TITLE: INCISION AND AGGRADATION IN PROGLACIAL RIVERS: POST-LITTLE ICE AGE LONG-PROFILE ADJUSTMENTS OF SOUTHERN ICELAND OUTWASH PLAINS

SHORT TITLE: INCISION AND AGGRADATION IN PROGLACIAL RIVERS

Erwan Roussel[1], Philip M. Marren[2], Etienne Cossart[3], Jean-Pierre Toumazet[1], Marie Chenet[4], Delphine Grancher[4], Vincent Jomelli[4]

(1) Université Clermont Auvergne, CNRS, GEOLAB, F-63000 Clermont-Ferrand, France

(2) Department of Geography and International Development – University of Chester – Chester – United Kingdom.

(3) Université de Lyon (Jean Moulin, Lyon 3), UMR 5600 CNRS – Environnement Ville Société, Lyon, 69007, France

(4) Université Paris 1 Panthéon-Sorbonne, CNRS, UMR 8591, Laboratoire de Géographie Physique, 92195, Meudon, France

Abstract:

The retreat of glaciers in response to climate warming leads to substantial changes in meltwater and sediment yield. Glacial shrinkage also induces the emergence and growth of proglacial margin landforms which strongly affect water and sedimentary transfers from the glacier to the outwash plains. On a decadal-timescale, field observations show that outwash plains of retreating glaciers typically exhibit proximal incision which decreases in magnitude downstream and stops at an inflection point where aggradation begins. Nevertheless, there is a lack of knowledge about the rates and magnitude of this fluvial adjustment and the effects of the proglacial margin configuration on the temperance or the aggravation of this fluvial adjustment to glacier retreat. This paper investigates the proglacial rivers of 14 retreating glaciers in southeast Iceland over a post-Little Ice Age timescale, combining fluvial deposits mapping, lichenometric dating and long-profile measurements of proglacial fluvial terraces. Our results demonstrate that: (1) proximal incision, associated with distal aggradation and downstream migration of the inflection point is the dominant pattern of proglacial river response to post-LIA glacier retreat in Iceland; (2) estimated mean rates of downstream migration of the inflection point range between 5 and 46 m.a^{-1}; (3) the downstream migration rate of the inflection point is positively correlated with the proportion of proglacial lakes within the glacier foreland. These findings suggest that proglacial margins dominated by proglacial lakes intensify the rates of proximal incision and inflection point migration.

Keywords: proglacial river, glacial retreat, proglacial landforms, fluvial geomorphology, paraglacial processes

1. INTRODUCTION

Since the end of the Little Ice Age, rising temperatures have elevated the equilibrium lines altitude of many glacial systems and intensified ice melting (Barry, 2006; Vaughan *et al.*, 2013). This implies an adjustment of glacier geometry, which is usually reflected by the retreat of the glacier front. Glacial retreat leads to the emergence and growth of proglacial margin landforms at the interface between the glacier and the fluvial system (Heckmann *et al.*, 2016; Carrivick & Heckmann, 2017). The water discharge of proglacial rivers is largely supplied by meltwater and they are directly connected to the glacier, such that sediment released by glacial melt is immediately available to be carried as fluvial sediment load. The supply of both water and sediment is highly sensitive to changes in glacier mass balance. The primary hydrological impact of glaciers on downstream river systems is to modulate the timing and seasonality of hydrological fluxes (Röthlisberger & Lang, 1987). A shift to deglacial conditions leads to marked changes in the seasonality of river flow (e.g., a shift of the hydrograph to an earlier peak flow in spring months) and the amount of annual glacier runoff (Bliss *et al.*, 2014; Milner *et al.*, 2017). Current trends in annual glacial runoff differ among world regions and glacier hypsometry but there is a consensus on the typical sequence of meltwalter changes during catchment deglaciation (Bliss *et al*. 2014): at the beginning of deglaciation, when the glacier is still sufficiently voluminous, the glacial meltwater tends to increase for a period lasting several decades, before decreasing when the ice stock runs out (Braun *et al.*, 2000; Jansson *et al.*, 2003; Marren, 2005; Huss *et al.*, 2008). The successive growth and decay of the glacial meltwater supply during deglaciation periods is both supported by palaeohydraulic studies (Maizels, 1986) and simulations of meltwater discharge (Bliss *et al.*, 2014). In Iceland, modelling approaches predict increased meltwater runoff until 2050. According to these simulations, glacier runoff declines afterwards (Flowers *et al.*, 2005; Jóhannessons *et al.*, 2006; Bliss *et al.*, 2014). A shift to deglacial conditions also affects the sediment supply to proglacial rivers in three ways: (1) The erosive capacity of a glacier is a function of basal sliding such that there is a dependence of erosion rate on the ice thickness which rely on the ice mass balance driven by climate conditions. Thus, in the long term, the loss of glacier thickness and volume limits their own ability to produce glacigenic material (Hallet *et al.*, 1996; de Winter *et al.*, 2012); (2) Deglaciation produces a lagged paraglacial sediment pulse caused by the reworking of stored englacial, subglacial and marginal sediments (Church & Ryder, 1972; Church & Slaymaker, 1989; Ballantyne, 2002; Mercier, 2008). Paraglacial sediment supply is highest during the initiation of glacial retreat and declines as sediment sources are exhausted or stabilized; (3) However, glacier retreat also induces the emergence and growth of proglacial margin landforms, like moraine ridges or proglacial lakes, which trap and store sediments and decrease the sediment connectivity in the glacier foreland (Cossart, 2008; Cossart & Fort, 2008; Heckmann & Schwanghart, 2013; Carrivick & Tweed, 2013; Cossart & Fressard, 2017; Cordier *et al*, 2017, Lane et *al.*, 2017). As a result, the downstream sediment flux of retreating glaciers may be significantly reduced, particularly where lasting proglacial lakes develop (Geilhausen *et al.*, 2013 ; Bogen *et al.*, 2015, Staines *et al.*, 2015). Nevertheless, sediment connectivity may temporarily or durably be restored by rare and extreme floods triggered by rainstorms (Marren & Toomath, 2013 ; Owczarek *et al.*, 2014), glacial outbursts (Westoby *et al.*, 2014; Worni *et al.*, 2014) or iceberg-jam failures (Roussel *et al.*, 2016).

Changes in meltwater and sediment fluxes from retreating glaciers lead to substantial geomorphic adjustments of proglacial rivers. Understanding the behaviour of proglacial rivers during periods of rapid change of glacier mass balance is especially important at present, given that climate change projections indicate that most glaciated regions will undergo a rapid decrease in glacier extent over the coming century (Marzeion *et al.*, 2012; Radić *et al.*, 2014), coupled with major hydrogeomorphological changes (Bliss *et al*., 2014; Milner *et al.*, 2017). Reduced sediment flux from a retreating glacier usually induces incision of the upper reach of the proglacial river (proximal outwash plain) and the formation of a single deep channel (Fahnestock, 1969 ; Maizels, 1979, 1983 ; Germanoski & Schumm, 1993; Gurnell *et al.*, 1999; Marren, 2005 ; Wilkie & Clague, 2009, Owczarek *et al.*, 2014). In Iceland, this fluvial response has been observed on the proglacial margins of the southern glacial outlets of the Vatnajökull icecap, especially on the proglacial rivers of Skaftafellsjökull and Svínafellsjökull, (Thompson & Jones, 1986 ; Thompson, 1988 ; Marren, 2002; Roussel *et al.*, 2008 ; Marren & Toomath, 2013, 2014). During the incision phase, fluvial bars tend to merge into larger sedimentary units, which are not in equilibrium with the water and sedimentary flux (Germanoski & Schumm, 1993). The incision of the proximal river results in the establishment of glacio-fluvial terraces whose morphosedimentary features reflect the former hydrological conditions before the incision (Marren, 2005). Flume experiments show that the long profile of an incising braiding river is usually characterized by an inflection point, downstream of which aggradation is dominant (Germanoski & Schumm 1993). Indeed, the reworked material of the incising upper reach is transported and deposited downstream induces an increase of the braiding intensity of the lower reach of the river. When the sediment delivery deficit continues, flume experiments have shown that the inflection point, which separates the incision from the aggradation reach, tends to migrate downstream (Germanoski & Harvey, 1993). At the same time, glacial retreat promotes the formation of proglacial lakes that trap the coarser fraction of the sediment flux and may strengthen the proximal river incision and the aggradation of the distal river reach (Embleton & King, 1968; Chew & Ashmore, 2001; Benn *et al.*, 2003; Gardarsson & Eliasson, 2006; Schomacker, 2010; Carrivick & Tweed, 2013; Bogen *et al.*, 2015). Nevertheless, this pattern of fluvial response to glacier retreat can vary over space and time due to topographical constrains induced by marginal landforms: *(1)* the spatial and temporal evolution of the overall sediment connectivity within the proglacial margin can lead to damming and localised river aggradation (Cossart, 2008 ; Cossart & Fort, 2008); *(2)* Channel pattern changes can be controlled by moraine ridges between which the river is forced to find a course (Marren & Toomath, 2013, 2014).

Field observations report that most proglacial rivers associated with retreating glaciers in Iceland exhibit proximal incision which decreases in magnitude downstream and stops at an inflection point where aggradation begins (Marren, 2005 ; Roussel *et al.*, 2008 ; Marren & Toomath, 2013). However, no studies have measured longitudinal profile adjustments of this type on a large sample of proglacial rivers and over a sufficiently long time period to determine whether this behaviour is a representative response to glacial retreat. Furthermore, there is a lack of information about the rates and magnitude of this fluvial adjustment and the effects of the glacial retreat rate and the proglacial margin configuration on the temperance or the aggravation of the fluvial response to glacier retreat.

Therefore, this paper investigates the proglacial landscapes and the outwash plains of 14 retreating glaciers in southeast Iceland over a post-Little Ice Age timescale. It combines quantitative assessment of the glacier retreat, the landform assemblage of the proglacial margin and the post-LIA outwash plains changes through lichenometric dating and longitudinal profile measurements of active outwash and fluvial terraces. The aims of this paper are threefold: (1) to test the representativeness of the outwash plains response to glacier retreat described above over the time-scale of centuries; (2) to assess the rates of this fluvial response through the reconstruction of the downstream migration of the inflection point; (3) to explore the influence of the landforms assemblage of the proglacial margin on the rate of the outwash plains response.

