Geologos, 2010, 16 (4): 201–222
doi: 10.2478/v10118-010-0005-y

Glacial Terminations II and I as recorded in NE Iceland

Brigitte Van Vliet-Lanoë[1], Agust Guðmundsson[2], Hervé Guillou[3], Soléne Guégan[1], A.J. (Tom) van Loon[4], François De Vleeschouwer[5]

[1]UMR 6538 CNRS Domaines Océaniques, IUEM, Place N. Copernic, 29480 Plouzané, France; e-mail: brigitte.vanvlietlanoe@univ-Brest.fr
[2]JFS Geological Services Raudagerdi 31, 108 Reykjavik, Iceland; e-mail: agust@geoice.net
[3]Laboratoire des Sciences du Climat et de l'Environnement, Domaine du CNRS, Bât. 12, Avenue de la Terrasse, 91198 Giff sur Yvette, France; e-mail: Herve.Guillou@lsce.ipsl.fr
[4]Geological Institute, Adam Mickiewicz University, Maków Polnych 16, 61-606 Poznań, Poland; e-mail: tvanloon@amu.edu.pl
[5]Department of Ecology and Environmental Sciences, Umeå University, Sweden; e-mail: fdevleeschouwer@gmail.com

Abstract

Volcanism in eastern Iceland has controlled the changes in glacier- and river-drainage patterns and the sedimentary budget, particularly during the Middle and Late Pleistocene. The glacial extent in NE Iceland appears to be related to the impact of volcanic activity, not only on the ice-stream dynamics, but also on the sedimentary successions. Analysis of the Jökuldalur and Jökulsa á Brù records results in a new interpretation of the changes in ice extent and flow direction for at least the last two glaciations. From MIS 8 onward, the development of the Snæfell volcano apparently forced the ice stream that derived from the Vatnajökull ice cap to take another course; it also affected the offshore sedimentary budgets at the new outlet at Vopnafjördur. The MIS 6 ice sheet was thick and extensive, and associated with an ice-stream diversion to the North. The thick sedimentary complex of palaeolake Halslón was formed close to an outlet of the Vatnajökull, the Brùarjökull, during Termination II and a part of the MIS 5e interglacial.
The deposits formed during MIS 5e record two climate optima interrupted by two successive glacial advances correlated with the mid-Eemian cooling. The deposits of the Weichselian deglaciation (Termination I) are much more limited in thickness. During the Last Glacial Maximum and the Late Glacial, glaciers also seem to have been restricted in the Jökulsa á Brù area. Valley glaciers issued from the Brùarjökull re-advanced several times in the Jökuldalur only during at least the Older Dryas, the Younger Dryas and the Preboreal. NE Iceland has undergone considerable deglaciation since the Bølling. In contrast to the conclusions of previous studies, the results presented here are consistent with data on the glaciations in other Nordic regions and can increase the understanding of the mid-Eemian cooling.

Keywords: ice-stream patterns, palaeolake, deglaciation, Last Glacial Maximum, mid-Eemian cooling, Iceland

Introduction

Iceland is an important site for obtaining a better insight into climate change in the middle of the Northern Atlantic. The main ice sheet, the Vatnajökull, has shown a cyclic evolution since at least the Pliocene (Eiríksson & Geirsdóttir, 1991). The ice sheet grew rapidly at the end of interglacials and during shorter events like the mid-Eemian cooling (Van Vliet-Lanoë et al., 2007) or the Neoglacial (Guðmundsson, 1997); glaciers totally vanished during the cli-

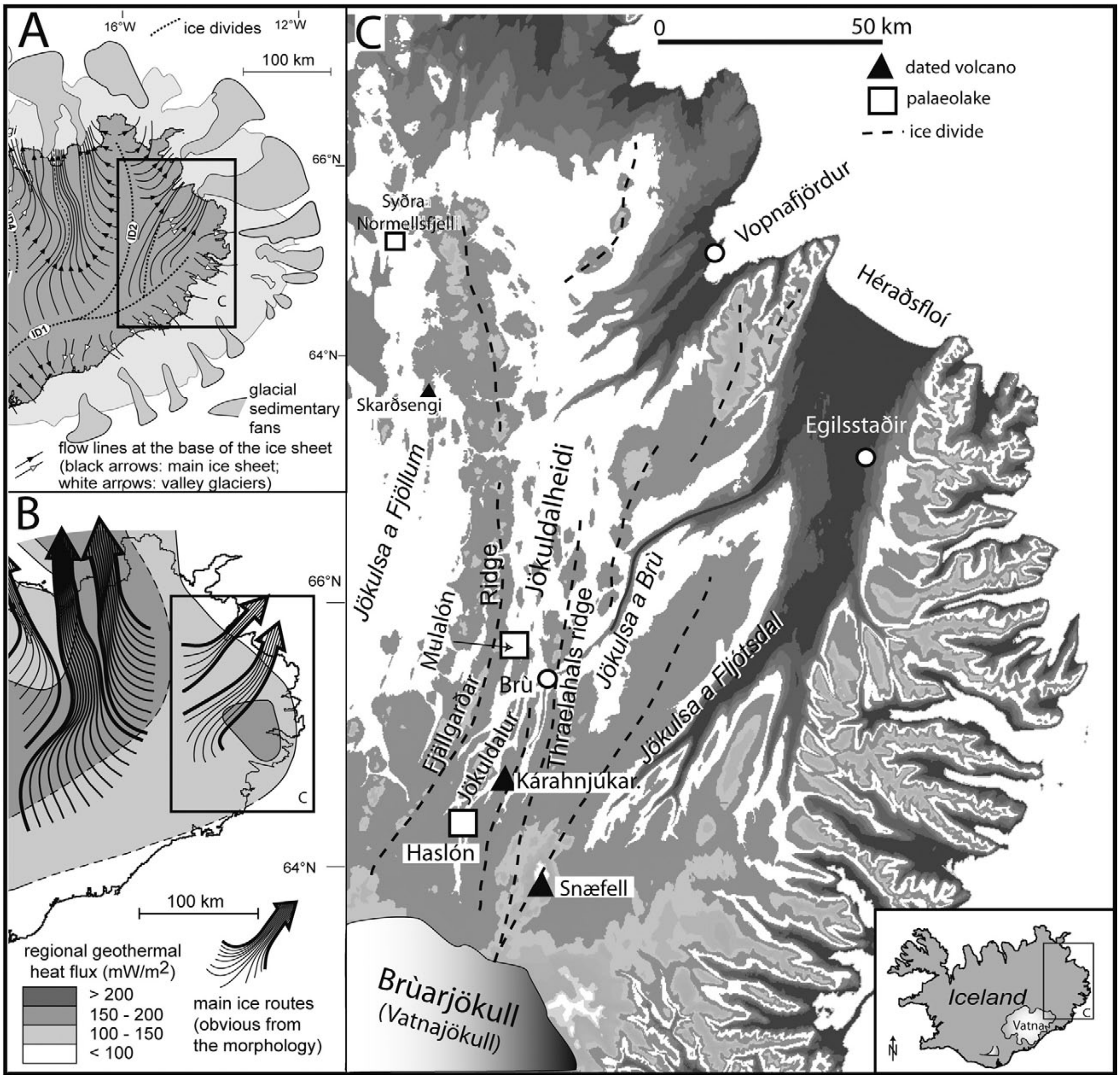

Fig. 1. Location of the study area and ice-stream characteristics.
A – Ice-stream flow lines and deep-sea fans. Compiled by Bourgeois et al. (2000) after Vogt et al. (1980); **B –** Geothermal gradients (from Anorsson, 1995) and glacier flow lines (from Bourgeois et al., 2000); **C –** Global morphology of NE Iceland. The white area indicates the lowlands, and the lighter grey indicates uplands and ridges. Dashed lines indicate the Weichselian ice divides.

matic thermal optimum (Ingólfsson & Norðdahl, 1994; Van Vliet-Lanoë et al., 2007).

For Iceland, several authors have postulated the existence of a glacial extent that reached the outer border of the island platform or shelf (Fig. 1A) during the Weichselian (e.g., Thoroddsen, 1911; Norðdahl & Halfidason, 1992; Ingólfsson et al., 1997). A later deglaciation, with the ice retreating from the present-day coast line during the Younger Dryas or the Preboreal (Norðdahl & Halfidason, 1992; Ingólfsson et al., 1997), is proposed by most researchers. A tephra deposit (Skógar) found at coastal sites, which was correlated with the Vedde Ash, is commonly used to date this limited deglaciation (Norðdahl & Halfidason, 1992). Móberg (subglacial table volcanoes) data are also used to reconstruct the thickness of the ice sheet and to map the boundary of the ice during the Last Glacial Maximum (LGM) (Walker, 1965). Thoroddsen

