16,18	16,78	12,71
H1	201403170	675	bedrock	7364650	428914	3,5	1,000	0,89	1,82	7,2	343022	2,7	62,19	36,67	38,32	39,31	**41,63**	50,25	43,29	33,51
H2	201403171	630	erratic	7364425	428559	3,5	1,000	0,89	1,76	5,1	154700	2,6	27,29	16,94	17,3	17,49	**17,9**	19,1	19,5	14,78
H3	201403172	585	erratic	7365235	420978	3,5	1,000	0,89	1,68	15,8	120645	2,8	20,96	13,62	13,76	13,85	**14**	14,58	15,3	11,97

4.2 Sedimentological logs from Reykjarfjörður

Six 1-5 m high river cut sections, section 1-6, were logged and described along the main river in Reykjarfjörður (Figs 6-8). The sedimentary facies are described and interpreted in Table 2. The sections are located centrally in the valley about 2-3 km from the coast (Fig. 9). Over a 4 km distance along the river, from the coast to the LIA maximum moraine, the valley floor is a completely flat and mostly vegetated outwash plain. An about 5-10 m high rock ledge, which the river has cut through, interrupts the flat valley floor approximately 2 km from the coast. To each side of the ledge, rock hillocks and ice-sculpted bedrock occur. Their surfaces are partly covered with till and vegetation stretching to the mountain slopes on each side of the valley (Brynjólfsson et al. 2014). Proximal to the rock ledge, the flat outwash plain becomes narrower and less vegetated as it continues further up-valley.

Section 1 (Fig. 6) cuts through an un-dated end-moraine about 1 km inside the rock ledge (Fig. 9; Brynjólfsson et al. 2014) whereas sections 2-6 (Figs 6, 7 and 9) are located along the river 50-300 m inside the rock ledge. Six different sedimentary facies were identified in the sections (Table 2). Unfortunately, no datable material was observed.

Table 2. Description and interpretation of sedimentary facies in Reykjarfjörður. The given references apply to interpretation of similar sediments.

Facies	Location and description	Interpretation	Reference
1	Sections 1, 2, 3, 4, 5, and 6 Up to 2 m thick. However, the lower contact was nowhere observed. 0.2-2 cm brown and grey coloured clay to silt laminae. Horizontally laminated in section 2, 3, and 4. The laminae in section 3 dip 6-9° towards 65°N. Occasionally interrupted by fine to medium grained sand layers and lenses. Heavily deformed and thrusted in section 1. Ripples and soft sediment deformation occur in the lower part of section 4 and 5. Dropstones and load structures occur in sections 5 and 6. Cohesive, often hard to excavate.	Low energy environment. Horizontal laminae represent lacustrine or lagoonal conditions. Dropstones suggest glacier calving, occasional loading structure indicate temporarily high sedimentation rate. Ripples formed by bottom currents in shallower parts. Most deformation is furthest up-valley, caused by a glacier advancing into the lagoon. The intact horizontal laminae furthest down-valley has not been affected.	Benn and Evans, 2010; Evans and Benn, 2004.
	Sections 1, 2, 4, 5, and 6 Typically massive fine to medium sand.	Fluvial sand. Low to moderate discharge or bottom currents allows	Evans and Benn, 2004.

2	Deformed in section 1 and ripple cross laminated in section 2, and partly horizontally laminated in section 6.	formation of ripple cross laminae and horizontal laminae. Deformed in section 1 by an advancing glacier into the lagoon.	
3	Sections 1, 2, 4, and 5 Medium to coarse, massive gravel, except in section 5 where fine gravel.	Fluvially deposited gravel, the massive and relatively coarse appearance indicates high water discharge. In section 1, unit 3 indicate that the lagoon gradually became an outwash delta before the glacier advance. In section 2 and 4 unit 3 is part of a delta built into the lagoon	Benn and Evans, 2010.
4	Section 4 Layer of sub-angular to sub-rounded boulders, up to 1.2 m in diameter.	Fluvial deposited under high discharge conditions, perhaps flood event.	Maizels, 1997; Benn and Evans, 2010.
5	Sections 3 and 6 20-40 cm thick reddish-brownish soil, plenty of roots, vegetation on top. Silt to sand in grain size.	Aeolian, fine fluvial sediments and organic material built up to soil layer.	
6	Section 1 1-2 m thick diamict. Rich in boulders in the upper part. Matrix-supported, medium grained sandy matrix. Sub-angular to sub-rounded clasts. Thinning down-valley.	Subglacial till deposited near the margin of a glacier advancing over a proglacial fluvial plain. The contact between the till and the gravel below (unit 3) is gradual.	Evans and Benn, 2004; Benn and Evans, 2010.

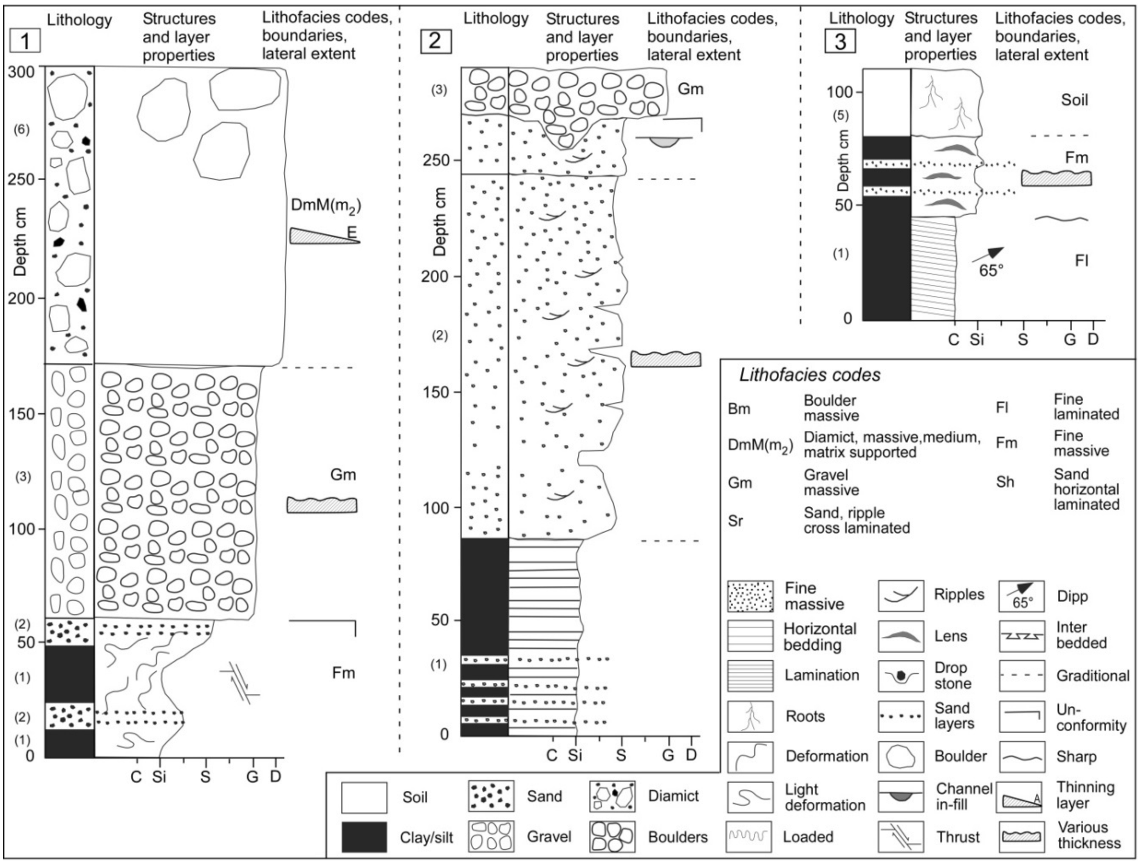

Fig. 6. Sedimentological logs of sections 1-3 in Reykjarfjörður. The section number is indicated in a box at the top of each section, and facies numbers, described in Table 2, are in brackets to left of the log. The logs follow the data chart of Krüger and Kjær (1999).

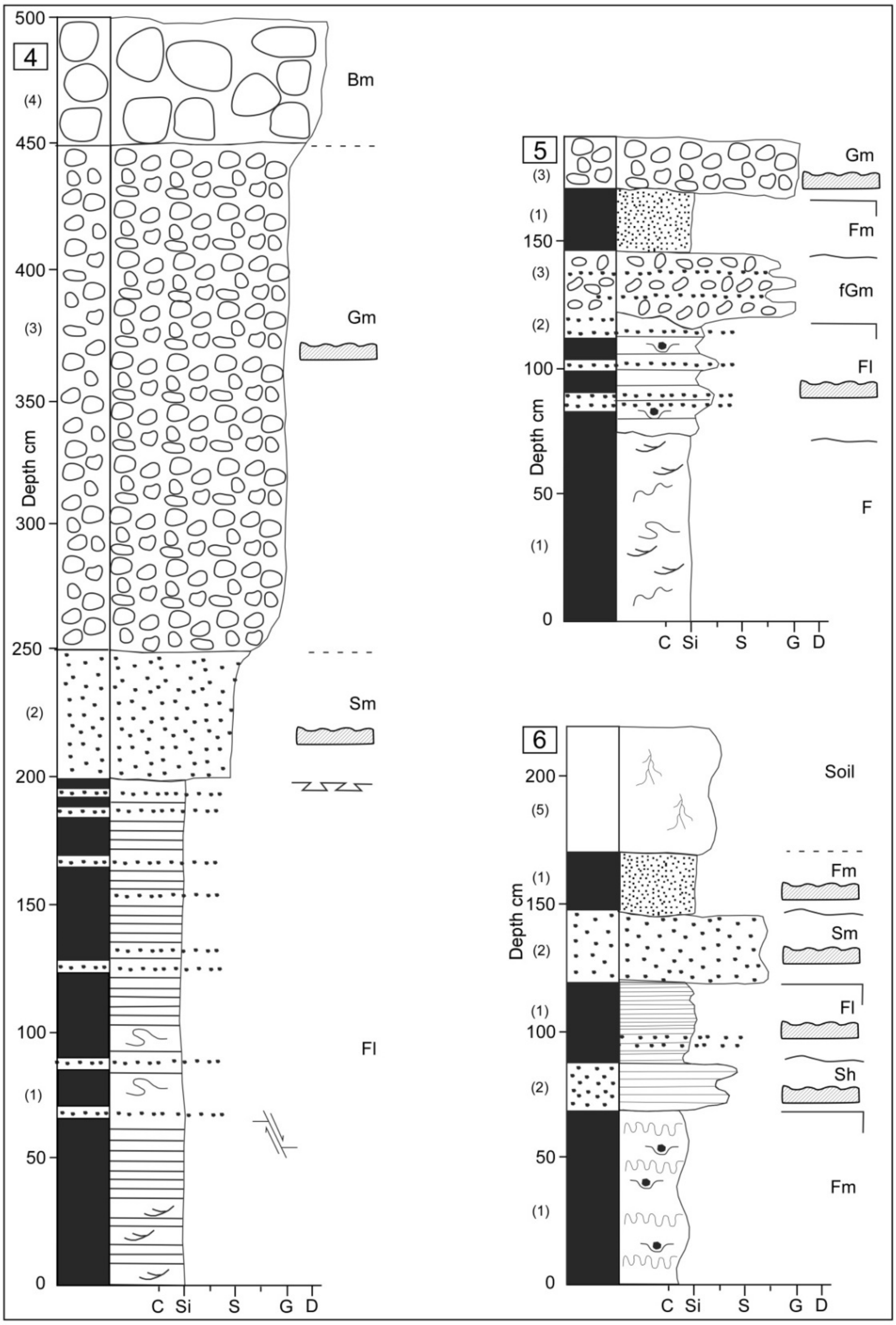

Fig. 7. Logs of sections 4-6, section numbers are indicated in boxes at the top of each log and the facies numbers, described in Table 2, are in brackets to the left of the logs. See Figure 6 for legend.

During deglaciation, the glacier retreated inside the rock ledge and left a depression which filled up with meltwater. The water level was controlled by the elevation of the rock ledge, which gradually lowered as the river eroded its way through it.

According to the evidence from the lacustrine sediments in the sections 1-6, a glacial lagoon formed in a topographic depression within the rock ledge. Till, gravel, laminated and deformed fine sediments with dropstones were observed in the six river-cut sections. The fine sediments, their structures and the dropstones indicate deposition in a glaciolacustrine environment. We consider that a fjord-setting can be excluded because raised beaches have only been observed approximately 5 m above present sea level in Reykjarfjörður, which is a few meters below the rock ledge (Fig. 8; Principato, 2008; Brynjólfsson et al., 2014). Furthermore erratic boulders and till cover the rock ledge and parts of the valley floor (10-30 m a.s.l) adjacent to the reconstructed lagoon (Fig. 9).

We consider that the strong deformation of facies 1 below the till (facies 6) in section 1 demonstrates an outlet glacier advancing into the lagoon after 9.4 ka BP. The dropstones and more gentle deformation of facies 1 in sections 5 and 6 confirm the proximity of a calving outlet glacier further up-valley. Layers and lenses of silt and sand in facies 1 could represent a higher energy sedimentary environment related to pulses of meltwater or floods coming into the lagoon, perhaps related to a glacier advance. On the other hand, the well-preserved horizontal laminae near the rock ledge are considered to exclude any glacier advance further down-valley after the deposition of the horizontally laminated sediments.

The depth and size of the lagoon is unknown and the base of facies 1 was not observed anywhere. Assuming calving ice-bergs floating around, it was at least a few meters deep. The reconstructed up-valley extent (Fig. 9) is limited by the observation of facies1 in section 1. There, about 1 m thick fluvial gravel (facies 3) occurs between the till (facies 6) and the fines (facies 1). This indicates that at this location the lagoon had already filled up with fine and fluvial sediments and gradually became an proglacial outwash plain before the glacier advanced and formed an undated moraine about 500 m beyond the LIA maximum moraine.

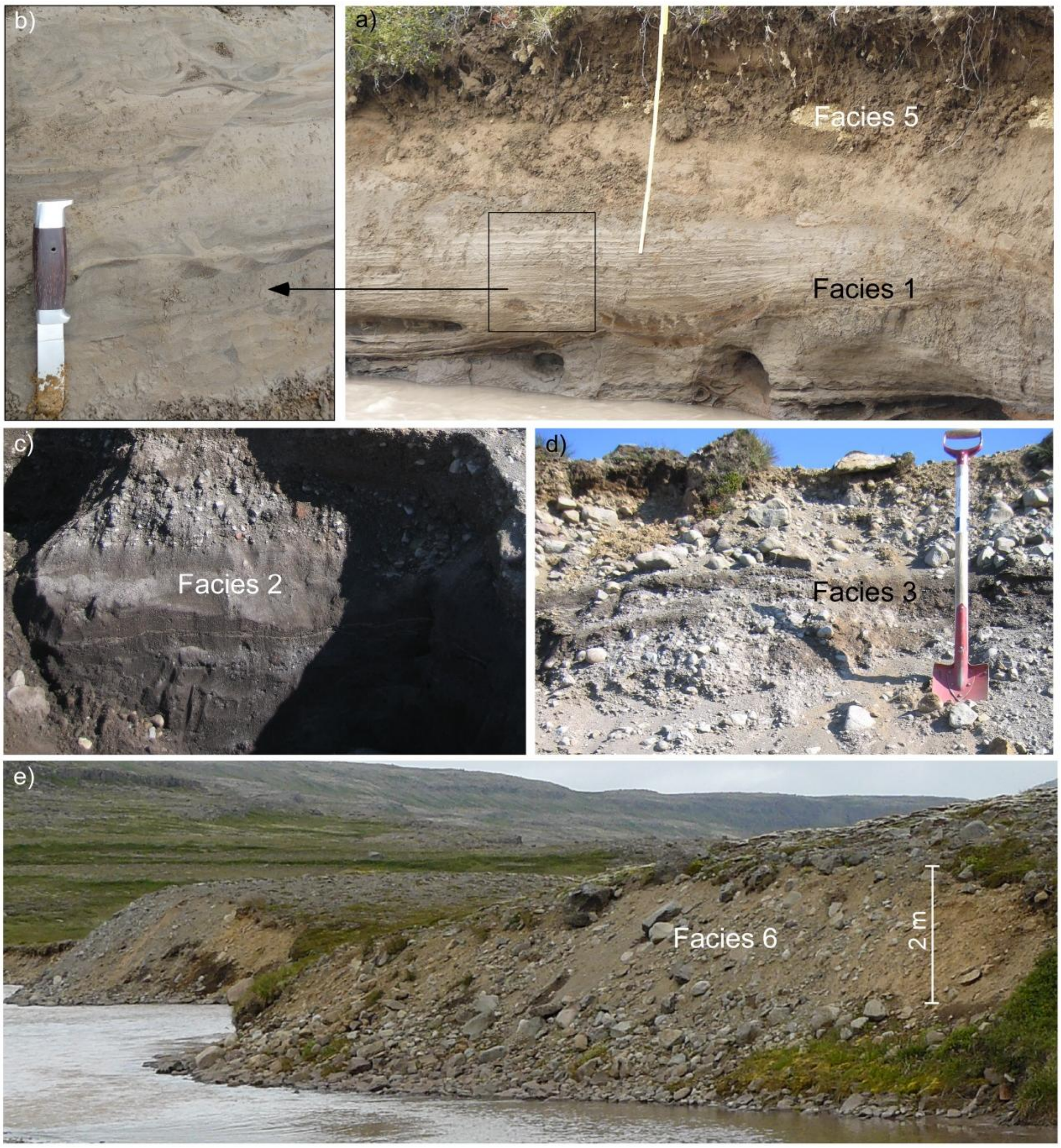

Fig. 8. Examples of the sedimentary facies identified in the sections.

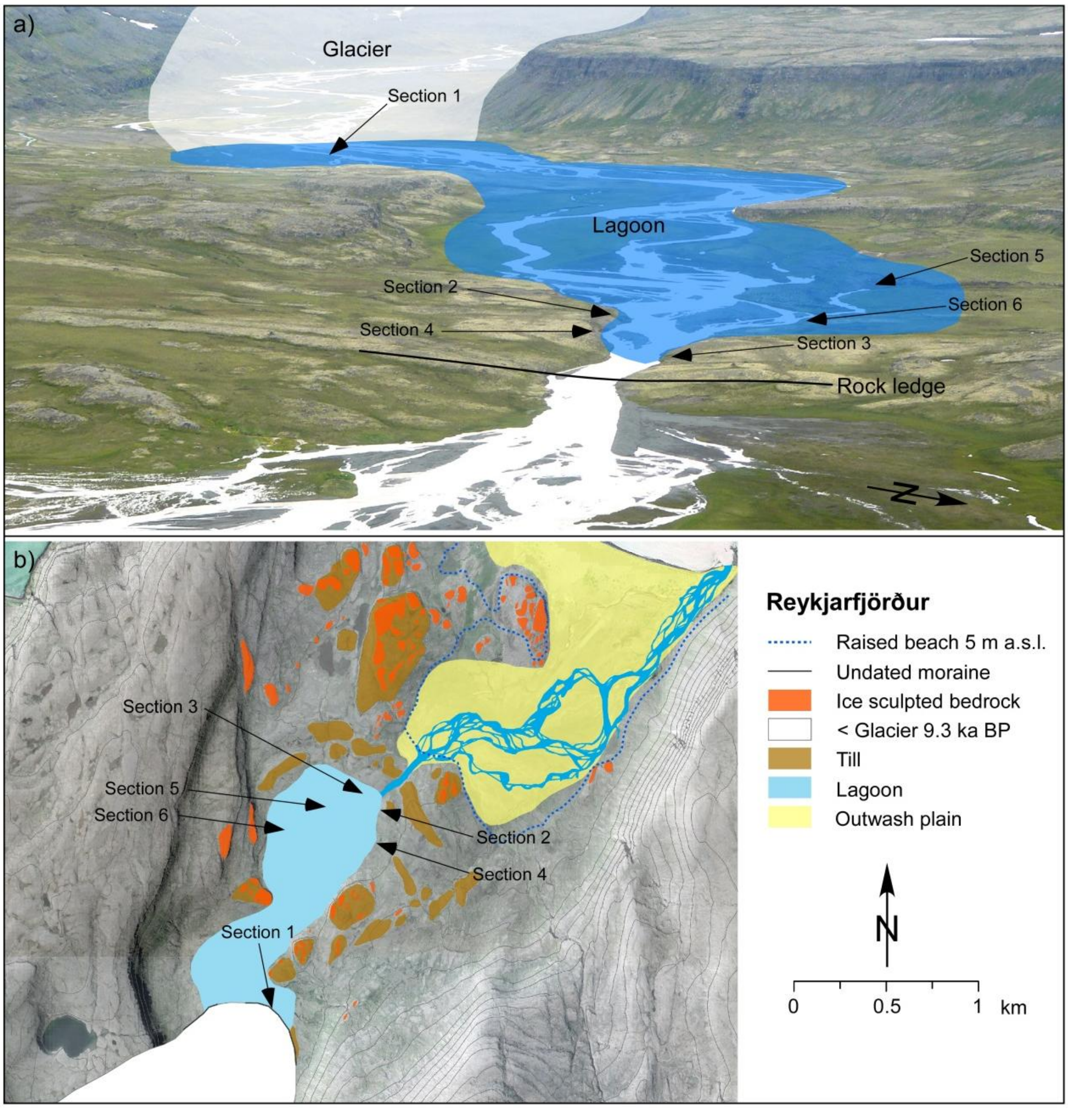

Fig. 9. Environmental reconstruction of Reykjarfjörður following the initial deglaciation of the valley, based on the sediment studies.

5. Discussion and interpretation

5.1 ^{36}Cl exposure ages of the study areas

5.1.1 Reykjarfjörður – northeast of Drangajökull

The scattered ages of the Reykjarfjörður samples are challenging to interpret and demonstrate the importance of collecting a group of samples from specific landforms or localities (Balco et al., 2011; Stroeven et al., 2011). The three youngest samples, R1 9.4 ±

0.4 ka, R6 10.6 ± 0.5 ka and R7 10.5 ± 0.5 ka correlate with ages obtained in Leirufjörður (sample number L1-L6). Those three samples are considered to represent the deglaciation age. They suggest deglaciation of Reykjarfjörður between 10.6 ka and 9.4 ka BP (Fig. 4).
Snow shielding, sediment and vegetation cover could contribute to younger exposure ages than the actual deglaciation age of a locality (Balco, 2011; Stroeven et al., 2011). However, shielding factors are considered negligible for the Reykjarfjörður localities, except potential snow cover for sample R7 in the 440 m high mountain pass, Svartaskarð, which we corrected for in our calculations.
The 15-25 ka old bedrock samples, R2, R3, R4 and R5 from Reykjarfjörður are best explained by ^{36}Cl inheritance. Bedrock exposure ages occasionally suffer from nuclide inheritance. Striated and ice sculpted bedrock do not necessarily confirm a sufficient erosion to reset the pre-glaciation nuclide inheritance within the bedrock, although it clearly indicates sub-glacial erosion (Stroeven et al., 2002; Briner et al., 2005, 2006; Phillips et al., 2006). To remove any inheritance, the bedrock has to be eroded by >2 m (Ivy-Ochs and Kober, 2008; Balco, 2011). Therefore, our old bedrock samples might indicate less than 2 m subglacial erosion of the bedrock during the last glaciation. That agrees with observations of a generally thin and non-coherent cover of glacial sediments around the Drangajökull ice cap, probably indicating limited sub-glacial erosion and sediment production (Brynjólfsson et al., 2014).
Samples for exposure dating have been noticed to commonly contain inherited nuclides and indicate limited sub-glacial erosion proximal to the ice sheet margins of the northwest Svalbard-Barents Sea ice sheet and the Laurentide ice sheet in eastern Arctic Canada. Young erratic boulders superimposed on older bedrock and un-scoured terrain, indicate transportation from thicker, warm based inland ice and deposition by lateral flow yielded from internal ice deformation of slow cold-based ice in marginal and coastal areas (Briner et al., 2005; Hormes et al., 2011; Gjermundsen et al., 2013). Considering that the eastern Vestfirðir peninsula was a marginal area of the last IIS, thin and slow, partly cold based, ice could explain a relatively little subglacial erosion in the area during the last glaciation.

5.1.2 Leirufjörður – northwest of Drangajökull

The group of moraines at the mouth of Leirufjörður indicate a fluctuating ice margin during the deglaciation of the fjords and valleys in the Vestfirðir peninsula. The samples L1-L4 from the outermost moraine yielded an average age of about 9.3 ± 0.8 ka (Table 1). A few of the outermost moraines in Leirufjörður extend into the fjord indicating glacier calving at that time (Fig. 4). Relatively sea level that was at least few meters lower than at present (Norðdahl and Pétursson, 2005) could have prevented the glacier from calving.
The high ages of 26.2 ± 1.5 ka and 76.5 ± 1.99 ka, of the two erratic samples L7 and L8 superimposed on the blockfield, argues for cold based ice conditions on the plateau

mountain Leirufjall. The 26 ka exposure age of the sample L7 can be considered as minimum deglaciation age of Leirufjall. That could suggest a glaciation maximum about 5-6 ka earlier than previously considered (Ingólfsson, 1991; Norðdahl, 1991). However, if actual erosion and snow cover was more extensive than we corrected for, the calculated exposure age is too low and the deglaciation time would be earlier. The 76.5 ka age of sample L8 must be explained by nuclide inheritance from earlier exposure.

We obtained surprisingly low ages of 7.2 ± 0.6 and 7.9 ± 0.7 ka (samples number L9 and L10) from two erratic boulders located in the 330 m high mountain pass, Skorarheiði, north of the ice cap (Fig. 1). The results do not correlate with earlier deglaciation reconstructions for Drangajökull nor the IIS in general (Principato et al., 2006; Norðdahl et al., 2008; Geirsdóttir et al., 2009). Furthermore, they do not agree with other samples in this study (Figs 4 and 5, Table 1). Thus, despite consistent age of this pair of boulders, we do not consider them to represent the deglaciation time of this locality.

5.1.3. Kaldalón – southwest of Drangajökull

The samples K1-K4 located 400 – 650 m a.s.l. on the mountains around Kaldalón yielded a 14.3 ka average exposure age. These ages corroborate with a 11.3 ka old erratic located 352 m a.s.l on the outermost part of the southern mountain in Kaldalón (Fig. 5; Principato et al., 2006). The ages demonstrate ice thinning and initial deglaciation over the higher areas around Kaldalón and subsequent deglaciation of lower areas. However, two bedrock samples also provided by Principato et al. (2006) located about 350 m a.s.l. at the outermost part of the mountain yielded 19.9 ka and 20.9 ka (Fig. 5). Similar to the Reykjarfjörður bedrock samples, we consider they most likely indicate insufficient erosion of the bedrock, rather than the exposure age. The samples from Principato et al. (2006) were processed in the early 2000s, before the ^{36}Cl cosmogenic production rate for Iceland by Licciardi et al. (2008) became available. A re-calculation of the ages with the local production rate and the age calculator of Schimmelpfenning et al. (2009) used in this study would most likely yield somewhat different results. Unfortunately, we have not been able to obtain all the raw ^{36}Cl AMS and lab data needed to re-calculate ages of Principato et al. (2006).

5.1.4. Hornstrandir

The high age of the bedrock sample H1 (41.6 ± 1.1 ka) is most likely due to nuclide inheritance, preserved from earlier exposure under non-erosive ice during the last glaciation, rather than representing the actual exposure age of the mountain. This interpretation is supported by the younger erratic samples H2 (17.9 ka) and H3 (14.0 ka), which indicate ice cover over plateaux and mountains in Hornstrandir prior to or until the collapse of the shelf based part of the IIS about 15 ka BP (Andrews et al., 2002; Geirsdóttir et al., 2002; Ingólfsson et al., 2010). However, the peak on which H1 was sampled is 45-90 m higher than samples H2-H3, and therefore we cannot exclude that the peak protruded through the ice at LGM. No glacial sediments were observed at these high-altitude localities. Considerable local weathering and no indications of recent glaciation except the occasional erratic boulders together with the 41.6 ka old bedrock sample H1 suggest cold-based ice conditions over the Hornstrandir mountains.

5.1.5. Were the end moraines in Leirufjörður formed during a 9.3 ka climatic event?

A 9.3 ka average age of the outermost moraine at the mouth of Leirufjörður, correspond with formation of moraines in Greenland and Baffin Island (Briner et al., 2009; Young et al., 2011, 2013), and glacier advances in Scandinavia at a similar time (Nesje, 2009). These advances have been correlated with the 9.3 ka meltwater pulse to the North Atlantic from Lake Agassiz (Briner et al., 2009; Nesje, 2009; Murton et al., 2010; Young et al., 2013). Furthermore, series of other meltwater pulses from Lake Agassiz and perhaps Siberian lakes occurred approximately from 9.5 ka to 8.2 ka ago (Fleitmann et al., 2008). Glacier advances in North America and Europe have been correlated to cold reversals associated with these meltwater pulses (Kleiven et al., 2008; Solomina et al., 2015) A set of high-resolution palaeoclimate records from across the Northern Hemisphere indicates a widespread and significant climatic anomaly during the 9.3 ka and 8.2 ka events (Fleitmann et al., 2008; Nesje, 2009; Solomina et al., 2015).

Therefore, the 9.3 ka exposure age of the moraine in Leirufjörður suggest a formation by a re-advancing glacier as response to climatic cooling during the 9.3 ka event. To our knowledge, this is the first observation from Iceland to suggest glacier response related to the climatic deteriorations in the period 9.5-8.2 ka BP (Fleitmann et al., 2008). Furthermore, we consider that the formation of other recently mapped, but still undated moraines, located beyond the LIA maximum moraines in the eastern Vestfirðir peninsula could have formed during similar glacier re-advances in the period 9.5-8.2 ka BP (Brynjólfsson et al., 2014; Hole, 2015).

Simultaneously, a rapid deglaciation and ice caps of similar size as at present has been suggested in the central highlands of Iceland as early as 10 ka BP (Geirsdóttir et al., 2009; Larsen et al., 2012). Potentially, the 9.3 ka glacier re-advance was confined to the Vestfirðir peninsula. This could have been favoured by the geographical conditions of the Vestfirðir peninsula, which extends towards cooler surface waters near Greenland and is exposed to the open sea towards the east, north and west. Sea ice cover is expected to increase and regional temperature to drop during enhanced advection of cooler surface waters around the peninsula, potentially favouring glacier advances (Bergþórsson, 1969; Ogilvie, 1984; Eiríksson et al., 2000).

Climatic deterioration in Iceland related to the 8.2 ka event, and over a period about 8.5-8 ka BP, has been interpreted from marine and lacustrine proxies (Alley et al., 1997;

Eiríksson et al., 2000; Geirsdóttir et al., 2009; Larsen et al., 2012). Sediment cores from Lake Hvítárvatn, which receives meltwater from the Langjökull ice cap in central Iceland, indicate an environmental perturbation between 8.7 and 7.9 BP. Two approximately 200 year long periods, about 8.5 and 8.2 ka BP, were the coldest in this period. Signs of landscape destabilization and possible glacier activity in the catchment are seen in the sediment cores (Larsen et al., 2012). However, the absence of varved sediments in the cores suggests a smaller Langjökull than at present (Larsen et al., 2012). Geirsdóttir et al., (2009) pointed out the absence of any direct glaciological and geological record related to the 8.5 and 8.2 ka events in Iceland, and discussed that potential geological evidence had perhaps been overridden by subsequent glacier advances during the late Holocene. We suggest that the group of moraines in Leirufjörður demonstrate advances of the Drangajökull ice cap due to the meltwater forced climatic cooling events during the period 9.5-8.2 ka (Fleitmann et al., 2008; Nesje, 2009).

5.1.6. LGM to Early Holocene glacial history of NW Iceland

Our upland erratic boulders suggest that the ice sheet started thinning at least ~ 26 ka BP. Ice-free conditions on mountain Leirufjall by that time indicate that the LGM ice sheet over the Vestfirðir peninsula started thinning 5-6 ka earlier than previously suggested (Norðdahl, 1991; Ingólfsson, 1991; Ingólfsson and Norðdahl, 2001).

The erratic boulders at Leirufjall, in Hornstrandir and on the Kaldalón mountains, except L8 which was probably deposited during an earlier ice advance, were deposited during the Late Weichselian by an ice sheet with relatively thin and cold based sectors over the uplands (Fig. 10). The erratic boulders were transported laterally by internal ice deformation while the ice remained mostly inactive and frozen at its substratum, leaving the blockfield undisturbed over the plateaux and the high uplands in the eastern Vestfirðir peninsula (Sugden, 1977; Hjort et al., 1985; Briner et al., 2005; Davis et al., 2006; Landvik et al., 2012; Brynjólfsson et al., 2014). The erratic boulders are neither angular nor of the local blockfield lithology, therefore they must have been transported and deposited on the blockfields by glaciers. Their ages, except sample L8, suggest deposition during the Late Weichselian. Similar observations of young erratic boulders sitting on top of intact block fields surfaces have been interpreted as evidence of transport by non-erosive cold based branches of ice sheets in Arctic Canada, Scandinavia, and Svalbard (Briner et al., 2003, 2005, 2014; Fjellanger et al., 2006; Hormes et al., 2011; Landvik et al., 2012).

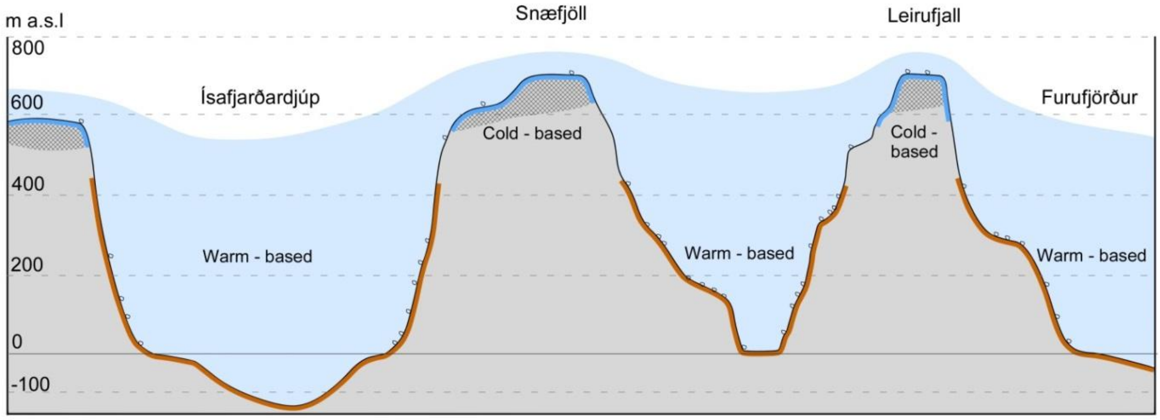

Fig. 10. A schematic cross-section of the Vestfirðir peninsula during the LGM. The glaciation was characterised by topographically confined fjord/valley-type ice streams or glaciers which flowed faster than the grounded ice domes over the mountains on either side. A thin, non-erosive and cold based ice, covered plateaux and some uplands in the eastern Vestfirðir peninsula, while warm-based ice filled valleys and fjords during the LGM.

It is not possible to exclude nuclide inheritance in the upland erratic boulders, and in case of inheritance, the exposure ages would be higher than the actual deglaciation time (Stroeven et al., 2011), which is demonstrated in the 76.5 ka old sample L8. If the mountains in eastern Vestfirðir remained ice free during the last glaciation, the erratic boulders would be suspected to include considerable amounts of ^{36}Cl nuclides. Thus, one would expect most of the erratic boulders to be of much higher ages than we obtained. Therefore, the upland erratic samples, L7, H2, H3 and K1-K4 are considered to represent the exposure age of their localities. On the other hand, ice sculpted and striated bedrock indicates warm-based ice, sliding over its substratum at lower altitudes in the fjords and valleys (Fig. 10). Such overall pattern of warm based, dynamic ice occupying lowland fjords and valleys, and thin cold based ice over plateaux and uplands suggests topographically confined fjord/valley-type glaciers which flowed faster than the grounded ice on either side (Fig. 10; Paterson, 1994; Briner et al., 2006) during the last glaciation of the Vestfirðir peninsula.

Applying ^{36}Cl exposure dating on the block field surfaces could provide important new data to the reconstruction of the last glaciation in Vestfirðir and Iceland in general. Blockfield boulders from Norway, Svalbard, and Arctic Canada have remained intact since before the last glaciation and yielded much higher exposure ages than the superimposed erratic boulders (Briner et al., 2003, 2014; Fjellanger et al., 2006; Hormes et al., 2011; Landvik et al., 2012).