2. Materials and Methods

Site conditions and history

Three southern Icelandic marginal areas were selected: the Sólheimajökull area in the south of the Mýrdalsjökull icecap, the Öræfajökull and southeastern Vatnajökull areas, both located on the southern margin of the Vatnajökull icecap (Figure 1). These three study areas present a sample of 14 glacier tongues which feed 12 proglacial rivers (Table 1). Two pairs of formerly coalescent glaciers feed proglacial rivers which merge within the proglacial margin (Hrútárjökull/Fjallsjökull and Skálafellsjökull/Heinabergsjökull). None of the studied glaciers exhibit surge dynamics (Thórarinsson, 1969; Björnsson *et al.*, 2003; Ingólfsson *et al.*, 2016). Three glaciers (Sólheimajökull, Kotárjökull and Heinabergsjökull) are known to be jökulhlaup outlets (Thompson & Jones, 1986; Sigurðsson, 1998; Björnsson, 1992; Roberts *et al.*, 2003; Russell *et al.*, 2003, 2010). For convenience, the investigated glaciers and outwash plains are called afterward by their ID number (Table 1) instead of their Icelandic name. Additional information on site conditions and history is available in the appendix.

Methods

In order to address the aims of this study, we followed three stage methodological workflow (Figure 2) including field surveying and GIS mapping of proglacial landforms, lichenometric dating of outwash terraces, and topographic surveys of the active proglacial rivers and outwash terraces.

Proglacial landform mapping

The proglacial margin plays an essential role in sediment connectivity and may impose sediment flux discontinuities into the glaciofluvial continuum. Therefore, characterizing the landform assemblage of the proglacial margins is essential to fully assess the fluvial response of outwash plains to post-LIA glacial retreat. The mapping work is based on the recognition of 11 typical classes of proglacial landforms in the field (Figure 2-A): Glacier front, fresh proximal till deposits (including dirty cones, hummocky debris, dead-ice with supraglacial debris), moraine ridge, active and abandoned intra-morainic plain, intra-morainic lake, proglacial lake, alluvial fan, rocky outcrop, active outwash and fluvioglacial terraces. Artificial dams and embankments built in the proglacial margins were also mapped as they may act on the sediment connectivity and marginal landforms evolution (e.g. inhibiting basal erosion of morainic ridges). LIA moraines mark the maximum glacier snout position during the LIA and their locations are well known for most of the glaciers we investigated. They have already been examined and mapped by several authors (Thórarinsson, 1956; Evans *et al.*, 1999, 2017; Bradwell, 2004; Chenet *et al.*, 2010, Hannesdóttir *et al.*, 2015). In the field, the locations of LIA moraines and outwash terraces were collected using a handheld GPS (Trimble Geoexplorer XH). Post-processing of the GPS signal enabled a planimetic precision of 1.5 m on average. The fieldwork phase ensured the robustness of the subsequent mapping work based on the photo-interpretation of recent photographs with GIS software. Coupling this field database to 2009 and 2012 georeferenced orthophotos (2 m resolution on average) provided by DigitalGlobe® within ArcGIS software (10.4), the landform assemblages in the proglacial margin, the outwash terraces, and the active outwash were mapped. The proglacial margin areas provide a quantitative assessment of the deglaciated surface since the LIA. An average linear retreat of the glacier fronts was computed by dividing the proglacial margin area by the glacier front length for each glacier. A manual measurement of the linear retreat of glacier fronts was also performed, along the centerline of the glacier tongues, between the maximum LIA moraines and the glacier front in 2009.

Lichenometric dating

Lichenometric dating of outwash terraces (Figure 2-B) was performed using the generalized extreme value approach (GEV) described in Jomelli *et al.* (2007) for the statistical treatment of the measurements of the longest axis of the thalli of *Rhizocarpon Geographicum*. This procedure has already been successfully applied in Iceland to date the LIA moraines of the south Vatnajökull (Chenet *et al.*, 2010). This method provides a more statistically robust estimate of the age of the deposits because it takes into account the bias induced by the strategy of data collection (i.e. the selection of the largest thalli whose distribution is similar to the model GEV). In addition, this method provides confidence intervals to assess the quality of the estimated age.

In the field, measurement of lichen long axes was carried out according to the procedure recommended by Jomelli *et al.* (2007): *(i)* a random selection of 50 boulders distributed over the whole surface of the deposit; *(ii)* on each boulder, the long axis of the largest thallus of *Rhizocarpon Geographicum* was measured with an accuracy of 0.5 mm. Coalescing lichens and thalli with irregular shape are systematically rejected. Consistent with all lichenometric methods, two datasets of thalli long axis were collected in the field: on already dated surfaces and on the deposits requiring dating. Eight well-known dated surfaces in an area of ca. 45 km2 in south foreland of Vatnajökull were used to ensure that climatic conditions do not vary excessively (Chenet *et al.*, 2010). Dated surfaces included jökulhlaup deposits, lava flows, rockfalls, dams, and moraines; all made of basalt to ensure lithological homogeneity. Surfaces were dated using historical descriptions, aerial photographs (available since 1945), and personal communications from the Skaftafell Park staff. Additional consideration of the statistical treatments applied to the lichenometric database are available in several papers (Cooley *et al*., 2006; Naveau *et al.,* 2007; Rabatel *et al.,* 2008).

Topographic survey

Longitudinal topographic profiles of active outwash and outwash terraces were surveyed in the field. Immediately downstream of LIA moraines of each investigated glacier, valley transverse profiles were also undertaken in order to capture the nesting of outwash terraces (Figure 2-C). Following the main direction of flow, longitudinal profiles of the active outwash and terraces were surveyed from the LIA moraines to the downstream edge of the fluvial deposit (confluence with a tributary or the shoreline). Transverse topographic profiles located just downstream of LIA moraines enable the computation of the relative altitude of each outwash terraces in respect to the active outwash. They are essential to align longitudinal profiles between them and to capture the downstream migration of the inflection points separating the upstream incising reach from the downstream aggrading reach of the proglacial rivers. These profiles were achieved using a laser telemeter (Impulse 200LR from Laser Technology) with a centimeter accuracy and a tripod target. The average distance between two measurements was 20 meters. Two measurements by stations (foresight and backsight) were carried out in order to systematically check that the difference between measurements were below the telemeter accuracy (0.03m) and to limit error propagation. As the objective was to assess the topography of the outwash plains, measurements were carried out between topographic stations located on the heads of fluvial bars. Therefore, fine topography at the channel scale was not captured. Each topographic station was located using a handheld GPS (Trimble Geoexplorer XH).

Reconstruction of the downstream migration of the inflection point (IP)

Inflection points (*sensu* Germanoski & Harvey, 1993) separate upstream incising reaches from downstream aggrading reaches of a river. The reconstruction of changes in the location of the inflection point (IP) characterizes the evolution of the extent of proximal outwash affected by incision. IP changes were reconstructed based on the identification of intersection points of the longitudinal profiles of two successive (dated) terrace generations. These intersections are the former IP location separating the incision and the aggradation reach of the outwash plain at the date of the youngest of the intersecting terrace profiles (Figure 2-D). For each reconstructed IP, its location (distance from LIA moraines and elevation) was validated in the field and recorded with the handheld GPS (Trimble Geoexplorer XH). The initial IP location (distance = 0 m) at the LIA glacial maximum coincides with the contact between the LIA moraines and the oldest and highest outwash terrace (T1), if lichenometric dating of the latter is younger than LIA moraine ridges. In other words, at the LIA glacial maximum, aggradation of the outwash is likely predominant and there is no upstream incision because there is no outwash upstream to the LIA moraines. The reconstruction of the former location of the inflection point when T1 was incised and the IP location observed in 2007 enabled the computation of average rates of longitudinal migration (increasing distance of the IP from the LIA moraines) and altitudinal lowering (decrease in the IP elevation) of the IP through time. A synthetic metric of average rate of downstream migration of the inflection point was also computed as the Euclidean norm of the vector (square root of the sum of the square of coordinate values) defined by the two values of longitudinal migration and altitudinal lowering of the inflection points. Note that this average rate of IP downstream migration does not necessarily reflect a gradual progression of the inflection point as IP migration may occur suddenly during extreme events. Finally, a stepwise multiple linear regression was performed in order to test the effect of proglacial landform assemblage on the downstream migration rates of the inflection point.

3. RESULTS

Proglacial landform assemblages

The proportion of each landform unit within the post-LIA proglacial margin has been quantified (Table 2 and Figure 3). Glacigenic deposits, combining moraine ridges and fresh proximal till deposits, are the most abundant proglacial landforms (respectively 44.8 % and 5.2% on average of the investigated proglacial margins). The marginal landscape is obviously dominated by glacial deposits associated with seasonal fluctuations of glacier fronts and the continuous re-advances of the 1970s and 1980s. Proglacial margins of the steepest, Öræfajökull glaciers (glaciers 2 to 10), show a higher proportion of moraine deposits. Steep proximal topography is associated with a large amounts of moraine deposits that may be related to large production rates of glacigenic sediments and/or limited fluvial reworking. Abandoned and active intra-morainic plains form respectively around 16.8 % and 8.9 % on average of the proglacial landforms. The abundance of these two types of landforms testify the involvement of fluvial processes in the morphogenesis of the proglacial margin. Rocky outcrops (11.4% on average) are only present in the proglacial margins of piedmont glaciers (Glaciers 1, 11, 12 and 13) whose tendency to overdeepening would generally exhume more resistant areas of the glacial bed, and in the foreland of glacier 6 where steep proximal relief tends to rapidly clean the glacial bed and remove till layers via gravity and runoff processes. Proglacial and intra-morainic lakes (respectively 10.4% and 1.5% on average) are totally absent from the proglacial margins of the steepest glaciers of the Öræfajökull sector (glaciers 6 and 7), most likely due to the extreme steepness of the ablation zone and the glacier forefield. Alternatively, proglacial lakes are common on the margin of piedmont glaciers of the southeastern sector of Vatnajökull where their large frontal lobes probably favored overdeepening of the marginal areas during the LIA glacier advance. Alluvial fans (0.6 % on average) are only present in steep marginal settings (glaciers 7 and 9). This association seems logical since development of alluvial fans is strongly dependent on the presence of steep marginal slopes. Embankments (0.3 % on average) are only present on the margins of two piedmont glaciers (glaciers 3 and 13). Artificial embankments are associated either with tourist trails running through the proglacial margin (in the case of glacier 13), or with consolidation structures of the main Icelandic road (road number 1) built through the LIA moraines (glacier 3).