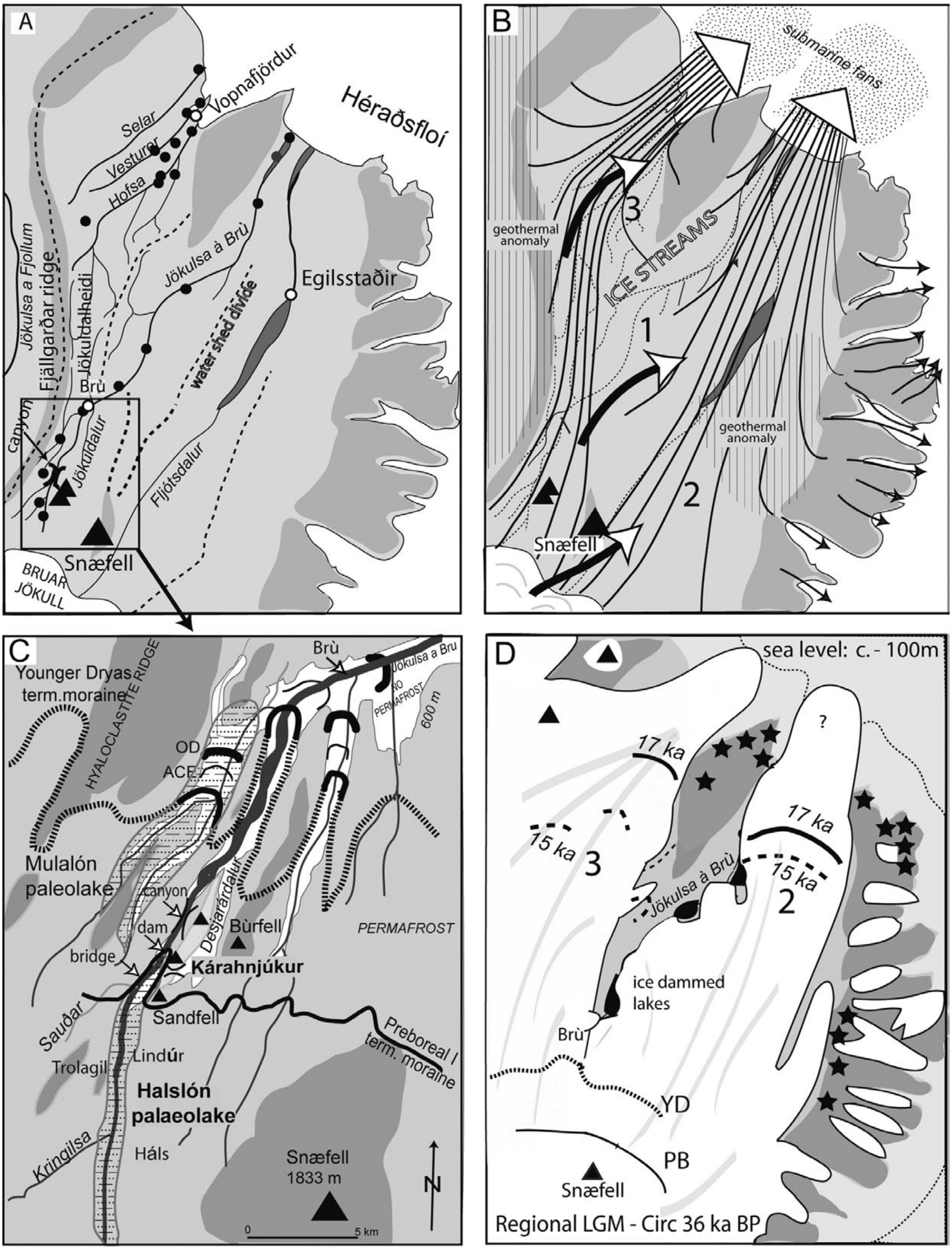

Fig. 2. Schematic evolution of the ice-sheet drainage patterns during MIS 6 and the Weichselian. The position of the terminal moraines is based on field data. Glacial advances: OD = Older Dryas; ACE = Allerød Cold Event; YD = Younger Dryas-Bùrfell; PB = Preboreal.

A – Positions of the present-day valleys showing the valley system north of Brù. Black dots indicate sites of sedimentological observations; **B –** Positions of the main ice streams (1, 2 and 3); **C –** Position of the Late Glacial terminal moraines within the Jökuldalur valley; **D –** Ice extent during the regional Late Glacial Maximum.

(1911) has suggested that some of the higher mountains were ice-free during the LGM.

Guðmundsson et al. (1999) discovered huge palaeolacustrine deposits (Figs. 1C and 2C), attributed to the Halslón palaeolake (64°57′14″N; altitude 550 m) and close to the Brùarjökull, assigning an age of 20–150 ka to the volcanic damming of the lake by the formation of a ridge after a subglacial eruption, known as the Kárahnjúkur (Fig. 2C). The lacustrine deposits were studied later by Knudsen & Marren (2002). The subsequent retreat of the Brùarjökull glacier to behind the volcanic dam led to the formation of a proglacial lake during the last deglaciation. The failure of the volcanic dam during the Late Glacial to Holocene, which caused the drainage of Halslón Lake, led to an incision into the underlying bedrock of approx. 200 m deep (Knudsen & Marren, 2002).

Subsequent dating of basalts (Helgason & Duncan, 2003) has revealed a more complex history. The original incision, which was approx. 120 m deep, has been suggested to have been formed earlier, i.e., during the last 200 ka, or since the birth of the Kárahnjúkur subglacial volcano (Guðmundsson & Helgason, 2004). Our observations suggest that apparently even older deposits exist. Thus, the sedimentary record of the region required a more detailed investigation, and the excavations for the Kárahnjúkar hydropower station (exposed until fall 2006) offered an opportunity to study it. The glaciolacustrine deposits trapped by the damming of the Jökuldalur (Fig. 2C) provided high-resolution interglacial successions in the vicinity of the Vatnajökull ice cap and its local outlet glacier (the Brùarjökull). Therefore, it allowed us to date the events recorded upstream and to correlate them with downstream sections in the Jökulsa á Brù valley.

An initial survey (2001) based on the mapping of glaciolacustrine and glaciofluvial terraces and of terminal moraines suggested that only a limited Weichselian ice cover existed in the Jökulsa á Brù. In order to achieve a better insight into the possible differences of flow patterns between ice streams and river networks in the region, we have documented in detail the characteristics of the preserved sediments in the entire Jökuldalur valley. This investigation included an analysis of the geomorphology, $^{40}Ar/^{39}Ar$ and K/Ar datings of the basalts that may have been responsible for the damming-off of the Halslón palaeolake, tephrostratigraphy and analysis of the deposits in the upper Jökuldalur, the valley downstream of Halslón Lake (Van Vliet-Lanoë et al., 2007).

The complexity of the sedimentary successions suggests a long history. The present contribution presents a new interpretation of the ice dynamics in NE Iceland for the last two glaciations and of the impact of volcanic activity on ice-stream dynamics. Dating the moment of maximum ice extent in northern Iceland during glaciations and establishing the relationship with rift activity is of great importance, as it would increase our knowledge of changes in oceanic circulation and climate in the northern Atlantic region.

Methods

The ice-sheet extents, as well as the active and fossil rock glaciers, were mapped from 1985 to 1999 by Guðmundsson (2000). We carried out additional mapping in 2001 and 2006 to locate and characterise the terminal moraines. Some recent terminal moraines crop out closer to the Brùarjökull, whereas another one (that was formed during the Little Ice Age) crops out at only 4 km from the present-day ice front. In 2005, we also investigated sections of the Jökuldalur valley and on the Jökuldalheidi plateau (Fig. 1C). Outcrops along the river and fresh road cuts were studied, and analyses of satellite images (Landsat cover and SPOT 5) and of topographic maps also contributed to the investigation.

The chronology in the present study is based on $^{40}Ar/^{39}Ar$ dating (Helgason & Duncan, 2003) and Ar-K dating (Guillou et al., 2009). Fresh rocks or rocks with minimal traces of alteration were selected for the K/Ar dating, following the methodology presented by Guillou et al. (1998), Charbit et al. (1998) and Scaillet & Guillou (2004).

Tephra were analysed by SEM with an EDX microprobe analyser on single, clean grains at the PBDS laboratory in Lille, France. The results of the tephra analyses, including the Skógar results, were reported previously by Van Vliet-Lanoë et al. (2007). Laser-ICPMS analyses of trace elements were performed at Brest (Guégan, 2010).

Geological setting

Regional geochronology

North-eastern Iceland is characterised by Neogene basaltic volcanism that is considered to be related to a N-S-trending Neogene rift that has been affected by transtension since at least the Mio-Pliocene (Garcia et al., 2003). The basalts form the basement of the lower Jökulsa á Brù river valley, with the oldest rocks located near its outlet into the Héraðsfloí stream and the youngest rock at its origin at Vatnajökull. The middle Jökulsa á Brù river flows on a cover of Late Pliocene to Early Pleistocene basalts deformed by strike-slip faults (approx. 1.6 Ma: Usagawa et al., 1999; 2.2 ± 0.5 Ma: Garcia et al., 2003). In the upper valley, the Jökuldalur Middle to Late Pleistocene basalts crop out as lava flows and hyaloclastite ridges. Most of these ridges show activity since approx. 1 Ma (Helgason & Duncan, 2003), but they were mostly built during the Middle Pleistocene Transition between 800 and 700 ka, forming a major discontinuity in the stratigraphy on an old, glacially abraded surface. On these ridges, subglacial volcanoes (móbergs) occur locally.

The study area is bordered on the West by the high (850–1035 m) hyaloclastite ridge which is known as the Fjällgarðar ridge (Figs. 2A and 2C). This ridge constitutes an ice divide and extends outside the active rift to the East under a thick ice cover (e.g., Bourgeois et al., 1998). The rocks underlying the base of the recent Kárahnjúkur ridge date from 740 ± 27 ka to 685 ± 22 ka (Guillou et al., 2009). This ridge is nearly contemporaneous with the Bùrfell and Thraelahals ridges, which probably lie unconformably on the top of the Plio-Pleistocene plateau. The last regional lavas that were spread at the surface of the Jökuldalheidi plateau by the Brù-Háls event, have been dated at 560–480 ka (Usagawa et al., 1999; Helgason & Duncan, 2003). Riftward displacement of the Fjällgarðar ridge took place afterwards (0.4 ± 0.1 Ma; Skarðsengi ridge, Fig. 1C: Garcia et al., 2003). A major incision took place after the Brù-Háls event; the resulting palaeovalley became filled with Kárahnjúkur subglacial volcanic deposits. This glacial valley was approx. 3.5 km wide, with a valley floor reaching a depth of 460 m at the hydroelectric-dam site (Guðmundsson & Helgason, 2004).