We consider the 10.5 ka old erratic sample from the mountain pass Svartaskarð, and the c. 11 ka old moraine in valley Dynjandisdalur as indications of thick outlet glaciers occupying the fjords and valleys at that time. This agrees with the distinct IRD signal interpreted as evidence of calving glaciers in the fjords until 10.2 ka ago, and with sea level curves indicating substantial ice cover over the Vestfirðir peninsula at Younger Dryas time (Andrews et al., 2002; Geirsdóttir et al., 2002; Castañeda et al., 2004; Andresen et al., 2005; Norðdahl and Pétursson, 2005). It remains unresolved if the glacier was confined to the valley between 10.2 ka BP and 9.3 ka BP, or if it retreated considerably, as have been suggested by Castaneda et al. (2004) and Geirsdóttir et al. (2009).
Six undated moraines in Furufjörður beyond the LIA maximum extent (Fig. 4) indicate stepwise final deglaciation possibly at the same time as in Leirufjörður and Reykjarfjörður. Central Iceland and the larger peninsulas Tröllaskagi, Eastern Iceland and Snæfellsnes peninsula have been considered completely deglaciated, or hosting glaciers of similar size as at present when Leirufjörður and Reykjarfjörður were still occupied with outlet glaciers about 9.3 ka BP (Kaldal and Víkingsson, 1990; Stötter et al., 1999; Caseldine et al., 2003; Norðdahl and Pétursson, 2005; Larsen et al., 2012). Furthermore, new cosmogenic ages of moraines in the central Tröllaskagi peninsula suggest that some valley and cirque glaciers had already retreated to a similar size as at present about 15 ka BP (Palacios et al., 2015). Hence, a much more extensive late glacial ice cap was preserved over the eastern Vestfirðir peninsula than elsewhere in Iceland.

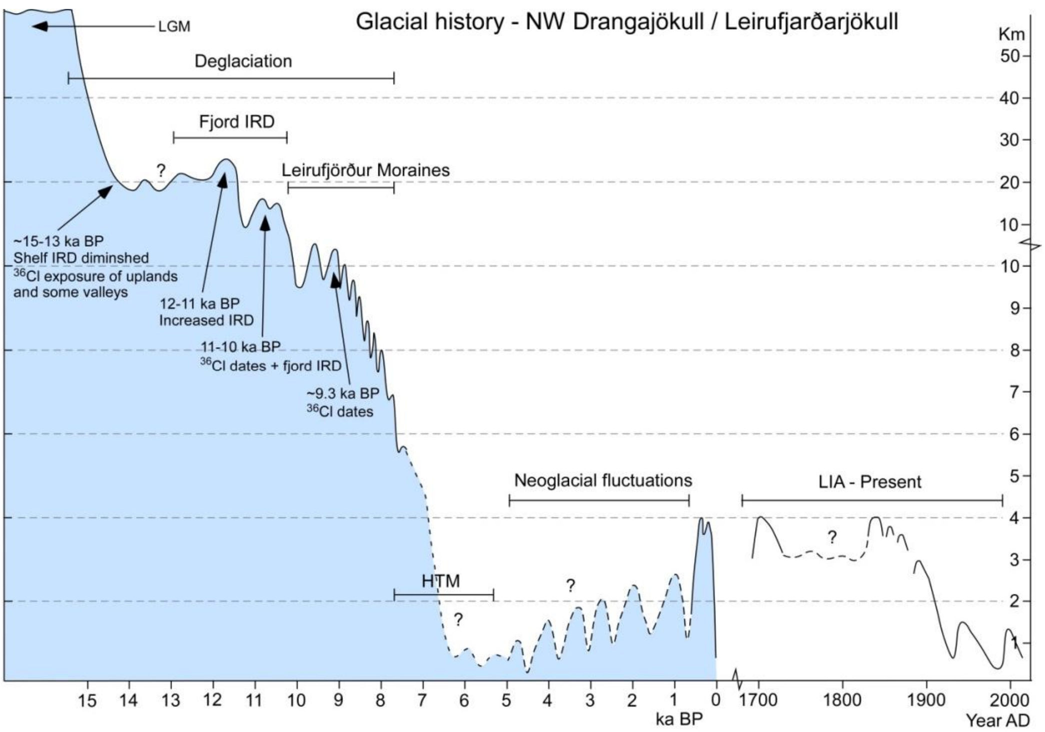

Fig. 11. Glacier changes of the north-western Drangajökull and later the Leirufjörður outlet glacier. Note that the axes (age horizontal, distance vertical) are truncated. The evolution 15-10 ka BP is based on exposure ages, marine sediment data and sea level curves (Andrews et al., 2002; Geirsdóttir et al. 2002; Norðdahl and Pétursson, 2005; Norðdahl et al., 2008). The frequent fluctuations around 9 ka BP are based on exposure dating of terrestrial moraines in Leirufjörður. The Neoglacial fluctuations are based on lacustrine sediment data from central Iceland (Geirsdóttir et al 2009; Larsen et al., 2012), and do not necessarily represent fluctuations of Drangajökull ice cap. The LIA fluctuations are from the recently reconstructed surge history of Leirufjarðarjökull (Brynjólfsson et al., 2015).

5.1.7. Deglaciation

The exposure age of sample L7 at Leirufjall indicates that the ice sheet deterioration started at least 26 ka BP. That precedes typically suggested LGM timing for Iceland by 5-6 ka (Ingólfsson, 1991; Norðdahl and Ingólfsson, et al., 1991; Ingólfsson et al., 2010), but correlates better with a general timing c. 26-20 ka BP of the LGM in the North Atlantic (Clark et al., 2009). Our exposure samples, L7, H2-H3 and K1-K4, from the plateaux and uplands indicate a deglaciation pattern where ice thinning preceded areal reduction of the Vestfirðir ice sheet. Shelf areas and fjords were stepwise deglaciated 15-10 ka BP, following the ice sheet break up about 15 ka BP (Fig. 11; Eiríksson et al., 2000; Andrews et al., 2000, 2002; Ingólfsson and Norðdahl, 2001; Geirsdóttir et al., 2002).

Glacier retreat was not synchronous between different inlets, valleys and fjords (Figs 4 and 5). This is well demonstrated by the contrasting exposure dates from the eastern Vestfirðir peninsula, 9.3 ka and 11 ka old moraines in Leirufjörður, c. 10 ka average exposure age in Reykjarfjörður, 11.7 ka old moraine in Kaldalón and a 14.6 ka old moraine in Laugardalur (Principato et al., 2006). Furthermore, a deglaciation age of 12 ka BP of central Vestfirðir was suggested from lacustrine sediments (Geirsdóttir et al., 2002). The 10.2 ka old Saksunarvatn tephra observed in several geological sections along the coasts of Hornstrandir also indicate ice-free conditions (Hjort et al, 1985; Principato et al., 2006; Hole, 2015).

After the last glaciers retreated on land by 10.2 ka BP (Andrews et al., 2000, 2002; Geirsdóttir et al., 2002), a stepwise deglaciation of the valleys that still hosted the major outlet glaciers of Drangajökull was going on about 9 ka BP (Fig. 11). However, the older moraines, 14.6 ka in Laugardalur and 11.7 ka in Kaldalón, demonstrate that some valleys in the area were already partly ice free or completely deglaciated 2-5 ka earlier (Figs 4 and 5; Principato et al., 2006).

6. Conclusions

We dated 24 bedrock, moraine and erratic boulder samples from the Vestfirðir peninsula with ^{36}Cl exposure dating. The new exposure ages combined with data from sedimentological logs, provide insight into the Weichselian to early Holocene glaciation history of eastern Vestfirðir. We conclude that:

- Bedrock and boulder samples from high altitudes yielding ages of 76.5 ka and 41.6 ka together with 14-26 ka old erratic boulders superimposed on blockfield characterised plateaux and uplands, indicate transport and deposition of the erratics by internal deformation of an ice sheet with relatively thin cold based, non-erosive ice sectors over plateaux and uplands in the eastern Vestfirðir peninsula during the last glaciation.

- Ice-sculpted bedrock and striations together with 9.3-11.5 ka old exposure dated samples indicate warm based glacier conditions in the lowland areas of fjords and valleys. However, the higher ages of bedrock samples in valley Reykjarfjörður, interpreted as nuclide inheritance, demonstrate subglacial erosion rates <2-3 m during the last deglaciation. The limited subglacial erosion corresponds with observations of a thin and patchy till cover around the ice cap.
- The 14-26 ka exposure samples, L7, H2-H3 and K1-K4, from the plateaux and uplands indicate a deglaciation pattern characterised by ice thinning before areal reduction of the Vestfirðir ice sheet, which subsequently was followed by deglaciation of shelf areas and fjords 14-15 ka BP.
- Glacier retreat was not synchronous between different valleys and fjords. While some valleys were partly deglaciated 11.7 and 14.0 ka BP (Principato et al., 2006). Leirufjörður and Reykjarfjörður were still occupied by glaciers extending to the sea at least until c. 9 ka BP.
- We suggest that the 9.3 ka old moraine in Leirufjörður formed by a glacier re-advance in response to a cooler climate forced by reduction in the Atlantic Meridional Overturning Circulation.
- Series of concentric moraines in Leirufjörður indicate stepwise deglaciation following formation of the 9.3 ka moraine. The moraines formed by glacier advances in response to the climatic deterioration events about 9.5-8.2 ka BP.
- A proglacial lagoon formed inside a rock ledge centrally in the Reykjarfjörður valley following the deglaciation about 9.4 ka BP. Dropstones and deformed lacustrine sediments in river cut sections demonstrate an outlet glacier calving and advancing into the lagoon.

- An extensive ice cap was preserved over the eastern Vestfirðir peninsula at least until c. 9 ka BP. It is very possible that the Drangajökull ice cap has been present throughout the Holocene.

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Appendix IV

Draft manuscript, to be submitted

Elevation and volume changes during glacier surges of outlets from the Drangajökull ice cap, northwest Iceland, 1994 to 2011

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Abstract

Surface elevation and volume changes of the Drangajökull surge-type glaciers, Reykjarfjarðarjökull and Leirufjarðarjökull, were studied by differencing digital elevation models that pre-date and post-date their most recent surges. Average ice velocities were also estimated on the basis of annual glacier-frontal measurements during their last surge. The observations show a distinct ice discharge, most from the upper reservoir areas, down to the receiving areas during the surges. The surface draw-down in the reservoir areas was usually 10-30 m during the surges, while the thickening of the receiving areas was significantly more variable, in order of 10-120 m. Despite a negative geodetic net mass balance derived from the digital elevation models, the reservoir areas have been gaining mass since the surge terminations. This surface thickening along with a considerable ablation of the receiving areas will most likely return the glacier surface profiles to the pre-surge stage. According to the surface evolution since the surge termination, this process could take about 45-65 years assuming a constant climate. Our results indicate that (a) greatest surface thinning in the upper reservoir areas and (b) development of Drangajökull surges that resembles Svalbard surge-type glaciers rather than Vatnajökull surge-type glaciers, could be explained by differences in glacier geometry, topography and substratum of the Drangajökull and Vatnajökull surge-type glaciers.

Keywords: Digital Elevation Models, surge-type glaciers, quiescent phase, active phase, reservoir area, receiving area

1. Introduction

Differencing of Digital Elevation Models (DEMs) is a well-established methodology to quantify volume changes of glaciers (Magnússon et al., 2005; Schomacker et al., 2012; Schomacker and Kjær, 2007, 2008; Sund et al., 2009, 2014, Abermann et al., 2010; Kjær et al., 2012; Tómasson et al., 2013). Time series of DEMs and other remotely sensed data are also commonly used to identify glacier surges and quantify their velocity, surface, volume and areal changes during the surges (Fischer et al., 2003; Magnússon et al., 2005; Frappé and Clarke, 2007; Sund et al., 2009, 2014; Quincey et al., 2011, 2014).

All Icelandic glaciers are considered to be warm-based and therefore any surge mechanism related to a thermal transition has been ruled out (Björnsson et al., 2003). However, an almost total lack of any glacial geomorphological imprints in areas that have been deglaciated since the Little Ice Age (LIA) about 500-650 m a.s.l. around the southern perimeter of Drangajökull might indicate polythermal conditions there during the LIA and perhaps at present (Brynjólfsson et al, 2014).

Notably, while the two recent most surges of the Drangajökull surge-type outlet glaciers lasted 5-7 years (Sigurðsson, 1998: Björnsson et al., 2003; Brynjólfsson et al., 2015), the active phase of the large Icelandic surge-type glaciers usually last for only a few months or 1-2 years (Sigurðsson, 1998; Björnsson et al., 2003). This relatively long surge phase of Drangajökull resembles a 3-10 years long active phase of Svalbard surge-type glaciers (Dowdeswell et al., 1991: Hamilton and Dowdeswell, 1996; Murray et al., 2003) and surge-type cirque glaciers in Iceland (Brynjólfsson et al., 2012).

Recent studies of Drangajökull have focused on geomorphology, glacial history since the Last Glacial Maximum (LGM), surge dynamics and recent areal changes of the ice cap (Principato, 2003, 2008; Principato et al., 2006; Þrastarson, 2006; Brynjólfsson et al., 2014, 2015, under review). Shuman et al. (2009) compared a GPS derived Digital Elevation Model (DEM) with series of repeated satellite profiles across Drangajökull, indicating up to 1.5 m a^{-1} surface lowering at the location of the satellite profile in the years 2003-2007. However, ablation stake measurements indicate positive mass balance of the whole ice cap in 2005-2007, indicating that the satellite profile is not representative for the whole ice cap (Shuman et al., 2009).

Comparison of a DEM since c. 1990 and from 2011, indicate about 8 m average surface lowering of the ice cap in the period 1990-2011. Obvious ice discharges during the recent most surges of the three outlet glaciers are reflected as much more thinning of their reservoirs and distinct thickening of the receiving areas of each outlet glacier as earlier described by Jóhannesson et al. (2013).

Surge-type glaciers can be warm-based or polythermal, and they cluster in certain areas, most commonly where climatic conditions are bounded with mean annual temperature of 0-10 °C and annual precipitation of 200-2000 mm (Sevestre and Benn, 2014), indicating that climate alone does not control their location (Meier and Post, 1969; Raymond, 1987; Hamilton and Dowdeswell, 1996; Jiskoot et al., 2000; Murray et al., 2003; Sevestre and Benn, 2014).

One surge cycle consists of an active phase and a quiescent phase. During the active phase, the glacier undergoes a dramatic change in geometry and morphology (Þórarinsson, 1964, 1969; Meier and Post, 1969; Sharp, 1988; Harrison and Post, 2003). The ice velocity during surge is commonly in the order of 2-3 magnitudes higher than during the quiescent phase. Ice from the reservoir area of the glacier is discharged down-glacier to the receiving area during the active phase (Dowdeswell et al., 1995; Björnsson et al., 2003; Aðalgeirsdóttir et al., 2005; Quiencey and Luckman, 2014). During the quiescent phase, the ice is stagnant or flowing at a velocity lower than required to maintain the glacier size. This gradually contributes to a steeper surface profile which is considered fundamental to return the glacier surface to the pre-surge state and subsequently enable a new surge (Raymond, 1987; Sharp, 1988; Dowdeswell et al., 1995; Eisen et al., 2001, 2005; Harrison and Post, 2003).

Considerable volumes of ice can be transported down-glacier during a surge, which often contributes to surface lowering in the order of 20-100 m in the reservoir areas and surface thickening of similar amounts in the receiving areas (Björnsson et al., 2003; Aðalgeirsdóttir et al., 2005; Sund et al. 2014). Such surges usually contribute to marginal advances of several hundreds of meters or even kilometres (Meier and Post, 1969; Raymond, 1987; Björnsson et al., 2003; Murray et al, 2003; Sund et al., 2009).

The aim of this study is to quantify the ice elevation and volume changes related to the most recent surges of the three major outlet glaciers of Drangajökull ice cap. We use DEMs that capture the recent most surges of the Drangajökull outlets.

2. Setting

The Drangajökull ice cap is located c. 100-915 m above sea level (a.s.l.) on the eastern Vestfirðir (Westfjords) peninsula in northwest Iceland (Fig. 1). Since the LIA, the glaciated area has decreased from about 190-216 km^2 (Sigurðsson et al, 2013; Brynjólfsson et al., 2015) to 142 km^2 in 2011 (Jóhannesson et al., 2013). The mean equilibrium line altitude (ELA) is 550-650 m a.s.l. which is about half the ELA at the major ice caps in south and central Iceland (Björnsson and Pálsson, 2008). In contrast to most glaciers in Iceland, ablation stake measurements show positive mass balance of Drangajökull from 2004-2007 (Björnsson and Pálsson, 2008; Shuman et al., 2009; Pálsson et al., 2012). However, geodetic mass balance measurements indicate negative net mass balance, about -0.35 m w.e. a^{-1}, of Drangajökull over a longer period, 1990-2011 (Jóhannesson et al., 2013).

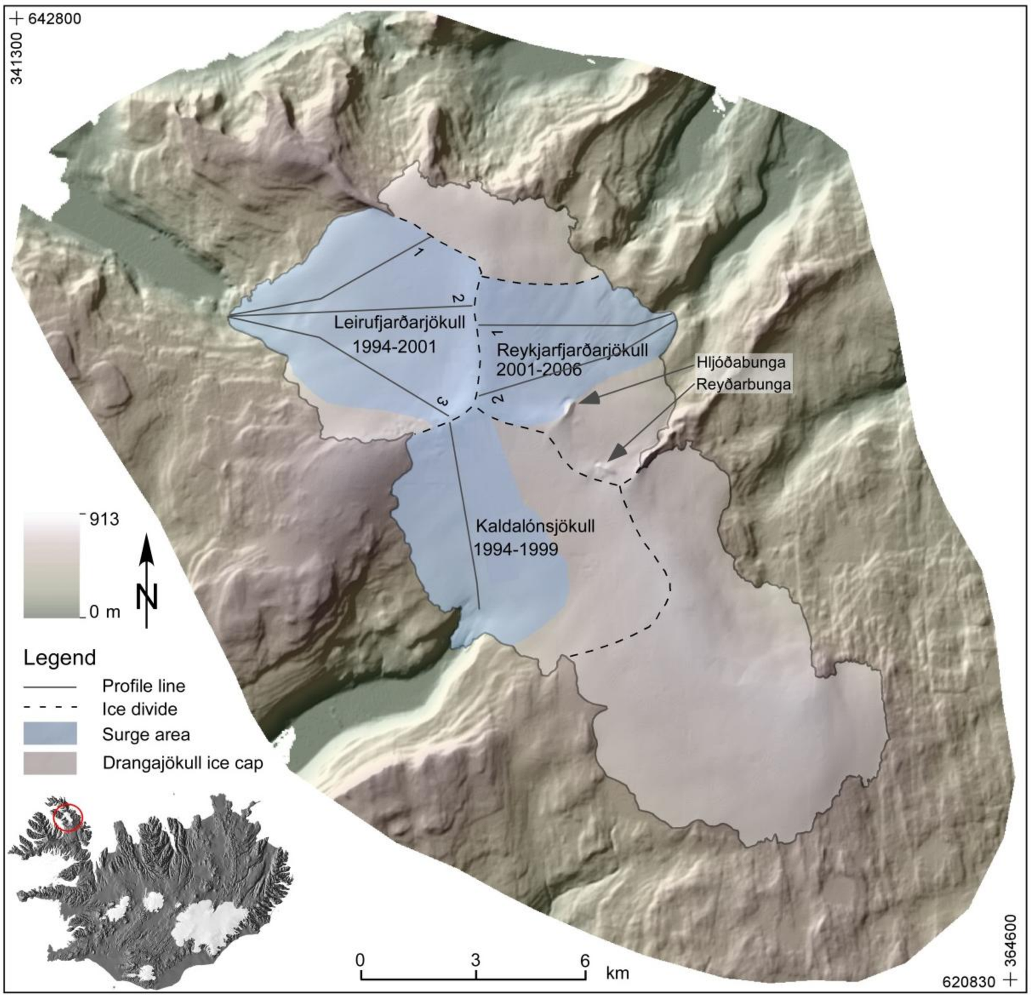

Figure 1. Overview of the Drangajökull ice cap in northwest Iceland. The catchments of the three surge-type outlet glaciers are marked with dashed lines, and the area affected by their last surge is shaded with blue colour. The grey lines at each glacier indicate positions of the surface profiles in Figure 5.

Located at the eastern Vestfirðir peninsula, proximal to the cold East Greenland Current and the warmer Atlantic Ocean, it has an open ocean from west to east favouring abundant precipitation in combination with a relatively cold, sub-polar climate (Eyþórsson, 1935; Bergþórsson, 1969; Eiríksson et al., 2000; Brynjólfsson et al., 2015). The regional lowland climate around Drangajökull is characterised by a mean summer temperature of 6-8 °C (June-September) and a mean annual temperature about 2.5-4 °C (Einarsson, 1976; Hanna et al., 2004). Average annual precipitation over the ice cap has been modelled to 2000-3000 mm/year (Crochet et al., 2007). The nearest weather stations are located 40-45 km from

Drangajökull, with a mean annual precipitation about 850 mm in the study period 1994-2011 (The Icelandic Met Office, 2014).

The three main outlets of Drangajökull; Reykjarfjarðarjökull, Leirufjarðarjökull, and Kaldalónsjökull are all surge-type glaciers (Fig. 1; Þórarinsson, 1969; Sigurðsson, 1998; Björnsson et al., 2003). Two nunataks, Reyðarbunga and Hljóðabunga, occur in the reservoir area of Reykjarfjarðarjökull (Fig. 1). Recent correlation of historical data and geomorphological data revealed at least 5-7 surges of each of the surge-type outlets, and a surge periodicity from c. 10 years to at least 60 years, and perhaps 140 years (Brynjólfsson et al., 2015).

Spatial and temporal details of the last surges of the Drangajökull outlets were recorded with measurements of annual marginal fluctuations that started in the year 1931 (Eyþórsson, 1935, 1963; Sigurðsson, 2011). The outlet glacier in Leirufjörður surged in 1995-2001, the glacier in Kaldalón surged in 1995-1999, and the outlet glacier in Reykjarfjörður surged in 2002-2006. The marginal advances lasted 5-7 years during those surges. This is several years longer than the few months or 1-2 years long active phases of the large surging outlet glaciers in central and south Iceland (Sigurðsson, 1998; Sigurðsson and Jóhannesson, 1998; Björnsson et al., 2003; Brynjólfsson et al., 2015).

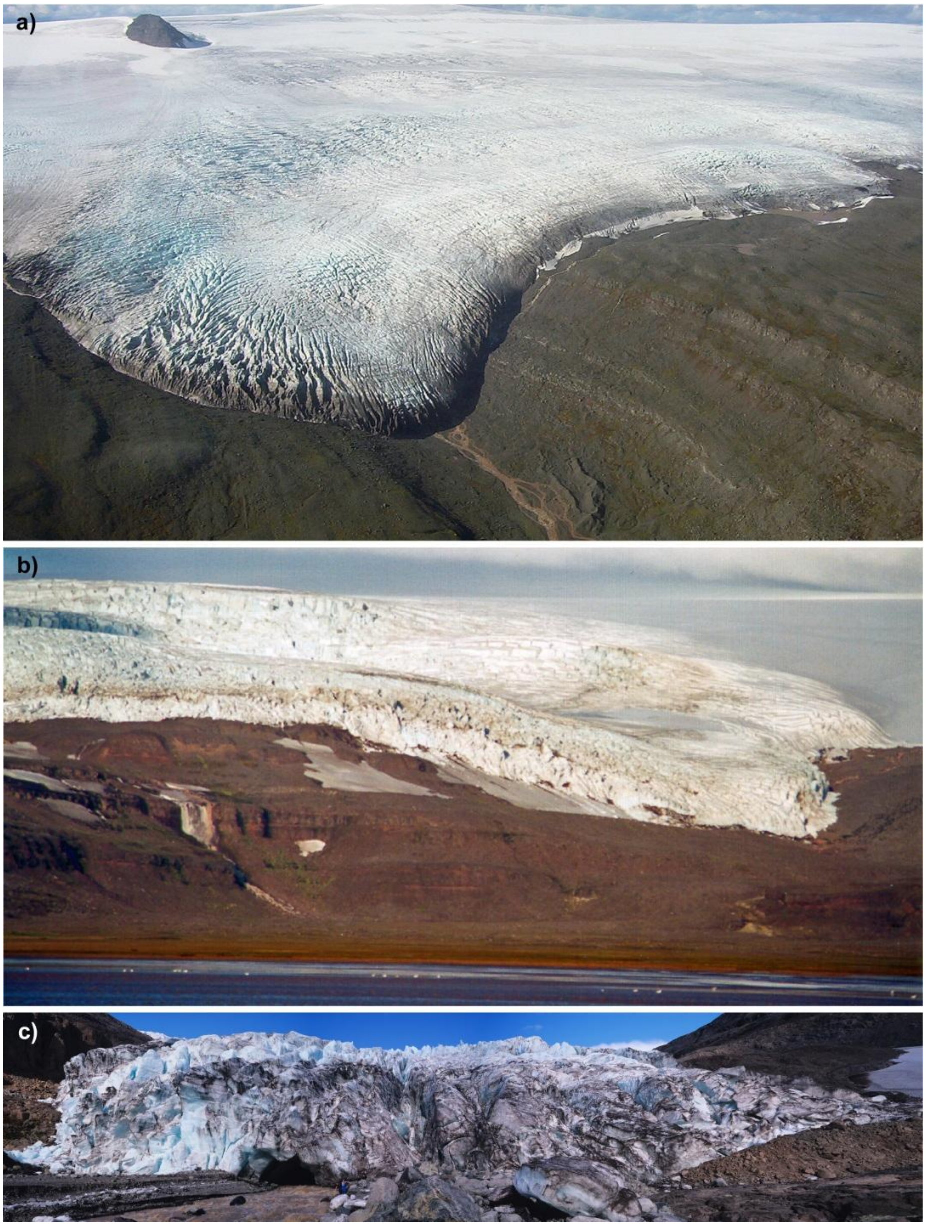

Figure 2. Photographs showing the three surging outlets of Drangajökull ice cap during their last surges. a) Reykjarfjarðarjökull 3rd of September 2004, photograph by Þröstur Jóhannesson. The Hljóðabunga nunatak is seen in the background. b) Leirufjarðarjökull in early August 1996, photograph by Hulda Björg Sigurðardóttir. c) Kaldalónsjökull, 5th September 1998, photograph by Oddur Sigurðsson. Note person in blue jacket for scale just left to the middle of the image.

3. Data and methods

3.1 Aerial photographs and Digital Elevation Models

Two sets of aerial stereo-photographs and derived DEMs from 1994 and 2005 in addition to an airborne Light Detection and Ranging (LiDAR) derived DEM from 2011, were used in this study (Table 1). The 1994 aerial photographs were supplied by the National Land Survey of Iceland. Unfortunately they do not cover the southernmost part of the ice cap including parts of the Kaldalón outlet glacier which is, therefore, partly missing on the 1994 DEM. Orthorectified aerial photographs and a DEM with 5 m ground resolution were produced by stereophotogrammetry. A digital photogrammetric workstation, using BAE Systems SocetSet 5.6 and BINGO 6.4 was used for the image processing. Coordinates for the ground control of the 1994 adjustment was obtained from repeat 1994 DEM co-registration to the 2011 LiDAR DEM of an 800x800 m window around each Ground Control Point, which are time-homologous points outside the ice cap measured in the 2011 LiDAR DEM. After final coordinates were determined, we adjusted and produced the 5 m 1994 DEM, which is thus coherent with the 2011 LiDAR DEM.

The aerial orthophotographs from 2005 and the derived DEM were provided by Loftmyndir ehf. The 2005 orthophotographs have ground resolution of 0.5 m and the derived DEM was delivered with 20 m ground resolution. A digital photogrammetric workstation, using Trimble Inpho software, was used to produce the 2005 DEM which has Route Mean Square error about 2-3 m.

The Drangajökull ice cap and its nearest surroundings were measured with airborne LiDAR on the 20th of July 2011. The LiDAR derived DEM, provided by the Icelandic Met Office, has a vertical and horizontal accuracy <0.5 m, and a ground resolution of 5 m (Jóhannesson et al., 2013). No aerial photographs were recorded during the LiDAR measurement. Because the LiDAR data were collected in the middle of the summer, several snow fields still existed in the forefields of the ice cap.

Table 1. Overview of the aerial photographs and DEMs used in this study. The Root Mean Square (RMS) error for the 1994 DEM is 1.2 m, for the 2005 DEM is estimated to less than 3 m (Loftmyndir, pers. comm., 2015). The accuracy of the 2011 LiDAR DEM is estimated to be better than 0.5 m both in position and elevation (Jóhannesson et al., 2013).

Date	Recorded by	Flying altitude (m a.s.l.)	DEM cell size (m)	Product	DEM RMS error (x,y,z in m)
29.08.1994	Landmælingar Íslands	5486	5	Orthophoto +DEM	1.2
2005	Loftmyndir ehf	c. 3000	20	Orthophoto + DEM	~ 3
20.07.2011	Icelandic Met Office	c. 2500	5	LiDAR DEM	0.5

3.2 Analysis of aerial photos and DEMs

The three DEMs were analysed and handled in the geographical information system ESRI ArcGIS 10.2. The comparison of the DEMs was carried out with a 3D extension of ArcGIS which enabled quantification of surface changes in order of metres, i.e. it provided DEMs of difference (DoDs). In order to quantify the discharge and volume changes during surges, ice thinning/thickening was calculated for the periods 1994-2011 and 1994-2005. Investigating the same parameters for the period 2005-2011 enabled us to estimate the surface changes of the glaciers during the current quiescent phase. Following the definition of Sorge's law, the volume mass balance (ice thickness changes) was converted to change of mass by multiplying the volume mass balance by constant ice density (Bader, 1954; Paterson, 1994). Density of ice varies from 830-917 kg m^{-3} depending on how much air it contains (Paterson, 1994). For the calculations we use a density of 917 kg m^{-3}. The volume changes represented in Table 2 were calculated by multiplying the mean thickness changes (m) and the size (km^2) of each study area. It should be noted that the DEMs are based on data recorded in different time of the summer, this has not been corrected for in the mass balance measurements presented in table 2.

4. Results

4.1 Ice surface and volume changes 1994-2011

The three individual DEMs (Table 1) enable calculations of ice surface and volume changes for the period 1994-2011, 1994-2005 and 2005-2011 (Table 2). Because the DEM from 1994 does not completely cover Kaldalónsjökull, nor its forefield (Fig. 3), Kaldalónsjökull was mostly excluded from this study. However, according to the 1994-

2011 DoD (Fig. 3) it was only the western and northwest part of Kaldalónsjökull that contributed to its last surge in 1995-1999. The 1994-2011 DoD covers 102 km^2 of the Drangajökull ice cap, excluding the southern perimeter (Fig. 3). The average surface elevation change for this area in 1994-2011 was -9.8 m which contributes to a loss of 1 km^3 and negative mass balance of -0.53 m w.e. a^{-1} (Table 2; Fig. 3). The 2005-2011 DoD covers the whole ice cap, 142.5 km^2 in 2011. The average surface elevation change was -1 m during this period. The volume change was -0.143 km^3 and the mass balance -0.15 m w.e. a^{-1} (Table 2; Fig. 3). Despite average total surface thinning and negative mass balance the reservoir areas of the three surging glaciers are gaining mass and increasing their thickness in the period 2005-2011 (Fig. 3).

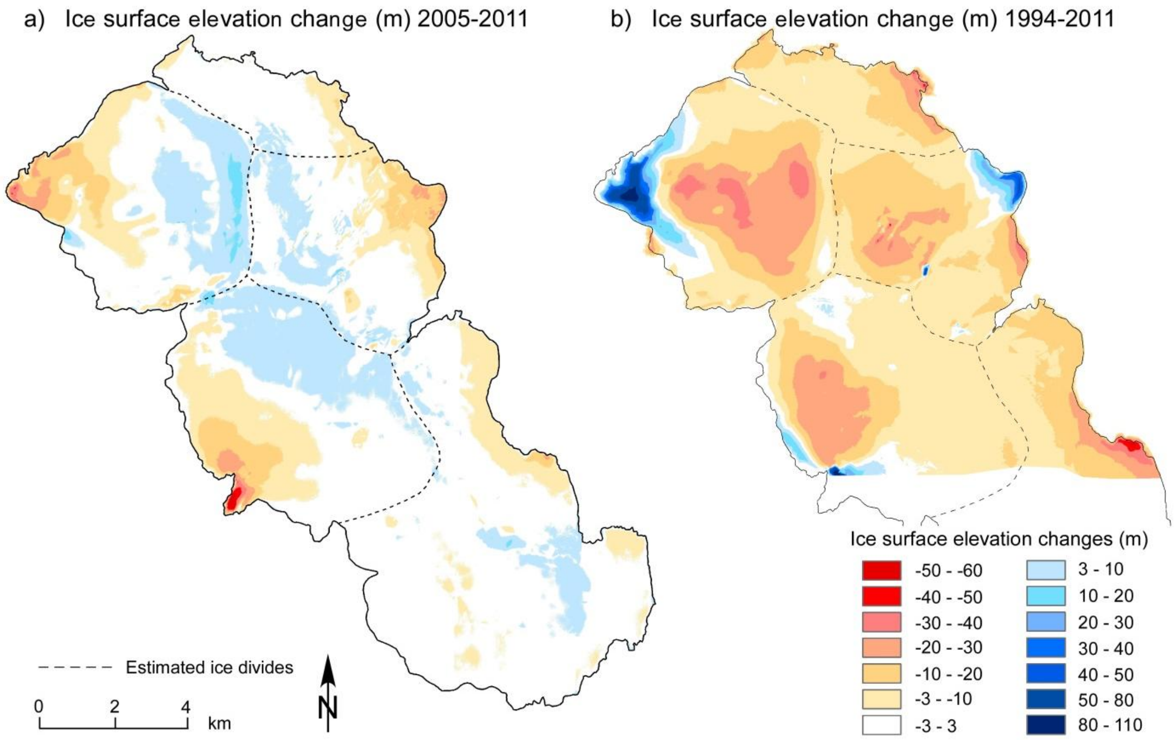

Figure 3. Surface elevation changes and evolution of Drangajökull and the surging outlet glaciers in the period 2005-2011 and 1994-2011.

Table 2. Overview of ice changes, Reyk = Reykjarfjarðarjökull, Leir = Leirufjarðarjökull, Dran-N = Drangajökull northern part, Dran-W = Drangajökull whole ice cap. The mean thinning/thickening are average values of the areas that showed thinning/thickening during each study period. The mean change is the average surface elevation change of the whole surging glacier. The volume changes are the corresponding changes given in km^3, and the average annual net balance is represented by B_n.

1994-2011	Glacier area (km^2)	Mean thinning - (m)	Vol loss (km^3)	Mean thickening + (m)	Vol gain (km^3)	Mean change (m)	Vol diff. (km^3)	B_n w.e. a^{-1} (m)
Reyk 1994	14.5							
2011	14.9	-13.1	-0.176	14.8	0.023	-10.2	-0.152	-0.54
Leir 1994	26.6							
2011	28.4	-15.3	-0.365	20.5	0.095	-9.5	-0.269	-0.51
Dran-N 2011	99.8	-12.1		16.0		-9.8	-0.978	-0.53
1994-2005								
Reyk 1994	14.5							
2005	14.9	-14.7	-0.175	18.7	0.054	-8.1	-0.121	-0.68
Leir 1994								
2005	28,7	-17.2	-0.384	25.0	0.152	-8.2	-0.235	-0.69
Dran-N 2005	100.2	-12.2		19.2		-8.4	-0.842	-0.7
2005-2011								
Reyk 2011	14,9	-7.1		2.8		-1.4	-0.021	-0.22
Leir 2011	28.4	-6.6		4.0		-1.1	-0.031	-0.18
Dran-W 2005	142.5							
2011	142.0	-4.3		2.8		-1.0	-0.142	-0.15

4.1.1 Reykjarfjarðarjökull

The margin of Reykjarfjarðarjökull advanced 227 m during the surge in 2002-2006 (Sigurðsson, 2011; Brynjólfsson et al., 2015). Though the 1994-2011 DoD spans several years before and after the surge, the ice surface changes related to the surge are clear (Fig. 4a-c). Ice structures and crevasses on the 2005 aerial photos and the elevation changes yielded from the DoD show that only the northern part of the glacier contributed to the surge in 2002-2006, while the part south of the nunatak Hljóðabunga was not affected (Fig. 1).

The area affected during the surge of Reykjarfjarðarjökull increased from 14.5 km^2 in 1994 to 14.9 km^2 in 2005 and 2011 was also 14.9 km^2 (Table 2). The 1994-2005 DoD indicates ice thinning up to 39 m, but most commonly in the order of 10-30 m in the reservoir area. The average surface elevation change of the glacier was -8.1 m, yielding a -0.121 km^3 net volume loss in the period or a net mass balance of -0.68 m w.e. a^{-1}. At least 0.054 km^3 of ice were discharged from the reservoir area down to the receiving area which is about 31% of the 0.175 km^3 volume loss of the reservoir area in the period 1994-2005 (Table 2).

The 1994-2011 DoD indicates a 10.2 m average surface lowering of the glacier, yielding a 0.152 km^3 volume loss which is equal to a net mass balance of -0.54 m w.e. a^{-1} in this period (Table 2). The ice discharge from the reservoir area down to the receiving area is well demonstrated in both DoDs (Fig. 4a-c). The main surface thinning in the order of 10-30 m occurs above the ELA in the upper reservoir area. According to the negligible surface lowering detected near the ice divide in the northern and westernmost parts of the reservoir area (Fig. 4b), those areas were almost unaffected by the surge.