Outwash terrace mapping and dating

Outwash terrace mapping and lichenometric dating (Figure 3) emphasizes four types of fluvial processes affecting the proglacial river responses to post-LIA glacier retreat: (1) post-LIA transient aggradation; (2) incision and contraction of the proximal outwash; (3) shifts in the location of the fluvial activity due to stream capture occurring within the proglacial margin and (4) outwash incision prior to LIA glacial maximum. Based on the analysis of glacier foreland mapping and lichenometric dating of outwash terraces, an overview of successive aggrading and incising stages of proximal outwash response to post-LIA glacial retreat has been produced (Figure 4). The outwash terraces of the proglacial rivers 6 and 7 could not be dated due to the absence of lichens: the oldest terrace (T1) of the outwash plain 6 is the historic jökulhlaup deposit of 1727 and is exclusively composed of fine materials devoid of lichens. The terrace (T1) of the outwash plain 7 is lightly vegetated and its abandonment by proglacial flows might be recent and artificial, relating to the upstream embankment structure (Figure 3). The youth of this fluvial deposit may explain the absence of lichens on the proximal surfaces.

Post-LIA transient aggradation of proximal outwash

First generation outwash terraces (T1 on Figure 3) are adjacent to the moraines ridges marking the maximal glacier extent during the LIA. They are interpreted as fill terraces built up by the aggrading proglacial river during the LIA glacier advances. Dating reveals that proximal aggradation, or at least the maintenance of the outwash plain in a steady state, has persisted after the glaciers reached their maximum LIA extent and their fronts likely initiated their retreat or stagnation phase. This post-LIA aggradation stage (or steady state phase) affected the proximal outwash of the proglacial rivers 1, 2, 5, 8, 9, 11, 13 and 14. The duration of this aggradation or steady state stage (computed as the difference between the T1 lichenometric dating and the age of LIA moraines) is highly variable and ranges between 5 (±14) and 147 (±15) years for the outwash 11 and 5 respectively (Figure 4 and Table 3). The outwash 7 is the only one which seems to persist in the aggradation stage (Figure 4). Nevertheless, as its proximal reach is affected by an artificial embankment, its fluvial response is not fully controlled by natural drivers. Based on the other investigated outwash surfaces, aggradation may be considered as a transient stage of post-LIA outwash response.

Contraction and incision of the proximal outwash

The post-LIA contraction of the proximal outwash width is detectable in the successive abandonment of outwash terraces (T2 to T5 on Figure 3) by proglacial rivers. These fluvial deposits are interpreted as nested cut terraces and reveal the incising dynamic affecting proximal outwash. This glaciofluvial evolution follows the transient aggradation stage and is still currently the dominant behaviour in most of the rivers. The outwash plain 7 is the only one which appears to still be aggrading or remaining in a steady state, likely due to the artificial constraints described above (Figure 4). On the other outwash plains, the proximal fluvial activity was gradually concentrated downstream of a single breach in the LIA moraines. Currently, the active outwash plain is framed by multiple paired or unpaired terraces whose preservation is dependent on the lateral mobility and/or the rate of proximal river incision (Charlton, 2008).

Stream capture and shift in the outwash plain location

The mapping of glacier forelands (Figure 3) and the sequence of outwash terrace abandonment (Figure 4) also shows a typical marginal fluvial process: stream capture within the proglacial margin leading to a shift of the outwash plain location. This process affected the outwash plain 3 in 1939 (±3), 5 in 1891 (±3), 11 in 1914 (±8) and 1959 (±5), and 13 in 1918 (±7). Stream capture seems to preferentially occur within proglacial margins where two ice lobes were coalescent at the LIA glacial maximum (Figure 3). This is the case for the glaciers 3, 11 and 13 (Figure 3) whose glacial fronts were respectively coalescing with those of 4, 12 and the western arm of 13 at the LIA glacial maximum. Stream capture and location shift of the proximal outwash plain can occur slowly, and follows a sequence of proximal outwash narrowing: In the foreland of glaciers 11 and 12(Figure 3), three outwash terraces (T1, T2 and T3) were formed due to the contraction of the proximal braiding belt of the outwash plain shared by the two coalescing glaciers. In 1914, about 50 years after the LIA glacial maximum, a first stream capture occurred within the glacier 11 proglacial margin causing the abandonment of the outwash plain T3 and the development of the outwash plain T4. After 45 years of activity (during which T4 may have eroded the outwash terrace T3), a new stream capture occurred in the proglacial margin of glacier 11. T4 is abandoned in 1959 in favour of glacier 11 intra-morainic plains and the outwash plain which is still currently active. Ultimately, proximal incision and stream capture processes do not exclude each other. Moreover, according to our results, stream captures occur exclusively in the incising stages of the proximal outwash.

Proximal outwash incision prior to the LIA maximum

The lichenometric dating of the first terrace generation (T1) of the proglacial rivers 3 and 4 predates the LIA glacial maximum (M1). According to these lichenometric ages, the abandonment of outwash T1 downstream of the glacier 3 may have occurred 16 years (±25 yr.) before the LIA maximum. But, as the absolute uncertainty exceed the difference between the ages of M1 and T1, proximal outwash incision and LIA maximum could also have occurred synchronously between 1864 and 1873. Dating results are much more surprising for the river 4, since the abandonment of the oldest terrace occurred almost two centuries (179 ±21 years) before the LIA maximum (Figure 3 and Figure 4). Thus, proximal incision, or at least lateral contraction, of the outwash of the rivers 3 and 4 began before the glacier reached its maximum LIA extent. Early incision also affected the outwash 6 whose T1 terrace (Figure 3) was built by the 1727 jökulhlaup. The incision of this deposit may be independent of the fluctuations of the front of the glacier 6 and is instead related to a progressive return of the proglacial river to pre-jökulhlaup conditions (Thompson & Jones, 1986).

Outwash long profile adjustment

Outwash longitudinal profiles reveals a dominant pattern of fluvial adjustment characterized by a proximal incision often accompanied by distal aggradation (Figure 5). The inflection point (IP) connecting the incising upper reach and the aggrading lower reach tends to migrate downstream through time.

Proximal incision, slope decrease and distal aggradation

Post-LIA proximal incision can be quantitatively assessed for the rivers 1, 2, 8, 9 and 14, which were not affected by stream captures or early incision prior to the LIA glacial maximum. These proglacial rivers present several nested outwash terraces, the oldest of which, (T1) is always backed by the LIA moraine ridges (Figure 5). More recent outwash terraces can be disconnected from these moraines. This is the case for the rivers 8, 9 and 14. This observation reflects either a downstream migration of the proximal outwash contraction or the destruction of the upstream part of recent terraces due lateral erosion and avulsion of the active river channel. Following the transient aggradation stage (or steady state stage), the initiation of proximal incision implies the abandonment of the oldest outwash terrace (T1) by river flows. As proximal degradation continues, proglacial rivers abandon younger terraces. This is particularly the case for the rivers 2 and 14 where six and four generations of terraces are distinguishable, respectively. The rate of proximal outwash incision, computed over the period from the abandonment of the oldest terrace until 2007, averages 0.073 ± 0.03 $m.a^{-1}$and ranges from 0.045 ± 0.006 $m.a^{-1}$ to 0.131 ± 0.012 $m.a^{-1}$ for the rivers 2 and 14 respectively (Table 3).
The slope of the proximal rivers has decreased almost systematically since the LIA maximum. This is qualitatively detectable on Figure 5 by a visual comparison of the long profiles of the first generation of post-LIA outwash terraces (T1) with those active in 2007 (A). The mean slope of the proximal rivers (the first 500 meters from the maximum LIA moraines) were calculated using a simple linear regression: results indicate that proximal slope decreased by 0.011 $m.m^{-1}$ (SD =0.009 $m.m^{-1}$) on average between the first-generation of post-LIA outwash terraces and the active rivers in 2007. However, there is a high variability in the post-LIA lowering of the proximal slope. For example, the proximal slope values of the active and the outwash terraces of the proglacial river 7 are identical, in agreement with the absence of proximal incision (Table 3). Conversely, the proximal slope of the river 9 decreased by 0.0311 $m.m^{-1}$, dropping from 0.0339 $m.m^{-1}$ in 1891 (79 years after the LIA maximum) to 0.0028 $m.m^{-1}$ in 2007 (195 years after the LIA maximum PAG). For the river 9, the

comparison of the old (terrace T1) and current (2007) proximal slopes is only based on the first 200 meters from the LIA maximum moraines due to the narrowness of the conserved fragment of T1 terrace. Consequently, the slope measurements on this outwash plain reflect more closely the proximal concavities and are difficult to compare with the other rivers.

Decrease of the proglacial rivers slopes is mainly due to the proximal incision of the outwash plains. However, proximal incision is often associated with a distal aggradation of the outwash plain, which contributes to the overall decrease of the proglacial river slope. Indeed, nine of the twelve studied outwash plains present this type of post-LIA adjustment of their longitudinal profile: these include the outwash plains of glaciers 2, 3, 4, 5, 8, 9, 11, 13 and 14 (Figure 5).

Downstream migration of the inflection point (IP)

Post-LIA changes in the location of the inflection point (IP) were reconstructed for the proglacial rivers 2, 3, 4, 5, 8, 9, 11, 13 and 14 (Figure 5). The longitudinal profiles of terraces and active outwash of the proglacial rivers 1, 6 and 7 do not exhibit inflection points. Results demonstrate the post-LIA downstream migration of the inflection points, indicating a longitudinal increase of the extent of proximal outwash area affected by incision (Figure 4). This evolution may indicate a perennial sediment deficit in the marginal area leading to a downstream progression of the proximal incision. According to these results, the sediment recharge of the proglacial rivers is primarily carried out by the reworking of bed material (proximal incision) rather than lateral erosion of the former proximal outwash terraces (proximal contraction).