The Snæfell volcano was formed more recently and is considered to have been active until the last deglaciation. Our dating places this volcano, a móberg, at the MIS 10 to MIS 8 transition (324 ± 12 ka until 265 ± 6 ka), ending with MIS 8 (SN02: 253 ± 6 ka) (Guillou et al., 2009).

The Kárahnjúkur prolongates the Snæfell activity in time with dates ranging from 279 ± 152 (Helgason & Duncan, 2003) to 241 ± 12 ka (Guillou et al., 2009). The southern Kárahnjúkur summit formed between 160–200 ka (Helgason & Duncan, 2003). The southernmost móberg, the Sandfell, formed 150 ± 9 ka ago (Guillou et al., 2009). Thus, the Kárahnjúkur ridge activity persisted later under the large ice sheets of MIS 8 and MIS 6. The valley was potentially dammed by the Kárahnjúkur hyaloclastites between approx. 250 and approx. 150 ka and was partly filled downstream of the present-day dam site. Late dyke volcanism on the same ridge occurred approx. 103 ka ago at the Sandfell volcano (Helgason & Duncan, 2003).

Morphology and ice-flow lines

Today, most of the valleys present a typical U-shaped glacial morphology. The large Jökuldalheidi plateau, with parallel shallow valleys draining to the North (Figs. 1C and 2A), constitutes the logical area where the ice stream must have flowed. Even the topography, with valleys heading towards the NE, appears to be very old for both valley systems (Fljótsdalur and Jökulsa á Brù valleys). The geothermal gradient of this zone is medium high (Anorsson, 1995). Basal melting of the ice masses due to the geothermal heat allowed the development of temperate-base ice streams (Bourgeois et al., 1998).

The present-day morphological evolution of the region began during the Middle Pleistocene Transition, about 900 ka ago. Above the Brù farms, the upper valley (the Jökuldalur) is disconnected from the Jökulsa á Brù valley by an old ridge of hyaloclastite, the Bùrfell ridge, which runs from the Bùrfell to immediately north of Brù. The Jökulsa á Brù valley downstream of Brù is V-shaped and narrow, with limited glacial abrasion. The Jökuldalur, which is directly fed by the Brùarjökull and which was occupied by the Halslón palaeolake, also represents the upstream prolongation of the Jökuldalheidi plateau (Fig. 1B and 1C). The Holocene drainage on the Jökuldalheidi plateau forms an intricate, but hardly organised, network (Fig. 2A) with traces of several Late Glacial lakes.

During the Early Pleistocene and a part of the Middle Pleistocene, most of the ice from the Brùarjökull apparently flowed to Egilsstaðir and Héraðsfloí (Fig. 2B, routes 1 and 2). At this potential ice-stream outlet, the glacial U-shaped valley and drumlin morphologies are very well developed, but middle to late Quaternary deposits are limited in the lower Jökulsa á Brù. At the surface of the Jökuldalheidi plateau, two sets of glacial striae were discovered by Thorarinsson et al. (1977) and further compiled by Norðdahl & Hjort (1995). The first one shows an eastward ice-flow direction related to the maximum ice extent (Fig. 2B, route 2), and the second set is parallel to the Fjällgarðar ridge system (Fig. 2B, route 3). Because the Brù-Háls lava flow, dated as 560–480 ka, is deeply incised by recent glacial activity, the diversion seems to have occurred more recently than MIS 13. Since the formation of the Snæfell volcano in the late MIS 10, which runs parallel to the Fjällgarðar ridge system (Fig. 2B), recent ice streams were apparently confined along this ridge. They flowed towards the North through the outlet of the Vopnafjördur, which is consistent with the compiled striae directions and drumlin fields, with the Brùarjökull being partly routed to the West (Fig. 2B, route 3).

The sedimentary record and its morphological setting

Palaeolake Halslón

Palaeolake Halslón was located about 5–13 km downstream of the Brùarjökull glacier in a wide incised valley system (Guðmundsson et al., 1999; Guðmundsson & Helgason, 2004) and filled this valley temporarily. The succession of palaeolake Halslón is preserved in the form of individual palaeo-delta deposits now forming terraces. The maximum altitude of the lake sediments is officially 560 m (Knudsen & Marren, 2002), while their base is at 460 m. Actually, however, the lake terraces with sparse thin deposits or beaches reach 570 m. The river now flows at an altitude of 450 m at the bridge site.

The sedimentary succession has been described by Knudsen & Marren (2002) as consisting of two major lacustrine units that show some glaciotectonic deformation, and that are separated from one another by moderate glacial erosion. A composite log (Fig. 3) based on sections from the main relict fans (issued from the Sauðar valley) indicates a much more complex stratigraphy than initially described; it will be re-interpreted here on the basis of our field observations.

The substratum of the sedimentary succession is faulted and consists of basalts and hyaloclastites with an age of 1.25 to 0.95 Ma and lithified sediments (unit A). They are all time-equivalent to one or more eruptions of the old subglacial volcanoes, and they are intercalated between tills. Laterally, unit A has been partially eroded by glacial activity that resulted in the formation of a drumlin, while it is buried (to the North) by the volcanites of two recent móbergs: the Kárahnjúkur (approx. 250 ka) and the Sandfell (approx. 150 ka). The latter móberg is situated about one kilometre south of the former (Fig. 2C). These Middle Pleistocene volcanites form unit B and are truncated and covered unconformably by the first lake unit described by Knudsen & Marren (2002). These have not been faulted tectonically.

The sedimentary complex of Halslón consists of two successions (a lower and an up-

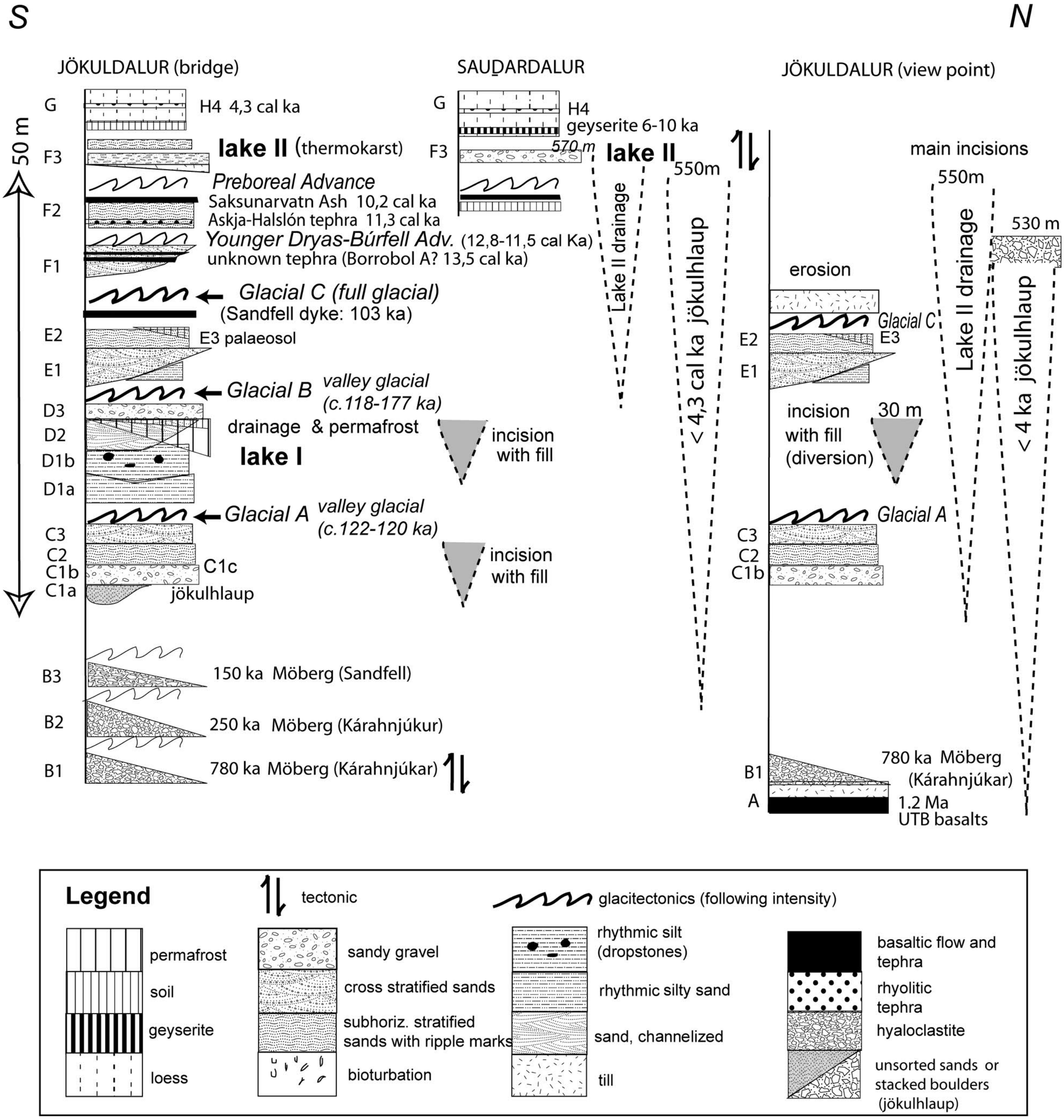

Fig. 3. Stratigraphical record at three sites (bridge, lower Sauðar valley, outlook point below the dam). V-shaped symbols show the intensity of valley incision; in grey, sediment-filled incisions.

per 'lacustrine' succession) of glaciofluvial to glaciolacustrine deposits (Fig. 3 and 4).