An average marginal advance rate of 0.12 m d^{-1} during the surge 2002-2006 was obtained from the ice-frontal measurements (Table 3; Sigurðsson, 1998, 2011). The fastest marginal advance of Reykjarfjarðarjökull during the surge was 0.2 m d^{-1} in the period 2003-2004. The accelerating and particularly the decelerating phase of the two recent most surges of Reykjarfjarðarjökull surges take several years (Table 3).

In the period 2005-2011, after the surge termination, an opposite pattern with distinct accumulation in the reservoir area and ablation in the receiving area of the surging glaciers and the whole ice cap was detected (Figs 3, 4c). The total average thickening in the reservoir area was 2.8 m, which is c. 0.5 m a^{-1}. According to this, it could take the reservoir area 55-65 years to reach the pre-surge stage assuming constant climate. At same time, the receiving area thinned by c. 1.2 m a^{-1} on average. This yields a negative net mass balance of 0.21 m w.e.a^{-1} of Reykjarfjarðarjökull in the period 2005-2011 (Table 2). However, this is a pattern that gradually returns the surface profile towards the pre-surge stage and most likely contributes to a new surge.

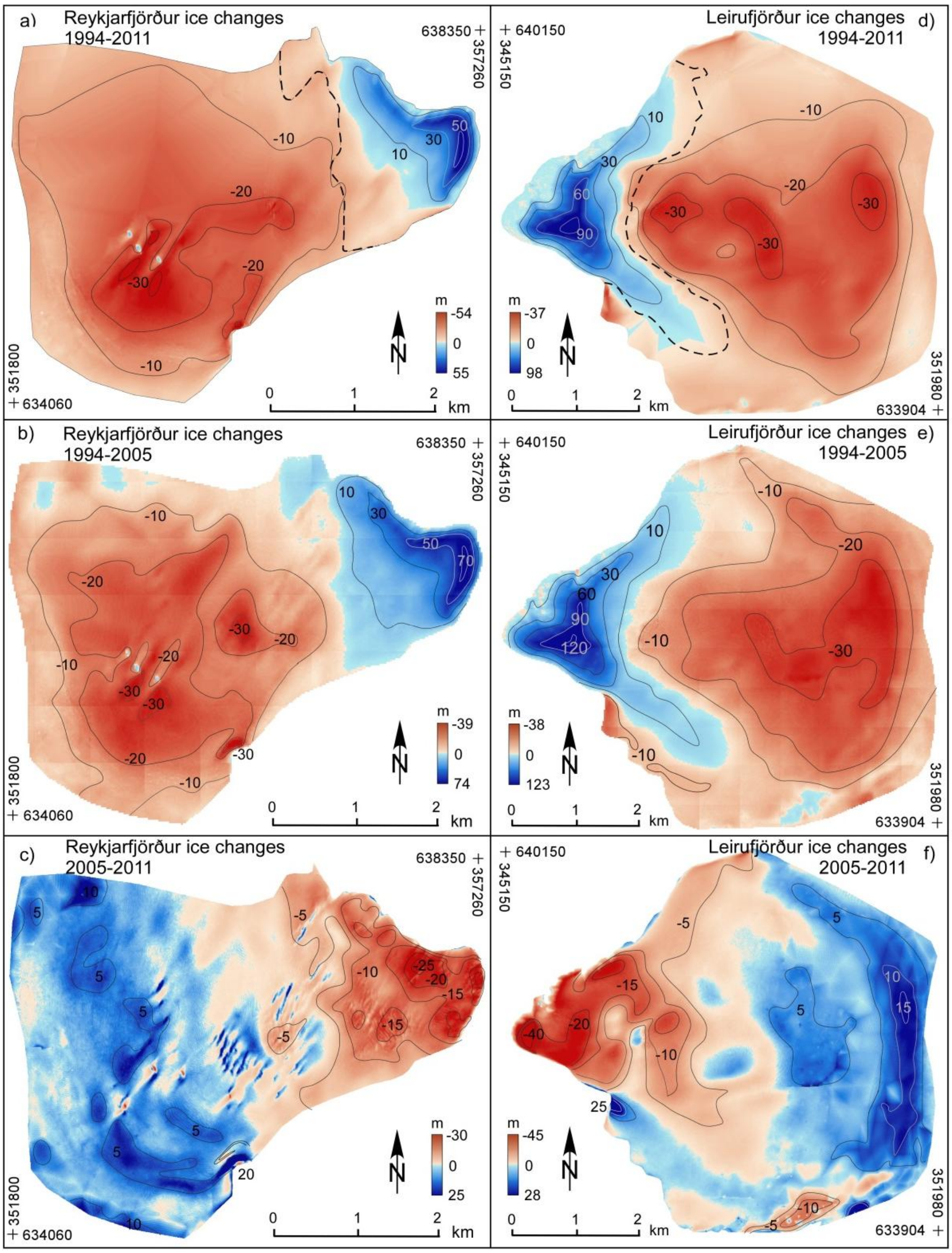

Figure 4. Ice surface elevation changes (m) related to the most recent surges of Reykjarfjarðarjökull (a-c) and Leirufjarðarjökull (d-f). The dashed lines in a) and d) indicate the zero surface elevation change line in the 1994-2005 DoD. Contour lines and numbers on the figures indicate the surface elevation change during each period. Coordinates in ISN93 WGS84 format.

4.1.2 Leirufjarðarjökull

During the recent most surge of Leirufjarðarjökull in 1995-2001, the margin advanced about 1150 m (Björnsson et al., 2003; Brynjólfsson et al., 2015). The total area of Leirufjarðarjökull was 26.6 in 1994, 28.7 in 2005 and 28.4 km^2 in 2011 (Table 2). The 1994-2005 DoD shows a small area along the southwest and south margin that had not been affected by the surge by the year 2005 (Fig. 4d and 4e). The surge bulge was located about 500 m up-glacier of the southwest margin at this time. The area proximal to the southwest margin thinned about 10 m and more in the period 1994-2005, while the marginal area thickened by 10-120 m elsewhere (Fig. 4d-e). The annual glacier-front measurements indicate that the surge and the marginal advance terminated in 2001. However, the increased surface elevation along the southwest margin of Leirufjarðarjökull that was observed on the 2005-2011 DoD (Fig. 4f) shows that the surge was still going on in this area of the glacier at least until 2005 and the surge lasted at least for 10 years.

Ice discharge from the reservoir area down to the receiving area is well demonstrated with thinning of up to 38 m, most commonly 10-30 m in the upper reservoir area and up to 123 m thickening in the receiving area (Fig. 4d-e). The 1994-2005 DoD suggests a -8.2 m average surface change of Leirufjarðarjökull, yielding volume change of -0.235 km^3 which corresponds to a net mass balance of -0.68 m w.e. a^{-1}. For the period 1994-2011, the mean surface change of Leirufjarðarjökull was -9.5 m and the volume change was -0.269 km^3 corresponding to a net mass balance of -0.51 m w.e. a^{-1} (Table 2). Less thickening of the receiving area was observed for the whole study period 1994-2011 compared to the period 1994-2005. This demonstrates ablation of stagnant ice in the receiving area since the surge termination in 2001. The ice thinning in the reservoir area of the northern part of the glacier is less in the period 1994-2011 compared to 1994-2005, demonstrating that snow has been accumulating in the reservoir area since the surge. These results are supported by the 2005-2011 DoD which suggests a mean snow accumulation of c. 0.7 m a^{-1} in the receiving area over this six-year period (Table 2; Fig. 4f).

Table 3. Overview of annual marginal fluctuations (in m) of the Drangajökull outlets during their last two surges (data from the database of the Icelandic Glaciological Society (www.spordakost.jorfi.is); Sigurðsson, 1998, 2011).

Year	**1934**	**1935**	**1936**	**1937**	**1938**	**1939**	**1940**	**1941**	**1942**					
Reyk	82	149	495	-27	57	2								
Leir						539	150	272	37	0				
Kald			39	107	41	2	2							
Year	**1995**	**1996**	**1997**	**1998**	**1999**	**2000**	**2001**	**2002**	**2003**	**2004**	**2005**	**2006**	**2007**	**2008**
Reyk								17	70	72	36	32	-1	4
Leir	39	737	169	75	94	36	10	0	3					
Kald	12	38	669	150	146	0								

Leirufjarðarjökull advanced 737 m from 1995-1996, yielding the highest average marginal advance rate of about 2 m d^{-1}. However, the average marginal advance rate during the whole active phase was 0.46 m d^{-1}. The acceleration of Leirufjarðarjökull surges can be quite abrupt, while the decelerating phase, as for Reykjarfjarðarjökull, can take several years (Table 3).

Similar to Reykjarfjarðarjökull, an opposite pattern was observed after the surge termination, during the period 2005-2011 (Fig. 4c and 4f). Net snow accumulation in the order of 0.7 m a^{-1} in the reservoir area and an average thinning by c. -1.2 m a^{-1} in the receiving area yields a slightly negative net mass balance (-0.18 m w.e. a^{-1}) of Leirufjarðarjökull in the period 2005-2011 (Table 2). However, according to this pattern, the surface profile gradually returns towards its pre-surge stage during this six-year period. Assuming an average net snow accumulation of c. 0.7 m a^{-1}, it might take 35-45 years to return the surface of the reservoir area to the pre-surge stage assuming a constant climate.

5. Discussion

5.1 Reykjarfjarðarjökull

The nunatak, Hljóðabunga (Fig. 1), seems to have acted as a barrier for ice flow during the surge of Reykjarfjarðarjökull 2002-2006. Longitudinal fractures and folded sediment bands within the ice can be traced from Hljóðabunga downglacier to the margin. Thus, the surge did not affect the glacier south of Hljóðabunga during the recent most surge in 2002-2006 (Fig. 1). Figure 3 indicates that most of the ice-volume discharge was from above the ELA in the upper-central area of Reykjarfjarðarjökull, north and northwest of Hljóðabunga, whereas the north-western part of the reservoir area was less affected by the surge.

According to the 1994-2005 DoD, at least 0.054 km^3 of ice were discharged from the reservoir area down to the receiving area of Reykjarfjarðarjökull, which is about 31% of the 0.175 km^3 volume loss of the reservoir area in this period. The 0.054 km^3 discharge is considered a minimum volume because the surge did not terminate until 2006. Therefore, an unknown volume of ice was discharged in this final year of the surge. For comparison, the ice discharge during Vatnajökull surges has been estimated to about 75% of the reservoir area volume loss (Aðalgeirsdóttir et al., 2005). The net mass balance of Reykjarfjarðarjökull was -0.68 m w.e. a^{-1} and contributed to a 0.121 km^3 total volume loss of the glacier in the period 1994-2005 (Table 2).

5.2 Leirufjarðarjökull

The ice volume loss of Leirufjarðarjökull was more evenly spread over the reservoir area than at Reykjarfjarðarjökull, except the southernmost part which seems to have been little or unaffected by the surge in 1994-2001 (Fig. 4d-e). The annual glacier-front measurements from the westernmost tip of the margin indicate surge termination in 2001 (Sigurðsson, 2003). However a minor advance was measured in 2003 and therefore the total surge duration could be seven years (Sigurðsson, 2004). The 2005-2011 DoD revealed a small area of surface thickening along the southwest margin in this period (Fig. 4f). We consider it to indicate that the area proximal to the southwest margin of the glacier was still surging in 2005. This suggests at least 10 years surge duration since the first marginal advances were observed in 1995. According to this observation, a small area proximal to the southwest margin continued surging while the surge seems to have terminated in other areas of the glacier in 2001 or at latest in 2003.

Similar to Reykjarfjarðarjökull, the surface thinning was largest in the upper reservoir area and most commonly in the order of 10-30 m. The ice discharge down to the receiving area during the surge of Leirufjarðarjökull was at least 0.152 km^3 which is about 40% of the 0.384 km^3 volume loss of the reservoir area from 1994-2005. The actual volume discharge was more than 0.152 km^3, an unknown amount of the ice mass was ablated in the receiving area in the four years that elapsed from the surge termination until the 2005 DEM was measured. The net mass balance of Leirufjarðarjökull was -0.68 m w.e. a^{-1} which resulted in a 0.235 km^3 total volume loss of the glacier in the period 1994-2005.

5.3 Surge dynamics of Drangajökull

Because the DEMs pre- and post-date the active phase of the surges, we are not able to make any direct observations of the surface velocity or how a potential surge bulge spread across the glaciers. However, according to the annual glacier-front measurements, the marginal advance rate can be roughly estimated. The highest annual average advance rate of the Leirufjarðarjökull margin was 2 m d^{-1} when the advance was measured to 737 m from 1995-1996. Reykjarfjarðarjökull advanced 75 m during the year 2003-2004 yielding about 0.2 m d^{-1} as the highest average advance rate. The 2 m d^{-1} average frontal advance of Leirufjarðarjökull in 1995-1996 is comparable to the lower average surface flow rates of

Vatnajökull and Svalbard surging glaciers. Flow rates in the order of 0.1-7 m d^{-1} during surges of some Svalbard surging glaciers have been observed (Dowdeswell et al., 1991; Murray et al., 2003; Kristensen and Benn, 2012), while the average surface flow rate of Vatnajökull surges vary from 2-22 m d^{-1} (Þórarinsson, 1969; Fischer et al., 2003; Aðalgeirsdóttir et al., 2005). The maximum frontal advance rate during the 1963-1964 surge of Brúarjökull was up to 125 m d^{-1} (Þórarinsson, 1969; Kjær et al., 2006). The average marginal advance during the whole active phase is significantly lower, 0.46 m d^{-1} for Leirufjarðarjökull and 0.12 m d^{-1} for Reykjarfjarðarjökull. According to the glacier-frontal measurements, the Drangajökull surges often reach a maximum flow rate and ice discharge over about one year (Table 3). On the other hand the accelerating and decelerating phases of the Drangajökull surges tend to takes several years (Table 3). This resembles surges of Svalbard glaciers (Dowdeswell et al., 1991; Murray et al., 2003; Sund et al 2009), rather than the Vatnajökull surges which most often switch on and off over short periods (Sigurðsson, 1998; Björnsson et al., 2003; Aðalgeirsdóttir et al., 2005).

Observations of surging outlets of the Vatnajökull ice cap in southeast Iceland have shown that a surge initiation is typically marked by an abrupt increase in ice velocity in the upper ablation area of the glacier. The velocity peak is confined to the ablation area and the main surface lowering occurs in the area of the high velocity. Subsequently, the high-velocity area spreads downglacier, leading to thickening of the glacier downstream of the velocity peak and crevasse formation and a slight surface lowering upstream of the high-velocity area due to extension (Björnsson et al., 2003; Fischer et al., 2003; Aðalgeirsdóttir et al, 2005).

Our observations from Drangajökull reveal that the maximum surface lowering occurs in the accumulation areas of the glaciers, at 600-780 m a.s.l. The maximum surface drawdown seems to correlate with the area of high accumulation during the quiescent phase (Fig. 4). The average ELA of Drangajökull is 550-650 m a.s.l. (Eyþórsson, 1935; Björnsson and Pálsson, 2008), whereas the mean ELA after the surge termination of the outlet glaciers, indicated by the transition zone of accumulation and ablation (Fig. 4c and 4f), is about 650 m a.s.l. in the years 2005-2011. This pattern of maximum surface lowering, and perhaps maximum surface velocity, at Drangajökull surging glaciers coincides with observations of Svalbard surging glaciers. There, the surge initiation, maximum surface lowering, and a high surface velocity have been observed in the uppermost accumulation areas of some Svalbard surging glaciers (Sund et al., 2009; 2014). These areas coincide with the areas of maximum accumulation during the quiescent phase (Sund et al., 2014).

The surface thinning of the surging glaciers reservoir areas at Drangajökull is usually in the order of 10-30 m, while the reservoir areas of Vatnajökull surge-type glaciers typically thins by 25-100 m during a surge (Björnsson et al., 2003). Therefore, Reykjarfjarðarjökull and Leirufjarðarjökull seem to show a different surface expression and perhaps pattern of ice flow velocity compared to the Vatnajökull surging outlet glaciers (Björnsson et al., 2003; Fischer et al., 2003; Aðalgeirsdóttir et al., 2005). This different surface expression, surge propagation and perhaps velocity patterns of Drangajökull surging glaciers and Vatnajökull surging glaciers could be related to the contrasting valley/mountain settings of northern Drangajökull compared with less broken and less constraining topography of the Vatnajökull surging glaciers. The Neogene plateau basalt substratum at Drangajökull, compared with the Pleistocene palagonite-dominated substratum at Vatnajökull, could also contribute to the different surge dynamics. Thermal regimes of the polythermal Svalbard surging glaciers, different geometry and type of substratum bedrock are also considered to affect dynamics of surge-type Glaciers (Dowdeswell et al., 1991; Hamilton and Dowdeswell, 1996; Jiskoot et al., 1998, 2000; Murray et al., 2003; Sund et al., 2009; Flowers et al., 2011). An almost complete lack of evidence for glacial erosion or deposition might indicate that the Drangajökull margins at 500-700 m a.s.l. were recently polythermal (Brynjólfsson et al., 2014).

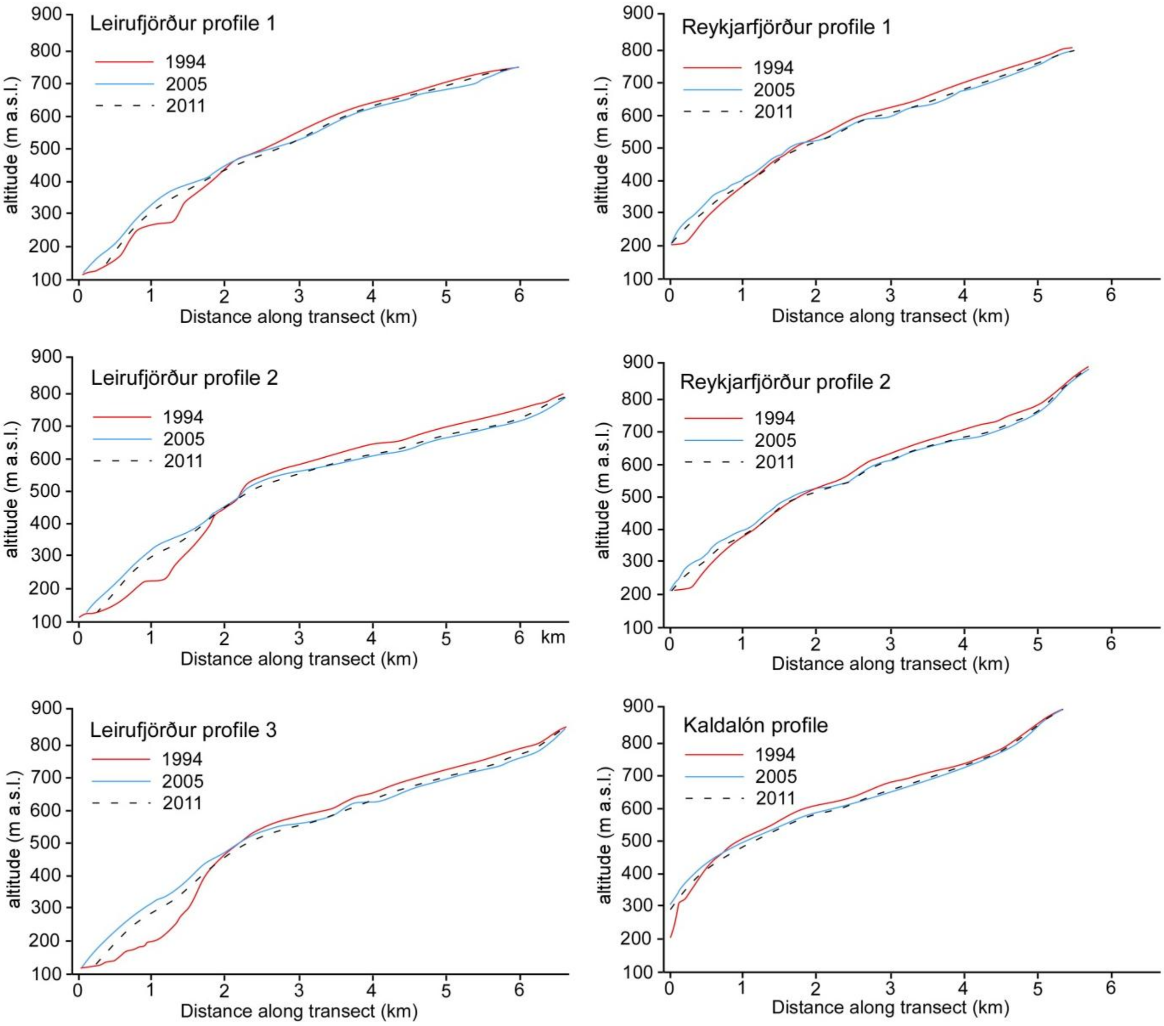

Figure 5. Terrain surface profiles of each surging outlet glacier of the Drangajökull ice cap. The red lines are profiles from 1994, at the beginning or prior to the surges, and the black profiles are from 2005, at the end or several years subsequent to the surges. For reference, profiles from 2011 are also shown. The location of the profiles is shown in Fig. 1.

Most of the reservoir areas of both Reykjarfjarðarjökull and Leirufjarðarjökull thinned in the order of 10-35 m in the study period (Figs 3 and 4). At the same time, the thickening of the receiving area was much more variable, c. 10-120 m. The variation in the thickening of the receiving areas is interpreted to reflect the high-relief topography that the surges have overridden. The ice surface thinning contributes to a volume loss of 0.117 km^3 from 1994-2005 and 0.145 km^3 from 1994-2011 for Reykjarfjarðarjökull, and 0.218 km^3 from 1994-2005 and 0.253 km^3 from 1994-2011 for Leirufjarðarjökull (Fig. 4; Table 2).

The ice discharge from the reservoir areas down to the receiving areas is well demonstrated on the surface profiles in Figure 5. The minimum ice volume transported to the receiving area is 31-40% of the reservoir area ice loss, while ice discharge during surge contributes to about 75% of the reservoir ice loss for some Vatnajökull surging glaciers (Aðalgeirsdóttir et al., 2005). Notably, this observation of Aðalgeirsdóttir et al. (2005) captured almost exactly the surge duration, while our observation captures several years before or after the surges.

We suggest that the larger part of the volume loss of the reservoir areas of Reykjarfjarðarjökull and Leirufjarðarjökull in 1994-2005, can be explained by negative mass balance or increased melting of the ice, enhanced by the surges. Thus, the ice discharge from the reservoir area to the receiving area during the surges explains c. 30-40% of the total volume loss of the reservoir areas in the period 1994-2005. However, because the 1994 DEM pre-date the surge of Reykjarfjarðarjökull and the 2005 DEM post-date the surge of Leirufjarðarjökull, some additional melting has been going on either prior to or subsequently to the surges covered by the 1994-2005 DoD. Therefore, we consider the ice discharge during the surges to be slightly underestimated in our observations.

The surface area increases and becomes much more crevassed during glacier surges (Fig. 2). This, along with transportation of ice from higher to lower altitudes, increases surface melting from turbulent heat exchange and solar radiation, both during and after surge (Björnsson et al., 2003; Aðalgeirsdóttir et al., 2005). Furthermore, additional melt occurs due to friction at the glacier bed during the surge (Aðalgeirsdóttir et al., 2005). These components that contribute to increased ice melting are considered to explain the 25% negative volume difference of the ice deposition in the receiving area and the volume loss in the reservoir area of the larger surging outlet glaciers of the Vatnajökull ice cap (Björnsson et al., 2003; Aðalgeirsdóttir et al., 2005).

5.5 Post-surge changes and surge periodicity

Comparison of the 2005 and the 2011 DEMs allow us to assess the present and potential future surface evolution of the surging glaciers. By dividing the average thickening of the reservoir areas by the six years covered by the 2005-2011 DoD, an average annual thickening of the reservoir areas can be calculated. Based on that, a potential time needed for the glaciers to reach pre-surge surface conditions in the reservoir area can be estimated. We have not accounted for snow compaction and firnification in this period, and therefore the potential time needed to reach the pre-surge stage should be regarded as minimum values.

The 2.8 m average total thickening in the reservoir area of Reykjarfjarðarjökull from 2005-2011 corresponds to 0.47 m of annual thickening. Given that the ice flow from the reservoir area to the receiving area is negligible or less than the balance velocity of the glacier during the quiescent phase, which is typical for surge type glaciers (Raymond, 1987; Björnsson et al., 2003; Harrison and Post, 2003; Murray et al., 2003), and the future pattern of accumulation in the reservoir area will be similar to the period 2005-2011, the glacier can be expected to need 55-65 years to reach the pre-surge stage. For comparison, the quiescent phase between the two recent most surges, 1933-1939 and 2001-2006, of Reykjarfjarðarjökull was 62 years (Brynjólfsson et al., 2015).

For Leirufjarðarjökull, the average total thickening of the reservoir area was 4 m, and the average annual thickening 0.67 m during the period 2005-2011. Given the same conditions as for Reykjarfjarðarjökull, it will take about 40-50 years for Leirufjarðarjökull to reach the pre-surge stage surface elevation. The quiescent period between the last two surges of the Leirufjarðarjökull was 52 years (Sigurðsson, 1998; Brynjólfsson et al., 2015).

The surge periodicity has been considered to be linked to climatically forced mass balance, and become shorter with more positive mass balance during colder periods and longer during warmer and less favourable mass balance conditions (Dowdeswell et al., 1995; Eisen et al., 2001; Striberger et al., 2011). Despite the slightly negative net mass balance of Drangajökull and the surging outlets, in the period 2005-2011 (Table 2), Leirufjarðarjökull and Reykjarfjarðarjökull seem to be recovering after their latest surges (Figs. 3 and 4). Mass accumulation in the reservoir area and melting of the stagnant ice in the receiving area contribute to the development of a steeper surface profile which gradually brings the glacier surfaces to the pre-surge stage (Fig. 5). Direct mass balance measurements of the Drangajökull ice cap revealed a positive mass balance of the ice cap in the years 2005-2007 (Shuman et al., 2009). However, the long-term geodetic mass balance of the ice cap is negative with 0.35 m w.e. a^{-1} from 1990-2011 (Jóhannesson et al, 2013) and about -0.5 m w.e. a^{-1} in the period 1994-2011 (Table 2). The applied density is a source of uncertainty, and therefore the mass change calculations have to be considered as estimations because we don´t know how much of the volume changes constitute snow or ice gain/loss. Despite that, we expect the surging outlet glaciers of Drangajökull to reach pre-surge surface elevation in about 45-65 years if the accumulation and mass balance conditions observed in the period 2005-2011 remain similar in the coming decades (Table 2; Fig. 5).

Conclusions

By DEMs from 1994, 2005 and 2011 we have quantified ice surface elevation and volume changes that relate to the recent most surges of the surging outlet glaciers, Reykjarfjarðarjökull and Leirufjarðarjökull.

- The surface thinning of the reservoir areas of the glaciers is most generally in the order of 10-30 m. The largest surface thinning occurs in the upper reservoir areas, mainly above the ELA, between 600-780 m a.s.l. The maximum surface draw-down correlates with the areas of maximum accumulation during the quiescent phase.
- The thickening of the receiving areas is in the order of 10-120 m. The ice discharge, 0.054 km^3 for Reykjarfjarðarjökull and 0.152 km^3 for Leirufjarðarjökull, to the receiving areas contributes to 30-40 % of the volume loss in the reservoir areas in the period 1994-2005. As the DEMs pre- and post-date the surges, the ice discharge is minimum estimates.
- According to glacier-frontal measurements, the highest annual average marginal advance during the last surges was 2 m d^{-1} during the surge of Leirufjarðarjökull. The average marginal advance of the total surge duration is much lower, 0.46 m d^{-1} at Leirufjarðarjökull and 0.12 m d^{-1} at Reykjarfjarðarjökull.
- Despite the negative mass balance, the surging outlet glaciers accumulated mass and thickened in the reservoir areas since the last surges terminated until 2011. This and distinct thinning in the receiving areas will bring the surface elevation of the glaciers to a pre-surge stage in about 45-65 years.
- The DoDs reveal that the upper reservoir areas experienced most surface thinning, in order of 10-30 m, during the last surges.
- The DoDs reveal surface thinning in the order of 10-30 m, that occurred mainly above the ELA in the upper reservoir areas, and that the surge of Leirufjarðarjökull went on for at least 10 years.

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Appendix V

(Earth Science Reviews – in review)

Surging glaciers in Iceland – research status and future challenges

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Abstract

Surging glaciers are thought to be modern analogues in understanding terrestrial ice streams, which have been shown to have a determining role in ice-sheet behaviour, and a highly dynamic response to climate change. The temperate surging glaciers of Iceland probably provide the best modern analogue we have to terrestrial ice streams in the geological record. The geomorphic signatures left by the Icelandic surging glaciers vary and range from glaciotectonic end moraines formed by folding and thrusting, hill-hole pairs, crevasse-fill ridges, concertina eskers, drumlins, and fluted forefields to extensive dead-ice fields and even drift sheets where fast ice-flow indicators are largely missing. We outline some outstanding research questions and review case studies from the surge-type outlets of Brúarjökull, Eyjabakkajökull and Tungnaárjökull (Vatnajökull ice cap), Múlajökull and Sátujökull (Hofsjökull ice cap), Hagafellsjökull and Suðurjökull (Langjökull ice cap), Kaldlónsjökull, Leirufjarðarjökull and Reykjafjarðarjökull (Drangajökull ice cap), as well as surging cirque glaciers in northern Iceland. We review the current understanding of the factors that trigger surging and discuss how rapid ice flow is sustained through the surge, the processes that control the development of the surging glacier landsystem and the geological evidence of surges found in sediments and landforms. We also examine if it is possible to reconstruct past surge flow rates from glacial landforms and sediments and scale-up present-day surging glaciers processes/landforms/landsystems for applying to past ice streams. Finally, we also examine if there is a climate/mass-balance control on surge initiation, duration and frequency.

Keywords: Surge-type glacier, Iceland, glaciotectonism, landsystem models, surge history

1. Introduction

A basic definition of a surge-type glacier identifies it as an outlet glacier that periodically has major fluctuations in velocity over timescales that range from a few years to several decades or centuries (Benn and Evans, 2010). Surging glaciers experience distinctive changes in morphology and activity over a surge cycle, where the phase of rapid motion over a few months to several years is described as the surge or active phase, and the period of slow flow or stagnation for tens to hundreds of years between surges is described as the quiescent phase (Harrison and Post, 2003b; Kamb et al., 1985; Raymond, 1987). It has been estimated that less than 1% of Earth's glaciers surge (Jiskoot et al., 1998), and they have been shown to be unevenly distributed around the world's glaciated regions and cluster in certain areas, notably Alaska (Kamb et al., 1985), Arctic Canada (Copland et al., 2003), Greenland (Murray et al., 2002), Iceland (Björnsson et al., 2003), Svalbard (Dowdeswell et al., 1991; Hagen et al., 1993), Novaya Zemlya (Grant et al., 2009), as well as in the Caucasus, Karakoram, Pamir and Tien Shan mountain ranges (Benn and Evans, 2010; Hewitt, 2007; Quincey et al., 2011). This clustering implies that certain environmental factors control the location of surge-type glaciers, but despite a number of studies that have investigated the possible constraints at a regional scale the reasons for this remain poorly understood (Clarke, 1991; Clarke et al., 1986; Hamilton and Dowdeswell, 1996; Jiskoot et al., 1998; Jiskoot et al., 2000; Meier and Post, 1969; Murray et al., 2003) and as yet there exists no unifying theory for explaining the surge mechanism (Rea and Evans, 2011). While both temperate and polythermal glaciers exhibit surging behaviour, it has been suggested the highest densities of surge-type glaciers occur in a relatively narrow climatic band bounded by mean annual temperatures of ca. 0 to -10°C and mean annual precipitation of ca. 200-2000 mm (Sevestre and Benn, in press). With its maritime cold-temperate to low-arctic climate and numerous temperate glaciers, Iceland largely lies within this climatic band.

All major ice caps in Iceland have surge-type outlet glaciers, and glaciological studies and historical records have revealed at least 26 surging outlet glaciers in Iceland that occur almost exclusively as outlets of the major ice caps (Figs. 1 and 2) (Björnsson, 1998; Björnsson and Pálsson, 2008; Björnsson et al., 2003; Thorarinsson, 1964b, 1969). Glacial geological studies have confirmed at least 2 additional surge-type glaciers (Evans, 2011; Larsen et al., 2015). Over 80 surge advances have been recorded, ranging from tens of metres up to 10 km, and systematic observations over the last several decades have allowed for a detailed description of several surges (Björnsson, 1998, 2009; Björnsson et al., 2003).

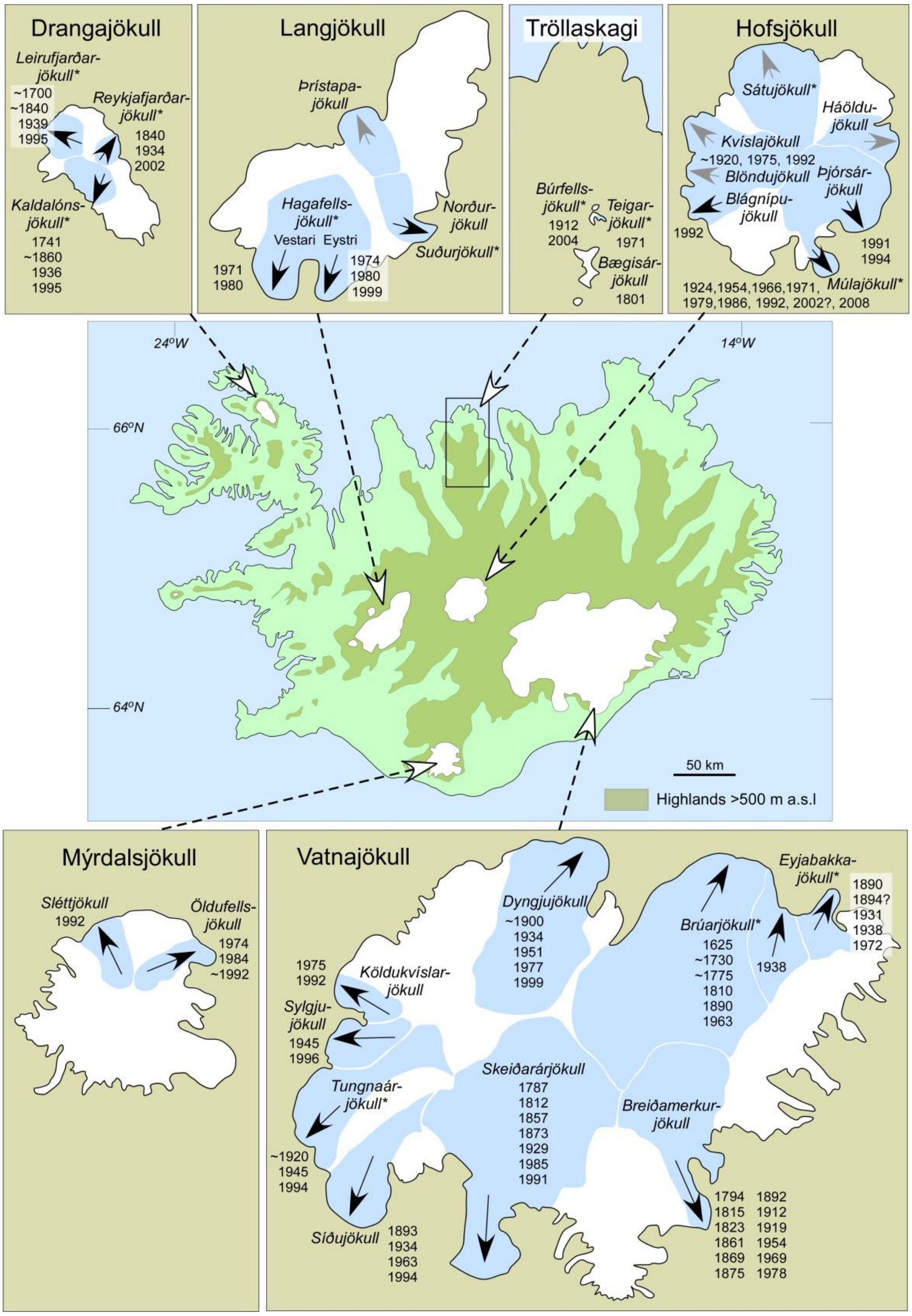

Fig. 1. Surge-type glaciers in Iceland. Glaciers marked with * are discussed in the text, and poorly constrained surges are indicated by grey arrows. Modified after Björnsson et al. (2003) and Björnsson and Pálsson (2008), and updated according to literature discussed in the text.