Average rates of longitudinal migration, elevation lowering and downstream migration of the inflection points were also computed for the proglacial rivers 2, 3, 4, 5, 8, 9, 11, 13 and 14 (Table 3). Rates of IP longitudinal migration average 20.4 m.a^{-1} (SD=13.6 m.a^{-1}) and range from 5.2 m.a^{-1} for the river 5 to 46. 4 for the river 14. Altitudinal lowering rates of the inflection points average 0.3 m.a^{-1} (SD=0.15 m.a^{-1}) and range from 0.09 m.a^{-1} for the river 9 to 0.5 m.a^{-1} for the river 14. The synthetic rates of IP downstream migration are similar to longitudinal migration rates (average = 20.4 m.a^{-1}, SD=13.6 m.a^{-1} and values range from 5.2 m.a^{-1} for the Virkisá to 46. 4 for the Hoffellsá).

A category of proglacial rivers in which the downstream migration of the inflection point is rather slow can be identified. These are rivers 4, 5, 8, and 9, whose post-LIA average rates of IP longitudinal migration and elevation lowering are respectively lower than 15 m.a^{-1} and 0.2 m.a^{-1}. A second category of proglacial rivers (rivers 2, 11 and 14), exhibit the highest rates of IP longitudinal migration and elevation lowering, exceeding 28 m.a^{-1} and 0.40 m.a^{-1} respectively. Between these two categories, the rivers 13 and 3 display intermediate rates of IP longitudinal migration (16.85 and 26.5 m.a^{-1} respectively) and IP elevation lowering (0.25 and 0.32 m.a^{-1} respectively). These results suggest that the proximal sediment deficit of the rivers 4, 5, 8 and 9 is less severe than for the rivers 2, 11, and 14, whose inflection points, and the spatial extent of proximal incision, rapidly migrated downstream.

Linear regression analysis (Figure 6-A) indicates that average rates of longitudinal migration and altitudinal lowering are significantly and positively correlated (R^2=0.791, P=0.001). It suggests that the overall process of post-LIA downstream migration of proximal incision involves the same proportion of longitudinal migration and elevation lowering of the IP whatever the intensity of the downstream migration.

The stepwise multiple linear regression procedure test the dependency of the IP downstream migration rate on the landform assemblage of the proglacial margin and the post-LIA glacial retreat. The linear stepwise procedure only selected the proportion of proglacial lakes within the marginal landscape as a significant model parameter ($t=3.527$, $p=0.01$). The downstream migration rate of the inflection point is positively and significantly correlated with the percentage of proglacial lakes within the proglacial margin ($R^2=0.640$, $P=0.010$). According to this result (Figure 6-B), the development of large proglacial lakes in the glacier foreland increases the rate of that the IP migrates downstream.

Atypical outwash adjustments

The outwash plains of the glaciers 1, 6 and 7 exhibit different post-LIA adjustments of their longitudinal profile (Figure 5). The outwash plain of glacier 1 has experienced an incision of its proximal part since the LIA maximum. However, no evidence of distal aggradation was found in the field. In addition, the post-LIA decrease of the proximal slope remains extremely low (0.013 $m.m^{-1}$ for T1 terrace versus 0.012 $m.m^{-1}$ for the river in 2007). The deep incising efficiency of recurrent volcanic jökulhlaups which affect the Glacier 1 glacier foreland may produce this uniform degradation of the entire outwash plain and inhibits post-LIA distal aggradation. The glacier 6 is also affected by volcanic jökulhlaups. Moreover, its outwash terrace T1 (Figure 5) constitutes the deposit of the historical jökulhlaup of 1727. The incision of this deposit may be independent of the fluctuations of the glacier front and is rather related to a progressive return of the river 6 to the pre-jökulhlaup conditions (Thompson & Jones, 1986). Alternatively, the case of the outwash plain 7 is more surprising: no incision of the abandoned and vegetated outwash plain was observed on the field. As mentioned above, the vegetated outwash plain of glacier 7 seems extremely recent (absence of lichens). Its abandonment by proglacial flows seems artificial and related to the embankment structure located on the right bank of the upstream reach of the river 7 (Figure 3). As a consequence, the abandoned outwash plain of the river 7 and the decrease in the lateral extent of the fluvioglacial dynamic cannot be considered as a purely natural response of the proglacial river to glacial retreat. Alternatively, large alluvial fans located within the proglacial margin of this steep glacier may provide an abundant sediment supply that mitigates the marginal sediment deficit and inhibits proximal incision.

Local controls on the long profile of active outwash in 2007

Long profiles of the active outwash of the glacier 4, 8, 10 and 13 (in 2007) reveal the local controls exerted by embankment structures or rocky outcrops on the efficiency of sediment transport. Indeed, these long profiles display significant decrease of local river slope values in the upstream vicinity of bridgeworks (related to the Icelandic Ring Road) or rocky outcrops. These zones force sediment deposition, related to dam effects and lateral constraints exerted by embankment and outcrops, and can hinder the downstream migration of sedimentary waves and mitigate the standard pattern of post-LIA adjustment of the outwash long profile by moderating proximal incision. Alternatively, the large alluvial fan located on the left bank of the river 14 seems to exert a control over the river long profile. A local increase in the slope value can be observed in the vicinity of the fan toe, probably due to lateral sediment contribution coming from the alluvial fan (Knighton, 1998). Conversely, the alluvial fan located on the left bank of the distal part of the outwash plain 2 does not seem to exert any control over the slope values of the active river in 2007. Nevertheless, it appears that presence of large active alluvial fans can also modify the post-LIA adjustment of the outwash long profile by exerting control over local slope values.

4. DISCUSSION

The dominant sequence of post-LIA outwash adjustment: a two-stage descriptive model

According to our results, the dominant response of Type I (*sensu* Maizels, 1993) Icelandic proglacial rivers to post-LIA glacier retreat follow two successive stages: (1) a first stage of transient aggradation, or steady state, of the proximal outwash, (2) a second stage characterising by proximal incision, distal aggradation and the downstream migration of the inflection point (IP) which separates the upstream incising reach and the downstream aggrading reach of proglacial rivers.

Stage I: Outwash transient aggradation (or in balance), a paraglacial signature?

During the first stage, even if the post-LIA glacial retreat is initiated, the outwash plain is still aggrading or, at least, kept in a steady state (Figure 7-A). Two explanations of this transient aggrading stage are possible: (1) The glacial front may have registered a slight retreat during stage I and experienced a steady period with minor seasonal fluctuations that maintain high glacigenic sediment supply to proglacial rivers in which aggradation continues (or remains in its LIA maximum state); (2) Alternatively, if the ice front registered a pronounced and continuous post-LIA retreat, the transient aggradation stage can only be related to efficient recharge of the marginal sediment stock by paraglacial processes and / or greater rates of subglacial sediment evacuation. In particular, in the early stages of deglaciation, thermokarstic processes, related to the dismantling of dead-ice moraines, may be able to compensate for the decrease of glacigenic sediment supply induced by glacial retreat for a time (Mercier *et al*., 2009, Irvine-Fynn *et al.*, 2011). When the efficiency of subglacial sediment evacuation decrease, thermokarstic processes cease and the moraine ridges are stabilized, the paraglacial recharge of the proglacial fluvial system is no longer sufficient to maintain high sediment supply and stage I ends.

According to lichenometric dating of the first generation outwash terraces (T1), the duration of the transient aggrading stage is highly variable: it averages 61.8 years (SD= 40.48 years) and ranges between 5 (±14) and 147 (±15) years. However, the range of duration of post-LIA transient aggradation stage of the South Icelandic fluvioglacial plains is in agreement with previous studies on the duration of paraglacial processes as effective sediment providers for the proglacial fluvial system. On the proglacial margin of the Small River Glacier in Canada, marginal deposits and mostly abandoned intra-morainic plains have been shown to cease to function as sediment sources after several decades (Orwin and Smart, 2004). Our results tends to support this hypothesis of a stabilization of morainic ridges and intra-morainic deposits within half a century or even a century. This period of time is too short to encompass all of what is commonly referred as the so-called paraglacial period (Ballantyne, 2002, Mercier, 2010, Cossart, 2005). More likely, the duration of stage I is only the time period during which the effectiveness of paraglacial processes is sufficient to, at least, keep the outwash plain in balance. As the marginal landscape differs widely from one glacier to another, the duration of

the transient aggrading period also varies accordingly. This variability may be related to the amount of the morainic sediment stock in the proglacial margin and its sensitivity to potential reworking processes.
In particular, the presence of dead ice in the moraine deposits may play a major role in the initiation of debris flows eroding moraine ridges inducing the recharge of the proglacial fluvial system with paraglacial sediments (Church & Ryder, 1972, Church & Slaymaker 1989, Mercier, 2009).

Stage II: proximal incision, distal aggradation and downstream progression of incision wave

The incision stage of the proximal outwash plain is initiated when the sediment stock of the proglacial margin is deficient and / or disconnected and the efficiency of paraglacial remobilization of the morainic and intra-morainic deposits is no longer sufficient to maintain the outwash plain in balance (Figure 7-A). The erodibility of the marginal sediment stock and the continuation of glacial retreat results in the lowering of the local base level which promotes stream captures within the proglacial margin. The occurrence of stream captures implies the abandonment by proglacial flows of former intra-morainic plains, often to the benefit of a single moraine breach which constitutes the favoured outlet of water and sediment flows to the outwash plain.
This post-LIA response of the marginal channel network was also observed in Spitsbergen on the Lovén glacier margins (Griselin 1982, Mercier, 2001, Mercier & Laffly, 2005). The response of south-Icelandic outwash plains to post-LIA glacial retreat is the consequence of this marginal sediment deficit and the concentration of water and sediment flux at a single moraine breach: The sediment recharge of the outwash plain is primarily carried out by the incision of the proximal fluvioglacial plain, resulting in the abandonment of the previous aggrading outwash plain and the contraction of the proximal fluvial activity width. The production of glaciofluvial sediments in the upstream incising reach caused further aggradation in the downstream reach. As the marginal sediment deficit continues, the degradation of the proximal outwash progresses downstream as an incision wave (Germanosky & Harvey, 1993). Rather than a uniform degradation of the outwash plain, the successive formation of different terrace generations reflect the crossing of geomorphological thresholds and the nonlinearity of the fluvioglacial response to fluctuations of water an sediment fluxes. These crossing of geomorphological thresholds may be linked to abrupt variations in the efficiency of the marginal sediment connectivity caused especially by moraines dams and their failures (Cossart, 2008, Cossart & Fressard, 2017).
Proglacial lakes within the marginal landscape also play a key role as they interrupt the delivery of water and sediment to the outwash plain (Schomacker, 2010; Carrivick & Tweed, 2013; Geilhausen, 2013; Bogen *et al.*, 2015, Staines *et al.*, 2015). According to our results, the abundance of proglacial lakes within the proglacial margin speed up the downstream progression of the incision wave affecting proximal outwash. They clearly reduce the marginal sediment connectivity in trapping the coarsest part of the glacigenic sediment production. Thus, in a similar way to artificial dams, proglacial lakes deliver at their outlets “hungry water” prone to channel incision (Kondolf, 1997) and bed armoring (Vericat et al., 2006). High-magnitude/low-frequency events emerging from proglacial lakes as outburst floods (Westoby *et al.*, 2014; Worni et al., 2014), rainstorm floods (Marren & Toomath, 2013) and iceberg jam floods (Roussel *et al.*, 2016) may be able to temporarily reconnect the sediment connectivity between the proglacial lake and the outwash plain.