Description of the lower succession (palaeolake I)

The lower succession comprises three units (C, D, E). It is usually consolidated and is rich in hyaloclastite debris. The three distinct units are separated by levels with evidence of glaciotectonics. The section described underneath is located at the drowned bridge, at the southern face of the main accumulation, downstream of the Sauðar valley.

Unit C — This unit begins with an erosional sandy (clast-supported) conglomerate (subunit C1b) of 2–3 m thick; the scourings in potholes (1 m deep) at its base are filled with pure basaltic coarse sands (subunit C1a). Subunit C1c, which is exposed only locally, consists of several metres of cross-stratified sands, incised by gullies

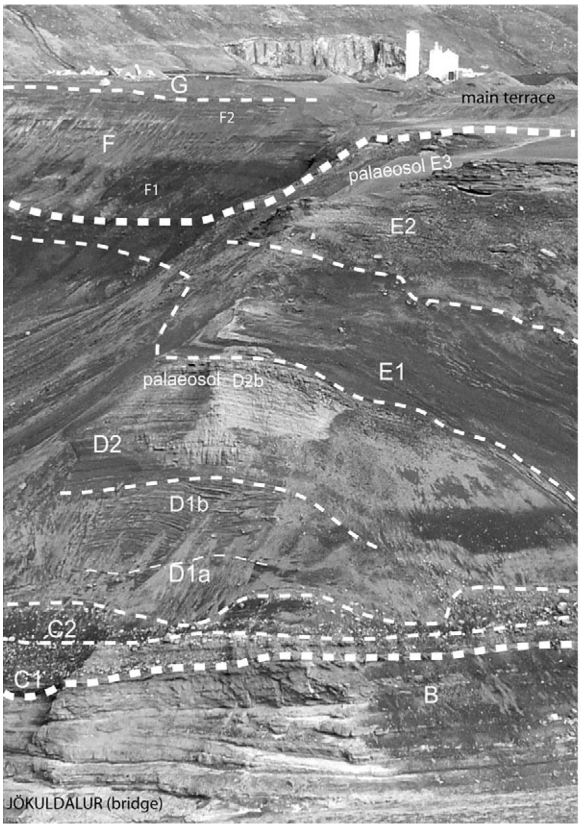

Fig. 4. Stratigraphy of the main terrace in the Jökuldalur. The picture shows the bridge section (details in Fig. 7).

unconformably onlapped by laminated sandy silts that fine towards the top. Downstream, subunit C1c is present close to the foot of the Kárahnjúkur and probably forms a lacustrine delta (Fig. 5C); this delta has been deformed by glaciotectonism. Subunit C1c is partially eroded and overlain by subunit C2.

Subunit C2 is a deformed, stratified, fine-grained diamicton with reworked cobbles that has later become injected by sedimentary dykes enriched in cobbles and gravels (Fig. 6C).

Unit D – This unit unconformably overlies unit C. It is mostly bent by creep and strongly fractured by transtension for the upper 15 m. It consists of three subunits (D1-D3).

Subunit D1 is a varved succession (up to 40 m thick) with dropstones (Figs. 3 and 7B). In its lower part (D1a), it consists of laminated fine silty sand, but it becomes silty (D1b) towards the top (approx. 58 layers). The couplets range from 20 to 60 cm, with three thicker layers (coarse part of the couplets) with large dropstones (15 cm). Some shallow channels are present at the transition between the two subunits.

Subunit D2 is coarsely bedded, silty medium-sized sand, interstratified with some gravels and cobbles (approx. 32 layers, 60–80 cm thick). The top is oxidised, truncated and has a bent coarse prismatic structure; the prisms are 3 m high and 10–30 cm in diameter. This prismation resembles a deeply penetrating desiccation network and is associated with consolidation of the prisms. The uppermost part is fragmented. Subunit D2 extends to Lindúr, upstream in the valley. Large deformations form metre-scale popdowns (between Lindúr, and Trolagill) (Fig. 2C and 6D), a large transpressional flower structure (50–30 m high in the sector between the Kárahnjúkur and the Sandfell) and, at the foot of the Sandfell, listric transpressional faulting (Fig. 6B).

Subunit D3 is a deformed stratified gravely sand, and is preserved only downstream of the main terrace (Fig. 5C).

Unit E – Unit E consists of medium-grained hyaloclastite sands, up to 15 m thick, with cross-bedding ending with ripple lamination, pointing to a transport mainly towards the North (subunit E1). It covers the older units unconformably. Its base is strongly erosional, especially upstream at Tröllagil and Lindúr. Climbing ripples are locally present (Knudsen & Marren, 2002). This subunit also fills a lateral pass (40 m deep) to the valley between the Sandfell and the Kárahnjúkur implying a river diversion to the Desjarárdalur (Fig. 2C).

To the North, subunit E1 passes downstream into the varve-like fine sands of subunit E2, up to an altitude of 545 m. This subunit occurs downstream of the concrete dam, incising the bedrock (unit B) for about 30 m. It has undergone weathering as shown by a bioturbated sandy marshy palaeo-andosol with some root-related iron nodules (subunit E3). The bioturbations are present as tubes bent by the water current and stained with iron oxides (Fig. 7C). Plant remains, bacterial iron, taecamoebian and diatom tests and amorphous clays are visible in thin section. This subunit is also found upstream, but has not been recognized by Knudsen & Marren (2002, their fig. 4) as representing a palaeosol. This subunit is faulted and folded by more recent glaciotectonism, mostly in the form of normal faulting with limited transpres-

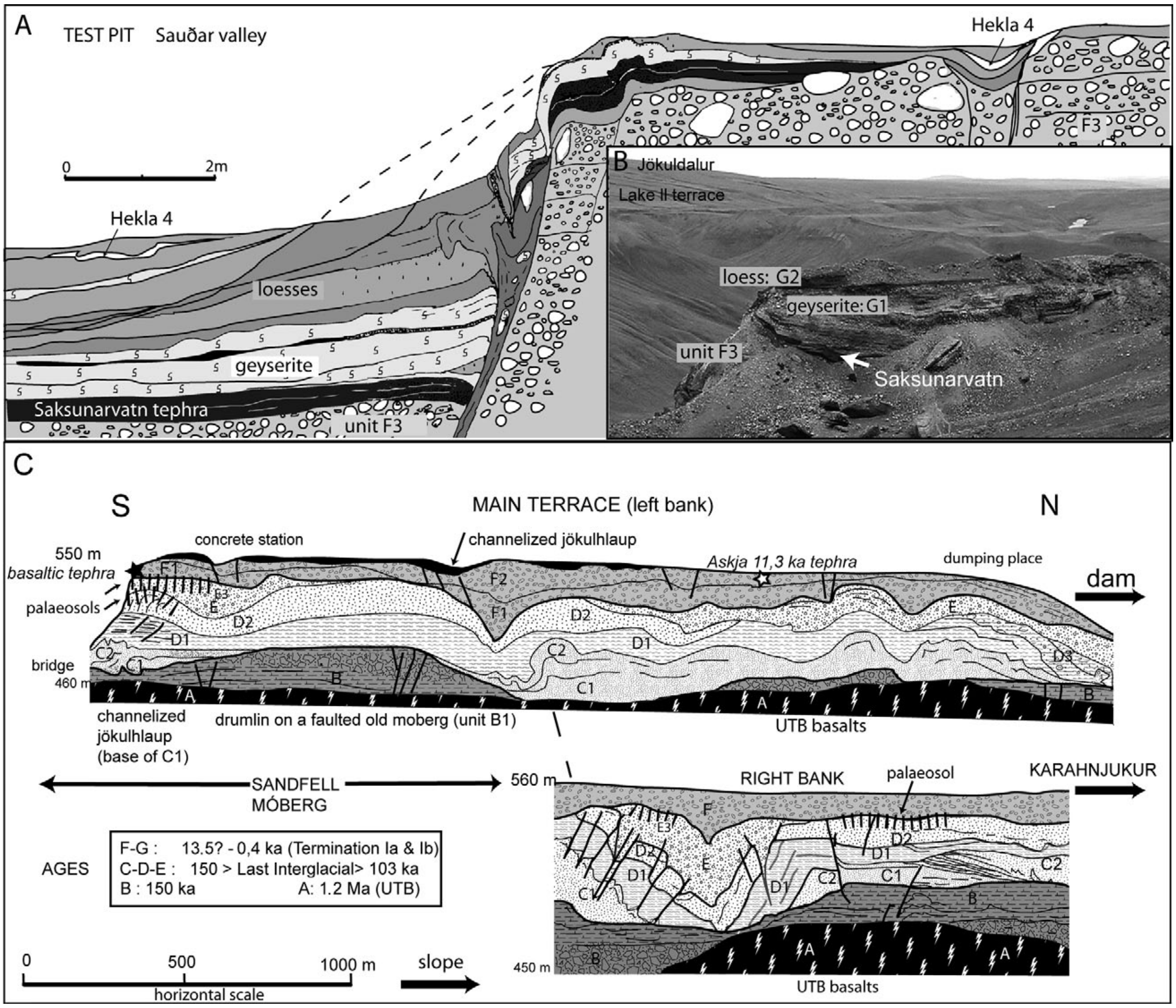

Fig. 5. Deformation structures.
A – Early Holocene tectonics (Sauðar valley); **B –** Glaciotectonic deformations in the Saksunarvatn tephra in the Jökuldalur close to the entrance of Sauðar valley; **C –** Field sketch of the valley sediments between the bridge (South) and the dam (North). Notice the drumlin form of unit B, in the southern part, and the giant glaciotectonics between the Sandfell and the Kárahnjúkur, within the right bank.

sion also affecting unit D along the right bank (Fig. 5C).