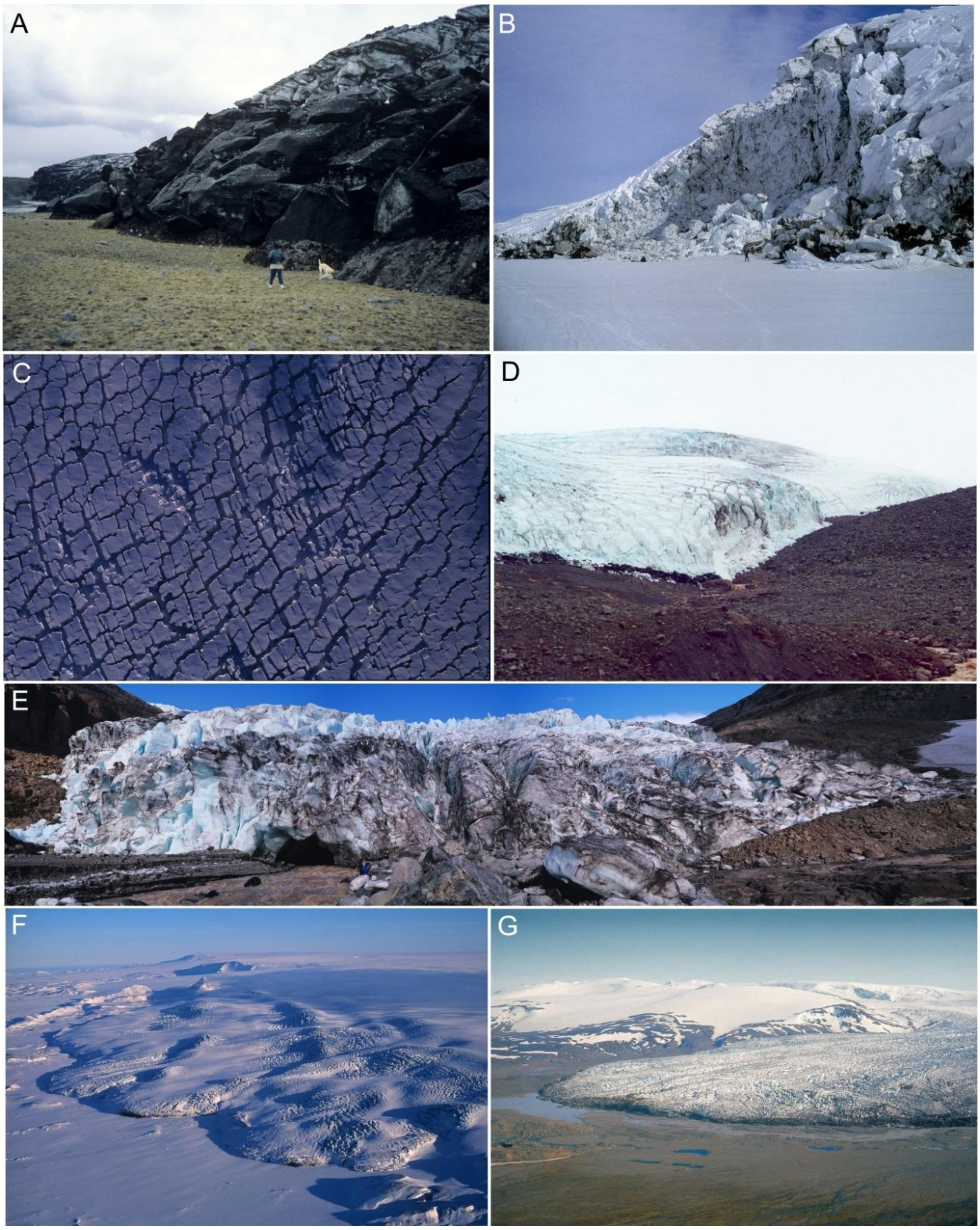

Fig. 2. Examples of Icelandic glaciers in surge: A) Skeiðarárjökull ice front during the 1991 surge; B) Síðujökull very steep ice front during its surge in 1994; C) Glacier surface behind the front of Síðujökull 1994 surge broken up into seracs formed by intersecting crevasses; D) Steep and

crevassed front of Leirufjarðarjökull during its 1995 surge; E) Kaldalónsjökull ice front during its 1995 surge; F) Tungnaárjökull ice front and crevassed surface during the 1994 surge; G) Eyjabakkajökull surging in 1973. Photos A-F courtesy of Oddur Sigurðsson; photo G by Richard S. Williams jr.

Combining the historical records of ice-front variations and glaciological field research, Björnsson et al. (2003) and Björnsson and Pálsson (2008) summarized the geographic distribution of surging glaciers, their subglacial topography, the frequency and duration of surges, changes in glacier surface geometry during the surge cycle, and measured velocity changes compared to calculated balance velocities. Björnsson et al. (2003) pointed out that all major outlets of the Vatnajökull ice cap are surge-type glaciers, and that surge-affected areas of Vatnajökull occupy approximately 75% of the ice cap (Fig. 1). They also recorded the indicators of surge onset and described changes in ice, meltwater and suspended sediment fluxes during a surge. They show that surge-type glaciers in Iceland are characterized by gently sloping surfaces and that they move too slowly to remain in balance given their accumulation rate, and that surge frequency was neither regular nor clearly related to glacier size or mass balance. Aðalgeirsdóttir et al. (2005) found that the mass transport during surges of Vatnajökull outlets could be up to 25% of the total ice flux of individual glaciers, and that this could affect the whole ice cap, the location of the ice divides, the flow field and the size and shape of the ice cap. Fischer et al. (2003) suggested that a surge cycle on Sylgjujökull and Dyngjujökull, outlets of the Vatnajökull ice cap (Fig. 1), spans several years, with slowly increasing motion over an extended area in the beginning, and more pronounced velocity changes during the active surge phase lasting 1-2 years. They further suggested that the most active surge phase lasted for about 1 year for these glaciers.

Surging glaciers are of great interest in glaciology because they can shed light on dynamic instabilities and threshold behaviour in glacier systems. Glaciological research in Iceland in general has focused on glacier distribution as an effect of climatic and topographical conditions, glacio-meteorology, glacier geometry (including extensive mapping of subglacial topography), as well as glacier mass balance, glaciohydrology, jökulhlaups and modelling glacier responses to climate change, whereas research on surging glaciers has specially focused on dynamic behaviour of the glacier during their most active surge phase, surge periodicity and meltwater production associated with surges (Aðalgeirsdóttir et al., 2005; Aðalgeirsdóttir et al., 2011; Aðalgeirsdóttir et al., 2006; Björnsson, 1982, 1998, 2009; Björnsson and Pálsson, 2008; Björnsson et al., 2003; Flowers et al., 2003; Gudmundsson et al., 2011; Magnússon et al., 2005; Marshall et al., 2005; Pálsson et al., 2012; Pálsson et al., 1991).

Glacial geological studies of surging glaciers in Iceland have had a different focus from the glaciological research (Benediktsson et al., 2009; Bennett et al., 2004a; Croot, 1987; Evans, 2011; Evans et al., 1999; Evans and Rea, 1999; Kjær et al., 2006; Sharp, 1985a; Sharp, 1985b). The motivation has been that ice streams and surging glaciers are dynamic constituents of the glacial system that influence and control form, flow, discharge and stability of present and former ice sheets and ice caps, and a better understanding of basal processes is particularly important for fast-flowing ice streams because of their crucial role in ice sheet dynamics (Boulton, 2010; Cofaigh and Stokes, 2008; Dowdeswell et al., 2004; Evans and Rea, 2003; King et al., 2009; Nelson et al., 2005). Research on surging glaciers is important for understanding the causal mechanisms of modern and past ice sheet instabilities and their contribution to sea level rise, particularly in changing climatic conditions, and studying the geomorphological and sedimentological products of surge-type can allow glacial geologists to better interpret evidence for past glacier activity and its climatic implications (Domack et al., 2005; Dowdeswell et al., 1995; Evans and Rea, 2003; Ottesen et al., 2008). Surge-type glaciers provide an opportunity to address important questions about the basal boundary conditions of fast flowing ice, in particular the significance of sediment deformation and sliding/subglacial decoupling (Clarke, 2005; Kjær et al., 2006; Murray, 1997). Icelandic surge-type glaciers have been intensely studied for better understanding glacier-induced stresses and ice-flow mechanism, basal temperature and hydrology, as well as the processes at work in the sub-marginal and ice-marginal zones (Andrzejewski, 2002; Benediktsson, 2012; Benediktsson et al., 2009; Benediktsson et al., 2008; Benediktsson et al., 2010; Bennett, 2001; Bennett et al., 2004a; Bennett et al., 2004b; Croot, 1988b; Fuller and Murray, 2000; Fuller and Murray, 2002a; Nelson et al., 2005; Russell et al., 2001; Schomacker and Kjær, 2007). Outstanding research questions include if there is a mass balance or climatic control on surges (Benn and Evans, 2010; Copland et al., 2011; Dowdeswell et al., 1995; Striberger et al., 2011), and how rapid ice flow can be sustained through a surge (Alley et al., 1987; Benediktsson et al., 2008; Engelhardt and Kamb, 1998; Kjær et al., 2006)? Between surge events, the glaciers retreat and landform associations and sediment successions re-emerge, imprinted with information on sub-glacial and ice-marginal driving processes (Benediktsson et al., 2008; Bennett et al., 2000a; Kjær et al., 2008; Kjær et al., 2006; Schomacker et al., 2014; Schomacker et al., 2006; Sharp, 1985b; Sharp and Dugmore, 1985; Waller et al., 2008). Outstanding research questions thus also concern the actual geological fingerprinting of surges, if different types of surging glaciers produce different sediment-landform assemblages, and what is the impact of surges on sediment distribution (Benn and Evans, 2010; Brynjólfsson et al., 2012; Evans and Rea, 2003; Schomacker et al., 2014)? A question raised by Bennett (2001) is what palaeoglaciological and environmental inferences, if any, can be made from the occurrence of large, often multi-crested push moraines in the geological record, and if they are a characteristic landform in front of surge-type glaciers (Croot, 1988a)? Studies of surge-type glacier landsystems (Evans and Rea, 1999) have highlighted research questions related to the genesis of e.g. crevasse fill ridges, concertina eskers and streamlined landforms, such as drumlins (Johnson et al., 2010; Jónsson et al., 2014), and if they might be indicative of fast flow? This review aims to give an overview of advances in our understanding of the surging glacier landsystem, processes and products that have emerged from the research in Iceland, as well as outlining future research challenges.

2. Case studies

2.1 Vatnajökull – All major outlet glaciers surge

Vatnajökull is the largest ice cap in Iceland, covering about 8000 km^2 and containing approximately 3100 km^3 of ice (Björnsson and Pálsson, 2008). All of Vatnajökull' s major outlets are surge-type glaciers (Fig. 1), and approximately 75% of the ice cap has been affected by surges (Björnsson et al., 2003). The magnitude of the surge impact is illustrated by Björnsson and Pálsson (2008) as they report that during the 1990's alone, ~3000 km^2 of Vatnajökull (38% of the icecap area) was affected by surges, which transported about 40 km^3 of ice from accumulation zones to ablation areas. This amounted to approximately 25% of the total ice flux to ablation areas during this period. The Vatnajökull case studies below include primarily recent glacial geological studies on surges and their sediment and landscape impacts.

2.2 Brúarjökull – extreme surge velocities, processes and landforms

Brúarjökull, a northern outlet of the Vatnajökull ice cap in eastern Iceland (Fig, 1) has experienced some of the largest and fastest surges known to have occurred, with major velocity fluctuations switching between active surging of a few months' duration and quiescent phases lasting from 70 to 90 years (Raymond, 1987; Thorarinsson, 1969; Todtmann, 1960). Historical surges of Brúarjökull occurred in 1625, ca. 1730, ca. 1775, 1810, 1890 and 1963–1964 (Björnsson et al., 2003; Thorarinsson, 1964a, 1969). During the two most recent surges, initiated in 1890 and 1963, the glacier advanced 8-10 and 9 km, respectively, affecting an area of more than 1400 km^2 (Guðmundsson et al., 1996; Thorarinsson, 1969). The peak velocity of ice front advance was above 120 m day^{-1} over a period of three months, which exceeds the fastest ice streams in Antarctica and Greenland (Echelmeyer and Harrison, 1990; Joughin et al., 2002; Scheuchl et al., 2012) and is three times the highest measured flow rate of Jakobshavns Isbræ in Greenland (Scheuchl et al., 2012). Presently, Brúarjökull is in its quiescent phase with the ice margin retreating up to 250 m yr^{-1} (Kjær et al., 2006) and down-wasting generating thin cover of sediments from emerging and disintegrating crevasse-squeeze ridges and debris bands in the ice (Kjær et al., 2006; Schomacker and Kjær, 2007).

It has long been reasoned that deformation of subglacial sediment is an important contributor to the flow of ice streams and warm-based outlet glaciers (Boulton and Hindmarsh, 1987; Evans et al., 2006; Piotrowski et al., 2001; van der Meer et al., 2003) and that it may impact the periodicity of surge-type glaciers (Boulton and Hindmarsh, 1987; Evans and Rea, 1999; Murray et al., 2003). The favoured explanations of how rapid ice flow velocities were reached by ice streams or surging glaciers have been through modes of basal motion where (Fig 3A) decoupling of a glacier from its bed enables fast ice flow through enhanced basal sliding across the ice/bed interface or very shallow subglacial deformation (Engelhardt and Kamb, 1998), or (Fig. 3B) where fast ice flow is sustained by deformation of water-saturated subglacial sediment that is strongly coupled to the glacier (Alley et al., 1989; Bennett, 2003).

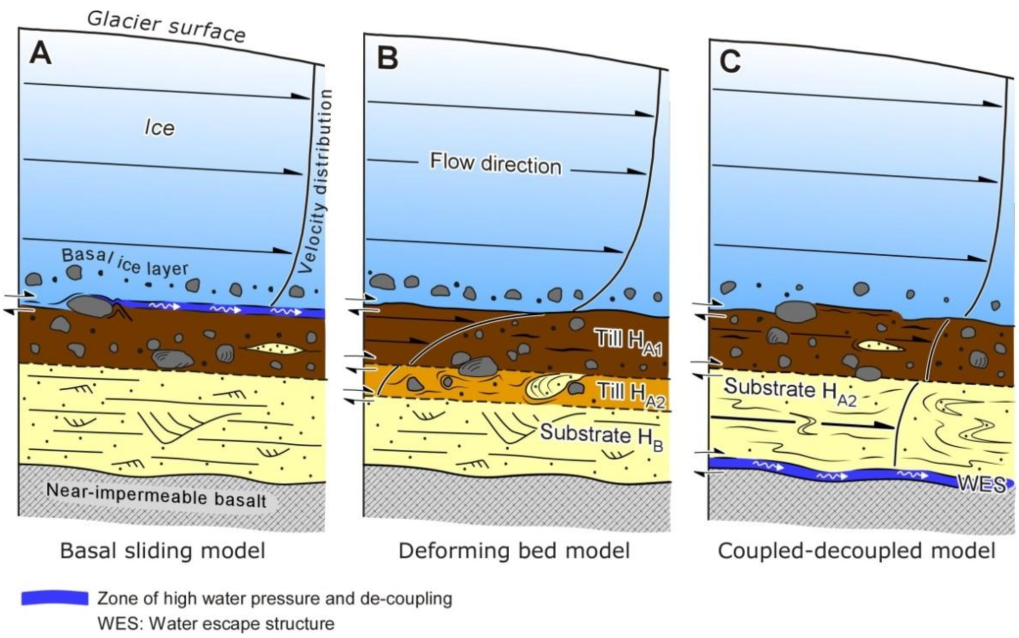

Fig. 3. Basal motion models associated with ice streams and surging glaciers. (A) Decoupling is sustained by enhanced basal sliding across the glacier–till interface with limited or no subglacial deformation. (B) The glacier is coupled to its bed and fast ice flow is sustained through subglacial deformation of water-saturated sediment with a low effective pressure. A horizon with fast deforming sediment and high strain rates (HA1) overlies a horizon with more slowly deforming sediment and low strain rates (HA2) that is superimposed on stable horizon without deformation (HB) (C) The new model of Kjær et al. (2006) from Brúarjökull where the glacier is coupled to its bed as expressed in slow subglacial deformation and densely fluted till plane, while the substrate is decoupled from the bedrock leading to fast ice flow and a substantial dislocation of subglacial

sediments. Water escape structures indicate that water and sediment were forced along a near-impermeable bedrock surface leading to substrate separation due to over-pressurized water. Slightly modified from Kjær et al. (2006)

2.2.1 Mechanisms of rapid basal flow and sediment dislocation during Brúarjökull surges

Kjær et al. (2006) suggested a third mechanism to explain motion of surging glaciers where subglacial deformation was one primary mechanism for sustaining rapid ice flow and high sediment discharge during Brúarjökull surges. They presented data that support a model where the extremely rapid ice flow observed during the 1963-1964 surge (>120 m day^{-1}) was sustained by over-pressurized water causing decoupling at the bedrock beneath a thick sediment sequence that was coupled to the glacier (Fig. 3C). In the model of Kjær et al. (2006) the glacier is coupled to its bed as expressed in slow subglacial deformation, while the substrate is decoupled from the bedrock leading to fast ice flow and a substantial dislocation of sediments. Water escape structures indicate that over-pressurized water and sediment were forced along a near-impermeable bedrock surface leading to substrate separation. The decoupling between the bedrock and a sediment sequence that is strongly coupled to the ice is a mechanism that is dependent upon a range of factors: (i) the subglacial bed and the foreland must be comprised of low-permeable fine-grained sediments that can support very high porewater pressures, (ii) the rate of water input to the sediments from basal melting or upstream sources must be in excess of the permeability of the sediments in order to raise porewater pressures, and (iii) the presence of a weak horizon or stratigraphic discontinuity at depth is essential for the decoupling and associated hydrofracturing. The physics and mechanisms of subglacial sediment deformation and detachment of sediments are debated, and the deformation mode remains one of the most controversial elements of glacier dynamics (Benn and Evans, 2010; Boulton and Hindmarsh, 1987; Clarke, 2005; Cuffey and Paterson, 2010; Damsgaard et al., 2013). The model of Kjær et al. (2006) that attributes detachment of sediments to elevated pore-water pressures at the base of the deforming layer has caught considerable attention. However, there is no single unifying theory that explains large-scale dislocation/rafting of subglacial sediments. Rafts may e.g. be dislocated as a result of being frozen to the base of cold-based glaciers; or be the result of detachment associated with elevated pore-water pressures along a basal décollement; or along water-rich décollement surfaces within the subglacially deforming layer; or as a consequence of subglacial hydrofracturing by forceful upward dewatering (Aber, 1985; Benn and Evans, 2010; Hart, 1995b; Moran et al., 1980; Phillips et al., 2008; Ruszczynska-Szenajch, 1987; Vaughan-Hirsch et al., 2013).

2.2.2 – Surge sediments and landforms

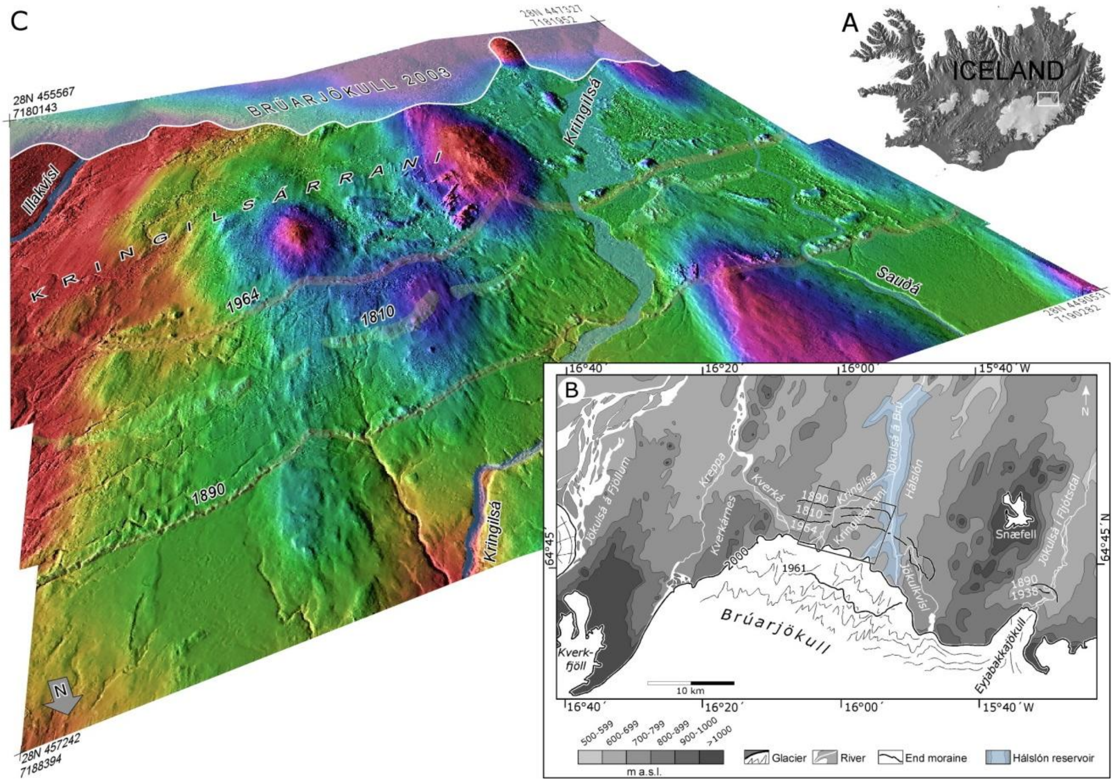

Fig. 4. The Brúarjökull forefield. A) Location of Brúarjökull (white square) draining Vatnajökull ice cap towards north; B) Brúarjökull setting and end moraine systems. Black square indicates study area; C) Terrain-Shaded Relief draped over a digital elevation model based on a 3m grid generated from orthorectified 1:15,000 aerial photographs recorded in 2003. Modified from Benediktsson et al. (2008).

As a result of repeated surges, the forefield of Brúarjökull consists of a 6–7m thick sediment sequence overlying basaltic bedrock. Despite its relatively remote and inaccessible location, the geomorphology of the Brúarjökull forefield is comparatively well known. A detailed pre-1963-1964 surge survey of the forefield was provided by (Todtmann, 1960). A broad geomorphological map from the central part of Brúarjökull was delivered by Evans and Rea (1999) and Evans et al. (2007) mapped the entire margin of Brúarjökull at the scale of 1:30.000. In connection with a major research effort over three field seasons in 2003-2005 about 64 km^2 of the Brúarjökull central forefield was carefully mapped at a scale 1:16.000 by Kjær et al. (2008), including more than 20,000 landforms. The forefield is glacially streamlined with end-moraine ridges, ice-cored landforms and ice-free hummocky moraines, crevasse fill ridges, eskers and flutings located in shallow basins between widely spaced elongated bedrock hills (Kjær et al., 2006, 2008; Schomacker et al., 2006). Three distinct end moraines, originating from the three last surges, are present in the glacier forefield (Fig. 4). Evans et al. (2007) and Kjær et al. (2008) suggested that the distribution of landforms on the Brúarjökull forefield has close resemblance to landform assemblages of paleo-ice streams, and that the present terrain surface at Brúarjökull is the cumulative result of multiple landform generations because each surge has superimposed a new association of landforms on older surfaces (Fig. 5). The simplest landscape architecture occurs outside the 1810 ice margin, where the 1890 surge advanced over hitherto un-deformed sediments. Proximal to the 1810 ice margin, the landscape have been transgressed by two overriding glaciers (in 1890 and 1964). The most complex landscape architecture is found proximal to the 1964 ice margin, where the impact of four surges is evident. Kjær et al. (2008) found no obvious pattern in the distribution between drumlins and flutes, but suggested the formation of drumlins to be tied to re-moulding of pre-existing landforms.

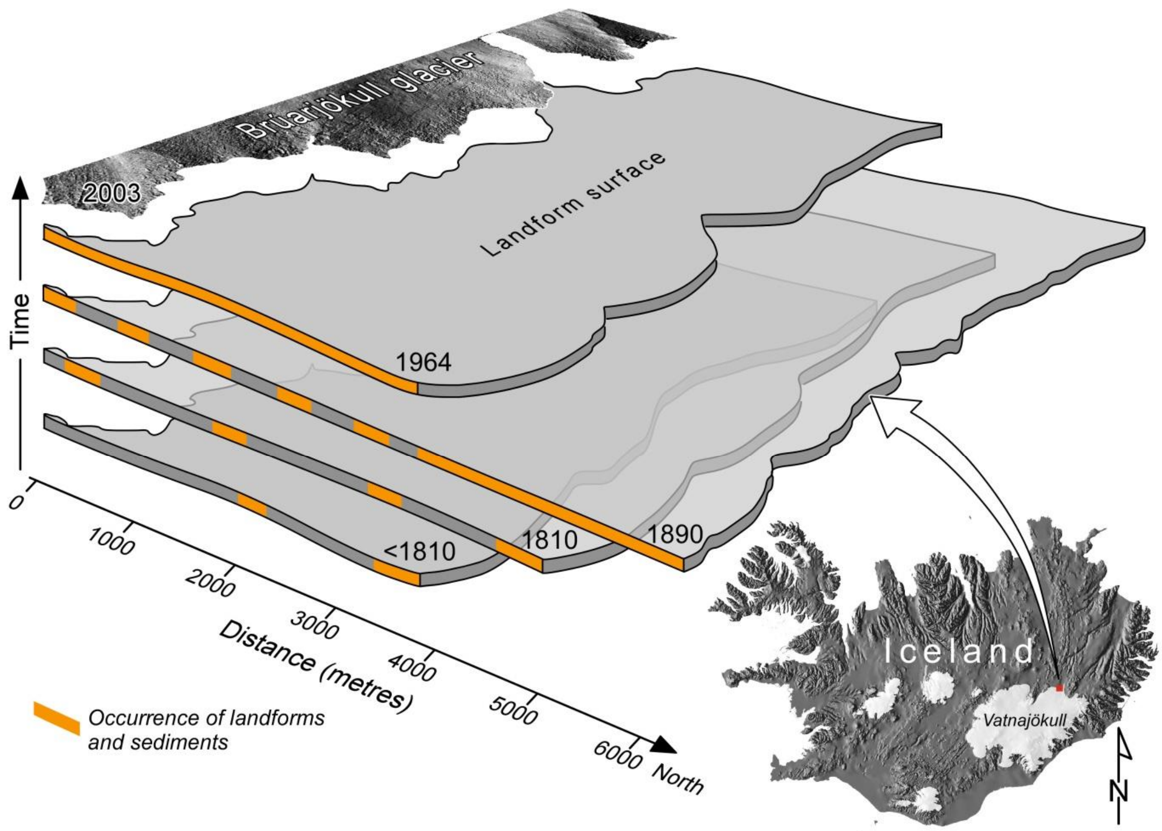

Fig. 5. The present terrain surface of Brúarjökull is the cumulated result of at least four surge events: pre-1810, 1810, 1890 and 1963-64. Illustration from Kjær et al. (2008).

The Brúarjökull 1890 and 1963-64 surges left major terminal moraines that can be followed across the forefield (Fig. 4). It was demonstrated by e.g. Croot (1988b) and Bennett et al. (2004a) that large end-moraine ridges formed by surging glaciers in Iceland chiefly originated from glaciotectonism, primarily as a result of ice pushing into a pre-existing foreland wedge. The landsystem model for surging glaciers (Evans et al., 1999; Evans and Rea, 1999; Evans and Rea, 2003; Evans et al., 2007), where large thrusted and stacked terminal moraines are recognized as a distinguishing component of the geomorphological fingerprinting of surge-type glaciers, is fundamentally based on observations from Brúarjökull and other Icelandic surging glaciers. Glaciotectonic end moraines have for long been understood to be the result of thrusting associated with an imbricate thrust stack formed in front of an advancing glacier (Bennett, 2001; Boulton et al., 1999; Krüger, 1985; Moran et al., 1980; Ruszczynskaszenajch, 1987), but it has neither been fully understood what controls the morphological and structural characteristics of surging-glacier glaciotectonic end moraines, nor how their properties are related to the glacier dynamics (Fig. 6). A study by Benediktsson et al. (2008) investigated the morphology, sedimentology and tectonic architecture of the large 1890 Brúarjökull surge moraine. The work resulted in a sequential model that illustrates the stepwise formation of a surging-glacier end moraine (Fig. 7).

Benediktsson et al. (2008) and Benediktsson (2012) showed that as a result of substrate/-bedrock decoupling during the 1890 surge, subglacial sediment was dislocated across the bedrock surface, conforming with the model of Kjær et al. (2006), and deformed compressively (Fig. 7). This led to gradual substrate thickening and the formation of a sediment wedge in the marginal zone. A drop in subglacial porewater pressure at the very end of the surge led to substrate/bedrock coupling and a stress transfer up into the sediment sequence causing brittle deformation of the substrate. Simultaneously, the glacier toe ploughed into the topmost part of the marginal sediment wedge initiating the moraine-ridge construction. Fine-grained and incompetent sediment deformed in ductile manner, resulting in a narrow moraine dominated by rooted folds, while coarse-grained and competent sediment deformed in brittle fashion, resulting in a wider moraine dominated by thrust blocks. Thus the ice-marginal position of the 1890 surge is marked by a twofold end-product: a sedimentary wedge in the marginal ca. 500m and the end-moraine ridge as a surface expression of this wedge. By comparing this to observed ice-flow velocities during the 1963-64 surge, Benediktsson et al. (2008) concluded that the sedimentary wedge is thought to have formed within approximately 5 days and the moraine ridge in about 1 day. Therefore, they termed this an “instantaneous end moraine”.

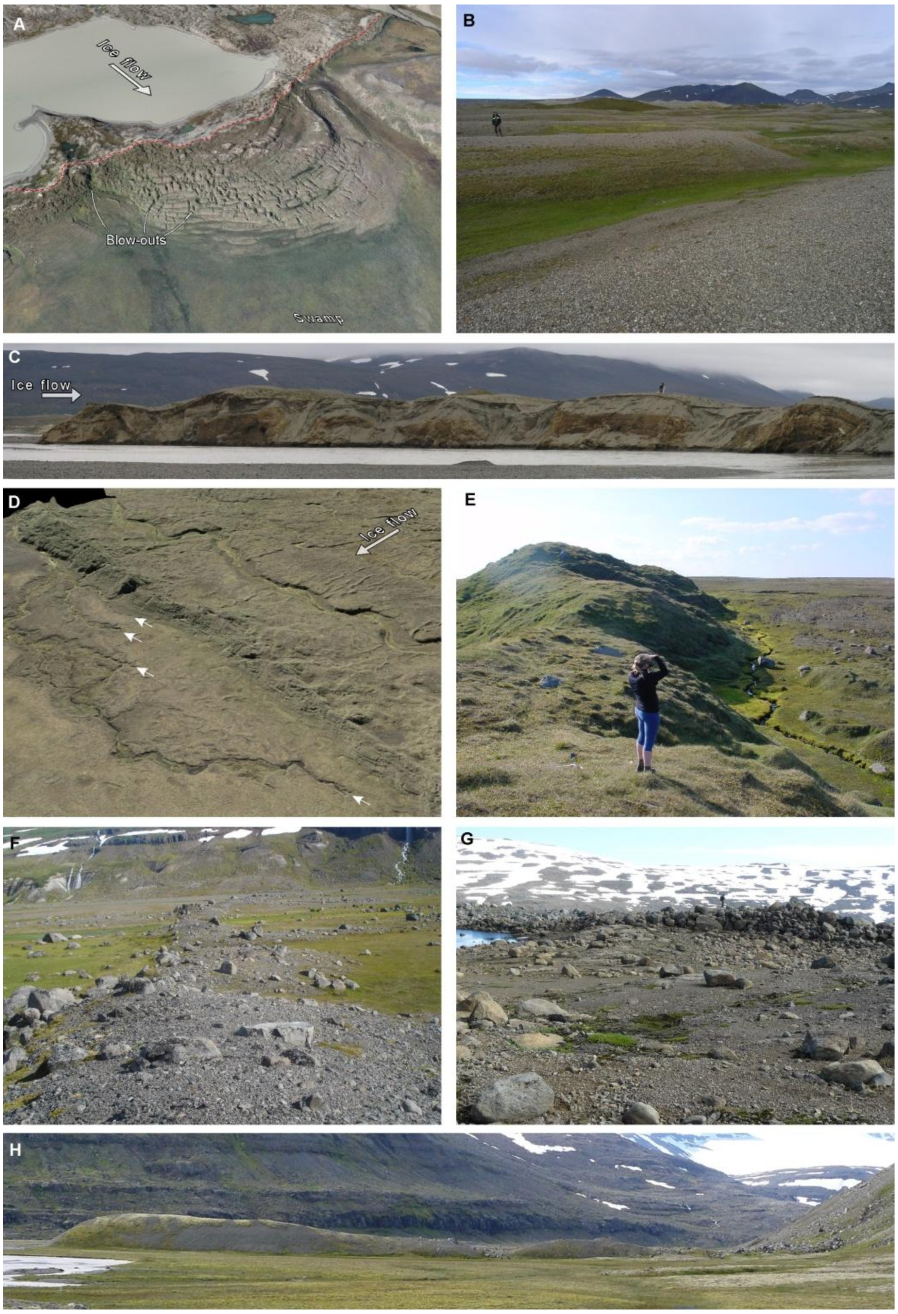

Fig. 6. A) Large ridges in the central part of the 1890 end moraine at Eyjabakkajökull, consisting of multiple closely spaced, asymmetric narrow crests. A number of semi-circular blow-out depressions and associated channels dissect the foreslope. The ridge is c. 280 m wide. B) The eastern part of the Eyjabakkajökull 1890 moraine with multiple regularly spaced symmetric crests. Ice flow was from upper left to lower right. C) A section through the most proximal part of the ridge shown in B. Note person for scale. D) 3D aerial view of the Brúarjökull 1890 end moraine in Kringilsárrani (central

forefield). The moraine is about 40-80 m wide and up to 20 m high (upper left). Semi-circular blow-out depressions at the abrupt head of channels are arrowed. E) Ground-view of the moraine visible in the upper-left corner of photograph D. Ice flow was from left to right. F) A low relief end moraine eroded by the glacial river in Leirufjörður, formed c. 1885. About 1-4 m high and 10-15 m wide constituting of coarse-grained till and boulders. G) Bouldery part of the 1846 terminal moraine at Reykjafjarðarjökull, reflects restricted sediments supply at its location. H) The 1846 terminal moraine at Reykjafjarðarjökull, approximately 10-15 m high and 15-25 m wide, constituted of gravel pushed up from the outwash plain.

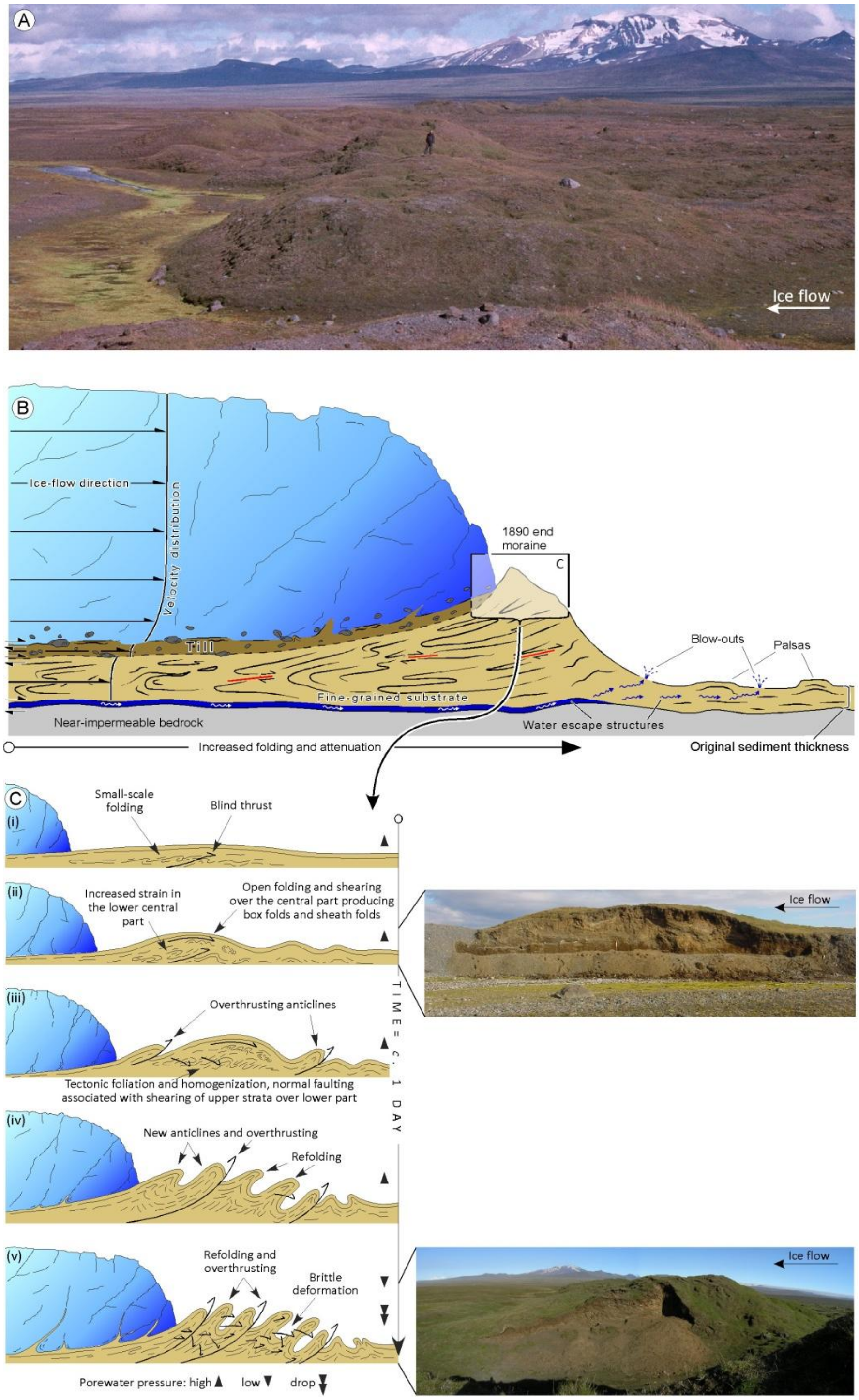

Fig. 7. Sequential model of the Brúarjökull 1890 surge sediment-wedge and terminal moraine formation. A) View along the 1890 end moraine. B) Conceptual model illustrating the formation of the ~500m long sediment in the marginal zone. Note the velocity distribution indicating the greatest

displacement at the substrate-bedrock interface. The displacement at this stratigraphic boundary results in net sediment flux into the marginal zone and the formation of the sediment wegdge. C) Model showing the structural evolution of the actual end moraine that formed on top of the wedge at the last day of the 1890 surge. The inset photographs show sections through the moraine at different stages of the end-moraine continuum. The section by stage ii cuts across the whole moraine ridge while the section at stage v only covers the distal slope. Modified from Benediktsson et al. (2008) and Benediktsson 2012.