Singularities in the dominant outwash adjustment model

The dominant sequence of post-LIA response of the South Icelandic type I outwash developed above has to be qualified on the basis of field observations. The reported singularities are not major divergences from the overall logic of the dominant outwash plain adjustment. They rather reflect the effects of individual marginal landscape assemblages on the response of outwash plains to post-LIA glacial retreat.

Coalescent glacier fronts at the LIA maximum and stream capture

The post-LIA response of the outwash plains 3, 5, 11 and 13 demonstrate that processes of marginal stream capture and large-scale shifts of the outwash location appear to be associated with a particular glacial configuration: Two coalescent lobes at the LIA maximum (Figure 7-B). As the contact between two glacier lobes is a preferential outlet for water and sediment fluxes (Thompson & Jones, 1986), the active proximal outwash plain is typically located downstream of the junction of the glacier fronts at the LIA glacial maximum. When the ice fronts of coalescent glaciers were disconnected following the post-LIA glacial retreat, the outwash plain shared by coalescent lobes was abandoned in favour of the intra-morainic plains of each of the two forefields which then developed two distinctive outwash plains. Downstream progression of incision waves and stream capture processes do not exclude each other. They may succeed each other in the history of outwash plain response to post-LIA glacial retreat.

Outwash incision prior to the LIA maximum

Evidence of proximal outwash plain incision prior to the LIA maximum (proglacial rivers 3 and 4) suggests two alternative interpretations: *(1)* a lateral contraction and/or a stream capture and a shift in the location of the fluvial activity may occur during periods of glacial advance or glacial stagnation. In periods of glacial advance, this kind of adjustment may be related to the damming effect of proglacial moraine ridges which spatially constrain and channelize water and sediments flows; *(2)* A lateral contraction and/or a stream capture and a shift in the location of the fluvial activity may be caused by a transient retreat of the ice front during the LIA. The later advance of the ice front, up to the maximum glacial extension of the LIA, does not produce a sufficient aggradation or widening of the outwash plain to destroy or bury the glaciofluvial deposit built prior to the temporary LIA glacial retreat. These scenarios may explain the conservation of outwash terraces older than the LIA maximum.

Deviation to the dominant outwash adjustment model: Jökulhlaups outwash (Type III)

The glacial tongues of the glaciers 1 and 6 are known to be outlets of jökulhlaups triggered by subglacial eruptions (Sigurðsson, 1998). Our results show an alternative post-LIA response of the longitudinal profile of the outwash plains 1 and 6. The river 1 experienced an incision of its upper reach since the LIA maximum. However, no evidence of distal aggradation was found in the field. In addition, the post-LIA decrease in the proximal outwash slope remains extremely low (0.013 $m.m^{-1}$ for T1 and 0.012 $m.m^{-1}$ for the active outwash). It suggests that the strong incisional power of the volcanic jökulhlaups that affect the foreland of glacier 1 is responsible for this uniform degradation of the outwash plain and the absence of distal aggradation. The outwash plain 6 is also affected by volcanic jökulhlaups (the outwash terrace T1 is the major deposit of the 1727 historical jökulhlaup). Following Thompson & Jones (1986), we believe that the successive incisions of this inherited deposit are independent of the fluctuations of the glacier 6 and rather testify to the gradual return of the outwash plain to pre-jökulhlaup conditions. In more general terms, and as Nicholas & Sambrook-Smith (1998) suggest, in terms of morphogenesis and reworking of the fluvioglacial sediment stock, the outwash plains impacted by jökulhlaups (Type III) are insensitive to the range of meltwater discharge associated with the "normal" glacial ablation. The size of the material deposited during these catastrophic floods is beyond the scope of seasonal proglacial flows. Consequently, the re-shaping of fluvioglacial forms of type III outwash is totally independent of glacial fluctuations. Their fluvial morphogenesis seems to depend exclusively on the frequency and magnitude of volcanic jökulhlaups.

5. CONCLUSION

Mapping, lichenometric dating and long-profiles surveys of Icelandic proglacial margins, glacio-fluvial terraces and active proglacial rivers permit to address the threefold aims of this study: (1) Proximal incision, associated with distal aggradation and downstream migration of the inflection point, is the dominant pattern of proglacial river response to post-LIA glacier retreat in Iceland. The post-LIA contraction and incision of Icelandic proximal outwash plains usually occurs after a first transient aggradation or steady state stage of proximal outwash (duration ranges between 5 ±14 and 147 ±15 years) which may be related to efficient recharge of the marginal sediment stock by paraglacial processes and / or greater rates of subglacial sediment evacuation. Proximal incision, distal aggradation and downstream migration of the inflection point can precede or be preceded by stream capture phenomena which particularly affects marginal landscapes where two glacier fronts were coalescent at the LIA maximum. Outwash plain incision prior to the LIA maximum have been observed and may be related to damming effect of proglacial moraine ridges or transient glacial retreat during the LIA. Two proglacial rivers affected by historical jökulhlaups exhibit an alternative adjustment rather controlled by extremes events than by glacier retreat. (2) The mean rates of downstream migration of the inflection point average 20.4 $m.a^{-1}$ (SD=13.6 $m.a^{-1}$) and range from 5.2 $m.a^{-1}$ for the river 5 to 46. 4 for the river 14. IP migration results suggest that the proximal sediment deficit of the rivers 4, 5, 8 and 9 is less severe than for the rivers 2, 11 and 14, whose inflection points, and the spatial extent of proximal incision, rapidly migrated downstream. (3) The downstream migration rate of the inflection point is positively correlated with the proportion of proglacial lakes within the glacier foreland. These findings suggest that proglacial margins dominated by proglacial lakes intensify the rates of the fluvial adjustment to glacial retreat.

Acknowledgements

Research was supported by the GEOLAB-CNRS 6042, and the GDR "Mutations Polaires" (Coord. M. Griselin). The authors thank the Skaftafell National Park which provided assistance and Pr. Johannes Steiger for his helpful suggestions. The contribution to the manuscript has to be shared between authors as follows: the investigation was outlined by the first author which carried out most of the field measurement activity. All authors contributed to analyse the results and to write the paper.

REFERENCES

Ballantyne CK. 2002. Paraglacial geomorphology. *Quaternary Science Reviews* **21**: 1935–2017. DOI: 10.1016/S0277-3791(02)00005-7

Barry RG. 2006. The status of research on glaciers and global glacier recession: a review. *Progress in Physical Geography* **30**: 285–306. DOI: 10.1191/0309133306pp478ra

Benn DI, Kirkbride MP, Owen LO, Brazier V. 2003. Glaciated Valley Landsystems. In: Evans DJA (ed) *Glacial Landsystems*. London: Arnold, 372–406.

Bennett MR, Huddart D, McCormick T. 2000. The Glaciolacustrine Landform–Sediment Assemblage at Heinabergsjökull, Iceland. *Geografiska Annaler: Series A, Physical Geography* **82**: 1–16. DOI: 10.1111/j.0435-3676.2000.00107.x

Björnsson H. 1992. Jökulhlaups in Iceland: prediction, characteristics and simulation. *Annals of Glaciology* **16**: 95–106. DOI: 10.1017/S0260305500004882

Björnsson H. 2002. Subglacial lakes and jökulhlaups in Iceland. *Global and Planetary Change* **35**: 255–271. DOI: https://doi.org/10.1016/S0921-8181(02)00130-3

Björnsson H, Pálsson F, Sigurðsson O, Flowers GE. 2003. Surges of glaciers in Iceland. *Annals of Glaciology* **36**: 82–90. DOI: 10.3189/172756403781816365

Bliss A, Hock R, Radić V. 2014. Global response of glacier runoff to twenty-first century climate change. *Journal of Geophysical Research: Earth Surface* **119**: 717–730. DOI: 10.1002/2013JF002931

Bogen J, Xu M, Kennie P. 2015. The impact of pro-glacial lakes on downstream sediment delivery in Norway. *Earth Surface Processes and Landforms* **40**: 942–952. DOI: 10.1002/esp.3669

Bradwell T. 2004. Lichenometric dating in southeast Iceland: the size–frequency approach. *Geografiska Annaler: Series A, Physical Geography* **86**: 31–41. DOI: 10.1111/j.0435-3676.2004.00211.x

Braun LN, Weber M, Shulz M. 2000. Consequences of climate change for runoff from Alpine regions. *Annals of Glaciology* **31**: 19–25. DOI: 10.3189/172756400781820165

Carrivick JL, Heckmann T. 2017. Short-term geomorphological evolution of proglacial systems. *Geomorphology* **287**: 3–28. DOI: 10.1016/j.geomorph.2017.01.037

Carrivick JL, Tweed FS. 2013. Proglacial lakes: character, behaviour and geological importance. *Quaternary Science Reviews* **78**: 34–52. DOI: 10.1016/j.quascirev.2013.07.028

Charlton R. 2008. *Fundamentals of fluvial geomorphology*. Oxon: Routledge, 234 pp.