Interpretation of the lower succession (palaeolake I)

Unit C – Unit C marks a rapid deglaciation phase in a braidplain environment with some flashfloods, or jökulhlaup events, as shown by the erosional base of the unit. Apart from the initial erosion, accumulation prevailed, in a fully deglaciated plain. The deposits fine upwards. Subunit C2 reflects total deglaciation. This was followed by a short-lived surge of a warm-base glacier, defined here as Glacial A (Fig. 3), as indicated by the presence of glaciotectonic deformations, including sedimentary dykes (Fig. 6C).

Unit D – Sandy facies with rare dropstones have commonly been interpreted as ice-contact fans (a.o., Thomas & Connell, 1985); dropstones are, however, not necessarily derived from glaciers but may have been transported by lake-ice and onshore-ice rafts (a.o., Forbes, 1975; Dionne, 1998). Unit D followed the Glacial A advance and represents possibly an ice-contact glaciolacustrine fan (subunit D1). The lake was bounded by an obstacle (terminal moraine or landslide) that was further incised,

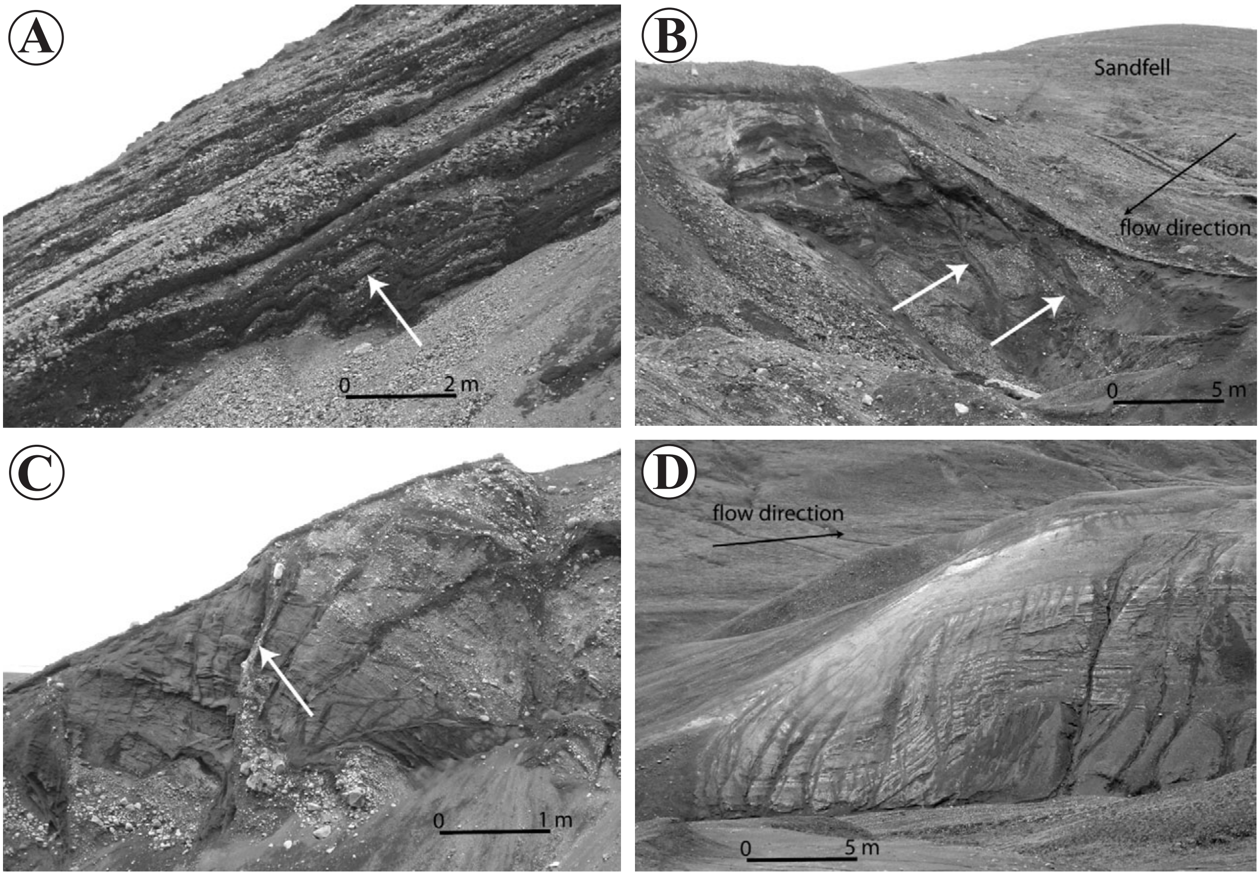

Fig. 6. Glaciotectonic deformations.
A – Glaciotectonic deformation (arrowed; folding due to transpression) of the oldest scree unit at the foot of the Kárahnjúkur; **B –** Listric glaciotectonic faults (arrowed) linked to glacial transpression (unit F), 500 m south of Figure 5C, right bank; **C –** Glaciotectonics and sedimentary dyke (arrowed) in unit C1 (bridge section); **D –** Glaciotectonic folding of subunits C2 and C3, 4 km upstream of the dam (close to Trolagil).

inducing a temporary spread upward of the stream (coarser subunit D2) (cf. Brodzikowski & Van Loon, 1991).

The Halslón palaeolake I probably drained rapidly, allowing some desiccation of the lake bottom, which became possibly affected by permafrost development, as indicated by deep prismation and mass oxidation (Fig. 7D). The lacustrine sediments became incised by a shallow valley (30 m) that can now be seen close to the river, and are now bent towards the valley by slow permafrost creep. They are also deformed by transtensional forces at the foot of the Sandfell volcano and 2 km upstream of the bridge (Fig. 6D). These deformed lacustrine sediments were derived from an old glaciofluvial fan (unit C1b). This is indicative of a significant re-advance (defined here as Glacial B: Fig. 3) of a temperate-base or a polythermal glacier.

Unit E – Unit E starts with an erosional base, scouring into both consolidated and soft sediments, suggesting the recurrence of jökulhlaups at the onset of its deposition; it probably eroded its way through the terminal moraines of unit D, as incision also exists downstream of the Kárahnjúkur at the base of unit E. Unit E consists in the main section mainly of fine-grained sands with horizontal, low-angle and climbing-ripple lamination, attesting to rapid sedimentary influx in a shallow river-fed lake, as described by Menzies (2002), with the foresets and bottomsets preserved – though strongly deformed – immediately upstream of the Kárahnjúkur. The temporary damming of the river may have resulted from a landslide on the western face of the Kárahnjúkur. The fluvial/lacustrine area shallowed and drained due to the prograding of a sandy delta on top of which a marshy palaeo-andosol developed with bioturbations (initially soft mucus tubes containing aquatic worms or larvae, stained by ferrihydrite), indicating clear-water conditions before its drainage and burial; the area reached full deglaciation, in the same way as reflected by unit C. A 50-cm thick andic palaeosol, present in several sections, ends this basal succession. It was deformed postdepositionally by Glacial C glaciotectonics (Figs. 3 and 5C), especially close to the outlet of the Sauðar valley.

Description of the upper succession (palaeolake II)

The upper succession (Fig. 5A) is mostly unconsolidated. It is thinner than the lower succession, mostly channelised, incising the lower succession up to 25 m deep. This has been attributed by Knudsen & Marren (2002) to a sandy ice-contact subaqueous fan giving way to a sandy braided-river system that coarsens towards the top. Our investigations showed that this succession is more complex than previously reported and does not contain here any ice-contact deposits. It consists of three regressive, mostly glaciolacustrine subunits separated by levels indicating strong glaciotectonic events (tension graben, local squeezing with metre-scale transpressional flowers, listric shear faults, etc.), and by locally preserved tills. Two main units (F and G) can be distinguished.

Unit F — This unit is composed of three subunits (F1-F3).

The lowest subunit, F1, is present on the main terrace. It consists of channelised coarse silty sands with somewhat gravelly layers and local cross-stratification. It lacks dropstones and is composed of reworked material from at least one basaltic tephra layer. This subunit is deformed by normal faults following a channelised pattern. Downstream, to the edge of the main terrace, some foresets are preserved; it has been eroded by subunit F2 (Fig. 5C, left bank).

The middle subunit, F2, which also occurs on the main terrace, consists of coarse silty sands prograding in channels. It also comprises reworked material from a rhyolitic tephra layer (Fig. 7A) from which the grains present a typical shard morphology. This subunit is deformed by normal faults following a channelised pattern and is irregularly eroded by subunit F3. Very locally, up to 1 m of matrix-supported silty/sandy diamicton is preserved. At the entrance of the Sauðar valley, on the side of the main valley, cross-stratified gravel is buried under fining-upward sands, with oblique stratifications. These are further strongly plastically deformed in stacked vertical slabs and finally truncated by subunit F3.

Subunit F3, the upper one, is preserved only locally. It consists mostly of sandy gravels that are laterally deformed by moderate glaciotectonism (normal faulting) at the left valley side (Jökuldalur; Fig. 4B), but not in the tributary Sauðar valley. This sediment aggradation has more gravelly layers and cross-stratification at its base. It includes at its top the classical early Holocene Saksunarvatn basaltic tephra (Fig. 8) from the Grimvötn volcano (Guðmundsson & Helgason, 2004). In the main valley, sediment aggradation continues to the 570 m terrace level. On the terrace, it is partly washed as a gravelly beach, followed by loess deposition. It is locally onlapped by subunit G1, but commonly further entrenched and scoured by recent fluvioglacial erosion.