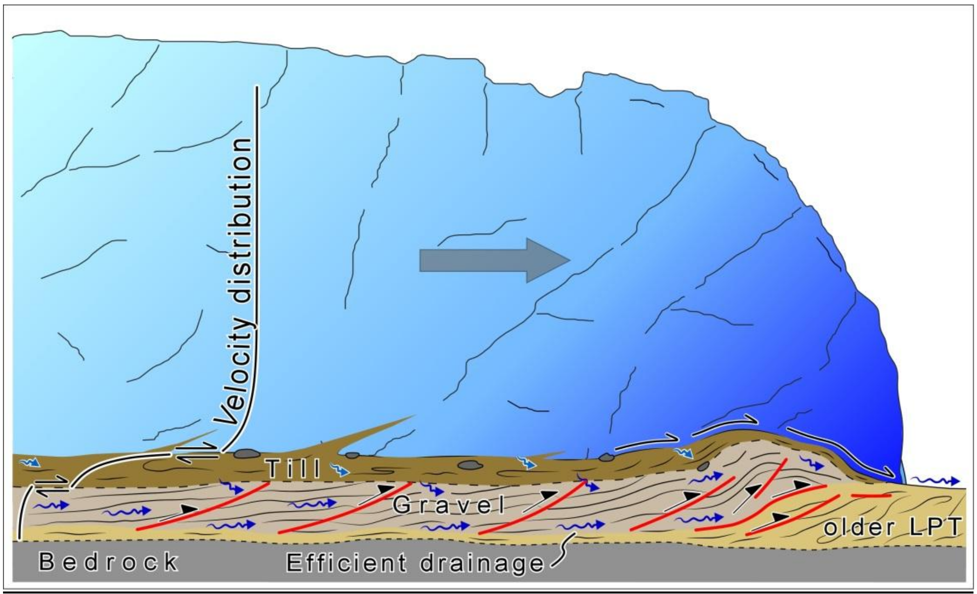

Fig. 8. A model explaining the formation of the 1964 submarginal end moraines at Brúarjökull. Due to high hydraulic conductivity (larger blue arrows) and shear strength of the gravel, it mainly deformed through thrusting while high porewater pressure in the till resulted in higher strain rates. The principal velocity (greatest displacement) was thus located within the till (see velocity profiles in Figs 2 and 5 for comparison), which allowed the glacier to override the gravel thrust sheets that constitute the end moraine at the very end of the surge. Modified from Benediktsson et al. (2009).

Benediktsson et al. (2009) studied the morphology, sedimentology and architecture of an end moraine formed by the ca. 9 km surge of Brúarjökull in 1963–64 (Figs. 4 and 5) and related to ice-marginal conditions at the surge termination. They found that different mechanisms had operated along different parts of the glacier margin during the 1963–64 surge, resulting in the formation of either proglacial or sub-marginal end moraines. In the eastern part of the study area, the 1964 moraine resembles the 1890 moraine in geometry, morphology and sedimentary composition. They suggested that the model of Benediktsson et al. (2008) of a marginal sediment wedge, forming as a result of substrate/bedrock decoupling and associated sediment influx to the marginal zone, and an end moraine forming on top of the wedge in response to a sudden drop in subglacial pore water pressure at the very end of the surge (Fig. 7), was applicable for the eastern part of the 1964 end moraine. The western part of the end moraine contrasts the eastern part in terms of morphology, sedimentary composition and position relative to the ice margin and thus required a different genetic explanation. There, the sedimentological and structural data suggest the end moraines formed in areas of a relatively high hydraulic conductivity, where subglacial traction till overlies glaciofluvial gravel. They suggested a model for sub-marginal end moraine formation (Fig. 8) where high strain rate in the till, favoured by high porewater pressure, facilitated the glacier advance through shear deformation. In contrast to the till, porewater pressure was low in the underlying gravel, resulting in higher shear strength and lower strain rates. Thus, the principal velocity component was located within the till, allowing the glacier to override the gravel thrust sheets that constitute the end moraine. Submarginal end moraines thus formed as a consequence of changes in the subglacial hydrology and ice-flow mechanism at the very end of the surge. It is known from the surge monitoring and aerial photos that the 1963-1964 surge terminal moraines formed at the closing stage of the surge (Thorarinsson, 1969), and this is also supported by the local origin of the sediment within the end moraine and the high ice flow velocities (100–120 m/day) of the surge. Benediktsson et al. (2009) suggested that submarginal end moraines only occur in the marginal zones of rapidly flowing glaciers.

2.2.3 Quantifying spatial distribution and volumes of sediment transport during the Brúarjökull 1963-1964 surge

Korsgaard et al. (in review) used time-series of Digital Elevation Models (DEMs) of the forefield of the Brúarjökull to quantify the volumes of material that was mobilized by the 1963-64 surge. The motivation for the study is that the mode and range of sediment mobility beneath ice-sheets has direct bearing on their stability and sensitivity to external forcing (Bougamont et al., 2014; Parizek et al., 2013; Smith et al., 2007). Because surge-type glaciers advance rapidly, stagnate and retreat within a few decades, remotely sensed data recorded before and after a glacier surge provide an opportunity to map the spatial pattern of erosion and deposition of glacial landforms for better understanding how fast ice flow impacts on glacial sediments and landforms. The DEMs were produced by stereo-photogrammetry on aerial photographs taken in 1961, before the 1963-64 surge, and aerial photographs taken after the surge, in 1988 and 2003. The analysis was performed on two DEMs of Difference (DoDs), i.e. a 1961-2003 DoD documenting the impact of the surge and a 1988-2003 DoD documenting the post-surge modification of the young surging glacier landsystem. Combined with a digital geomorphological map, the DoDs make it possible to quantify the impact of the surge on a landsystem scale down to individual landform scale. A total of 34.2×10^6 m^3 of material was mobilized in the study area as an impact of the last Brúarjökull surge. Of these, 17.4×10^6 m^3 of material were eroded and

16.8×10^6 m^3 were deposited (Fig. 9). More than half of the deposited volume was ice-cored. Interestingly, this study demonstrates that although the total mobilized mass volume is high, the net volume gain of material in the forefield as a result of the surge is low. Furthermore, deposition of new dead-ice from the 1963-64 surge constitutes as much as 64% of the volume gain in the forefield. The 1988-2003 DoD is used to quantify the melt-out of this dead-ice and other paraglacial modification of the recently deglaciated Brúarjökull forefield: at least 51-64%, of the volume deposited by the 1963-46 surge consists of dead-ice that is still melting out in the forefield. Since 1988, ~45% of the volume loss has been due to dead-ice meltout.

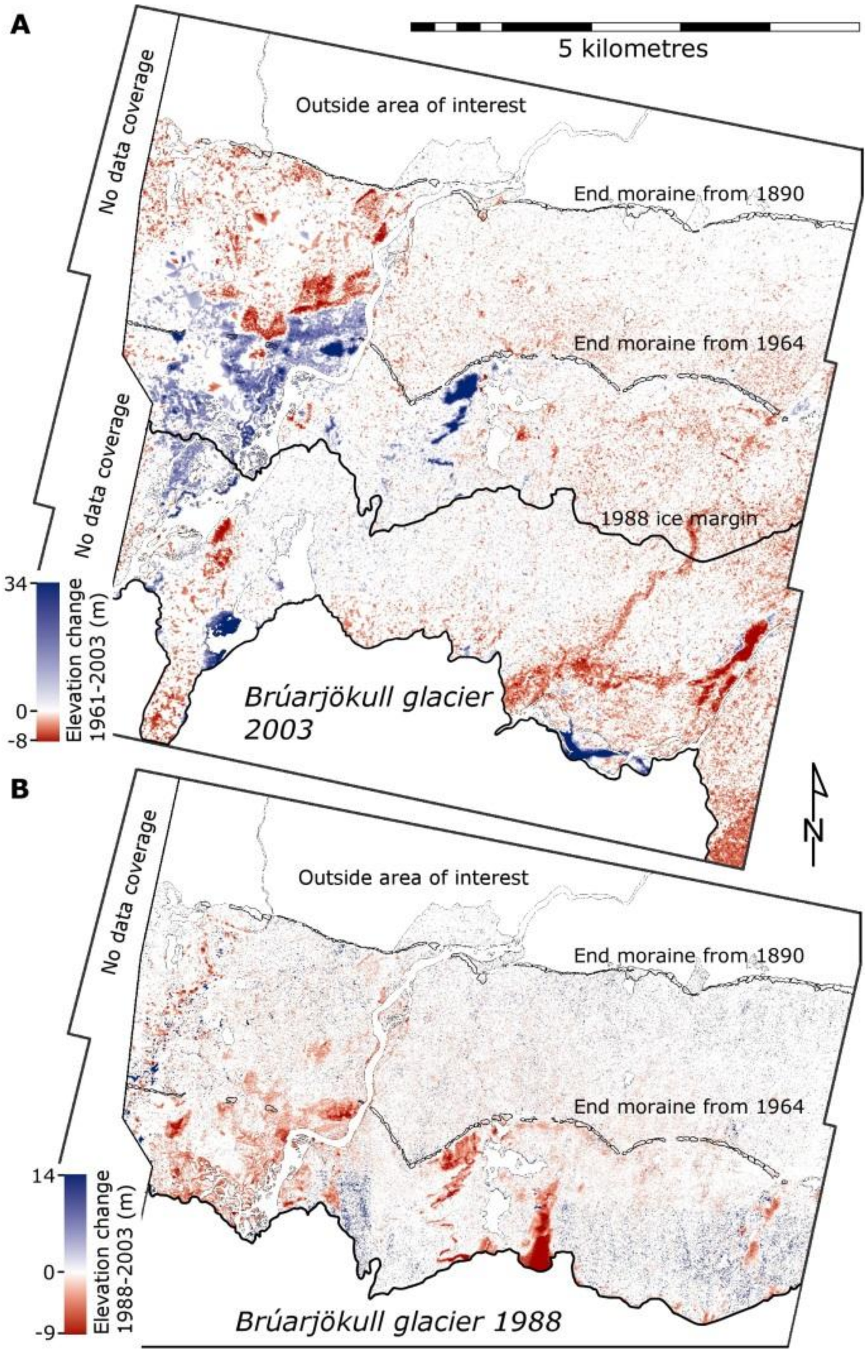

Fig. 9. Elevation change in the Brúarjökull forefield from (A) 1961 to 2003 and (B) 1988 to 2003 demonstrating the elevation gain (deposition, blue colours) and loss (erosion and dead-ice melting, red colours). For reference, the end moraines from 1890 and 1964 are indicated on the map, as are the Brúarjökull glacier margin positions for 1988 and 2003.From Korsgaard et al. (in review).

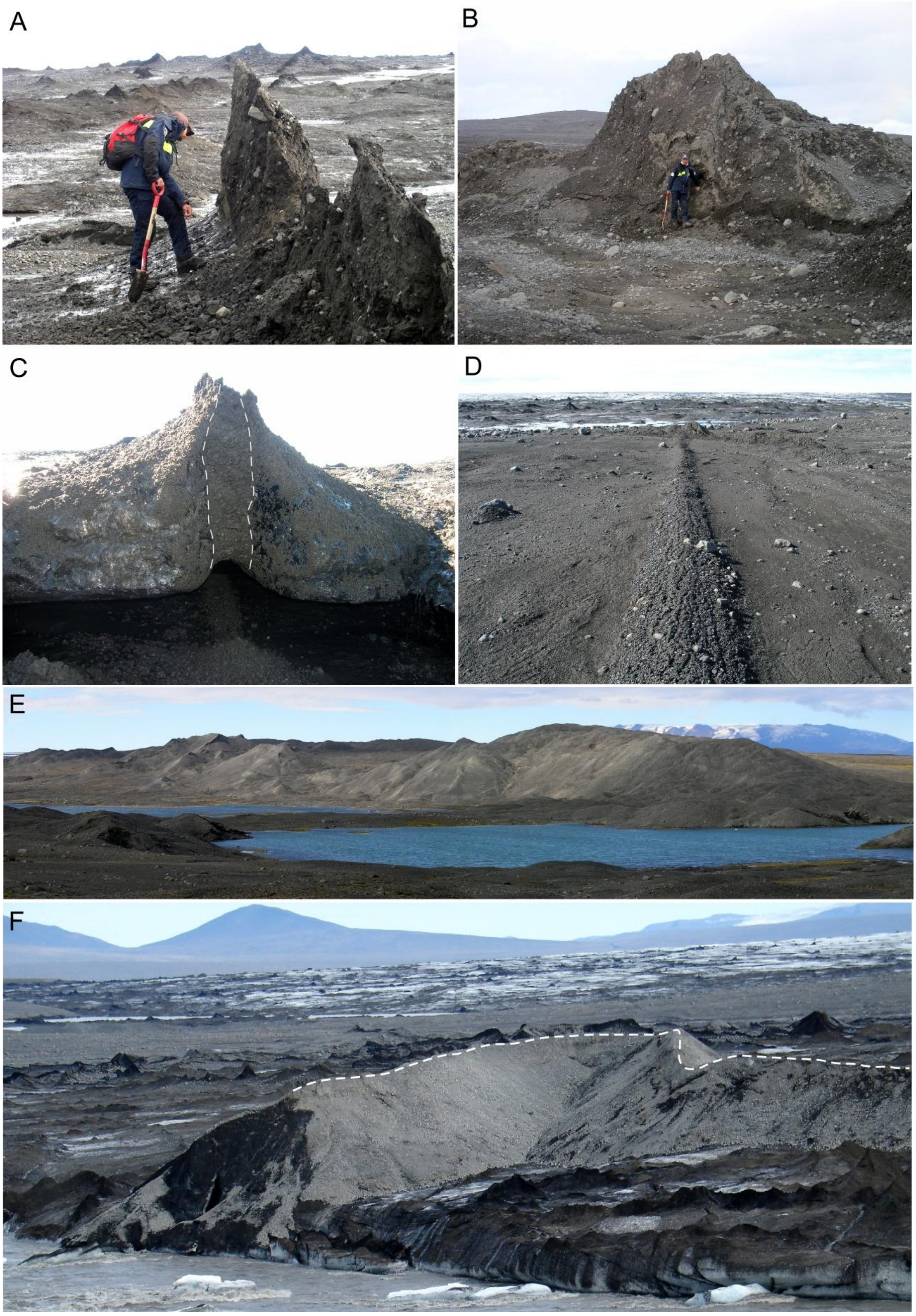
A
B
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Fig. 10. A) Crevasse fill ridge melting out of Brúarjökull ice front. B) Large crevasse fill ridge in the Brúarjökull forefield.C) Crevasse fill in ice at Brúarjökull. D) A flute in front of the Brúarjökull margin, with a crevasse fill ridge on top in the distance. E) Large concertina esker formed during in connection with the 1963-64 Brúarjökull surge. F) Concertina esker melting out of the ice at Brúarjökull. Zig-zag form of the esker ridge outlined with dashed lines. Photos A, B, C and E by I.Ö.Benediksson, D and F by Ó. Ingólfsson.

2.2.4 Genesis of crevasse fill ridges

Crevasse fill ridges (Sharp, 1985a) - also called crevasse squeeze ridges (Benn and Evans, 2010) - have been identified as a landform characteristic of surging glaciers, ever since Sharp (1985a) presented comprehensive analyses of the ridges from the foreland and ice margin of Eyjabakkajökull. The ridges are typically 1-2 m high and usually occur in fluted parts of the forefield. Eyjabakkajökull and Brúarjökull are classic localities for studying crevasse fill ridges, as both are retreating after surges in 1973 and 1964, respectively, exposing in their forefields and ice margins extensive fields of crevasse fill ridges (Fig. 10A-D) (Evans et al., 1999; Evans and Rea, 1999; Evans et al., 2007; Kjær et al., 2008; Schomacker et al., 2014). They have also been described from the forefields of other Icelandic surging glaciers, like Skeiðarárjökull after its 1991 surge (Waller et al., 2008), Tungnaárjökull after its 1994 surge (Evans et al., 2009), Hagafellsjökull Eystri (Bennett et al., 2003), Síðujökull (Kozarski and Szupryczinski, 1973), Múlajökull (Jónsson et al., 2014) and Sátujökull (Evans, 2011). There is a general consensus that crevasse fill ridges form as a consequence of sediment infilling basal surge crevasses from the bed upwards and subsequently melting out during quiescence. The mechanics of this process are, however, not entirely clear, mainly because of the lack of detailed sedimentological and geomorphological investigations. Rea and Evans (2011) assessed the potential for the formation of crevasses and their infilling with sediments, using linear elastic fracture mechanics approach and empirical data derived from the literature, for seven surging glaciers from Svalbard, Iceland, Greenland and Alaska. They concluded that crevasse squeeze ridges most likely resulted from the infilling of basal crevasses, driven for the most part, bottom-up, by high basal water pressures. Crevasse-fill ridges commonly occur with flutes, and they relate to each morphologically in three ways: a strike/slip displacement of the flute along the line of the ridge, undisturbed passage of the flute through the ridge, and a rise in the crest line of the flute to the ridge where they intersect (Sharp, 1985a). Bjarnadóttir (2007) found that crevasse squeeze ridges crossing flutes in the forefield of Brúarjökull had not been injected through the flutes as the flutes were undisturbed by the formation of the crevasse squeeze ridges. She proposed a model whereby sediment is injected into basal crevasses formed as the compressive surge-bulge passes, extended glacier flow sets in and the glacier thins. Till that previously was subject to high effective stress under the surge bulge was suddenly subject to lower pressure and injected up into basal crevasses, driven by pressure differences. Subsequently, the sediment is detached and transported englacially for some distance and deposited, forming a ridge on top of the fluted surface as it melts out of the dead ice during the quiescent phase (Fig. 10C,D). The crevasse fill ridge-fluted moraine relationships (Sharp, 1985a) suggest there might be more than one mode of crevasse fill ridge formation.

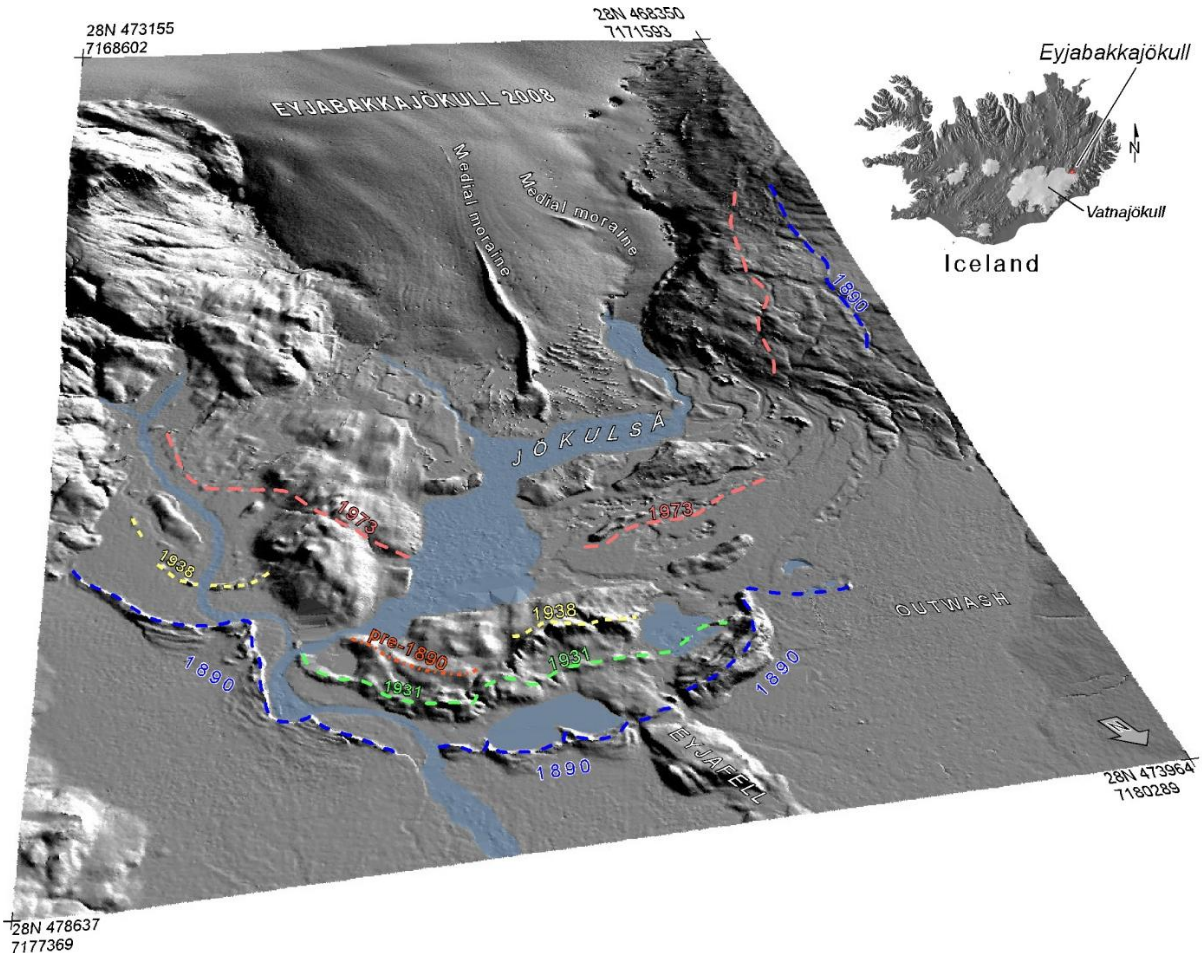


Fig. 11. The forefield of the surge-type glacier Eyjabakkajökull, eastern Iceland (modified from Benediktsson et al., 2010).

2.3 Eyjabakkajökull – surge history, geomorphic impact and an end moraine continuum

Eyjabakkajökull is a surge-type outlet glacier of the Vatnajökull ice cap (Figs. 1 and 11), composed of three distinct outlets from the main ice cap that descend from 1200–1500 m a.s.l. and combine to form a glacier tongue which is about 10 km long and 4 km wide where it terminates at around 700 m a.s.l. The recent surge history of Eyjabakkajökull documents surges in 1890, 1931, 1938 and 1972–1973 (Björnsson et al., 2003; Kaldal and

Víkingsson, 2000; Thorarinsson, 1938, 1943; Thoroddsen, 1914; Todtmann, 1953; Todtmann, 1960), with the most extensive advance (3–4 km) in 1890. The 1890 surge terminal position is marked by a conspicuous ridge complex (Figs. 11 and 12). The surges in 1931 and 1938 terminated ca. 250 and 750 m up-glacier from the 1890 moraines, respectively (Fig. 11) (Kaldal and Víkingsson, 2000), producing conspicuous glaciotectonic end moraines (Croot, 1987; Croot, 1988b). During the 1972–1973 surge, the debris-rich ice front advanced ca. 2 km, with maximum flow rates of 30 m day^{-1} (Björnsson, 1982; Clapperton, 1975; Sharp, 1985b; Sharp and Dugmore, 1985; Williams Jr., 1976).

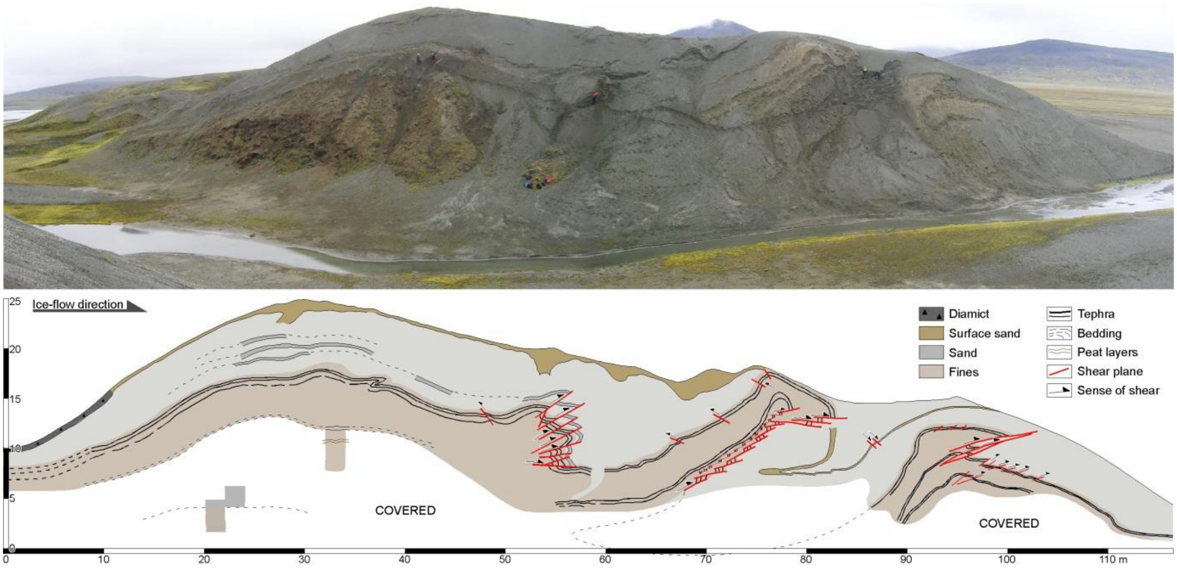


Fig. 12. A section through the western and largest part of the 1890 surge end moraine at Eyjabakkajökull. This part of the moraine complex is up to 500 m wide and 40 m high, and contains multiple widely spaced and broad symmetric crests in the central and distal zones of the ridge, but lower asymmetric crests on the proximal slope (not visible). The section on the photograph and in the scale diagram is 112 m wide and 25 m high covering the central and distal crest zones of the ridge. The architecture reveals both ductile and brittle deformation but is characterized by large-scale folding with intensive shearing in high strain areas. Section balancing revealed a total horizontal shortening of 39% and a décollement depth of 27 m. Note persons for scale in the section and a pile of backpacks in the center. Modified from Benediktsson et al. (2010).

2.3.1 The surge moraines

The Eyjabakkajökull surge moraines were described by Croot (1987, 1988) as ridges of two distinct elements: a subglacial extension zone with low-angle normal faults that are linked by a floor thrust to a proglacial compression zone signified by imbricate thrust sheets. The Eyjabakkajökull moraines have since been referred to as a classic example of the architecture of push moraines by e.g. Benn and Evans (2010), Aber et al. (1989), Bennett

(2001), and Aber and Ber (2007). Benediktsson et al. (2010) re-investigated the Eyjabakkajökull glaciotectonic moraines, focusing on what could explain the morphology, architecture and formation of the moraines and their relation to glacier dynamics (Fig. 12). Based on morphological, geological and geophysical data from terrain cross-profiles, cross-sections and ground penetrating radar profiles Benediktsson et al. (2010) demonstrated that three different qualitative and conceptual models were required to explain the genesis of the Eyjabakkajökull moraines. Firstly, a narrow, single crested moraine ridge formed at the distal end of a marginal sediment wedge that developed in response to decoupling of subglacial sediment from the bedrock and associated down-glacier sediment transport, following Benediktsson et al. (2008). Secondly, large lobate end-moraine ridges with multiple, closely spaced, narrow asymmetric crests formed by proglacial piggy-back thrusting. Thirdly, a new conceptual model (Fig. 13) shows that moraine ridges with different morphologies may reflect different members of an end-moraine continuum. This is true for the eastern and western parts of the Eyjabakkajökull moraines as they show similar morphological and structural styles that developed to different degrees. The former represents an intermediate member with décollement at 4-5 m depth and 27-33% shortening through multiple open anticlines that are reflected as moderately spaced symmetric crests on the surface. The latter represents an end member with décollement at about 27 m depth and 39% horizontal shortening through multiple high amplitude, overturned and overthrusted anticlines, appearing as broadly spaced symmetric crests. It is proposed that the opposite end member would be a moraine of multiple symmetric, wide open anticlinal crests of low amplitude. The data suggest that the glacier coupled to the foreland to initiate the end-moraine formation when it had surged to within 70-190 m of its terminal position. This indicates a time frame of 2-6 days for the formation of the end moraines, given an ice-flow velocity of ~30 m/day.

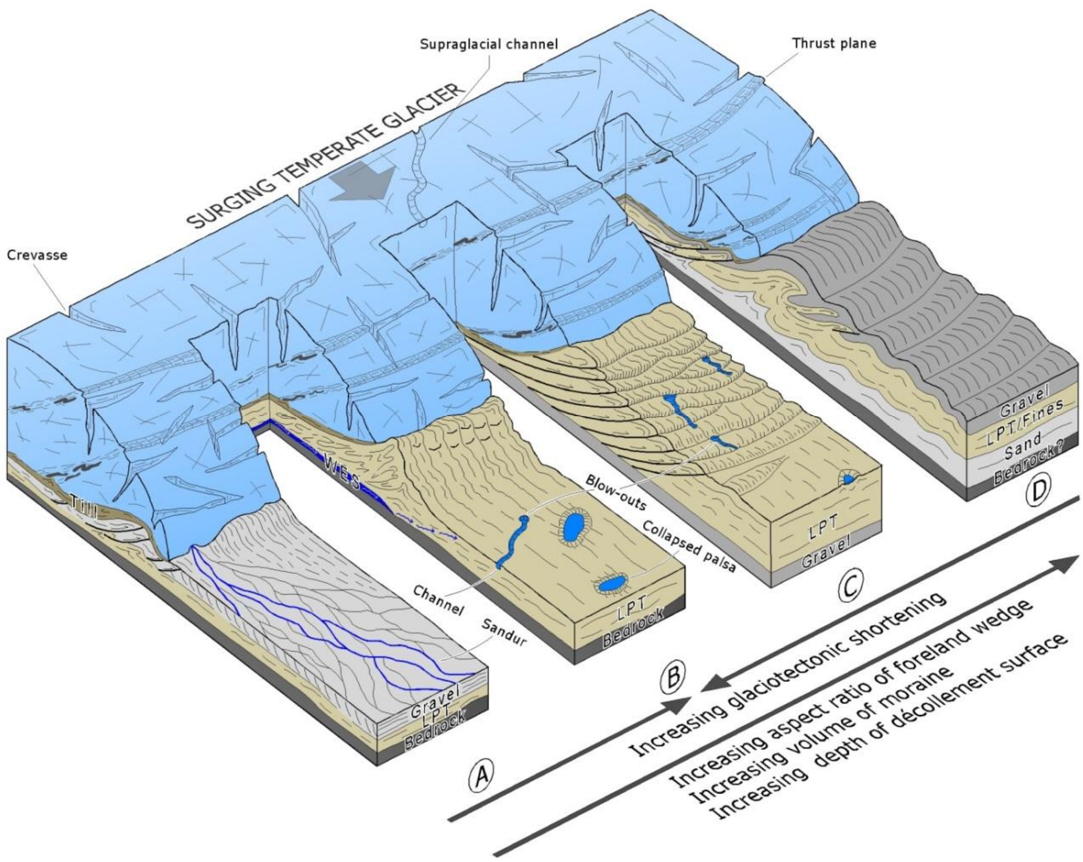


Fig. 13. A conceptual model summarizing the end moraine formation at Brúarjökull and Eyjabakkajökull. A) Where the glacier advances across glaciofluvial sediments, wedges of gravel and sand are thrusted and stacked beneath and in front of the ice margin to form a single- but broad crested end moraine. B) A single-crested moraine with multiple rooted folds in fine-grained sediments (loess-peat-tephra; LPT) forms on the distal end of a marginal sediment wedge. WES: water-escape structure. C) A wide moraine ridge with multiple asymmetric crests that represent thrust sheets in fine-grained material overlying gravel. D) A wide, multi-crested moraine with symmetric crests in the central and distal zones, reflecting large-scale folding of sand, fines and gravel, but asymmetric crests on the proximal slope, denoting submarginal thrusting. The horizontal shortening is greatest in moraine B (~60-80%), intermediate in moraines D and C (~40%) and smallest in moraine A (~20%). Modified from Benediktsson (2009).

2.3.2 Eyjabakkajökull landsystem model

Schomacker et al. (2014) concluded that the geomorphology of the Eyjabakkajökull forefield reflects the impact of multiple glacier surges and that the terrain resulted from several landform generations, in line with similar pattern identified at Brúarjökull by Evans et al. (2007) and Kjær et al. (2008). They identified individual landforms forming the landform assemblage that are typical of surging to include glaciotectonic end moraines, hummocky moraine, pitted outwash, flutes, crevasse-fill ridges, and concertina eskers, and published a modified landsystem model for Eyjabakkajökull (Fig. 14). They suggested that the landform distribution at Eyjabakkajökull agrees well with terrestrial paleo-ice streams and surge-type ice sheet lobes, and that Eyjabakkajökull provided a modern analogue with a high level of detail due to the well-preserved young landforms.

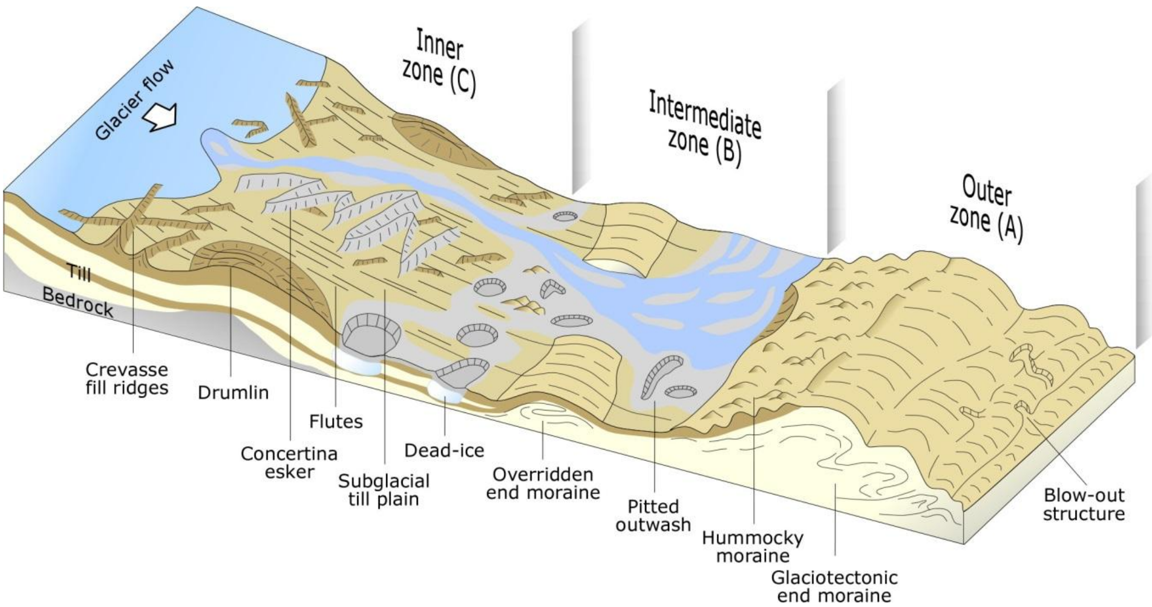


Fig. 14. Surging glacier landsystem model of Eyjabakkajökull. (A) Outer zone of glaciotectonic end moraines consisting of deformed pre-surge sediments. Hummocky moraine occurs on the backslope of the end moraine. (B) Intermediate zone of active, channeled outwash as well as inactive, pitted outwash deposited on top of stagnant ice. (C) Inner zone of subglacial till, flutes, drumlins, crevasse-fill ridges, and concertina eskers. Modified from Schomacker et al. (2014), partly based on Evans and Rea (1999, 2003).