Chenet M, Roussel E, Jomelli V, Grancher D. 2010. Asynchronous Little Ice Age glacial maximum extent in southeast Iceland. *Geomorphology* **114**: 253–260. DOI: 10.1016/j.geomorph.2009.07.012

Chew LC, Ashmore PE. 2001. Channel adjustment and a test of rational regime theory in a proglacial braided stream. *Geomorphology* **37**: 43–63. DOI: 10.1016/S0169-555X(00)00062-3

Church M, Ryder JM. 1972. Paraglacial sedimentation : a consideration of fluvial processes conditioned by glaciation. *Geological society of America Bulletin* **83**: 3059–3071. DOI: 10.1130/0016-7606(1972)83[3059:PSACOF]2.0.CO;2

Church M, Slaymaker O. 1989. Disequilibrium of Holocene sediment yield in glaciated British Columbia. *Nature* **337**: 452–454. DOI: 10.1038/337452a0

Cooley D, Naveau P, Jomelli V, Rabatel A, Grancher D. 2006. A bayesian Hierarchical Extreme Value Model for Lichenometry. *Environmetrics* **17**: 555–574. DOI: 10.1002/env.764

Cordier S, Adamson K, Delmas M, Calvet M, Harmand D. 2017. Of ice and water: Quaternary fluvial response to glacial forcing. *Quaternary Science Reviews* **166**: 57–73. DOI: 10.1016/j.quascirev.2017.02.006

Cossart E. 2005. Evolution géomorphologique du haut bassin durancien depuis la dernière glaciation (Briançonnais, Alpes française du Sud). Université Paris 7 - Denis Diderot, 382 pp

Cossart E. 2008. Landform connectivity and waves of negative feedbacks during the paraglacial period, a case study: the Tabuc subcatchment since the end of the Little Ice Age (massif des Écrins, France). *Géomorphologie : relief, processus, environnement* **14**: 249–260. DOI: 10.4000/geomorphologie.7430

Cossart E, Fort M. 2008. Sediment release and storage in early deglaciated areas: Towards an application of the exhaustion model from the case of Massif des Écrins (French Alps) since the Little Ice Age. *Norsk Geografisk Tidsskrift - Norwegian Journal of Geography* **62**: 115–131. DOI: 10.1080/00291950802095145

Cossart E, Fressard M. 2017. Assessment of structural sediment connectivity within catchments: insights from graph theory. *Earth Surface Dynamics* **5**: 253–268. DOI: 10.5194/esurf-5-253-2017

de Winter IL, Storms JEA, Overeem I. 2012. Numerical modeling of glacial sediment production and transport during deglaciation. *Geomorphology* **167**–**168**: 102–114. DOI: 10.1016/j.geomorph.2012.05.023

Dugmore AJ, Sugden DE. 1991. Do the anomalous fluctuations of Sólheimajökull reflect ice-divide migration? *Boreas* **20**: 105–113. DOI: 10.1111/j.1502-3885.1991.tb00299.x

Embleton C, King CAM. 1968. *Glacial and Periglacial Geomorphology*. London: Edward Arnold, 608 pp.

Evans DJA, Archer S, Wilson DJH. 1999. A comparison of the lichenometric and Schmidt hammer dating techniques based on data from the proglacial areas of some Icelandic glaciers. *Quaternary Science Reviews* **18**: 13–41. DOI: 10.1016/S0277-3791(98)00098-5

Evans DJA, Ewertowski M, Orton C. 2017. Skaftafellsjökull, Iceland: glacial geomorphology recording glacier recession since the Little Ice Age. *Journal of Maps* **13**: 358–368. DOI: 10.1080/17445647.2017.1310676

Fahnestock RK. 1969. Morphology of the Slims River. In: Bushnell VC and Ragle RH (eds) *Icefield Ranges Research Project, Scientific Results*. American Geographical society and Arctic Institute of North America, 161–172

Flowers GE, Marshall SJ, Björnsson H, Clarke GKC. 2005. Sensitivity of Vatnajökull ice cap hydrology and dynamics to climate warming over the next 2 centuries. *Journal of Geophysical Research* **110**: 1–19. DOI: 10.1029/2004JF000200

Gardarsson SM, Eliasson J. 2006. Influence of climate warming on Halslon Reservoir sediment filling. *Nordic Hydrology* **37**: 235–245. DOI: 10.2166/nh.2006.014

Geilhausen M, Morche D, Otto J-C, Schrott L. 2013. Sediment discharge from the proglacial zone of a retreating Alpine glacier. *Zeitschrift für Geomorphologie, Supplementary Issues* **57**: 29–53. DOI: 10.1127/0372-8854/2012/S-00122z

Germanoski D, Harvey MD. 1993. Asynchronous terrace development in degrading braided river channels. *Physical Geography* **14**: 16–38. DOI: 10.1080/02723646.1993.10642465

Germanoski D, Schumm SA. 1993. Changes in braided river morphology resulting from aggradation and degradation. *Journal of Geology* **101**: 451–466. DOI: 10.1086/648239

Griselin M. 1982. Les modalités de l'écoulement liquide et solide sur les marges polaires (exemple du bassin Loven Est, côte NW du Spitsberg). Université de Nancy II, 500 pp.

Gurnell AM, Edwards PJ, Petts GE, Ward JV. 1999. A conceptual model for alpine proglacial river channel evolution under changing climatic conditions. *Catena* **38**: 223–242. DOI: 10.1016/S0341-8162(99)00069-7

Hallet B, Hunter L, Bogen J. 1996. Rates of erosion and sediment evacuation by glaciers: A review of field data and their implications. *Global and Planetary Change* **12**: 213–235. DOI: 10.1016/0921-8181(95)00021-6

Hannesdóttir H, Björnsson H, Pálsson F, Aðalgeirsdóttir G, Guðmundsson S. 2015. Variations of southeast Vatnajökull ice cap (Iceland) 1650-1900 and reconstruction of the glacier surface geometry at the Little Ice Age maximum: Timing and reconstruction of the LIA maximum of SE-Vatnajökull. *Geografiska Annaler: Series A, Physical Geography* **97**: 237–264. DOI: 10.1111/geoa.12064

Hannesdóttir H, Guðmundsson S, Björnsson H, Pálsson F, Aðalgeirsdóttir G. 2010. Modelling the response of Vatnajökull´s southeastern outlet glaciers to climate change. *EGU General Assembly 2010*. Geophysical Research Abstracts

Heckmann T, McColl S, Morche D. 2016. Retreating ice: research in pro-glacial areas matters. Earth Surface Processes and Landforms 41: 271–276. DOI: 10.1002/esp.3858

Heckmann T, Schwanghart W. 2013. Geomorphic coupling and sediment connectivity in an alpine catchment — Exploring sediment cascades using graph theory. *Geomorphology* **182**: 89–103. DOI: 10.1016/j.geomorph.2012.10.033

Huss M, Farinotti D, Bauder A, Funk M. 2008. Modelling runoff from highly glacierized alpine drainage basins in a changing climate. *Hydrological Processes* **22**: 3888–3902. DOI: 10.1002/hyp.7055

Ingólfsson Ó, Benediktsson ÍÖ, Schomacker A, Kjær KH, Brynjólfsson S, Jónsson SA, Korsgaard NJ, Johnson MD. 2016. Glacial geological studies of surge-type glaciers in Iceland — Research status and future challenges. *Earth-Science Reviews* **152**: 37–69. DOI: 10.1016/j.earscirev.2015.11.008

Irvine-Fynn TDL, Barrand NE, Porter PR, Hodson AJ, Murray T. 2011. Recent High-Arctic glacial sediment redistribution: A process perspective using airborne lidar. *Geomorphology* **125**: 27–39. DOI: 10.1016/j.geomorph.2010.08.012

Jansson P, Hock R, Schneider T. 2003. The concept of glacier storage: a review. *Journal of Hydrology* **282**: 116–129. DOI: 10.1016/S0022-1694(03)00258-0

Jóhannesson T, Aðalgeirsdóttir G, Ahlstrøm A, Andreassen LM, Björnsson H, Woul M de, Elvehøy H, Flowers GE, Guðmundsson S, Hock R, Holmlund P, Pálsson F, Radic V, Sigurðsson O, Thorsteinsson T. 2006. The impact of climate change on glaciers and glacial runoff in the Nordic countries. Paper presented at the European Conference on Impacts of Climate Change on Renewable Energy Sources. Reykjavik, Iceland, June 5-9 2006, 31–34

Jomelli V, Grancher D, Naveau P, Cooley D, Brunstein D. 2007. Assessment study of lichenometric methods for dating surfaces. *Geomorphology* **86**: 131–143. DOI: 10.1016/j.geomorph.2006.08.010

Knighton D. 1998. *Fluvial Forms & Processes: a new perspective*. London: Arnold, 383 pp.