Unit G — Subunit G1 consists of horizontally stratified fine sands and gravels, including some strata of whitish geyserite, preserved in a kettle hole (Fig. 5B).

Subunit G2 mainly consists of loesses with several rhyolitic tephra, including a layer from the recent Hekla 4 outburst. In the Sauðar valley, one metre of geyserite (G1) crops below the loesses.

Interpretation of the upper succession (palaeolake II)

Unit F — Subunit F1 forms the upper part of an unconsolidated fluviolacustrine fan. The material was supplied mainly by glaciofluvial streams (cf. Menzies, 2002) and is hard to position stratigraphically by itself.

Subunit F2 represents a braided plain fed and filled by glaciofluvial streams. In the preserved outcrops, it is difficult to identify a lacustrine component, even though the sediment remains fairly fine-grained. The redeposited rhyolitic tephra have partly been ejected dur-

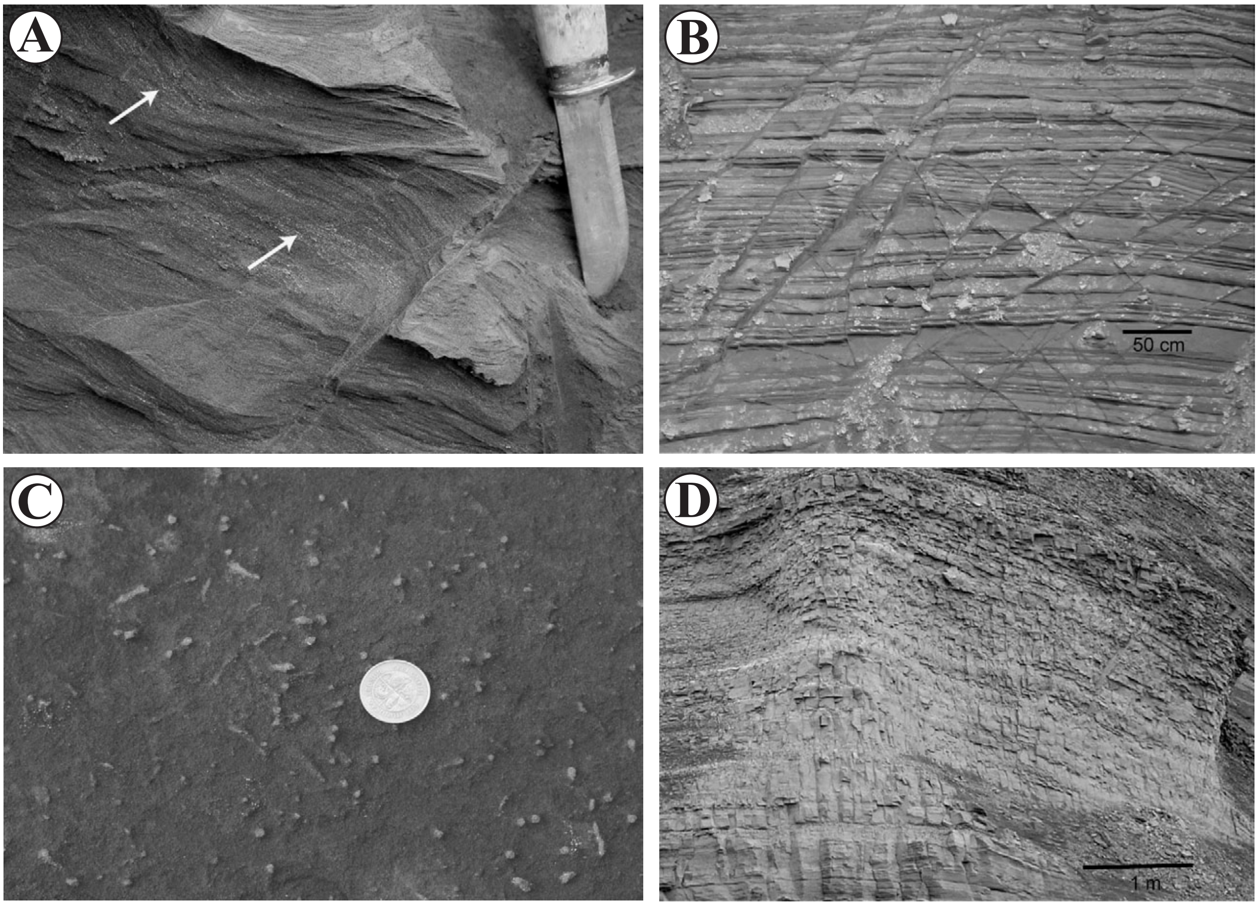

Fig. 7. Synsedimentary features in the main bridge section, Jökuldalur.
A – Reworked Askja (11.3 ka BP) tephra in unit F2 (arrow; note also the glaciotectonic faults); **B –** Glaciolacustrine rhythmites with dropstones, deformed by transtension fractures; **C –** Fresh-water bioturbations (burrows of insect larva's or worms, orientated in the current direction) at the top of subunit D3; **D –** Prismation of subunit D3, related to permafrost development.

ing the Askja '10 ka' eruption following the major and trace-element geochemistry (Guégan, 2010), and not the Skógar eruption as believed before (Van Vliet-Lanoë et al., 2007). This tephra has been dated as about 11.3 cal ka (Fig. 8) by Wastegård (2005) and was derived from the Dyngjufjöll caldera (Sigvaldason et al., 1992). For these reasons, subunit F2 should correspond to the deglaciation of the Younger Dryas–Búrfell Advance (Fig. 8). Therefore, subunit F1 must be correlated with the deglaciation of the Alleröd Cold Event as defined by Yu & Eicher (2002). The reworked basaltic tephra in unit F1 are probably derived from the basaltic component of the upper Borrobol tephra close to 13 ka (Wastegård, 2005), very similar to the 1553-m of the NGRIP ice core (Mortensen et al., 2005) and probably derived from the Snæfellsness.

Subunit F3 represents a glaciofluvial succession linked to deglaciation after a minor glacial advance restricted to the main valley, ending with the Saksunarvatn tephra (approx. 10,180 cal. years BP). It is probably connected to a local terminal moraine downstream (Fig. 2C). The highest lacustrine terrace (570 m) is incised in this subunit. The damming of the valley by the minor glacial advance of approx. 10.1 ka is probably responsible for the development of palaeolake II, which had a very brief existence, and for the successive abrasion terraces down to the main level at 550 m, before entire or partial erosion of the lowermost terrace took place by recent jökulhlaups. The latest deformations affecting this subunit, including the

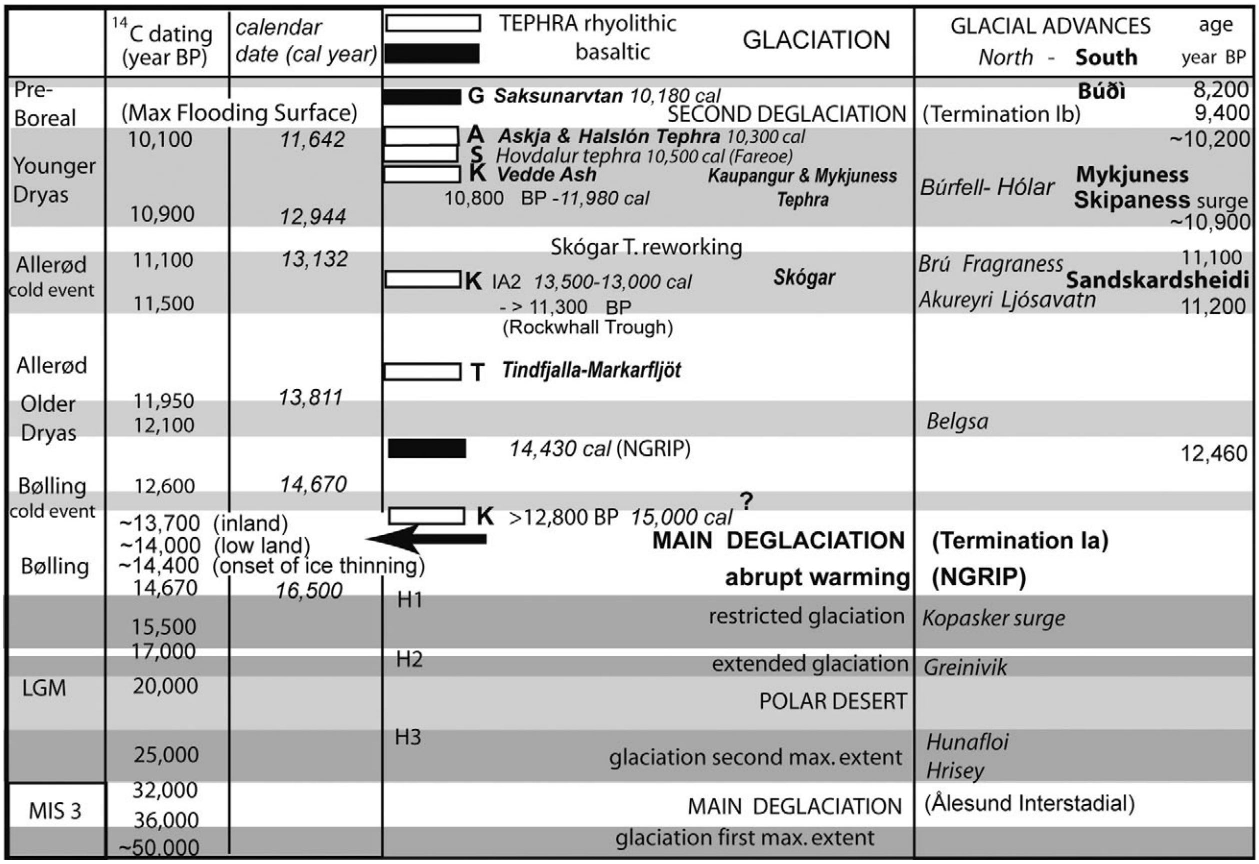

Fig. 8. Stratigraphy of the Late Glacial deglaciation in Iceland. Modified from Van Vliet-Lanoë et al. (2007). Tephra sources: A = Askja; K = Katla; T = Tindafjalla; Sn = Snaefell. H1-H2-H3 = Heinrich layers.