2.3.3 Eyjabakkajökull concertina eskers

Ólafsdóttir (2011) and Schomacker et al. (2014) described and discussed the concertina eskers at Eyjabakkajökull. The concertina ridge elements are sharp-crested with slopes standing at the angle of repose, 30–650 m long, up to 12 m high, and 15–85 m wide, with ridges arranged in an en echelon pattern (Fig. 10E,F). The concertina eskers most proximal to the retreating glacier are ice-cored, whereas dead-ice was not observed in the distal-most ridge segments. The concertina eskers drape subglacial landforms like flutes and crevasse-fill ridges, suggesting a supraglacial origin. Concertina eskers, also referred to as 'zigzag, eskers by Benn and Evans (2010), have been have been regarded as a unique landform left by surging glaciers since they were first described from Brúarjökull by Knudsen (1995) and included in the surging glacier landsystem model of Evans and Rea (1999). However, they are not common features in the foreland of surging glaciers, and have hitherto only been described from Eyjabakkajökull and Brúarjökull (Evans et al., 1999; Evans and Rea, 1999; Kjær et al., 2008; Knudsen, 1995), from two surging glaciers on Svalbard (Hansen, 2003) and Novaya Zemlya (Grant et al., 2009), as well as from a deglaciation setting at a presently submarine setting in the Baltic Sea (Feldens et al., 2013). Knudsen (1995) originally proposed that concertina eskers were formed by shortening of pre-surge sinuous eskers by compression in the glacier snout during surging. Evans et al. (1999) likewise suggested they were deposited in englacial meltwater conduits and were crumpled during the passage of a wave-like bulge, whereas Evans and Rea (2003) suggested they were deposited in crevasse systems inherited from the surge. The Eyjabakkajökull studies of Ólafsdóttir (2011) and Schomacker et al. (2014) also point out that the pattern and zigzag shape of concertina eskers (Fig. 10F) is similar to the crevasse pattern of the glacier, and suggests they form en- and supraglacially by sediment accumulation and reworking by water in large, linked crevasses at the termination and after surges. Similar glaciofluvial landforms were described from the forefield of Skeiðarárjökull by Bennett et al. (2000b). They thought these landforms to result from supraglacial infills of crevasses and conduits by surface drainage during the 1991 Skeiðarárjökull surge. The reason why concertina eskers are rarely identified at the margins of contemporary surging glaciers, or from Pleistocene glacial landscapes, could be that supraglacial deposition in crevasses and subsequent dead ice melting makes their preservation potential very poor (c.f. Evans and Rea, 1999).

2.3.4 Lake Lögurinn – mass-balance control on surge frequency and reconstruction of the surge history of Eyjabakkajökull beyond historical records

Lake Lögurinn (65°15'N, 14°25'W, 53 km^2) is situated centrally in the Fljótsdalur valley, eastern Iceland (Fig. 13), 55 km northeast of Eyjabakkajökull. The lake occupies a deep glacially eroded basin, with lake surface elevation at 20 m a.s.l., but with maximum water depths of 112, 72 and 42 m in its three sub-basins and up to about 100 m thick postglacial sediment sequence in its deepest basin (Hallgrímsson, 2005). The glacial river Jökulsá í Fljótsdal drains Eyjabakkajökull into Lake Lögurinn.

Striberger et al. (2011, 2012) examined a 17.8-m-long sediment sequence retrieved by coring in the northern sub-basin of Lögurinn, to understand Holocene meltwater variability of Eyjabakkajökull. Their studied revealed a stratigraphic sequence characterized by nearly continuous sedimentation that was deposited over the past 10400 years (Striberger et al., 2012) (Fig. 15). Their data show that Eyjabakkajökull receded rapidly during the final phase of the last deglaciation, and ceased to deliver glacial meltwater to Lake Lögurinn by 9000 years BP. The return of glacial meltwater transport to Lake Lögurinn, and thus a return of Eyjabakkajökull, is dated to ca 4400 years BP, suggesting an almost 5000 years long period during early and mid-Holocene when the Vatnajökull ice cap did not have an Eyjabakkajökull outlet. Eyjabakkajökull reformed, probably as result of the mid-Holocene general cooling and glacier expansion recorded in Iceland (Geirsdottir et al., 2009; Gudmundsson, 1997), and started delivering glacial meltwater to Lake Lögurinn ca 4400 years BP.

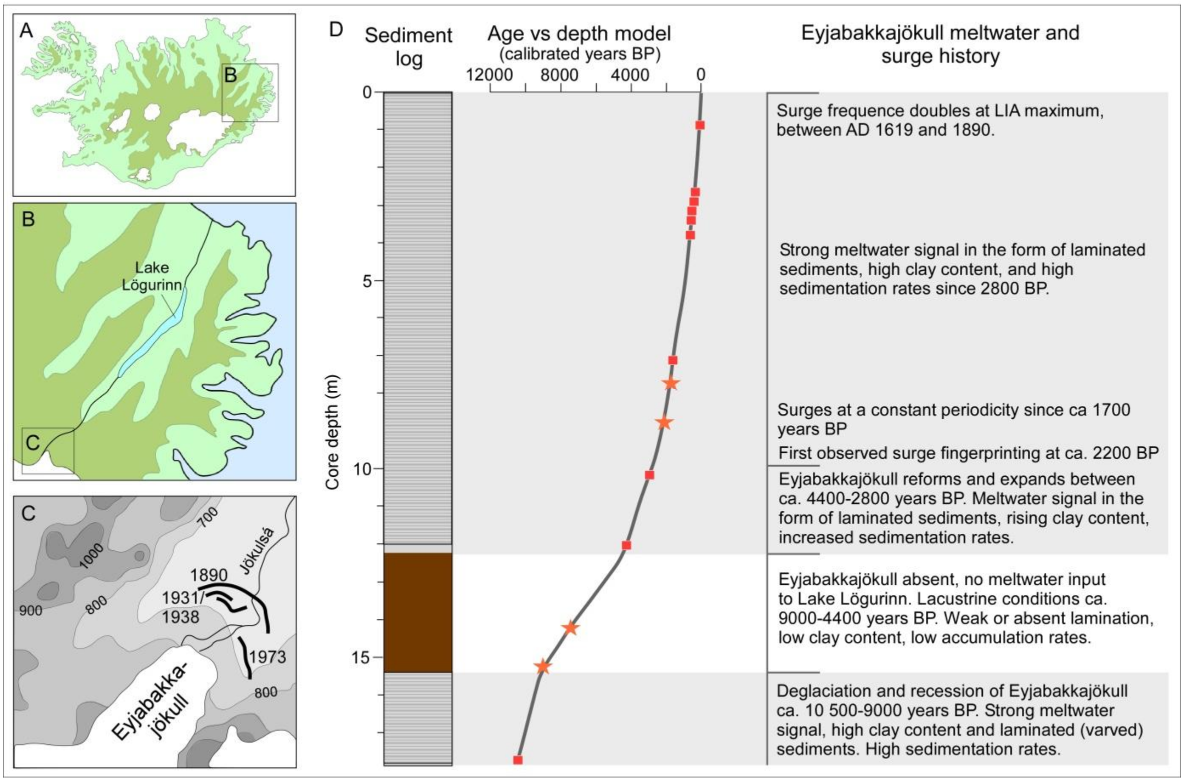


Fig. 15.The location of Lake Lögurinn in Eastern Iceland (A and B). (C) Eyjabakkajökull, with the surge terminal moraines of 1890, 1931/1938 and 1973 outlined. Jökulsá í Fljótsdal drains Eyjabakkajökull into Lake Lögurinn. (D) Sediment log and age vs depth model for the 17,8 m long core retrieved from Lake Lögurinn. Age determinations by tephrochronology (red squares) and ^{14}C (red stars). Modified from Striberger et al. (2001, 2012).

One specific aim of the study was to explore if if Eyjabakkajökull's surge history could be identified in the lake sediment record (Striberger et al., 2011). It is known that in lakes and fjords that are fed by meltwater from surge-type glaciers, sediment deposition rates increase dramatically during and immediately after surges due to increased erosion during surges and melting of excess ice in the ablation area after surges (Gilbert et al., 2002; Humphrey and Raymond, 1994; Sharp, 1988). As outlined above, the recent surge history of Eyjabakkajökull documents surges in 1890, 1931, 1938 and 1972–1973. The 1972–1973 surge of Eyjabakkajökull was accompanied by the highest recorded amount of suspended matter in Jökulsá í Fljótsdal during the period 1966–1995 (Pálsson and Vigfússon, 1996), and formed the thickest capping clay lamina recognized in the historical varved clay record in the sediment core. Following the 1972–1973 lamina, the clay lamina thicknesses gradually decreased until reaching normal thickness after about 3–5 years. Similar features were observed in the sediments following the 1890 and the 1930's surges. These sudden increases in glacial rock flour are reflected in the sediment sequence as recurring light-coloured intervals which Striberger et al. (2011) interpret as records of Eyjabakkajökull surges. The record suggests that Eyjabakkajökull began surging about 2200 BP, and approximately 1700 BP, the glacier started to surge at a uniform 34- to 38-year periodicity. This periodicity prevailed until the coldest part of the Little Ice Age (LIA) when it almost halved to 21–23 years. Since the late 1800's the surge periodicity of Eyjabakkajökull returned to a longer period of 35–40 years.

Striberger et al. (2011) suggested that the finding of evidence for "surge switch-on" of Eyjabakkajökull about 2200 BP, as well as LIA changes in periodicity of surges, suggest strongly that surge behaviour is linked to climatically driven mass balance changes and the ability of the glacier to transfer excess mass from the upper reservoir area to the lower ablation area. Their findings confirm with the notion of e.g. Eisen et al. (2001), Dolgushin and Osipova (1978) and Harrison and Post (2003a) that in some cases there is a clear connection between the interval between surges and the time needed to fill the reservoir area, associated with an increase in a glacier's net balance. Striberger et al. (2011) further suggested that surges of Eyjabakkajökull could become less frequent in a warming climate due to reduced rates of net mass input and lengthening of the quiescent phase. Ultimately, Eyjabakkajökull might even fail to re-enter its active phase and switch back to a non-surging mode, in line with the prediction of Dowdeswell et al. (1995) for Svalbard surging glaciers.

2.4 Surging glacier landsystem of Tungnaáarjökull, Iceland

Andrzejewski (2002) and Evans et al. (2009) studied proglacial sediments and landforms of Tungnaárjökull, a 17 km wide glacier lobe of the western margin of Vatnajökull (Fig. 1). Their mapping and interpretations were assisted by aerial photography from 1995, taken closely after the termination of the Tungnaárjökull 1994 surge (Fig. 2F). During this 10-month surge, the ice margin advanced up to 1200 m, with a maximum speed of 10 m/day (Sigurðsson, 1994). Their geomorphological maps (Andrzejewski, 2002; Evans et al., 2009) are at very different scales, approximately 1:60.000 and 1:30.000, respectively, which allows for pronounced differences in details. Yet, both studies identify zonation in the landsystem that confirms very well with the surging glacier landsystem of Evans and Rea (1999, 2003) and Evans et al. (2007). Diagnostic surge landforms identified on the foreland include thrust block and push moraines, overridden ice-cored thrust block moraines, hill-hole pairs, crevasse squeeze ridges, long flutings, hummocky moraine and ice-cored, pitted outwash. Andrzejewski (2002) studied the architecture and sedimentological and structural properties of glaciotectonically disturbed proglacial fan deposits and the end moraines that resulted from the 1994 surge. He concluded that spatial variations in the layout of glacial and glaciofluvial forms on the Tungnaárjökull forefield as well as in their geological structure were the result of a highly dynamic glacier margin during surge.

2.5 Langjökull surging outlets – studies highlight landscape impacts of surges and surge histories

Langjökull, a ~900 km^2 ice cap in the western highlands of Iceland, has at least three surging outlets, Hagafellsjökull (which splits into an eastern branch called Eystri-Hagafellsjökull Eystri, and a western branch called Vestari-Hagafellsjökull), Suðurjökull and possibly Þrístapajökull (Fig. 1) (Bennett et al., 2004b; Björnsson and Pálsson, 2008; Björnsson et al., 2003; Hart, 1995a; Larsen et al., 2015; Palmer et al., 2009; Sigbjarnarson, 1977; Theodórsson, 1980).

2.5.1 Surge sediments and landforms at Eystri-Hagafellsjökull Eystri

Bennett et al. (2000a) described the landform assemblage associated with the Neoglacial fluctuation of Eystri-Hagafellsjökull and an advance into the pro-glacial Lake Hagavatn (Fig. 16A). They provided a detailed landform/sediment assemblage model, where they found the landsystem to be dominated by three components: Firstly, a series of moraines formed both by thrusting at the glacier margin and by proglacial/subglacial deformation. The second component they identified is a morainal bank and associated ice-contact delta, and the third component is a series of lake-bed kame terraces formed by rapid sedimentation in a canal-like lake, formed as the ice retreated away from the push moraines. They concluded that this detailed case study provided an analogue for the interpretation of Pleistocene glaciolacustrine and glaciomarine landform/sediment assemblages and illustrated a range of styles of glaciolacustrine sedimentation and glaciotectonic deformation.

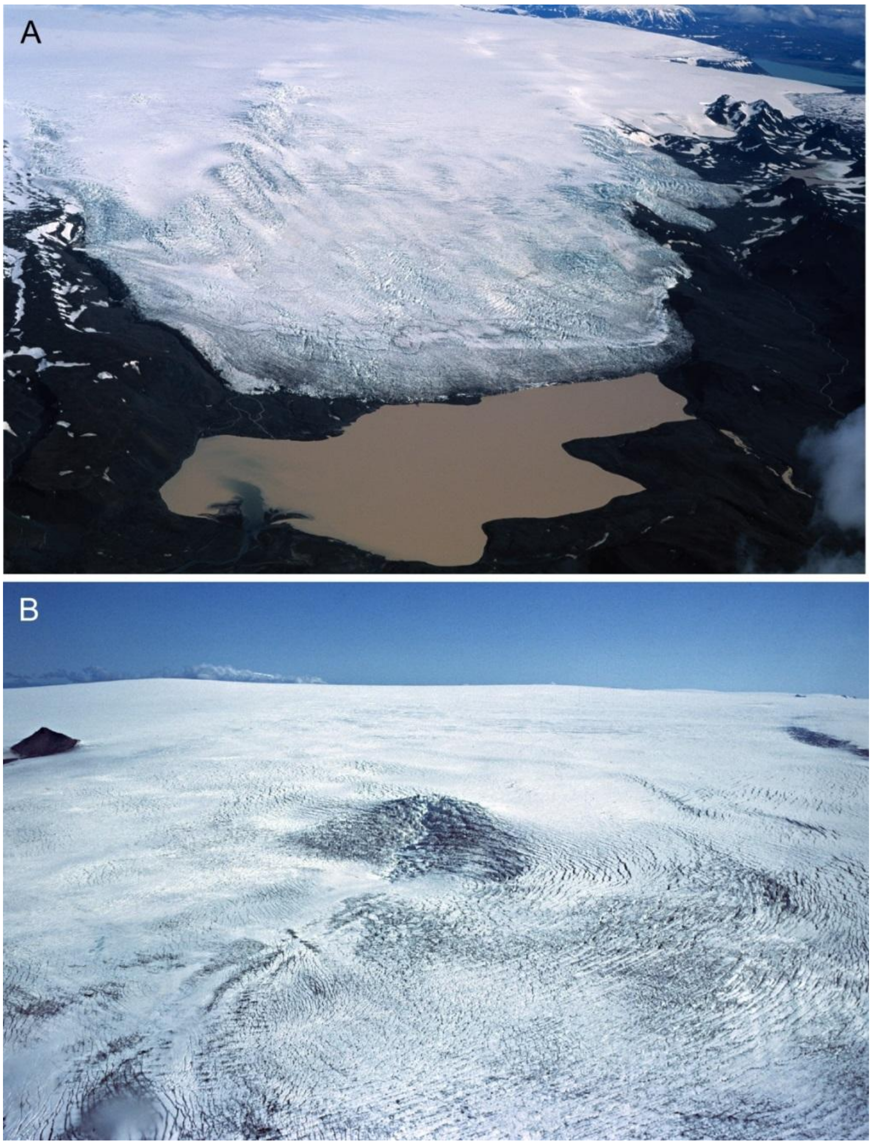


Fig. 16. A) Eystri-Hagafellsjökull surges into proglacial lake Hagavatn in 1999; B) Heavily crevassed surface of Vestari-Hagafellsjökull during its 1980 surge. Photos courtesy of Oddur Sigurðsson.

Bennett et al. (2004a) focused on the sedimentary and tectonic architecture of a large push moraine formed by a re-advance (perhaps surge) of Eystri-Hagafellsjökull between 1890 and 1929. They found that different tectonic architectures existed in two adjacent parts of the same push moraine complex. In one location, the ice advance pushed an ice-contact delta to form a prominent single-crested push moraine. In an adjacent location, the ice-margin advanced over the ice-contact delta to create a push moraine at the limit of the advance by subglacial gravity spreading. They discussed the causes of these two contrasting styles of deformation, and identified a range of possible controls. These included variation along the former ice margin and foreland in: (1) glacier–foreland coupling; (2) foreland shear strength; and (3) the frictional characteristics of the décollement. They suggested some combination of these variables the most likely cause.

Bennett et al. (2003) reported evidence of deformation at sub-freezing temperatures beneath Eystri-Hagafellsjökull, as the bed of a piedmont lobe that advanced during the 1998/1999 winter-spring surge comprised deformed blocks of glacier ice set within frozen sediment. This material had also been injected through overlying ice to form a network of crevasse-squeeze ridges. They concluded that there was evidence for two phases of deformation under contrasting rheological conditions: (1) deformation under relatively warm conditions, when the blocks of glacier ice acted as competent clasts within an unfrozen deforming matrix, and (2) subsequent deformation at sub-freezing temperatures when the ice blocks were attenuated into the surrounding frozen matrix along fractures and planar shears enriched with excess ice. This they took to suggest that the basal thermal regime of the advancing ice margin had changed from warm-based to cold-based during the surge event. In a separate study, Bennett et al. (2004b) described the internal architecture of the push moraine formed by the 1998/99 surge. The sedimentary architecture of thei moraine consisted of a folded multi-layered slab of glaciofluvial sediments that was displaced laterally by the advancing ice margin, whereas the crest and ice-distal face of the moraine consisted of sub-horizontal sediment sheets, and the ice-proximal face dipped steeply towards the ice margin. The core of the moraine consisted of frozen sediment, and thin slabs of glacier ice were embedded in its proximal face. The sediment slabs were characterized by both brittle and ductile styles of deformation. They argue on the basis of the observed variation in deformation style that a partially frozen glacial foreland was a key feature of the 1998/99 surge advance, as frozen foreland sediments would behave in a brittle fashion, while unfrozen sediments were likely to have deformed in a more ductile manner. Consequently, they concluded that the thermal regime of the foreland, and the timing of the ice advance, was of importance to the style of internal deformation found within ice-marginal push moraines.

2.5.2 Surging glacier landform and sediment association at Vestari-Hagafellsjökull – contrasting evidence on deforming bed or decoupling during surges

Vestari Hagafellsjökull (Figs. 1 and 16B) has two observed surges, in 1971 and 1980 (Björnsson et al., 2003; Sigbjarnarson, 1977; Theodórsson, 1980). The geomorphology of its forefield was studied by Hart (1995a), and she identified a diversity of streamlined features that included drumlins, flutes and glacial lineations; landforms that are accounted for in the surging glacier landsystem model of Evans and Rea (1999). She observed that the smaller bed forms (lineations, flutes) were superimposed upon the larger bed forms (drumlins, flutes), and suggested that the bed forms observed had formed due to subglacial deformation but in association with different thicknesses of the deforming layer. As the thickness of the deforming layer changed during ice retreat, one style of subglacial bed form superimposed upon another, resulting in a continuum of subglacial bed forms. Fuller and Murray (2002b) did a sedimentological study in the forefield of Vestari Hagafellsjökull to distinguish between pervasive sediment deformation and sliding as mechanisms of basal motion in surge. They found that ice retreat following the 1980 surge had exposed a two-layered discontinuous till cover and a suite of streamlined bed forms, including flutes and drumlins, and that the lower till layer had remained relatively undisturbed during the surge advance in both drumlinised and non-drumlinised areas. They suggested that decoupling between ice and substrate was important during the surge phase, and that the till did not weaken sufficiently for pervasive deformation to occur at depth. Thus they stated that although Vestari Hagafellsjökull overlies a soft bed, the 1980 surge advance did not occur by pervasive sediment deformation at depth, as suggested by Hart (1995a).

2.5.3 Surge history of Suðurjökull reconstructed on basis of proglacial lake sediments

Larsen et al. (2015) constructed a 300-year, high-resolution record of surges and terminus fluctuations of Suðurjökull and Norðurjökull, draining the Langjökull ice cap (Fig. 1). They used an innovative combination of varve counting, multibeam bathymetry, seismic imagery, and multiple sediment cores from the large proglacial lake Hvítárvatn. Langjökull achieved its maximum Neoglacial/LIA position, when two outlet glaciers, Norðurjökull and Suðurjökull, advanced into the lake and maintained active calving margins (Flowers et al., 2007; Larsen et al., 2011). Norðurjökull advanced into the basin ca. 1720, and remained at or near its maximum extension for most of the 19th century, whereas Suðurjökull underwent a quasi-periodic series of 8 surges. These are recorded in lake sediments as eight brief thick clay and increased IRD cycles, at AD 1828, 1838, 1855, 1866, 1885, 1905, 1917, and 1929, with a recurrence interval of 14 ± 4 yr. Each surge event resulted in fragmentation of the glacier terminus during advances of up to1.6 km that occurred in less than 2 years. Collapse of the expanded ice, iceberg melting, and re-establishment of the ice front at a near-shore grounding line occurred within 1–3 years of the surge. During the two largest surges (in 1855 and 1885) the glacier terminus advanced ~1600 m, amounting to minimum advance rates of 800-1600 m yr^{-1}. Larsen et al. (2015) could not establish any correlation between climate parameters and surge periods, although they acknowledge that the surge recurrence interval must be considered indirectly related to climate through changes in glacier mass balance where shorter quiescent periods are associated with an increase in a glacier's net balance (cf. Harrison and Post, 2003; Eisen et al., 2001; Striberger et al., 2011).

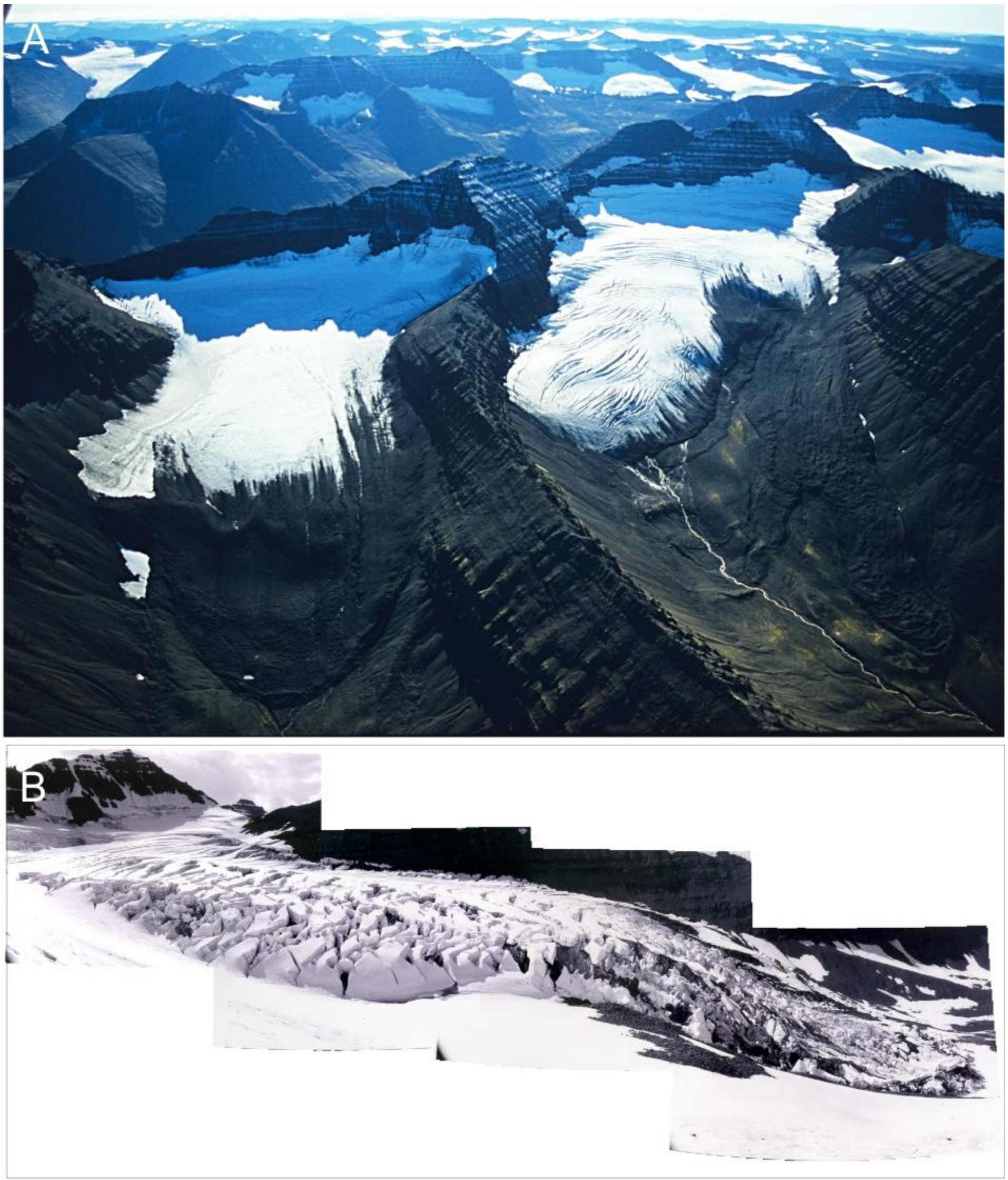


Fig. 17. A) The surge-type cirque glaciers Teigarjökull (left) and Búrfellsjökull (right). Búrfellsjökull was in surge in 1995 when this photo was taken. Photo: Oddur Sigurðsson 1995; B) The heavily crevassed Teigarjökull during the 1971 surge. The glacier length increased from about 1100 m to 1500 m during the surge. Photo: Helgi Hallgrímsson, July 1971.

2.6 Surge-type cirque glaciers in northern Iceland: unique surge fingerprinting and a new landsystem model

Brynjólfsson et al. (2012) described the geomorphology of the surge-type Búrfellsjökull and Teigarjökull cirque glaciers in the Tröllaskagi Peninsula, northern Iceland (Fig. 17A). The glacially sculptured landscape of Tröllaskagi Peninsula ranges in altitude from sea level to more than 1500 m a.s.l., with more than 150 small glaciers located in cirques and hanging valleys at altitudes between 700 and 1400 m a.s.l. (Sigurðsson and Williams, 2008). Although it is well known that surge-type glaciers vary greatly in size, shape and topographical and climate settings (Meier and Post, 1969), little is known about surging cirque glaciers and the landscape imprints their surges leave behind. The purpose of the Brynjólfsson et al. (2012) study was to explore the recent surge history of two small surge-type cirque glaciers and define a new landsystem model for surge-type cirque glaciers in alpine environments.

Four surges have been recorded in historic times on surge-type cirque glaciers, Búrfellsjökull, Teigarjökull (Fig. 17B) and Bægisárjökull (Björnsson, 1971; Björnsson et al., 2003; Hallgrímsson, 1972). Búrfellsjökull (1.45 km^2) and Teigarjökull (0.7 km^2), studied by Brynjólfsson et al. (2012), descend from about 1200 m a.s.l. to about 800 m a.s.l., with mean slopes of 11–14^o. A statistical analysis of an inventory of Icelandic glaciers shows that the median surface slope was about 7^o for non-surging glaciers and about 2^o for surge-type glaciers (Hayes, 2001). Thus, the high surface slope of Búrfellsjökull and Teigarjökull makes them the steepest surge-type glaciers in Iceland. The active phase of the last surge of Búrfellsjökull took four years, 2001–2004. This indicates that that the surging phase of these highland cirque glaciers has a longer duration compared to the surging outlets of Langjökull, Hofsjökull and Vatnajökull ice caps. Brynjólfsson et al. (2012) highlighted that surge-type cirque glaciers leave distinct impressions different from both non-surging cirque glaciers and large surge-type glaciers that drain out as broad lowland lobes from ice caps. The unique fingerprinting of the surge-type cirque glaciers in Iceland is that (a) the sediments are generally coarse, and at surface often characterized by angular supraglacial and englacial material considered to originate mainly from rock walls that surround the accumulation areas of glaciers; (b) the surging cirque glaciers also deposit subglacial till, but it is covered by englacially and supraglacially transported debris left by the down-wasting of dead-ice subsequent to the surge; (c) hummocky moraine is prominent inside the end moraines, and dead-ice is indicated by sinkholes, cracks, backslumping and collapse; (d) the surge end moraines are usually small and irregular, and constitute a low ridge at the front of a debris sheet; (e) annual (retreat) moraines do not occur in the forefield of these surging cirque glaciers, but (f) small crevasse-fill ridges can be recognized in the forefield as well as occasional poorly developed flutes. Brynjólfsson et al. (2012) defined a landsystem model for the surging cirque glaciers in northern Iceland (Fig. 18), that supplements the surge-type glacier landsystem model by Evans and Rea (1999, 2003) and Schomacker et al. (2014). Additionally, geomorphological mapping of the

Búrfellsjökull and Teigarjökull forfields recorded five end moraines that could not be related to known surges. Their geomorphology is similar to the end moraines formed by the recorded/observed surges and they were interpreted to indicate five additional surge events previously unrecognized (Brynjólfsson et al., 2012).

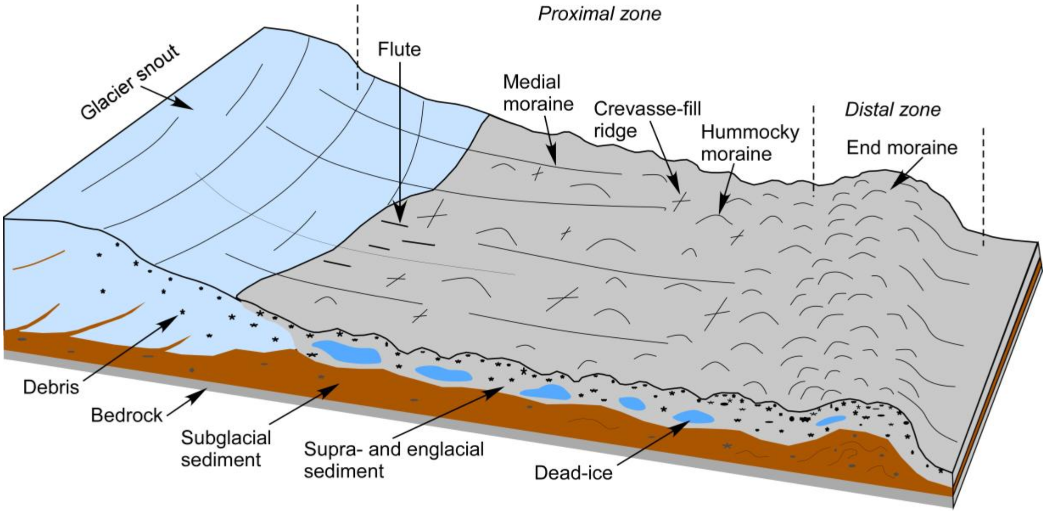


Fig. 18. Landsystem model for small surge-type cirque glaciers in Iceland. The distal zone, terminal moraine is formed of pre-surge sediment and dump of sediment of the surge front. The end moraines are small, and hummocky moraine often continues from their proximal slope. The proximal zone extends from the terminal moraine on to the glacier margin. Dead-ice and hummocky moraine are prevalent with rare occurrence of landforms such as crevasse-fill ridges, medial moraine remnants, and flutes. Modified from Brynjólfsson et al. (2012).

2.7 Hofsjökull – recent studies contribute to better understanding of the surging landsystem and surge histories

Hofsjökull (Fig. 1) is the third largest ice cap in Iceland, after Vatnajökull and Langjökull, situated in the central highlands and covering about 890 km^2 where it rests on the largest central volcano in Iceland (Björnsson, 2009; Björnsson et al., 2003; Sigurðsson and Williams, 2008). Hofsjökull carries seven surge-type outlets (Björnsson and Pálsson, 2008; Evans, 2011).

2.7.1 Múlajökull – splayed crevasses and repeated surges key to drumlin field occurrence

Múlajökull (Fig. 19) is a surge-type outlet glacier of Hofsjökull that drains through a 2-km narrow valley between Mt. Hjartafell to the west and Mt. Kerfjall to the east and forms a 4-km-wide piedmont lobe. Its present ice margin is about 620 m above sea level, ~2 km behind the LIA terminal moraine. The topography of the glacier bed, estimated from radio echo soundings from 1983, reveals an over-deepening beneath the 2-km wide outlet and the centre of the piedmont lobe, with its lowermost base being approximately 100 m lower than the present forefield (Björnsson, 1988a; 1988b). Johnson et al. (2010) published a pioneer study on the Múlajökull forefield, where they found that recent marginal retreat had revealed a drumlin field consisting of more than 50 drumlins. They described the drumlins as being composed of multiple beds of till deposited by lodgement and bed deformation. Further, they found that the youngest till layer truncates the older units with an erosion surface that parallels the drumlin form, and concluded that the drumlins were built up and shaped by a combination of subglacial depositional and erosional processes. Field evidence suggests each till bed to be associated with individual recent surges. They considered the drumlin field to be active in the sense that the drumlins were being shaped by the current glacial regime, and thus suggested the Múlajökull field is the only known active drumlin field on Earth, and therefore a unique analogue to Pleistocene drumlin fields. Drumlins are a geomorphological signature of ice-bed coupling, and studying them can give valuable information on processes controlling the flow of past and present ice sheets (Clark et al., 2009), which has motivated further studies of the Múlajökull drumlin field.

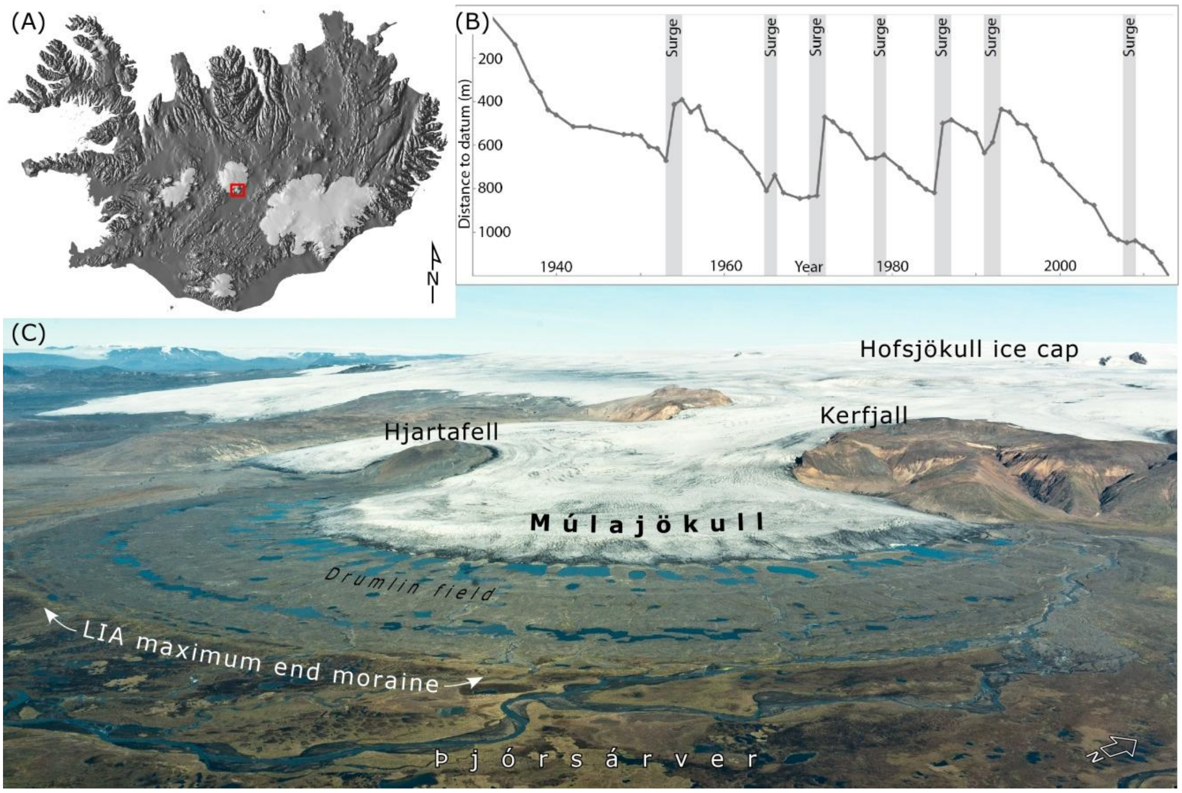


Fig. 19. A) Location of Múlajökull (red square) at the southern margin of the Hofsjökull ice cap; B) Ice margin and surge history of Múlajökull, modified from Johnson et al. (2010). C) Overview photo of the Hofsjökull ice cap, Mt. Hjartafell, Mt. Kerfjall, and the Múlajökull piedmont lobe. The vegetated wetlands of Þjórsárver in the foreground. Modified from Jónsson et al. (2014).