Kondolf GM. 1997. Hungry Water: Effects of Dams and Gravel Mining on River Channels. *Environmental Management* **21**: 533–551

Lane SN, Bakker M, Gabbud C, Micheletti N, Saugy J-N. 2017. Sediment export, transient landscape response and catchment-scale connectivity following rapid climate warming and Alpine glacier recession. *Geomorphology* **277**: 210–227. DOI: 10.1016/j.geomorph.2016.02.015

Maizels JK. 1979. Proglacial aggradation and changes in braided channel patterns during period of glacier advance : an alpine example. *Geografiska annaler* **61**: 87–101. DOI: 10.2307/520517

Maizels JK. 1983. Paleovelocity and paleodischarge determination for coarse gravel deposits. In: Gregory KJ (ed) *Background to Paleohydrology*. New York: John Wiley, 103–139

Maizels JK. 1986. Modeling of Paleohydrologic Change during Deglaciation. *Géographie physique et Quaternaire* **40**: 263. DOI: 10.7202/032648ar

Maizels JK. 1991. The origin and evolution of Holocene sandur deposits in areas of jökulhlaup drainage, Iceland. In: Maizels JK and Caseldine C (eds) *Environmental Change in Iceland: Past and Present.* Dordrecht, The Netherlands: Kluwer Academic Publishing, 267–302

Maizels JK. 1993. Lithofacies variations within sandur deposits: the role of runoff regime, flow dynamics and sediment supply characteristics. *Sedimentary geology* **85**: 299–325. DOI: 10.1016/0037-0738(93)90090-R

Maizels JK. 1997. Jokulhlaup deposits in proglacial areas. *Quaternary Science Reviews* **16**: 793–819. DOI: 10.1016/S0277-3791(97)00023-1

Marren PM. 2002. Glacier margin fluctuations, Skaftafellsjökull, Iceland: implications for sandur evolution. *Boreas* **31**: 75–81. DOI: 10.1111/j.1502-3885.2002.tb01057.x

Marren PM. 2005. Magnitude and frequency in proglacial rivers: a geomorphological and sedimentological perspective. *Earth-Science Reviews* **70**: 203–251. DOI: 10.1016/j.earscirev.2004.12.002

Marren PM, Toomath SC. 2013. Fluvial adjustments in response to glacier retreat: Skaftafellsjökull, Iceland. *Boreas* **42**: 57–70. DOI: 10.1111/j.1502-3885.2012.00275.x

Marren PM, Toomath SC. 2014. Channel pattern of proglacial rivers: topographic forcing due to glacier retreat. *Earth Surface Processes and Landforms* **39**: 943–951. DOI: 10.1002/esp.3545

Marzeion B, Jarosch AH, Hofer M. 2012. Past and future sea-level change from the surface mass balance of glaciers. *The Cryosphere* **6**: 1295–1322. DOI: 10.5194/tc-6-1295-2012

Mercier D. 2001. *Le ruissellement au Spitsberg. Le monde polaire face aux changements climatiques*. Clermont-Ferrand: Presses Universitaires Blaise Pascal, 278 pp.

Mercier D. 2008. Paraglacial and paraperiglacial landsystems: concepts, temporal scales and spatial distribution. *Géomorphologie : relief, processus, environnement* **14**: 223–233. DOI: 10.4000/geomorphologie.7396

Mercier D. 2010. La géomorphologie paraglaciaire : analyse de crises érosives d'origine climatique dans les environnements englacés et sur leurs marges (volume III). Clermont-Ferrand, Université Blaise Pascal, 262 pp.

Mercier D, Étienne S, Sellier D, André M-F. 2009. Paraglacial gullying of sediment-mantled slopes: a case study of Colletthøgda, Kongsfjorden area, West Spitsbergen (Svalbard). *Earth Surface Processes and Landforms* **34**: 1772–1789. DOI: 10.1002/esp.1862

Mercier D, Laffly D. 2005. Actual paraglacial progradation of the coastal zone in the Kongsfjorden area, West Spitsbergen (Svalbard). In: Harris C and Murton J (eds) *Cryospheric Systems: Glaciers and Permafrost*. Geological Society: London, 111–117

Milner AM, Khamis K, Battin TJ, Brittain JE, Barrand NE, Füreder L, Cauvy-Fraunié S, Gíslason GM, Jacobsen D, Hannah DM, Hodson AJ, Hood E, Lencioni V, Ólafsson JS, Robinson CT, Tranter M, Brown LE. 2017. Glacier shrinkage driving global changes in downstream systems. *Proceedings of the National Academy of Sciences* **114**: 9770–9778. DOI: 10.1073/pnas.1619807114

Naveau P, Jomelli V, Cooley D, Grancher D, Rabatel A. 2007. Modelling uncertainties in lichenometry studies with an application: The Tropical Andes (Charquini Glacier in Bolivia). *Arctic, Antarctic and Alpine Research* **39**: 277–288. DOI: 10.1657/1523-0430(2007)39[277:MUILS]2.0.CO;2

Nicholas AP, Sambrook-Smith GH. 1998. Relationships between flow hydraulics, sediment supply, bedload transport and channel stability in the proglacial Virkisa River, Iceland. *Geografiska annaler Series A: Physical geography* **80**: 111–122. DOI: 10.1111/j.0435-3676.1998.00030.x

Orwin JF, Smart CC. 2004. The evidence for paraglacial sedimentation and its temporal scale in the deglacierizing basin of Small River Glacier, Canada. *Geomorphology* **58**: 175–202. DOI: 10.1016/j.geomorph.2003.07.005

Owczarek P, Nawrot A, Migała K, Malik I, Korabiewski B. 2014. Flood-plain responses to contemporary climate change in small High-Arctic basins (Svalbard, Norway): Flood-plain responses to contemporary climate change, Svalbard. *Boreas* **43**: 384–402. DOI: 10.1111/bor.12061

Rabatel A, Jomelli V, Naveau P, Francou B, Grancher D. 2005. Dating of Little Ice Age glacier fluctuations in the tropical Andes: Charquini glaciers, Bolivia, 16°S. *Comptes Rendus Geoscience* **337**: 1311–1322. DOI: 10.1016/j.crte.2005.07.009

Radic V, Bliss A, Beedlow AC, Hock R, Miles E, Cogley JG. 2014. Regional and global projections of twenty-first century glacier mass changes in response to climate scenarios from global climate models. *Climate Dynamics* **42**: 37–58. DOI: 10.1007/s00382-013-1719-7

Roberts MJ, Russell AJ, Tweed FS, Knudsen Ó. 2000. Ice fracturing during jökulhlaups: implications for englacial floodwater routing and outlet development. *Earth Surface Processes and Landforms* **25**: 1429–1446. DOI: 10.1002/1096-9837(200012)25:13<1429::AID-ESP151>3.0.CO;2-K

Roberts MJ, Tweed FS, Russell AJ, Knudsen Ó, Harris TD. 2003. Hydrologic and geomorphic effects of temporary ice-dammed lake formation during jökulhlaups. *Earth Surface Processes and Landforms* **28**: 723–737. DOI: 10.1002/esp.476

Röthlisberger H, Lang H. 1987. Glacial hydrology. In: Gurnell AM and Clark MJ (eds) *Glacio-fluvial Sediment Transfer*. New York: Wiley, 207–284

Roussel E, Chenet M, Grancher D, Jomelli V. 2008. Processus et rythmes de l'incision des sandar proximaux postérieure au petit âge glaciaire (sud de l'Islande). *Géomorphologie : relief, processus, environnement* 249–260. DOI: 10.4000/geomorphologie.7416

Roussel E, Toumazet J-P, Marren PM, Cossart E. 2016. Iceberg jam floods in Icelandic proglacial rivers: testing the self-organized criticality hypothesis. *Géomorphologie : relief, processus, environnement* 37–49. DOI: 10.4000/geomorphologie.11229

Russell AJ, Tweed FS, Roberts MJ, Harris TD, Gudmundsson MT, Knudsen Ó, Marren PM. 2010. An unusual jökulhlaup resulting from subglacial volcanism, Sólheimajökull, Iceland. *Quaternary Science Reviews* **29**: 1363–1381. DOI: 10.1016/j.quascirev.2010.02.023

Russell AJ, Tweed FS, Roberts MJ, Knudsen Ó, Harris TD, Gudmundsson MT, Marren PM, Rice JW. 2003. The causes, characteristics and impacts of a volcanically-induced jökulhlaup, Sólheimajökull, Iceland. *Geophysical Research Abstracts* **5**

Schomacker A. 2010. Expansion of ice-marginal lakes at the Vatnajökull ice cap, Iceland, from 1999 to 2009. *Geomorphology* **119**: 232–236. DOI: 10.1016/j.geomorph.2010.03.022

Sigurðsson O. 1998. Glacier Variations in Iceland 1930-1995. From the database of the Iceland Glaciological Society. *Jökull* **N°45**: 3–25

Sigurðsson O. 2005. Variations of termini of glaciers in Iceland in recent centuries and their connection with climate. In: Caseldine C, Russell A, Hardardottir J and Knudsen O (eds) *Iceland - Modern processes and past environments*. Amsterdam: Elsevier, 241–255

Sigurðsson O, Jonsson T, Johannesson T. 2007. Relation between glacier-termini variations and summer temperature in Iceland since 1930. *Annals of Glaciology* **46**: 170–176. DOI: 10.3189/172756407782871611

Slomka JM, Eyles CH. 2015. Architectural–landsystem analysis of a modern glacial landscape, Sólheimajökull, southern Iceland. *Geomorphology* **230**: 75–97. DOI: 10.1016/j.geomorph.2014.11.006

Staines KEH, Carrivick JL, Tweed FS, Evans AJ, Russell AJ, Jóhannesson T, Roberts M. 2015. A multi-dimensional analysis of pro-glacial landscape change at Sólheimajökull, southern Iceland. *Earth Surface Processes and Landforms* **40**: 809–822. DOI: 10.1002/esp.3662

Thompson A. 1988. Historical development of the Proglacial Landforms of Svínafellsjökull and Skaftafellsjökull, Southeast Iceland. *Jökull* **38**: 17–30

Thompson A, Jones A. 1986. Rates and causes of proglacial river terrace formation in southeast Iceland: an application of lichenometric dating techniques. *Boreas* **15**: 231–246. DOI: 10.1111/j.1502-3885.1986.tb00928.x

Thórarinsson S. 1956. On the variations of Svínafellsjökull, Skaftafellsjökull and Kviárjökull in Öraefi. *Jökull* **6**: 1–15

Thórarinsson S. 1969. Glacier surges in Iceland, with special reference to the surge of Brúarjökull. *Canadian Journal of Earth Science* **6**: 875–882. DOI: 10.1139/e69-089

Thordarson T, Larsen G. 2007. Volcanism in Iceland in historical time: Volcano types, eruption styles and eruptive history. *Journal of Geodynamics* **43**: 118–152. DOI: 10.1016/j.jog.2006.09.005

Tweed FS. 2000. An ice-dammed lake in Jökulsárgil: Predictive modelling and geomorphological evidence. *Jökull* **48**

Vaughan DG, Comiso JC, Allison I, Carrasco J, Kaser G, Kwok R, Mote P, Murray T, Paul F, Ren J, Rignot E, Solomina O, Steffen K, Zhang T. 2013. Observations: Cryosphere. In: Stocker TF, Qin D, Plattner GK, Tignor M, Allen SK, Boschung J, Nauels A, Xia Y, Bex V, Midgley PM (eds) *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge, United Kingdom and New York, NY, USA: Cambridge University Press, 317-382.