Saksunarvatn tephra, are found in the upper Jökuldalur only (Fig. 5C). These deformations are probably linked to the burial of dead-ice and may date back to the Preboreal I (approx. 10,100–10,050 cal BP), as also observed close to the Jothunheim in Norway (Dahl et al., 2002); it seems somewhat earlier here, considering the vicinity of the Vatna ice cap.

Unit G — The G1 glaciofluvial deposits accumulated mostly in juxta-glacial position (kame terrace), partly on dead-ice (synsedimentary thermokarst), at the entrance of the Sauðar valley in connection with the palaeolake II terrace.

The G2 loesses formed, as in Northern Iceland, contemporaneously with the Neoglacial (Subboreal cooling) (Van Vliet-Lanoë et al., 1998).

Complementary stratigraphical observations

Looking for potential damming features, examination of the valley downstream of the main outcrop is important. In the right wall of the Jökuldalur valley, at the western foot of the Kárahnjúkur, 200 m above the present-day stream, a large scour occurs. It seems to be related to an old landslide that was later smoothed by glacial erosion, and it is partly buried by three generations of scree units that cover the eastern flank of the valley (at the eastern anchoring site of the concrete dam). The two lower scree units show wavy deformation (transpression, Fig. 7A), but the youngest one is still intact, with some infiltrated loess. Some erosional features such as stacked boulders have recently been affected again by jökulhlaup activity, especially at the level of the main terrace (dam's workplace; 550

m a.s.l.) and at the level of the outlook point in front of the Kárahnjúkur (also 550 m a.s.l.).

Immediately downstream of the concrete dam, the lower (palaeolake I) succession is partially preserved. Subunits E2 and E3 are topped here by a palaeosol, just like upstream. Subunits C2 and C3 are preserved here at 550–570 m a.s.l., resting immediately upon unit B. All these subunits are truncated by a lodgement till (unit F) and are often deformed by glaciotectonics. Unit F has been very strongly incised by sub-recent jökulhlaup activity, which is responsible for the columns, pillars and isolated large blocks shaping the local sedimentary outcrops perched at 150 m above the present-day stream (cf. Maizels, 1997). No Holocene loess cover is present.

In the Jökuldalur valley, downstream of Brù, the late deglaciation morphology is preserved only in the form of a series of deeply incised sediments formed in lakes (Fig. 3) dammed by terminal moraines belonging to the last deglaciation, which are preserved well in parallel valleys (Fig. 2C). Some sections on the left side of the valley show a similar succession as present now upstream in the Jökuldalur valley in the form of a terrace: (1) a lodgement till commonly deformed by glaciotectonics, covered (similar to Halslón unit B) by (2) finely stratified hyaloclastitic sands from a braided-river system, sometimes passing into lacustrine deposits (subunit E2), and (3) a weathered (75 cm thick) andosol (subunit E3). The entire succession is strongly deformed by glaciotectonics and 'recent' ice push.

Sedimentary deformations: glaciotectonics versus neotectonics

Most of the sedimentary deformations in the study area correspond to dynamic glaciotectonism induced by channelised glacier motion. Most of the deformations (Fig. 7) are extensional, due to loading (unit F) or transpressive forces (units G and E), but some of the glacitectonic deformations result from the sedimentary context (obstructive fans spreading from lateral valleys deformed by frontal and lateral glaciotectonism: Fig. 7B). Late Glacial scree units were also deformed by lateral pressure, probably of glacial origin (Fig. 7A).

There is very little evidence of neotectonics. The Halslón dam is also located outside the present-day zone of microseismicity defined by Einarsson (1994). Indications of true tectonic activity are limited, apart from faulting and seismites in the early Middle Pleistocene unit A (Fig. 3). The Halslón stratigraphic record (units C to F) does not show any evidence of synsedimentary macro- or micro-seismicity (load casts, liquefaction, rafting), as observed in the volcanic zone or in Eyjafjördur in the North. The only evidence of true tectonic activity has been found in a fresh fault scarp investigated with a test pit along the Sauðar River (Fig. 5A). The deformation with vertical offset and flower structure took place after the Saksunarvatn tephra had been deposited (subunit G2), during the accumulation of the geyserite. Geothermal activity persists nowadays upstream in the Jökuldalur. Seismic shocks affecting a reworked Saksunarvatn tephra layer are recorded at 60 km to the North close to the Snautasel farm, attesting to low-magnitude earthquake(s). The deformed sediments have been 'fossilised' at both sites by a cover of loess that includes rhylolithic Hekla 4 tephra, which have been dated as approx. 4310 BP (Wastegård, 2005). Some younger deformations must be ascribed to cryoturbation and hydrological piping, resulting especially from the Subboreal cooling.

Discussion

Sedimentary record and chronostratigraphy

Present-day observations in northern Iceland show that sandy sedimentation is common in gently sloping valleys and estuaries, particularly during phases of colonisation by plants. Thus, most of the sand bodies represent interglacial deposits. Pedogenesis often indicates the final drainage of alluvial plains before the onset of the next glacial; it needs several thousands of years to develop, much longer than any Weichselian interstadial. Sandy braided-river deposits in the study area may be interpreted as entirely accumulated during an interglacial, considering the proximity (< 20 km) of the Brùarjökull and their distinct pedological weathering. Moreover, the degree of consolidation of the whole formation, in combination with the small grain size of units D (apparent bulk density close to 2 or higher in silts), implies long-acting glacial loading and ultradesiccation as obtained under permafrost conditions (Van Vliet-Lanoë, 1998).

True lacustrine sedimentation is limited to subunits D1-D2 (palaeolake succession I), F1 and the end of F3 (palaeolake succession II); it is much less developed than claimed by Knudsen & Marren (2002). The entire record seems too complex for a ′Middle Pleniglacial Event′ and the tephrostratigraphy invalidates a single Late Glacial Event as proposed by Knudsen & Marren (2002). It is very possible that units C, D and E together represent a full interglacial succession prior to the Late Glacial.

The weathering of unit D and the maximum glaciotectonic disturbance prior to the intermediate incision, combined with the age of the youngest Sandfell móberg that is truncated by the valley infilling (150 ka, late dyke at 103 ka), suggests that the main Halslón infilling corresponds to the last interglacial, whereas the unconsolidated upper succession corresponds to the last deglaciation. The main ′lacustrine′ deposits (palaeolake succession I) are intercalated between sediments that accumulated during one important and one more ′restricted′ deglaciation phase, and that represent a brief but widespread event. The ′Lake II′ of Marren and Knudsen is commonly better recorded in the morphology than in the sedimentation, and is attributed here – on the basis of tephrostratigraphy – to the last deglaciation and classically to ′Termination I′ (e.g., Mangerud et al., 1998; Fig. 8).

Termination I

The upper succession (units F-G) is ascribed to Termination I. In the Jökulsa á Brù valley, the upper sedimentary succession (consisting of terminal moraines) also represents Termination I. Consequently, the base of the Halslón succession will be discussed in the following as belonging to glacial Termination II, whereas the main succession (units C-E) will be attributed to intra-MIS 5e glacial events, with most of the deformed sand bodies representing an interglacial complex. The last weathered material (unit E) was deformed during the local Early Glacial.

The sedimentary records from the Jökulsa á Brù and from the Jökuldalur/Halslón valleys are largely similar to those encountered in the NE rift zone and along the coast, despite their position upstream in another catchment area. The isotopic age of the intercalated basalts has been difficult to determine inside the NE rift zone, however, because of their low K content. Nevertheless, the base of the interglacial complex is, on the basis of K/Ar dating, younger than the 150 ka móberg at Halslón, and also younger than the little Petursey móberg at the southern edge of the Myrdal ice cap (ISLN–81: 126 ± 12 ka; Guillou et al., 2009), but older than a 80 ka lava flow close to Mörðrudalur, burying the interglacial formation sensu lato (Skarðsengi: Fig.1C; Van Vliet-Lanoë et al., 2005). The age of 103 ka for a dyke at the Sandfell volcano, the erosional unconformity and the large glaciotectonic deformations (Glacial C) (1) imply that a glacier covered both the valley and the plateau at that time, (2) indicate that the glacier was permanent at least from approx. 110 ka onward, and thus (3) imply that the entire lower Halslón sedimentary complex formed within MIS 5e.

Termination II

The penultimate glaciation, which is interpreted as Saalian II (MIS 6, 180–135 ka), was characterised in the study area by a major, temperate-base ice mass (Van Vliet-Lanoë et al., 2005) that probably covered most of the shelf. MIS 6 is associated with the storage of huge amounts of ice on lands around the North Atlantic (Ehlers, 1983; Mangerud et al., 1998; Svendsen et al., 2004), due to the mild conditions in the Atlantic Ocean (Funnell, 1995). Temperate-based glaciers use to have melting waters at their contact with the substratum, explaining the erosional capability of the ice masses. The thickness of the ice on Iceland also explains the importance of the subglacial hyaloclastites, which were further reworked by meltwaters during Termination II.