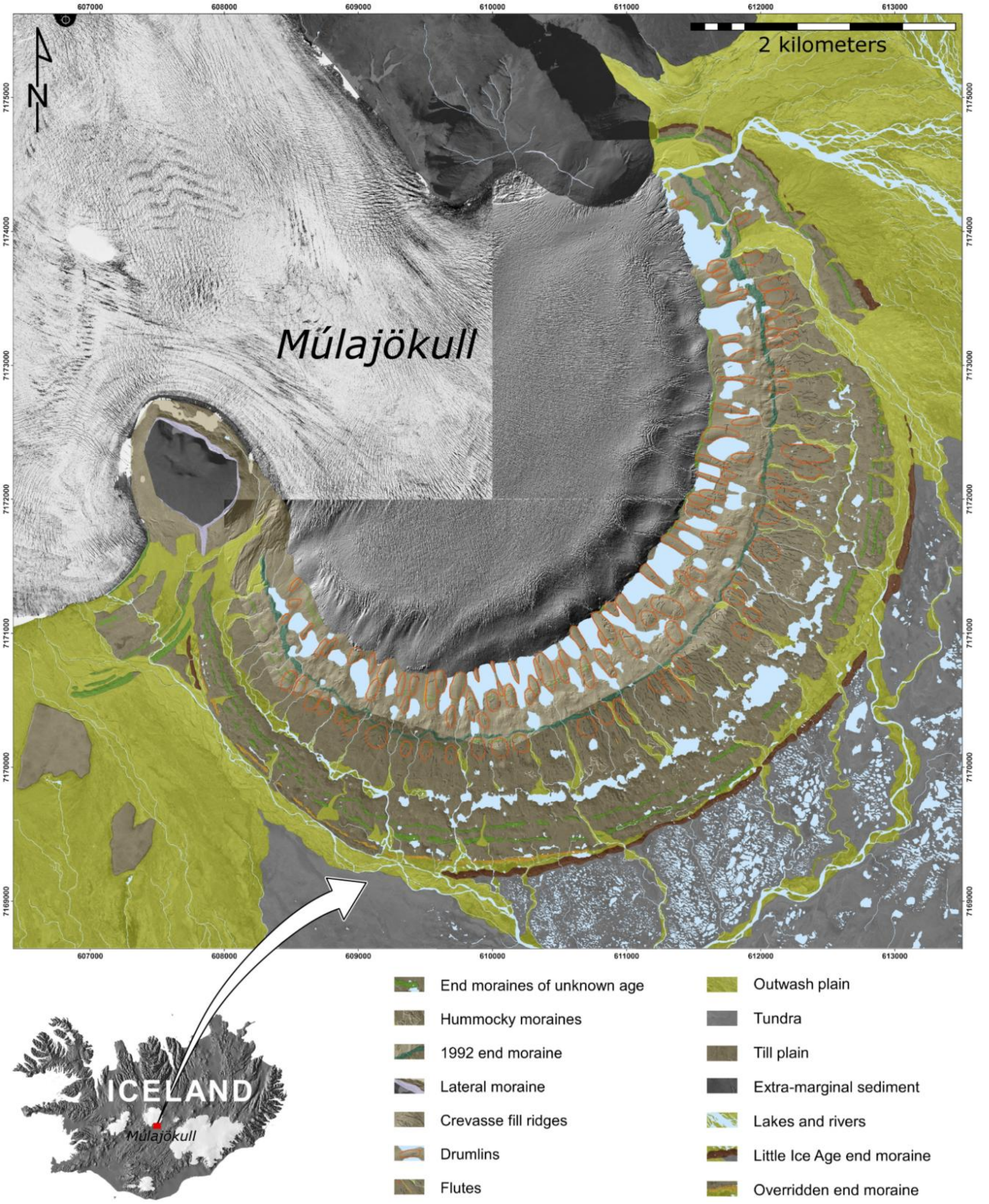


Fig. 20. Geomorphological map of the Múlajökull forefield. Map base is aerial photographs recorded in 1995 and LiDAR data from 2008. Scale 1:9000. From Jónsson et al. (2014).

2.7.2 The Múlajökull drumlin field

Jónsson et al. (2014) presented a geomorphological map of the active drumlin field and the forefield of Múlajökull, based on field investigations, aerial photographs taken in 1995 and LiDAR data recorded in 2008. They mapped subglacial, supraglacial, ice-marginal, periglacial, and glaciofluvial landforms, and concluded that the geomorphology of the Múlajökull forefield is similar to that of the forefields of other surge-type glaciers in Iceland: with a highly streamlined forefield, crevasse-fill ridges, and series of glaciotectonic end moraines. However, their mapping (Figs. 19 and 20) also revealed that an active drumlin field containing at least 110 drumlins, that have an average spacing of 94 m and an average areal density of around 10 drumlins/km^2 on the till plain, which is comparable to Pleistocene drumlin fields (Clark et al., 2009; Francek, 1991; Hess and Briner, 2009; Spagnolo et al., 2010). Drumlin orientation varies by nearly 180° as they occur in a splayed fan distribution in the forefield, so that the stoss end of drumlins in the north-eastern part of the forefield points toward NE whilst the stoss end points toward SW in the south-western part of the drumlin field (Fig. 20). The drumlins range from 70 to 380 m in length, 20 to 180 m in width, and 2 to 10 m in height and tend to be asymmetric in the sense that they are higher and wider up-glacier and taper down-glacier. The observation that the drumlins are wider and shorter in the distal part of the drumlin field and narrower and longer in the proximal part could suggest that the drumlins developed towards a more streamlined shape of the proximal landforms that had experienced more surges.

2.7.3 A conceptual model for the initiation and formation of the Múlajökull drumlins

It has been suggested by Clark et al. (2009) and Patterson and Hooke (1995) that ideas about drumlin formation needed to be developed into physically-based models. As a step towards this, Benediktsson et al. (in review-a) highlighted three questions regarding the Múlajökull drumlins that were raised by Johnson et al. (2010) and Jónsson et al. (2014): (1) are drumlins formed by a combination of erosion and deposition of till?; (2) does elongation ratio of drumlins increase with the number of surges they have experienced?, and (3) does the initial ice-front crevasse pattern control the location and nucleation of proto-drumlins?

They found that the drumlins at Múlajökull are composed of multiple till units, and suggested each till to represent one surge advance. The most recent till commonly truncates lower tills on the flanks and the proximal side of the drumlin. This implies that net deposition occurs on the drumlin crest and net erosion on the sides in every surge. They confirmed the observation of Jónsson et al. (2014) that drumlins proximal to the 1992 surge moraine are relatively long and narrow whilst drumlins distal to the moraine are wider and slightly shorter, and presented sedimentological and geomorphological observations that show that drumlins develop from broad and low 'proto-drumlins' to higher and narrower drumlins, resulting in increased drumlin relief and elongation ratio with time (number of surges or longer duration of ice flow). They show that there is a clear connection between drumlin location and the crevasse pattern of the ice front. They presented a model (Fig. 21) where radial crevasses formed in the glacier when it initially spilled onto the flat foreland. These crevasses led to spatial differences in normal pressure at the base of the glacier so that deposition was favoured beneath the crevasses and erosion between them. Consequently, the original crevasse pattern of the glacier controlled the location of proto-drumlins. Once the proto-drumlins were formed, a feedback mechanism was established leading to continued crevassing and increased sedimentation at the location of the proto-drumlins. The drumlins are then maintained and their relief and elongation ratio increases as the glacier erodes the sides and drapes a new till layer over the landform. Benediktsson et al. (in review-a) suggested that the drumlin mode of formation proposed above could explain the observations from the Múlajökull active drumlin field but the model needed to be tested in other areas.

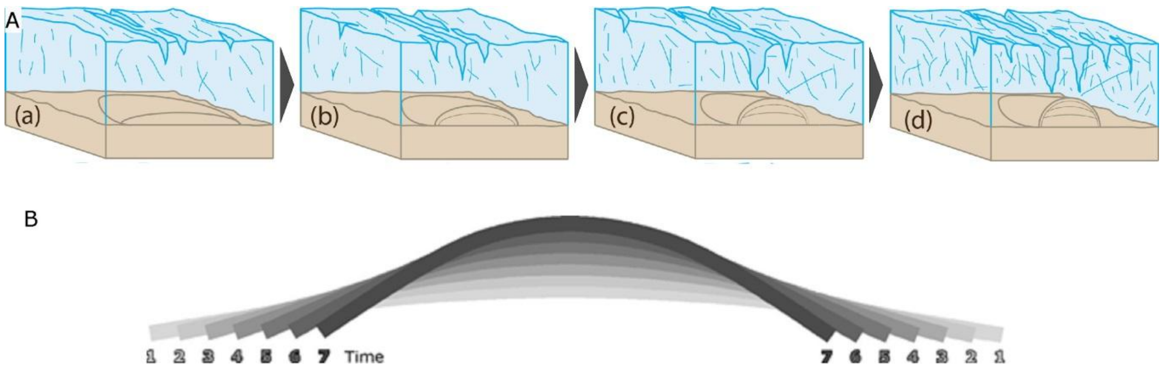


Fig. 21. Formation of drumlins at Múlajökull. A) (a) During a surge, longitudinal crevasses are formed due to lateral extension in the ice. This generates differences in basal stresses so that deposition is favoured beneath crevassed areas where 'proto-drumlins' will be formed. (b) During a following surge, a positive feedback mechanism is established that causes crevasses to form above the drumlins (because they are bumps on the bed). (c-d) Deposition is preferred on the drumlin crest, causing the drumlin to accrete, but while erosion is favoured on the drumlin sides and in the inter- drumlin areas. Thus, the drumlins get narrower and higher with time. B) The model explains the development of drumlin stratigraphy and morphology, and why there is an unconformity only below the uppermost (most recent) till but not below any of the lower ones, at any given time. The reason is that the lateral erosion that occurs with every surge simply removes older unconformities. An unconformity would always occur along the sides of a drumlin, but as a new surge arrives, it erodes the sides and thus removes the unconformity. Simultaneously, a new till layer is added on top – conformably in the centre but unconformably on the sides. Thus, theoretically, only the most recent unconformity exists at any time. From Benediktsson et al. (in review-a).

2.7.4 Early LIA terminal moraine

Benediktsson et al. (in review-b) studied the outermost end moraine ridge in the Múlajökull forefield (Fig. 22). The moraine is generally 4-7 m high and 50-100 m wide and composed of a sequence of loess, peat and tephra that is draped by till up to the crest. Steep, high-amplitude overturned folds and thrusts characterize the internal architecture in the crest zone while the distal slope is dominated by open, low-amplitude folds. Based on section balancing, Benediktsson et al. (in review-b) calculated that the total horizontal shortening of strata within the moraine was around 59% and that a basal detachment (décollement) occurred at a depth of 1.4 m. This implies that the glacier coupled to the foreland about 70 m short of its terminal position to initiate the formation of the moraine, most likely because of hydrological changes in the sub-marginal zone. The formation of the moraine commenced with low-amplitude open folding of the foreland that was followed by overfolding and piggyback thrusting. With radiocarbon dating and analysis of tephra layers, along with historical references, (Benediktsson et al., in review-b) bracket the formation of the moraine between AD 1717 and 1760, which suggests that Múlajökull had its LIA maximum and most extensive surge earlier than many other surge-type glaciers in Iceland.

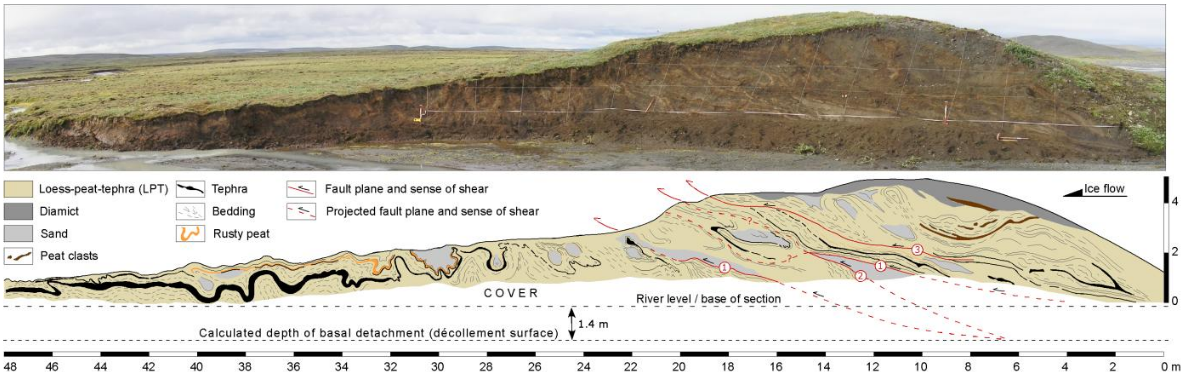


Fig. 22. A section through the outermost early Little Ice Age end moraine at Múlajökull. The photograph (upper) shows a 2x1 m grid that was used for accurate structural mapping. Note the spades for scale. The scale diagram (lower) shows the sediment facies architecture of the end moraine. Note the steep, overturned folds and thrusts in the crest zone (0-23 m) and open, low-amplitude folds in the distal part (23-48 m). Numbers in red indicate the interpreted relative age of thrusts. Modified from Benediktsson et al. (in review-b).

2.7.5 Sátujökull – surge deposits overprinting older polythermal landforms

Sátujökull drains the Hofsjökull ice cap towards north (Fig. 1). There are no historical or recent observations of surges in Sátujökull (Björnsson et al., 2003). Evans et al. (2010) and Evans (2011) mapped the Sátujökull forefield and found the surficial geology and geomorphology to suggest glacial landsystem overprinting as a result of complex glacier behaviour during the historical period. They defined two broad landsystems: Landsystem 1 comprises a wide arc of ice-cored moraine and controlled moraine ridges lying outside fluted and drumlinized terrain. This, they suggested, was strongly indicative of polythermal conditions, and recorded climatically driven glacier advance. These features are characteristic of LIA maximum limits on a number of glacier forelands in the arid interior uplands of Iceland, where environmental factors created polythermal conditions in glacier snouts during the LIA (Evans, 2010), and records a climatically driven glacier advance. Landsystem 2 contains most of the diagnostic criteria for the surging glacier landsystem and records two separate surges by the western margin of Sátujökull in the period since the attainment of the LIA maximum advance. The occurrence of Landsystem 2 is significant because Sátujökull has not been previously regarded as a surging glacier. Evans (2011) pointed out that landsystem overprinting, especially in response to changing thermal regimes and/or glacier dynamics, and particularly by different flow units in the same glacier, is rarely reported but is crucial to the critical application of modern landsystem analogues to Quaternary palaeoglaciological reconstructions.

2.8 Drangajökull – different surge fingerprinting

Drangajökull (Figs. 1 and 23) is in the class of small Icelandic ice caps, covering only about 143 km^2 (Jóhannesson et al., 2013). It reaches 915 m a.s.l. and its equilibrium line altitude at 550-600 m a.s.l. is about half the altitude of ELA on the other ice caps in Iceland, reflecting low summer temperature, short melting season, and high precipitation over the eastern Vestfirðir peninsula (Björnsson and Pálsson, 2008).

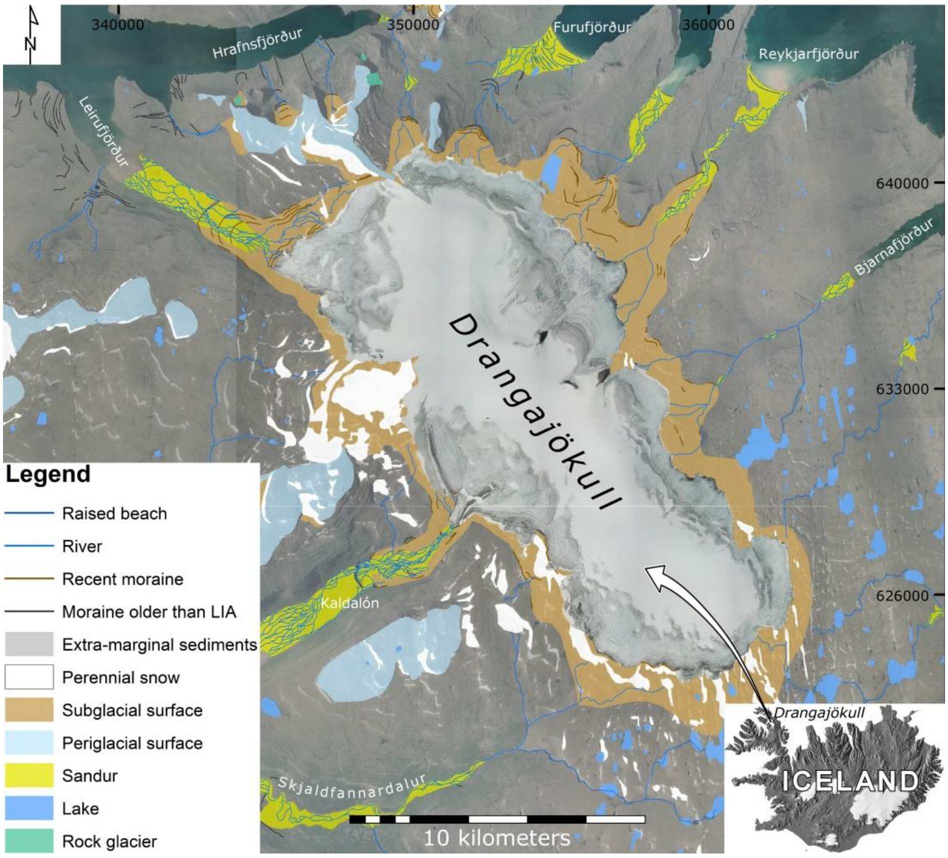
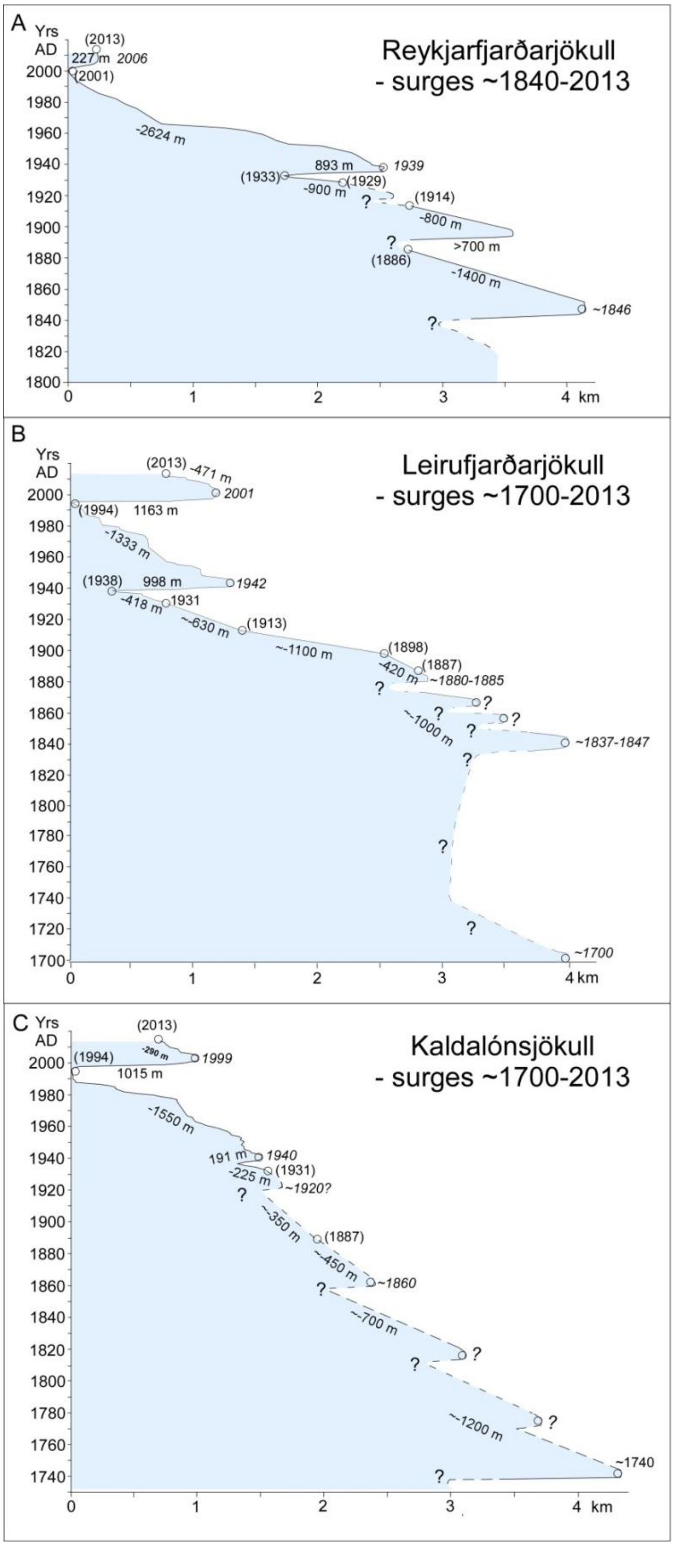


Fig. 23. An overview map of the geomorphology and sediments around Drangajökull. Map base: Aerial orthophotos from 2005-2006 by Loftmyndir ehf. Modified from Brynjólfsson et al. (2014).

Drangajökull' s three main outlets, Reykjarfjarðarjökull, Leirufjarðarjökull and Kaldalónsjökull, are all surge-type glaciers (Björnsson et al., 2003; Sigurðsson, 2005; Sigurðsson and Williams, 2008; Thorarinsson, 1969). A number of surges are known since the LIA (Eythorsson, 1935; Thorarinsson, 1943) (Fig. 24): Leirufjarðarjökull and Kaldalónsjökull surged around AD 1700 and 1740, and all three glaciers surged in the mid-nineteenth century and again in the 1940s. More recently, Leirufjarðarjökull and Kaldalónsjökull surged between 1995 and 2000 (Fig. 2D and 2E), and Reykjarfjarðarjökull between 2002 and 2006 (Sigurðsson, 1998; Sigurðsson, 2003; Sigurðsson and Jóhannesson, 1998). It is interesting to note that while the active phase of Icelandic surge-type glaciers usually lasts for few months to a year (Björnsson et al., 2003; Fischer et al., 2003), the last two surges of the Drangajökull outlets lasted about five years. This makes the surge activity of Drangajökull outlet glaciers unique for Icelandic ice caps, and resembles the surging of Icelandic cirque glaciers (Brynjólfsson et al., 2012) and Svalbard glaciers where the active phase typically lasts 3–10 years (Dowdeswell et al., 1991; Jiskoot et al., 1998; Lonne, 2014; Lønne, 2014; Murray et al., 2003).

2.8.1 Surge fingerprinting of Drangajökull outlet glaciers

Brynjólfsson et al. (2014) mapped landforms in front of the three surge-type outlets draining the Drangajökull ice cap (Fig. 22) and presented the first detailed geomorphological maps over the forefields of the surging outlets. The geomorphology is dominated by extensive sandur fields that cover the valley floors and end moraines. Landforms less common but present include fluted till plains, eskers, kame terraces, pitted sandur, and hummocky moraine. The till is generally coarse grained and rich in boulders. The terminal moraines of the three surging outlets are 10–15 m high and consist mainly of gravel pushed up from the outwash plain but also include some peat, fines and diamict. Other moraines are present but are lower in relief and somewhat indistinct. None of the landforms mapped are unique for surging glaciers nor is this suite diagnostic for the surging glacier landsystem as described by the models of Evans and Rea (1999, 2003) and Schomacker et al. (2014). The surging glacier landsystem models of Evans and Rea (1999, 2003) and Schomacker et al. (2014) recognize hummocky moraine, pitted sandur, crevasse fill ridges, and concertina eskers as important components of the surging glacier landsystem. These are minor or largely absent in front of the Drangajökull outlets. This could be owing to the thin basal till, generally coarse-grained till matrix, subglacial thermal- or hydrological conditions during the surges, or extensive fluvial erosion on the valley bottoms. The impermeable Neogene plateau basalt substratum at Drangajökull might also control subglacial hydrology and till composition to some degree.



A
Yrs AD
Reykjarfjarðarjökull
- surges ~1840-2013
(2013)
227 m
2006
(2001)
-2624 m
893 m
1939
(1933)
(1929)
-900 m
(1914)
-800 m
>700 m
(1886)
-1400 m
~1846
km
B
Yrs AD
Leirufjarðarjökull
- surges ~1700-2013
(2013)
-471 m
2001
(1994)
1163 m
-1333 m
(1938)
998 m
1942
-418 m
1931
~630 m
(1913)
~1100 m
(1898)
(1887)
-420 m
~1880-1885
~1000 m
~1837-1847
~1700
km
C
Yrs AD
Kaldalónsjökull
- surges ~1700-2013
(2013)
(1994)
1999
1015 m
-1550 m
191 m
1940
(1931)
-225 m
~1920?
~350 m
(1887)
~450 m
~1860
~700 m
~1200 m
~1740
km

Fig. 24. LIA and recent ice marginal fluctuations of (A) Reykjafjarðarjökull, (B) Leirufjarðarjökull, and (C) Kaldalónsjökull. Years in italics denote surges, and bracketed years are known ice marginal positions. Modified from Brynjólfsson et al. (2015).

2.8.2 Surge history of Drangajökull outlets

Brynjólfsson et al. (2015) reconstructed a 300 year surge history of the Drangajökull ice cap, based on geomorphological mapping, sedimentological studies and a review of historical records (Fig. 24). There is valuable historical information available on the surge frequencies over the past 300 years, because of the proximity of the surging outlets, Reykjarfjarðarjökull, Leirufjarðarjökull and Kaldalónsjökull to farms and pastures, as well as close monitoring of the glaciers since 1931. According to the historical data, each of the three surge-type outlet glaciers was recognized to have surged 2-4 times, but the geomorphological mapping of the glacier forefields revealed twice as many end-moraines as previously recognized surges. Accordingly, Brynjólfsson et al. (2015) reconstructed five surges by Reykjarfjarðarjökull, seven by Leirufjarðarjökull and six by Kaldalónsjökull. The surge interval varies between and within the outlets. About 50-60 years are between the last two surges of each outlet. The interval is generally shorter in the 19th century and in the beginning of 20th century. Surges of Leirufjarðarjökull occurred with about 10 year intervals during the period ~1840-~1898 AD. About 140 years elapsed with no surges recorded from Leirufjarðarjökull from ~1700-~1840 AD, possibly reflecting a lack of information rather than a long quiescent phase of the glacier. Individual outlets of Drangajökull reached their LIA maximum extent asynchronously during surges in the period ~ 1700 – 1846 AD. Leirufjarðarjökull reached its maximum extent about 1700 AD and again about 1840 AD. Reykjarfjarðarjökull reached its LIA maximum extent about 1846 AD, whereas Kaldalónsjökull reached its maximum around 1700 AD. Any clear relationship between the surge interval of the Drangajökull surge-type glaciers and climate has not been established. Surges were more frequent during the 19th century and the earliest 20th century compared to the relative cool 18th century and the warmer late 20th century, possibly reflecting a lack of information rather than a long quiescent phase of the glaciers.

Brynjólfsson et al. (2015) also estimated the areal and volumetric changes of these glaciers since the LIA maximum by making a digital elevation model (DEM) of the reconstructed glacier that could be compared with a modern DEM. As reference points for the digital elevation modelling they used the recently mapped lateral moraines and historical information on the exposure timing of nunataks in Drangajökull. During the LIA maximum surge events, the surging outlet glaciers extended 3-4 km further down-valley than at present, their ice volumes were at least 2-2.5 km^3 greater than in the beginning of the 21st century, and their size has diminished by 22-34% since the maximum extent in the LIA.

The Drangajökull ice cap decreased in area from ~216 km^2 during the LIA maximum extent (Brynjólfsson et al., 2015) to 143 km^2 in 2011 (Jóhannesson et al., 2013) .

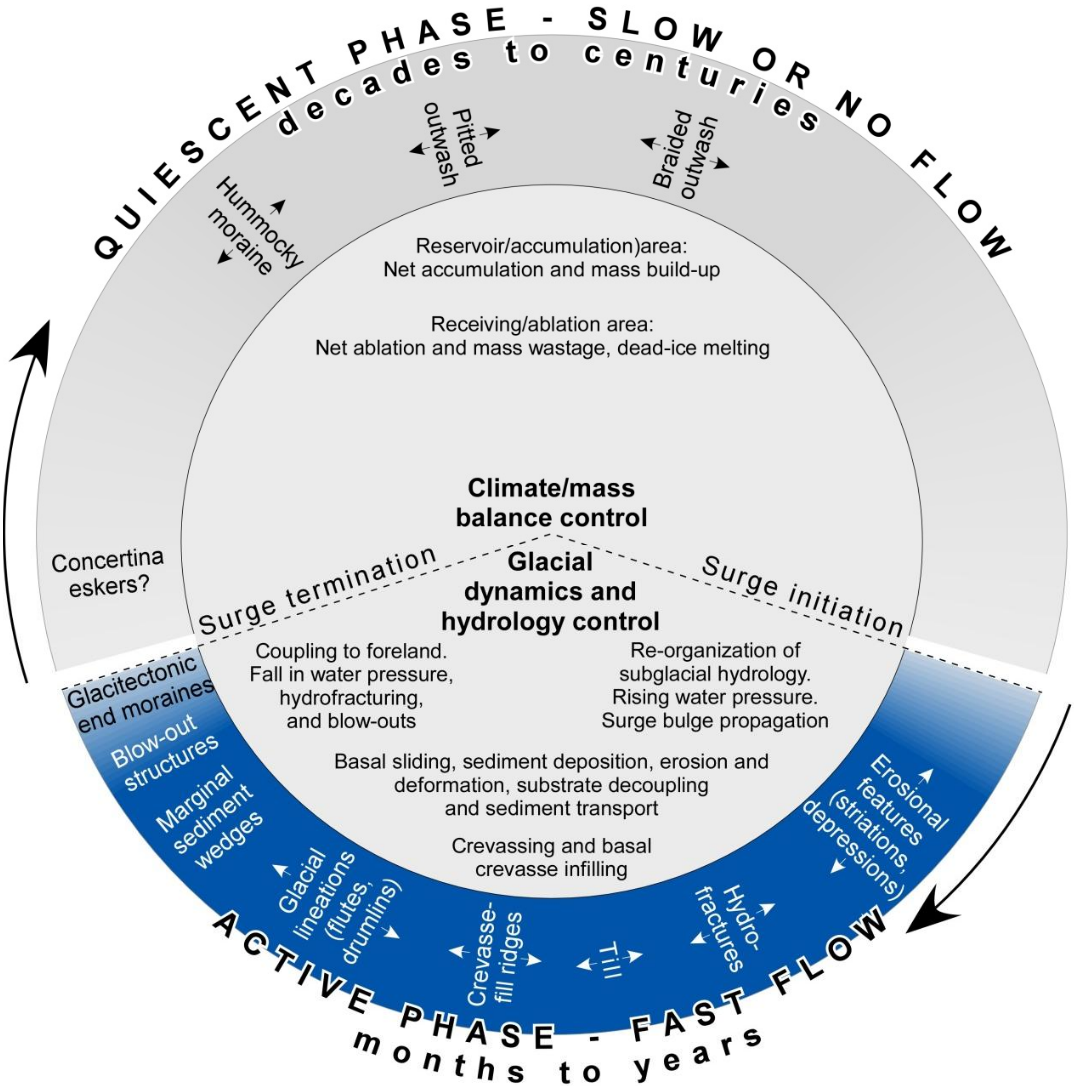


Fig. 25. Summary of controls and processes (inner circle) and products (outer circle) of surging glaciers as observed in Iceland. Most landforms (arrowed) are time-transgressive within the respective phase of the surge cycle, except marginal sediment wedges, blow-out structures and glacitectonic end moraines that are formed at surge termination. The timing of the formation of concertina eskers within the surge cycle is not clear, although they become exposed during the quiescent phase. Note the difference in duration of the quiescent and active phases of the surge cycle, and the contrasting number of active processes and products within the two phases.

Summary and discussion

Over the past decade, in depth studies of surge fingerprinting of a number of surge-type outlets draining the Icelandic ice caps have been undertaken, using state of the art methods in remote sensing and geomorphological, sedimentological and structural field mapping. This has allowed for unprecedented quantification of surge impacts on the subglacial and proglacial environment as well as advancing our understanding of the dynamics of surging glaciers and the processes operating during and after the surges (Fig. 25):

- The Brúarjökull studies resulted a new model (Fig. 3) for controls on rapid ice flow during surges (Kjær et al. 2006). Jiskoot et al. (2000) suggested that the most common property of surge-type glaciers is their setting on soft sediments and tills, rather than bedrock, and earlier models explained fast ice flow by two modes of basal motion largely dependent on ice/bed coupling (Cofaigh and Stokes, 2008; Fischer and Clarke, 2001). The basal sliding model proposes that decoupling of a glacier from its bed enables fast ice flow through increased basal sliding across the ice/bed interface or very shallow subglacial deformation (Engelhardt and Kamb, 1998), and the deformable bed model suggests that fast ice flow is sustained by deformation of water-saturated subglacial sediment that is coupled to the glacier (Alley et al., 1986; Alley et al., 1987; Boulton and Hindmarsh, 1987; Nelson et al., 2005). Both models associate the driving process with subglacial hydrological changes (Harrison and Post, 2003a). The new model of Kjær et al. (2006), involving a decoupling beneath a thick sediment sequence that was coupled to the glacier, provides a solution to the enigmatic observation that landforms and sub-surface sediments related to rapid ice flow often display depositional and deformational features that suggest strong ice/bed coupling, which is incompatible with enhanced basal sliding across the ice/bed interface. Subglacial sediment deformation is confirmed by a large body of evidence from studies of Pleistocene and recent tills (Benn and Evans, 1996; Boulton et al., 2001; Clarke, 2005; Cofaigh and Evans, 2001; Denis et al., 2010; Evans et al., 2006; Hart, 1995b; Nelson et al., 2005; van der Meer et al., 2003). The thickness of the deforming layer, however, and if spatial continuity of subglacial deformation is able to sustain fast flow, are still controversial (Larsen et al., 2004; Piotrowski et al., 2001), while arguments for a thinner and discontinuous deformable bed are provided from observations at modern glaciers and laboratory experiments (Fuller and Murray, 2000; Piotrowski et al., 2004).The model of Kjær et al. (2006) does not claim to account for all instances of observed fast flow, which likely reflects a suite of processes resulting from complex interplay between subglacial thermal and hydrological conditions as well as the nature of the substrate and the subglacial topography. However, it adds to the ongoing discussion on the role of subglacial sediment deformation as a mechanism for the rapid flow of ice streams that are responsible for the bulk of ice discharged from large ice sheets. Ice streams, in turn, impact on global sea level, ocean circulation and climate, and therefore the importance of understanding subglacial bed conditions and processes extends beyond the confines of the ice sheet bed itself (Cofaigh and Stokes, 2008).The presence of subglacial till and the limitations of subglacial drainage during the surge as outlined above provide mechanisms for the surge velocities to be sustained throughout the period of surging.

- Ice marginal landforms and sediments provide information on processes operating at the margins of both advancing and retreating glaciers (Benn and Evans, 2010; Bennett, 2001; Kristensen et al., 2009; Krüger et al., 2010). Recent glacial geological research developments in Iceland have emphasized the importance of integrating morphological, structural and sedimentological data for obtaining a holistic view of sub-marginal and ice-marginal processes at work during glacier surging, as well as of end-moraine architecture and formation (Benediktsson et al., 2009; Benediktsson et al., 2008; Benediktsson et al., 2010; Bennett et al., 2004a; Bennett et al., 2004b). Bennett et al. (2004a, 2004b) attributed lateral variability in ice-marginal deformation and tectonic architecture of two surge moraine complexes at Hagafellsjökull-Eystri to pre-deformation geometry, thermal regime, and rheology of the foreland. Similarly, Benediktsson et al. (2008, 2009, 2010) explained lateral variability in morphology and architecture of the 1890 and 1963-64 moraines at Brúarjökull and 1890 moraines at Eyjabakkajökull with spatial variation in subglacial ice-flow mechanism and hydrology, and thickness and rheology of the foreland wedge. These studies, along with more recent ones (Benediktsson, 2012; Benediktsson et al., in review-b), have provided new information on sub-marginal and ice-marginal processes during glacier surging and demonstrated the importance of glaciotectonic end moraines for the understanding of the dynamics of fast-flowing glaciers. According to the model by Benediktsson et al. (2008) the sediment wedge discovered in the marginal zone of the 1890 Brúarjökull surge formed in response to substrate/bedrock decoupling (Kjær et al., 2006) and concomitant down-glacier transport and deformation of subglacial sediment. Thus, Benediktsson et al. (2008) ventured if the sedimentary wedge could stand as an analogue for the formation of ice stream grounding line wedges described from e.g. the Barents Sea, Greenland and Antarctica (Andreassen et al., 2014; Cofaigh et al., 2008; Dowdeswell and Fugelli, 2012; Jakobsson et al., 2012; Ottesen et al., 2008) where fast flow could potentially be facilitated by high porewater pressure in the substrate and associated decoupling at a stratigraphic discontinuity at depth. Similar asymmetric, till-wedge end moraines are common along the former margins of southern Laurentide lobes and their formation has been attributed to bed deformation advecting till towards the ice margin (Eyles et al., 2011; Johnson and Hansel, 1999). Additionally, many of these southern Laurentide ice lobes are thought to be the outlets of rapid, streaming glaciers (Jennings, 2005). The balancing of moraine cross-sections at Brúarjökull, Eyjabakkajökull and Múlajökull has shown that the total horizontal shortening of the foreland strata during moraine formation ranges typically from ~27% in moraines with multiple low-amplitude folds to ~39% in moraines with overturned folds and thrusts and ~76% where multiple overfolding and refolding occurs. By comparing the shortening distances to known ice-flow velocities during surges, a time-frame of only about 1 day was established for the formation of 5-20 m high, single-crested, fold-dominated moraines and 2-6 days for up to 40 m high, multi-crested, fold and thrust moraines. This new information demonstrates the dynamic nature of ice-marginal landform development during glacier surges and may aid future efforts of reconstructing ice-flow rates from the landform record.