Vericat D, Batalla RJ, Garcia C. 2006. Breakup and reestablishment of the armour layer in a large gravel-bed river below dams: The lower Ebro. *Geomorphology* **76**: 122–136. DOI: 10.1016/j.geomorph.2005.10.005

Westoby MJ, Glasser NF, Brasington J, Hambrey MJ, Quincey DJ, Reynolds JM. 2014. Modelling outburst floods from moraine-dammed glacial lakes. *Earth-Science Reviews* **134**: 137–159. DOI: 10.1016/j.earscirev.2014.03.009

Wilkie K, Clague J. 2009. Fluvial response to Holocene glacier fluctuations in the Nostetuko River valley, southern Coast Mountains, British Columbia. In: Knight J and Harrisson S (eds) *Periglacial and Paraglacial processes and environments*. London: The Geological Society, 199–218

Worni R, Huggel C, Clague JJ, Schaub Y, Stoffel M. 2014. Coupling glacial lake impact, dam breach, and flood processes: A modeling perspective. *Geomorphology* **224**: 161–176. DOI: 10.1016/j.geomorph.2014.06.031

Figures and Tables

Figure 1: Location of investigated glaciers and associated proglacial rivers

Figure 2: Methodological workflow: A - Field prospection and GIS mapping of proglacial and glaciofluvial landforms ; B - Lichenometric dating of outwash terraces ; C - Topographic survey: Cross section and long profile of active outwash and terraces ; D - Method for the determination of the Inflection Point migration

Figure 3: Outwash terraces mapping and dating

Figure 4: Overview of aggrading and incising stages in the post-LIA outwash adjustment

Figure 5: Long profiles of dated outwash terraces (T1 to T4) and active outwash in 2007 (A)

Figure 6: A - Linear regression between average rates of altitudinal lowering and longitudinal migration of the inflection points. B - Linear regression between average rates of IP downstream migration and the proportion of proglacial lakes in the marginal area.

Figure 7: A - The dominant sequence of post-LIA fluvial adjustment: a two-stages descriptive model. B - LIA coalescent glaciers specific case, with stream capture and shift in the outwash location.

Table 1: Main characteristics of glaciers and associated proglacial rivers investigated in this study.

	Glacier								River	
ID	Name	LIA max dating (cal. yr. AD)*	Confidence interval*	Glacier Area (Km²)	Glacier length (km)	Glacier Slope (°)	Elevation max (m a.s.l.)	ELA (m a.s.l.)**	name	Outwash Type***
1	Sólheimajökull	1851	(1839 - 1863)	46.7	15.4	5.1	1480	na	Jökulsá	Type III
2	Morsárjökull	1888	(1873 - 1903)	26.3	13.1	6.2	1421	1000-1130	Morsá	Type I
3	Skaftafelljökull	1878	(1864 - 1892)	87.0	19.0	6.6	1889	1000-1160	Skaftafellsá	Type I
4	Svínafellsjökull	1765	(1749 - 1776)	31.7	12.7	12.4	2104	1000-1120	Svínafellsá	Type I
5	Virkisjökull	1740	(1731 - 1755)	20.6	11.3	10.7	2076	na	Virkisá	Type I
6	Kotárjökull	1819	(1809 - 1828)	6.1	6.6	12.8	1840	1000-1130	Kotá	Type III
7	Hólárjökull	1844	(1838 - 1857)	5.2	6.9	12.9	1838	na	Holá	Type I
8	Kvíárjökull	1810	(1798 - 1824)	18.6	11.7	10.3	2010	1010-1130	Kviá	Type I
9	Hrútárjökull	1812	(1805 - 1820)	11.9	8.5	14.6	2027	880-910	Fjallsá	Type I
10	Fjallsjökull	1812	(1798 - 1833)	48.8	13.4	9.1	2040	870-960		
11	Skálafellsjökull	1865	(1856 - 1876)	136.2	32.4	3.3	1520	910-1020	Kolgrimá	Type I
12	Heinabergsjökull	1851	(1835 - 1865)	93.8	23.5	4.8	1520	990-1100		
13	Fláajökull	1821	(1811 - 1834)	213.1	29.1	3.3	1520	1060-1120	Hólmsá	Type I
14	Hoffellsjökull	1888	(1874 - 1898)	246.0	32.0	3.3	1512	1050-1120	Hoffellsá	Type I

* after Chenet *et al.* , 2010 and Roussel *et al.* , 2008 ; ** after Chenet *et al.* , 2010 and Hannesdóttir *et al.* , 2014 ; *** after Maizels, 1993, 1997 ; "na" for not available

Table 2: Post-LIA glacier retreat and landform assemblages of proglacial margins

ID	Glacier	Deglaciated surface since the LIA (km²)	Computed average linear retreat (m)	Measured linear retreat (m)	Alluvial fan		Rocky outcrop		Fresh proximal till deposits		Intra-morainic lake		Proglacial lake		Moraine		Active intra-morainic plain		Abandoned intra-morainic plain		Embankment		TOTAL
					m²	%	m²	%	m²	%	m²	%	m²	%	m²	%	m²	%	m²	%	m²	%	m²
1	Sólheimajökull	2.1	888	1211	0	0.0	256366	12.4	299121	14.4	13921	0.7	0	0.0	467737	22.6	569831	27.5	466563	22.5	0	0.00	2073538
2	Morsárjökull	2.6	2047	1643	0	0.0	0	0.0	202767	8.5	4981	0.2	334873	14.0	870060	36.4	114907	4.8	862098	36.1	0	0.00	2389686
3	Skaftafelljökull	7.1	2420	2044	0	0.0	0	0.0	237577	3.4	154871	2.2	233199	3.3	3528684	49.8	819307	11.6	2110658	29.8	2521	0.04	7086818
4	Svínafellsjökull	1.6	566	877	0	0.0	0	0.0	228231	14.1	18155	1.1	75034	4.6	1211262	74.7	77301	4.8	10581	0.7	0	0.00	1620564
5	Virkisjökull	1.2	827	1157	0	0.0	0	0.0	140826	11.9	3088	0.3	0	0.0	336390	28.4	98437	8.3	603721	51.1	0	0.00	1182463
6	Kotárjökull	0.5	404	587	0	0.0	99918	20.6	0	0.0	0	0.0	0	0.0	321017	66.1	25122	5.2	39893	8.2	0	0.00	485950
7	Hólárjökull	0.5	774	1258	70171	13.7	0	0.0	19702	3.8	0	0.0	0	0.0	176764	34.4	80554	15.7	166569	32.4	0	0.00	513760
8	Kvíárjökull	5.3	1199	1631	0	0.0	0	0.0	1009960	19.0	25342	0.5	172200	3.2	3135337	59.1	342038	6.4	624307	11.8	0	0.00	5309185
9	Hrútárjökull	2.3	1665	1950	319076	14.1	0	0.0	172327	7.6	10191	0.5	0	0.0	1059046	46.9	544421	24.1	152509	6.8	0	0.00	2257570
10	Fjallsjökull	7.3	1071	1601	0	0.0	0	0.0	203291	2.8	161030	2.2	2113291	29.0	3873712	53.1	434737	6.0	507985	7.0	0	0.00	7294046
11	Skálafellsjökull	9.5	3544	2121	0	0.0	5726805	60.3	220745	2.3	101078	1.1	112348	1.2	1709461	18.0	923917	9.7	708383	7.5	0	0.00	9502737
12	Heinabergsjökull	8.7	1810	3171	0	0.0	67878	1.3	186334	3.6	70993	1.4	1689050	32.6	2687400	51.9	30710	0.6	441328	8.5	0	0.00	5173693
13	Fláajökull	15.1	2562	2429	0	0.0	785263	5.2	236760	1.6	374913	2.5	1246571	8.2	7609144	50.3	1359487	9.0	3500009	23.1	13708	0.09	15125855
14	Hoffellsjökull	0.7	160	335	0	0.0	0	0.0	0	0.0	0	0.0	500736	72.8	186911	27.2	0	0.0	0	0.0	0	0.00	687647
				TOTAL	389247	0.6	6936230	11.4	3157642	5.2	938563	1.5	6477302	10.7	27172926	44.8	5420769	8.9	10194605	16.8	16229	0.03	60703512

Table 3: Outwash responses to post-LIA glacial retreat

ID	Glacier name	River name	Outwash Type*	Duration of aggradation stage (*yr*)	Post-LIA proximal incision (m.a-1)	Post-LIA decrease of proximal slope ($m.m^{-1}$)	Average *IP* longitudinal migration ($m.a^{-1}$)	Average *IP* altitudinal lowering ($m.a^{-1}$)	Average *IP* downstream migration ($m.a^{-1}$)
1	Sólheimajökull	Jökulsá	Type III	41 (±16)	0.047 (±0.002)	0,00057	*na*	*na*	*na*
2	Morsárjökull	Morsá	Type I	51 (±18)	0.045 (±0.006)	0,00584	28,7	0,44	28,68
3	Skaftafelljökull	Skaftafellsá	Type I	*na*	*na*	0,0206	26,5	0,32	26,47
4	Svínafellsjökull	Svínafellsá	Type I	*na*	*na*	0,0044	8,9	0,13	8,91
5	Virkisjökull	Virkisá	Type I	147 (±15)	*na*	0,0157	5,2	0,19	5,21
6	Kotárjökull	Kotá	Type III	*na*	*na*	0,0177	*na*	*na*	*na*
7	Hólárjökull	Holá	Type I	*na*	0	0	*na*	*na*	*na*
8	Kvíárjökull	Kviá	Type I	60 (±21)	0.076 (±0.004)	0,0122	6,6	0,20	6,57
9	Hrútárjökull	Fjallsá	Type I	79 (±25)	0.068 (±0.003)	*na*	14,0	0,09	13,97
10	Fjallsjökull								
11	Skálafellsjökull	Kolgrimá	Type I	5 (±14)	*na*	*na*	30,3	0,47	30,34
12	Heinabergsjökull								
13	Fláajökull	Hólmsá	Type I	*na*	*na*	0,0108	16,9	0,25	16,86
14	Hoffellsjökull	Hoffellsá	Type I	50 (± 20)	0.131 (±0.012)	0,0056	46,4	0,5	46,38

* After Maizels, 1993, 1997, "*na*" for not applicable