- The different structural style of the end moraines at Brúarjökull implies spatial variations in subglacial ice-flow mechanism (Benediktsson et al., 2009). Single- and sharp-crested moraines occur in areas of overall fine-grained sediments; they are dominated by ductile deformation and were formed as part of a marginal sediment wedge. Thus, they indicate high porewater pressure in the substrate during a surge and substrate/bedrock decoupling as the principal mechanism of ice flow (Benediktsson et al., 2008; Kjær et al., 2006). On the contrary, broad-crested moraines are dominated by stacked thrust sheets of generally coarser sediments, which could not support high porewater pressure in the substrate during a surge. Therefore, it is unlikely that substrate/bedrock decoupling was the principal component of ice-flow in these areas, but rather deformation of the bed. This may explain why Brúarjökull advances a few kilometres further where fine-grained sediments dominate (substrate/bedrock decoupling) than in overall coarse-grained areas (bed deformation) (Benediktsson et al., 2009; Guðmundsson et al., 1996). Thrusted and stacked end moraines (thrust-block moraines) have been primarily linked with surge-type glacier advances (Croot, 1988a; Croot, 1988b; Evans and Rea, 1999; Sharp, 1985b). The studies by Benediktsson et al. (2008, 2009, 2010) definitely support the notion that complex thrust moraines can be regarded as an indication of the dynamic state of the advancing glacier, both modern and ancient.
- Most known surges around the world are recent or historical, as expanding LIA glaciers frequently obliterated signs of earlier Holocene or Neoglacial surges. Striberger et al. (2011, 2012) give a unique insight into the Eyjabakkajökull surge

history way beyond historical records by studying lake sediment cores. They suggested that the switching on of surge behaviour around 2200 BP, about 2100 years subsequent to the Neoglacial reforming of Eyjabakkajökull, as well as the increased frequency of surges during the LIA, strongly indicated climate/mass balance control on surge initiation and frequency. There are several previous studies on how glacier mass balance controls surge frequency (Harrison and Post, 2003a), and how a more positive mass balance can result in a shorter period between subsequent surges (Eisen et al., 2001). Dowdeswell et al. (1995) suggested for Svalbard surging glaciers that a distinct reduction in surges between 1936 and 1990 was related to a clear decrease in glacier mass balance over the same period. Shift in mass balance could also prevent glaciers which surged in the past from accumulating sufficient mass to initiate future surge activity and thus removing them from the surge cycle, as has been proposed for Scott Turnerbreen and Midtre Lovénbreen in Svalbard (Hansen, 2003; Hodgkins et al., 1999). Reversely, Hewitt (2007) and Copland et al. (2011) reported a sharp increase in the number of new surges in the Karakoram Mountains since 1990, that was coincident with a period of increased temperatures and precipitation causing positive glacier mass balance in that region. This agrees with the suggestion of Dolgushin and Osipova (1978) and Harrison et al. (2008) that in some cases there appeared to be a simple connection between the interval between surges and the time needed to fill the reservoir area. Björnsson et al. (2003), Larsen et al. (2015) and Brynjólfsson et al. (2015) found little connection between mass balance and surge periodicity for Icelandic surge-type glaciers. This makes particularly interesting the prediction of Striberger et al. (2011) that surges of Eyjabakkajökull, and possibly similar surge-type outlet glaciers in comparable settings, will become less frequent in a warming climate due to reduced rates of net mass input and lengthening of the quiescent phase, and ultimately, Eyjabakkajökull may even fail to re-enter its active phase and switch back to a non-surging mode. Evans et al. (2010) and Evans (2011) presented evidence that Sátujökull, an outlet glacier draining the northern Hofsjökull ice cap in central Iceland, had switched from non-surging mode to surge-type behaviour since the LIA. This, Evans (2011) regarded as an example of changing thermal regimes and/or glacier dynamics in different flow units of the same glacier, thereby providing a modern analogue for similar spatial and temporal switches in glacier behaviour in Pleistocene landform-sediment assemblages.

- Landsystem models help identifying former surging glaciers with the ultimate aim of better understanding past ice sheet dynamics, and detailed interpretations of ancient glaciated terrains rely heavily on understanding of process–form relationships in contemporary glacierized basins (Clayton and Moran, 1974; Evans and Rea, 1999; Evans and Rea, 2003; Evans and Twigg, 2002; Evans et al., 2007; Stokes and Clark, 1999; Stokes and Clark, 2001). The studies at Brúarjökull and Eyjabakkajökull resulted in a refined landsystem model for warm-based surging glaciers (Fig. 14) (Schomacker et al., 2014) and the work of Brynjólfsson et al. (2012) produced the first landsystem model for small surge-type cirque glaciers (Fig. 18). Evans and Rea's (1999, 2003) landsystem model for surging glaciers was partly based on a geomorphological map of the Eyjabakkajökull forefield. Their model includes three zones (A, B, and C), each of which contains a characteristic assemblage of sediments and landforms. The refined landsystem model of Schomacker et al. (2014) largely agrees with the model by Evans and Rea (1999, 2003); however, with some modifications as to the forefield zonation and location of landforms. Nevertheless, boxing-in the sedimentological and morphological fingerprinting of surging glaciers is not the ultimate aim of these studies; they primarily provide means to identify tracks of fast-flowing ice and obtain information about basal processes, ice flow and ice disintegration on local, regional and temporal scales. Moreover, the landsystem models reflect landform assemblages that in most cases result from the impact of multiple surges (Fig. 5). As a result, the landsystem models may serve as analogues when identifying and separating different fast flow events or, alternatively, the switching-on and -off of fast flow in the geological record. Long-term projects at modern surge-type glaciers also provide opportunities to track landform development through time and improve conceptual models of their formation. For instance, Schomacker et al. (2006) described how ice-cored drumlins at Brúarjökull develop into areas of hummocky moraine as the ice-cores melt out in the quiescent phase. Ice-cored drumlins are transitional state landforms from the surging glacier landsystem that have a low preservation potential but contain important information about the surge process. Similarly, Ólafsdóttir (2011) used field observations and time-series of aerial photographs to propose a step-by-step depositional model for concertina ridges at Eyjabakkajökull. Such studies highlight the value of repeated field campaigns and the use of historical aerial photographs that reveal snapshots of the forefields during earlier times. Hence, they provide information about the processes and stages of formation of surge-related landforms and sediments and not only the end-product left in the geological record. Finally, using the landsystem approach in mapping glacier forefields, surge-fingerprinting/overprinting has allowed for identifying some Icelandic outlet glaciers as surge-type glaciers despite lack of historical or observational records (Evans, 2011; Evans et al., 2010) as well as revealing a number of surge episodes previously unknown at known surge-type glaciers (Brynjólfsson et al., 2012; Brynjólfsson et al., 2015).

- Drumlins and drumlin fields are prominent features in the landscapes of Pleistocene ice sheets, and they have been intensely studied for decades as indicators of palaeoglaciological conditions and glacier dynamics (Benn and Evans, 2010; Clark et al., 2009; Menzies, 1979; Patterson and Hooke, 1995). Despite this, there is currently no unifying theory on drumlin formation (Benn and Evans, 2006; Boulton, 1987; Clark et al., 2009; Johnson et al., 2010; Shaw, 2002; Spagnolo et al., 2014; Stokes et al., 2013) and modern analogues to the Pleistocene drumlin fields were unknown until Johnson et al. (2010) described the Múlajökull drumlin field. The results of Johnson et al. (2010), Jónsson et al. (2014) and Benediktsson et al. (in review-a) suggest that (a) initial formation and location of proto-drumlins are controlled by the splayed crevasse pattern of the ice front; (b) a positive feedback mechanism causes crevasses to form above the drumlins and the drumlins to accrete through till deposition; (c) drumlins form as a result of a combination of erosion on the sides and deposition on top; (d) there is a connection between the relief and elongation ratio of drumlins and rate and/or duration of ice flow, where their relief and elongation ratio increases as the glacier erodes the sides and drapes a new till layer over the landform. Kjær et al. (2008), on the other hand, suggested that the formation of drumlins in the Brúarjökull forefield was tied to re-moulding of pre-existing landforms.

3. Future research challenges

Future research challenges on surging glaciers and their geological signatures that can be outlined based on the status of the glacial geological research in Iceland include the following outstanding research questions and challenges:

- *Understanding better the relation between climate parameters and surges*

Surge-type behaviour has largely been thought to be controlled by internal glaciological mechanisms rather than external forcing, and therefore surging glaciers have not been thought be a reliable indicator of glacier response to climatic perturbations (Meier and Post, 1969). Striberger et al. (2011, 2012) showed that Eyjabakkajökull switched from being a non-surging glacier to a surge-type glacier about 2000 years ago, and suggested that the frequency of surges could be related to climate controlled mass balance changes. Sevestre and Benn (in press) suggest the existence of climatic envelopes conducive to surging implies that glaciers may change from ‘normal’ to surge-type and vice versa under cooling or warming climates. Are surge frequencies in Iceland generally going to decrease with increasingly more negative mass balance of glaciers and ice caps? Conversely, could an increase in winter precipitation and changes in the ability of the glacier to transfer excess mass from the upper reservoir area to the lower ablation area cause some of the outlet glaciers draining the Mýrdalsjökull ice cap or the high domes of Vatnajökull to switch to surge-type behaviour, as has happened in the Karakoram Mountains (Copland et al., 2011)? Likewise, will increased subglacial meltwater as a consequence of amplified surface melting in a warming climate facilitate basal deformation and/or sliding so ice flow becomes more efficient and switches off surging behaviour?

- *Can we scale up landsystem models for surging glaciers for better understanding ice-stream behaviour?*

Recent understanding of ice-stream dynamics highlight differences between terrestrial based ice streams and those entering the marine realm and feeding coastal ice shelves, but it is becoming increasingly clear that during deglaciation of the Pleistocene ice sheets there occurred major changes in ice-stream activity where ice streams oscillated or turned on and shut down (Andreassen et al., 2014; Bjarnadóttir et al., 2014; Cofaigh et al., 2010; De Angelis and Kleman, 2005; Evans et al., 1999; Evans et al., 2014; Kleman and Applegate, 2014; Stokes et al., 2009). One motivation for studying the geological effects of surging glaciers in Iceland and landsystem modelling has been to better understand processes and patterns of fast flowing ice. Surging ice lobes/ice streams during the deglaciation of the Laurentide Ice-Sheet (LIS) have been identified by e.g. Clark (1994), Clayton et al. (1985), Colgan et al. (2003) and Evans et al. (2014). Although Evans et al. (1999) suggested it was possible to use modern surge-type glaciers in Iceland as analogues for fast flowing ice/ice streams during the deglaciation of the LIS, recent studies by e.g. Colgan et al. (2003) and Evans et al. (2014) indicate that during deglaciation alternate cold, polythermal, and temperate marginal conditions sequentially gave way to more dynamic and surging activity during ice-sheet recession. It is, therefore, still a challenge if we can scale-up present-day surging processes and landsystem models for applying to Pleistocene ice streams, and if so, if they are restricted to depicting sediment and landform patterns developed during decaying stages of ice streams? Yet, individual landforms and landform assemblages within surging glacier landsystems may hold the key to the up-scaling from surge-type glaciers to ice streams. Sediment wedges formed in the marginal zones of surging glaciers may be important analogues to ice stream grounding line wedges, and could suggest that studies, which aim at understanding the ice-flow mechanism of ice streams and the processes responsible for the formation and stability of grounding line wedges (Alley et al., 2007; Anandakrishnan et al., 2007), should be directed towards their internal architecture and boundary to underlying sediment or bedrock. Similarly, the increased elongation ratio of drumlins with the number of surges at Múlajökull may aid our understanding of bedform assemblages beneath fast flowing ice and could be regarded as analogous to the development of megaflutes or mega-scale glacial lineations beneath ice streams (Briner, 2007; Stokes and Clark, 2002). The question still remains, though, whether highly elongate subglacial bedforms are indicative of fast flow over relatively short time or slow flow over longer time (Stokes and Clark, 2002), and also if subglacial bedforms with low elongation ratio represent short-lived periods of fast flow? Due to poor accessibility of modern ice stream beds, further studies and monitoring of surge-type glaciers are the most likely means to provided answers to these questions.

- *Extending and complementing historical and observed surge history by studying sediment cores from proglacial lakes*

Coring proglacial lakes that preserve signatures of surge activity in their sedimentological record has been very successfully carried out by Striberger et al. (2011) for Eyjabakkajökull and Larsen et al. (2015) for Suðurjökull. This has proven very valuable for identifying surge episodes in the past and extending the surge history of the respective glaciers. As proglacial lakes are fairly common at surge-type glaciers in Iceland (and elsewhere) this is a novel archive that needs to be further explored.

- *Can we reconstruct past surge flow rates from glacial landforms and sediments?*

Although highly elongate bedforms have been suggested to indicate fast ice flow (Stokes and Clark, 2002; Briner, 2007), past flow rates have, to our knowledge, not been successfully reconstructed from the sediment and landform record. Benediktsson et al. (2008) found a clear connection between dynamics of sediment wedge development, end moraine formation, and rate of ice flow/marginal advance (~120 m/day) during the Brúarjökull surges. However, glaciotectonic end moraines at other surge-type glaciers in Iceland (Eyjabakkajökull, Múlajökull, Leirufjarðarjökull) show similar architectural and morphological characteristics although the surge advances of those glaciers were one or two orders of magnitude slower than at Brúarjökull (Benediktsson et al., 2010, in review-b; Brynjólfsson et al., 2015). The architectural and morphological characteristics indicate that these landforms originate from fast advances and may thus be useful analogues in the identification of surging glacier or ice stream landsystems. They cannot, however, be used to infer actual flow rates and the models for their formation need to be tested by carrying out more case studies. Likewise, high elongation ratios of drumlins or MSGLs may be a strong indicator of fast flow but as yet, not sufficient proxies for reconstructing rates of ice flow. Hence, techniques have surely been developed for identifying landscapes of fast flowing ice but need yet to be developed for the reconstruction of actual flow rates and must be regarded as a future challenge in the studies of fast flowing glaciers.

- *Are drumlin fields indicative of fast flowing ice?*

It is quite clear from the results of Johnson et al. (2010), Jónsson et al. (2014) and Benediktsson et al. (in review-a) that drumlins in the recent Múlajökull drumlin field form by a combination of erosion and deposition and there exists a connection between duration of ice flow/number of surges and drumlin relief and elongation ratio. Individual drumlins observed at Brúarjökull and Eyjabakkajökull are similar in dimensions to drumlins within the Múlajökull drumlin field but are of different composition being ice-cored or rock-cored, respectively. They originate from the streamlining of dead-ice and bedrock, which likely involves both erosion of the core and deposition of a till carapace (Schomacker et al., 2006, 2014). Although these drumlins are formed by surging glaciers, drumlins have also been described from non-surging glaciers (Krüger, 1987; van der Meer, 1983), so drumlins as individual landforms cannot be attributed to fast ice flow alone. Drumlin fields, however, have widely been used to identify the tracks of past ice streams (e.g. Stokes and Clark, 2002; Briner, 2007; Maclachlan and Eyles, 2013) and the recent discovery of the active drumlin field at Múlajökull may suggest that fast flow is required for the formation of large fields of drumlins (Johnson et al., 2010; Jónsson et al., 2014). The challenge here is to test if the model of Benediktsson et al. (in review-a) can be used to generate quantitative predictions of drumlin length, width and relief, as well as of landform evolution and history, and test predictions against statistically valid observations of real drumlin fields (cf. Clark et al. 2009).

- *How do concertina eskers and crevasse fill ridges form?*

Although crevasse fill ridges and concertina eskers are uniquely associated with surging glaciers their formation is not fully understood. Cross cutting relationships between flutes and crevasse fill ridges have not been entirely explained and the processes operating between infilling of basal fractures by till and crevasse fill ridges becoming exposed on the foreland during the quiescent phase melting out of dead ice are not clear. Thorough sedimentological studies, linked to detailed observations of processes operating during surges (Kristensen and Benn, 2012) could highlight the chain of processes explaining the flute-crevasse fill ridge relationships. There are also still some aspects as to the formation of concertina eskers that are not well understood. The original explanation of Knudsen (1995) of concertina eskers being formed by compression and deformation of pre-surge englacial eskers has largely been abandoned in favour of them being the result of supraglacial or englacial meltwater deposition in linked cavities or crevasses during or immediately after the surge (Benn and Evans, 2010; Evans and Rea, 2003; Ólafsdóttir, 2011). Further sedimentological and structural studies of concertina eskers could clarify the processes of their formation in the course of a surge as well as better explaining their spatial pattern.

- *Modelling challenges*

Modelling ice stream and surge-type glacier behaviour has proved to be a challenge, as 3-D higher-order ice sheet models need to be coupled to physically based subglacial processes models (Bougamont et al., 2011; Kirchner et al., 2011; Van Pelt and Oerlemans, 2012), where mechanical properties of subglacial till bed evolve in response to dynamical, hydrological and thermal changes. A recent study by Gladstone et al. (2014) highlighted the importance of incorporating basal processes in ice flow models and stated that models of surge-type glaciers should be able to simulate subglacial hydrology and its impact on bed yield strength. They suggested that modelling the interaction between sediment properties and water pressure evolution was essential. Schoof and Hewitt (2013) point out numerous poorly understood thermomechanical feedbacks in ice flow and describe the current, often poorly constrained, state of models for ice-sheet sliding and subglacial drainage, as well as their role in ice-stream dynamics. They underscore the need for the models to accurately describe the coupling between ice dynamics, basal conditions and subglacial hydrology. Glacial geological research on surge-type glaciers in Iceland, emphasizing subglacial and ice marginal processes and products, can provide boundary conditions for improved ice stream and surge-type glacier models. Recent data highlight how mechanical and hydrological conditions of a deformable substrate have affected surging of Brúarjökull and Eyjabakkajökull and the resulting landforms and patterns of sediment distribution. One research challenge is to better understand ice dynamics and subglacial conditions during the last phase of the surge when sediment wedge and surge moraines develop, as well as the development of ice marginal crevasses and how they affect subglacial stresses and thereby drumlin formation (Benediktsson et al., in review-a).

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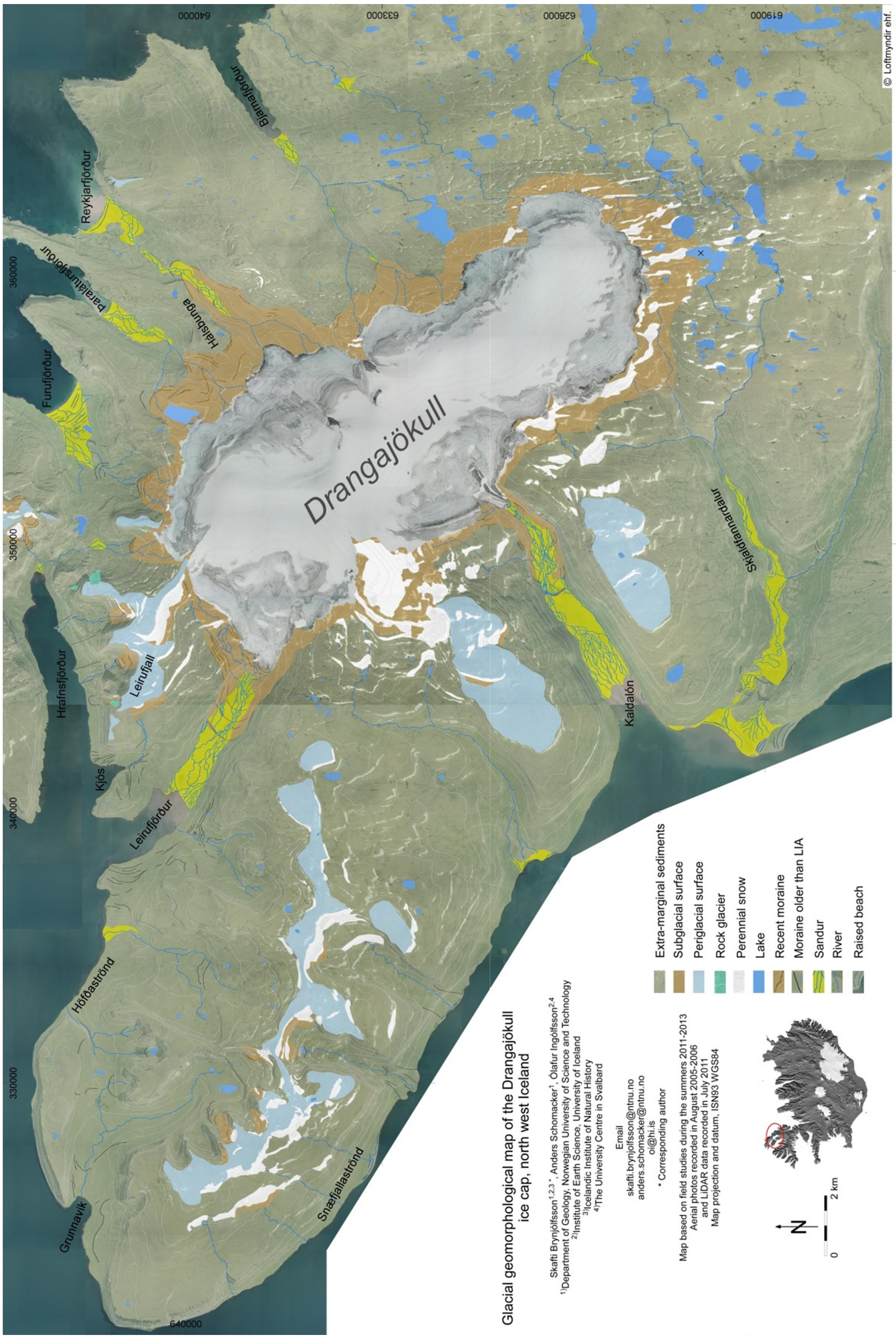
Appendix VI


Glacial geomorphological map of the Drangajökull ice cap, north west Iceland
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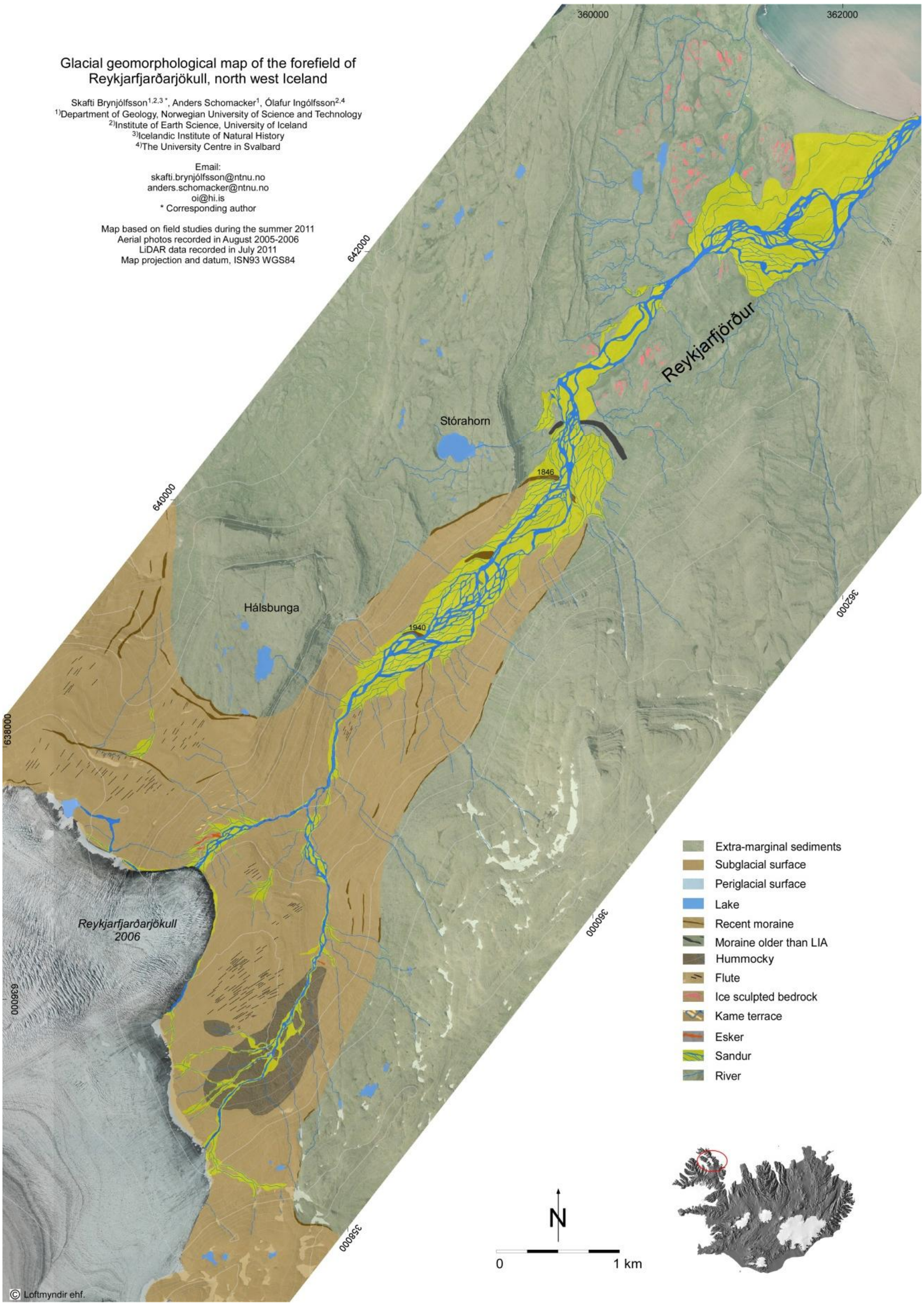
Map based on field studies during the summers 2011-2013
Aerial photos recorded in August 2005-2006
and LIDAR data recorded in July 2011
Map projection and datum, ISN93 WGS84
Extra-marginal sediments
Subglacial surface
Periglacial surface
Rock glacier
Perennial snow
Lake
Recent moraine
Moraine older than LIA
Sandur
River
Raised beach
N
0
2 km
© Loftmyndir ehf.
Drangajökull
Bjarnarfjörður
Reykjarfjörður
Þaralátursfjörður
Furufjörður
Hrafnsfjörður
Leirufjall
Kjós
Leirufjörður
Höfðaströnd
Grunnavík
Snæfjallaströnd
Kaldalón
Skjaldfannardalur
Hálsbunga
640000
633000
626000
619000
360000
350000
340000
330000


Glacial geomorphological map of the forefield of
Reykjarfjarðarjökull, north west Iceland
Skafti Brynjólfsson[1,2,3]*, Anders Schomacker[1], Ólafur Ingólfsson[2,4]
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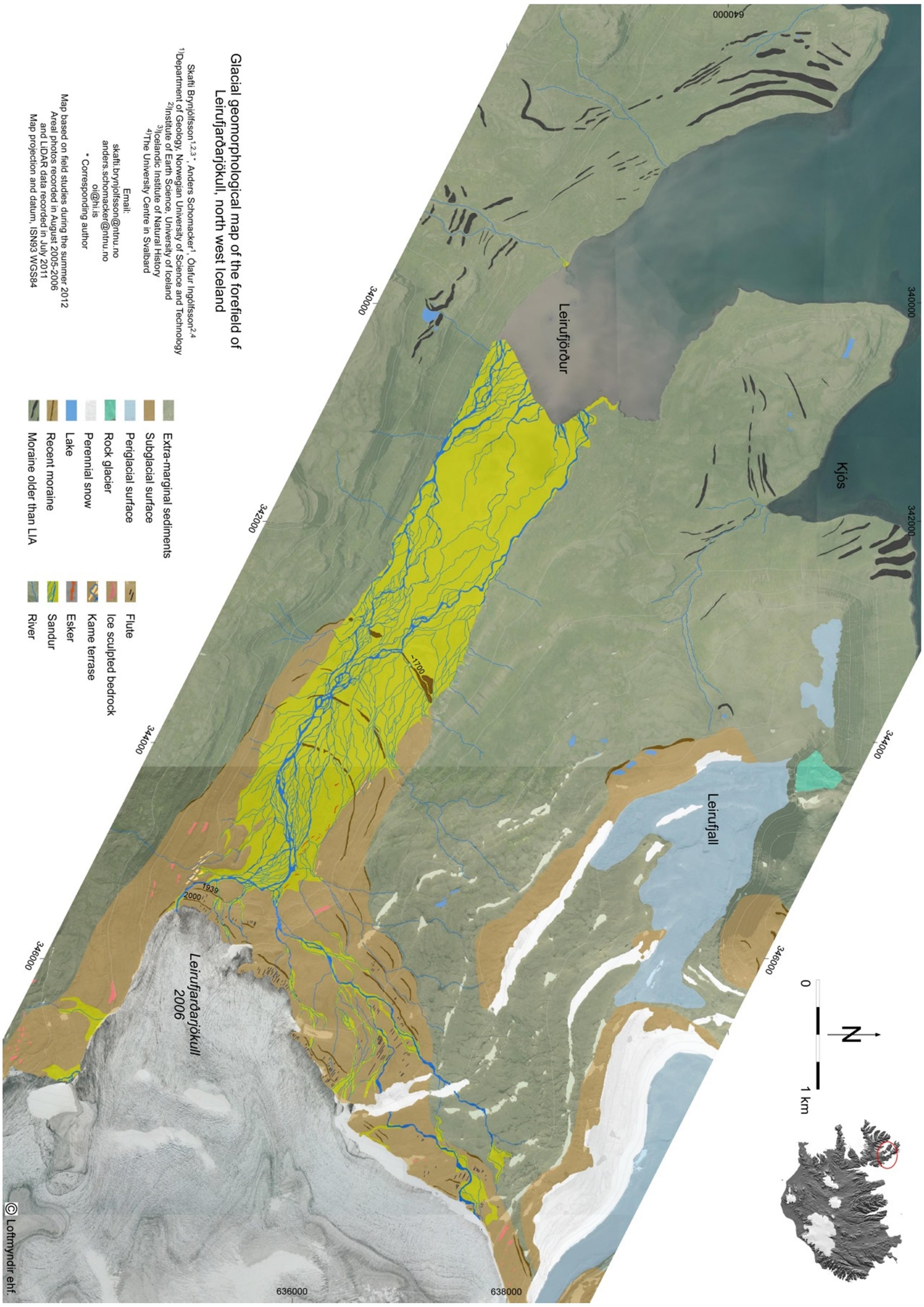
Map based on field studies during the summer 2011
Aerial photos recorded in August 2005-2006
LiDAR data recorded in July 2011
Map projection and datum, ISN93 WGS84
360000
362000
642000
640000
638000
636000
356000
358000
Reykjarfjörður
Stórahorn
1846
1940
Hálsbunga
Reykjarfjarðarjökull
2006
Extra-marginal sediments
Subglacial surface
Periglacial surface
Lake
Recent moraine
Moraine older than LIA
Hummocky
Flute
Ice sculpted bedrock
Kame terrace
Esker
Sandur
River
N
0
1 km
© Loftmyndir ehf.


Glacial geomorphological map of the forefield of
Leirufjarðarjökull, north west Iceland
Skafti Brynjólfsson[1,2,3*], Anders Schomacker[1], Ólafur Ingólfsson[2,4]
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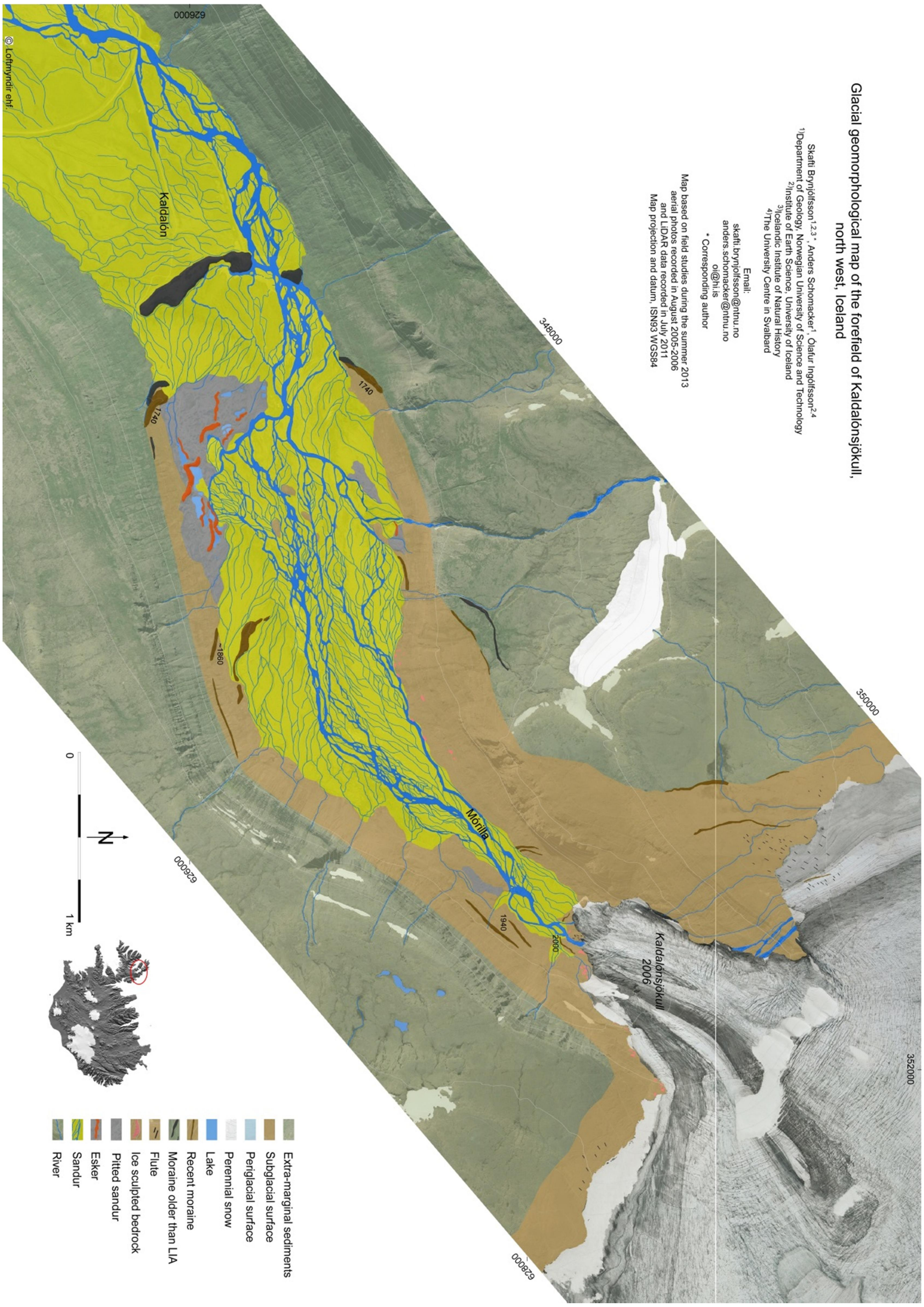
Map based on field studies during the summer 2012
Areal photos recorded in August 2005-2006
and LIDAR data recorded in July 2011
Map projection and datum, ISN93 WGS84
Extra-marginal sediments
Subglacial surface
Periglacial surface
Rock glacier
Perennial snow
Lake
Recent moraine
Moraine older than LIA
Flute
Ice sculpted bedrock
Kame terrase
Esker
Sandur
River
Leirufjörður
Kjós
Leirufjall
Leirufjarðarjökull
2006
1700
1939
2000
N
0
1 km
640000
340000
342000
344000
346000
636000
638000
© Loftmyndir ehf.


Glacial geomorphological map of the forefield of Kaldalónsjökull, north west, Iceland
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Map based on field studies during the summer 2013
aerial photos recorded in August 2005-2006
and LiDAR data recorded in July 2011
Map projection and datum, ISN93 WGS84
626000
348000
350000
352000
628000
Kaldalón
Mórilla
Kaldalónsjökull
2006
1740
1860
1940
2000
0
1 km
N
© Loftmyndir ehf.
Extra-marginal sediments
Subglacial surface
Periglacial surface
Perennial snow
Lake
Recent moraine
Moraine older than LIA
Flute
Ice sculpted bedrock
Pitted sandur
Esker
Sandur
River