Weichselian glaciation and Termination I-a

Glacial C (deformed subunit D3) had the widest extent of ice in the Jökuldalur, and its deposits were truncated by unit F1 at the local onset of Termination I-a. If our interpretation of Glacial C (110 ka) is correct, the Weichselian glaciation developed as early in northern Iceland (Van Vliet-Lanoë et al., 2001, 2005, 2007) as it did in Scandinavia, as well as in Siberia and Greenland (Funder et al., 1991; Mangerud et al., 1998; Svendsen et al., 2004). Only limited erosion occurred by the cold-based ice mass, which probably affected a substratum of preglacial continuous permafrost (unit C). The ice sheet retreated during the LGM (26–18 ka) due to the limited precipitation that was possible as a result of the wide extent of sea ice (Labeyrie et al., 2004; Van Vliet-Lanoë et al., 2007). This is consistent with the absence of deposits from 26 ka to 17 ka ago off NW Iceland, mentioned by Andrews et al. (2000) and Andrews & Helgadottir (2003), and with the Middle Weichselian transgression reconstructed by Svendsen et al. (2004) for the Scandinavian Arctic. A limited LGM glaciation could also explain the characteristics of the Termination I-a sediments, which are poor in hyaloclastites.

Termination I-a started early in NE central Iceland as shown (Fig. 8) by the occurrence of the Askja '10 ka' tephra (11.3 cal ka BP) and the Saksunarvatn tephra (10.2 cal. ka BP) close to the end of the Halslón sedimentation (Van Vliet-Lanoe et al., 2007; Guégan 2010). Termination I-a was extensive throughout Iceland from the Bølling onward (Geirsdóttir et al., 1997; Jennings et al., 2000, 2006; Principato et al., 2006; Licciardi et al., 2007). In the upper Jökuldalur, it started from the Allerød event; several terminal moraines of the Bølling, the Older Dryas and the Allerød are even preserved a few kilometres downstream, close to Brù (Fig. 2C).

In conclusion, an early ice extent, a limited LGM glaciation and the onset of deglaciation in NE Central Iceland prior to the Younger Dryas are major differences if compared to the interpretations presented in previous works about Iceland (e.g. Norðdahl & Halfidason, 1992; Ingólfsson et al., 1997; Knudsen & Marren, 2002).

Interglacial glaciations (Glacials A and B) and the Early Glacial

A widespread interglacial formation (Syðra Normellsfjell Formation, Fig. 1C; see also Van Vliet-Lanoë et al., 2001, 2005) is commonly present in the rift zone, in the North Volcanic Zone, and in several continental and shoreline deposits; it is intercalated between a lower, thick and consolidated glacial formation (MIS 6), and an upper, unconsolidated till or a deglaciation succession. The interglacial sands at Halslón are, as well as those in other regions of northern Iceland and Vopnafjördur, derived from hyaloclastites, not from classical tills. This is the signature of the Last Interglacial in northern Iceland (Van Vliet-Lanoë et al., 2001, 2005).

During this interglacial, the Halslón succession records limited glacial pulses, i.e., Glacial A and Glacial B (Fig. 3). The MIS 5e succession was rapidly followed (like in the NE rift) by a first ice advance, Glacial C, that could correspond to early MIS 5d (approx. 110 ka) in the vicinity of the Vatna ice cap with subglacial volcanic activity at 103 ka (Sandfell). This first major advance was probably followed by a more important one during MIS 5b (approx. 90 ka), probably reaching the coast in the north-eastern rift (Van Vliet-Lanoë et al., 2005), comparable to the development in northern Scandinavia (Mangerud et al., 1998).

Hyaloclastic interglacial sands representing two distinct marine highstands in a similar stratigraphical position, separated by a regression, also exist along the NW coast (Van Vliet-Lanoë et al., 2005). These events are also recorded offshore along the Reykjaness ridge (Eynaud et al., 2004). The occurrence of two distinct highstands during MIS 5e is common in western Europe (Van Vliet-Lanoë et al., 2000) and has been dated at <125 and >115 ka in Normandy (Van Vliet-Lanoë et al., 2006) and along the Mediterranean (Orszag-Sperber et al., 2001; Bardají et al., 2009; Mauz et al., 2009). The short regression probably corresponds to the lowstand at approx. 117 ka (Shackleton et al., 2003).

A limited glacial advance (Glacials A + B) between two highstands at the coast may correspond to MIS 5e and 5a as proposed, following the classical approach, by Funder et al. (1991) for the Thule region or MIS 5e and 5c, and as we have proposed for the Sydra Normellsfjell Formation, in NVZ, closer to the coastal zone. It may also correspond, however, to the mid-Eemian cooling (MEEC: a.o., McManus et al., 1994; Maslin et al., 1998; Shackleton et al., 2003). The MEEC actually consists of two cooling events: one at 122–120 ka and a second one at 118–117 ka. These coolings were orbitally forced (Occhieti, 1987; Maslin et al., 1998).

A controversy exists regarding the MEEC. Glaciers commonly develop at the onset of a cooling phase, when the sea level is still high (Ruddiman & McIntyre, 1981) and while active heat transport by sea currents still takes place (McManus et al., 2002). An NGRIP (North GReenland Ice core Project – NGRIP members, 2004) from Greenland (75°10′8″N) demonstrates a climatic cooling from 122 ka ago that was driven by orbitally controlled insolation, with a glacial inception at 118 ka. Although this is easily observed in Europe, with a second climate optimum recorded at approx. 116 ka, and a late cooling of the eastern Atlantic until 110 ka (Maslin et al., 1998; Müller & Kukla, 2004; Sanchez-Goni et al., 2005), no evidence of the MEEC has been observed in North America. Several arguments show evidence of early glacial advances there, however, although these have been dated only relatively. The main evidence is an early glacio-isostatic transgression in the ′Middle Sangamonian′, which could also be the MEEC or the early MIS 5d cooling. This transgression seems linked to the formation of ice sheets on the north-eastern territories and eastern Canada (Andrews et al., 1983; St-Onge, 1987; Richard et al., 1999; Kleman et al., 2010) under orbital forcing (Occhietti, 1987). In the eastern Canadian Arctic and north-western Greenland, the cooling set on from approx. 119 ka, and the ice cover apparently already reached its maximum approx. 114 ka ago at Thule (77°28′57″N) (Funder et al., 1991; Svendsen et al., 1989) and ice did not retreat inland during the MIS 3 in the vicinity of the Greenland ice cap (Funder et al., 1991). This was confirmed by the last NGRIP synthesis (NGRIP members, 2004); its high-resolution isotope record reveals a slow decline in temperature from the warm Eemian to cooler, intermediate values over 7 ka, from 122 to 115 ka BP. The end of the last interglacial in north-eastern America thus does not appear to have begun with an abrupt climate change, but with a long and gradual deterioration of the climate, interrupted by Dansgaerd-Oeschger (DO) events. Stadial 26 (~119–117 ka), recorded by NGRIP $\delta^{18}O$, fits the MEEC, and interstadial 25 (~116–113 ka) fits the second warm peak of MIS 5e. As most of the North Atlantic is free of ice-rafted debris from this time-span, this implies a largely deglaciated coast of Greenland with ice streams from the Greenland ice sheet that had retreated within the fjords (Keigwin et al., 1994; McManus et al., 1994; Bauman et al., 1995; Cuffey & Marshall, 2000).

Due to its inland position, close to the Vatna ice cap, Glacial A and Glacial B at Halslón, represented by sediments intercalated by a complete deglaciation succession, may be correlated with the MEEC, with a limited internal deglaciation (ice-contact lake). In contrast to this progressive cooling in northern Greenland and the north-western Atlantic (Rasmussen et al., 2003), Iceland's glacial history seems to react similarly to the ice core Dye 3 record (SE Greenland 65°N) (NGRIP members, 2004). The Vatna ice cap totally vanished during the two thermal peaks. The Halslón record (64°57′14″N) attests to a glacial advance in two pulses from a totally deglaciated area, at only 20 km north of the present edge of the Vatnajökull (Van Vliet-Lanoë et al., 2007). Moreover, an ice-dammed lake formed and permafrost developed at Halslón, between the two ice advances, indicating persistent cold conditions at an altitude of 550 m.

The extent of the Vatna, Myrdal and other ice caps during the MEEC and a glacio-isostatic subsidence of approx. 100 m of the Vatna southern margin should have allowed for a floating ice margin SW of Iceland. It is located precisely at the place from which the basaltic ice-rafted debris found in Brittany in MIS 5e beach-sediments seems to have originated (Hallégouët & Van Vliet-Lanoë, 1989). It should also fit the anomalous record of a continuous supply of ice-rafted debris SE off Iceland (MD95–2009 core: Eynaud, 1999), indicating an early extent of sea-ice to southern Brittany, which should imply a NW-SE surface drift from the East Icelandic Current with the disappearance of the Irminger Current (Eynaud et al., 2004).

As for Greenland, this allows also for Iceland the interpretation of an important deglaciation during MIS 5a, as supported by the observation, 85 km North of the Vatna ice cap, of a 80 ka subaerial lava burying the interglacial formation at Skarðsengi (Van Vliet-Lanoë et al., 2005).

Volcanism, ice streams and drainage

The Kárahnjúkur volcano probably forms part of the Snæfell system. During early deglaciation, tension faults opened up in formerly glaciated areas because the disappearance of the ice cover resulted in a decrease of the vertical stress, promoting volcanic activity (Sigvaldson et al., 1992). The Kárahnjúkur ridge system was reactivated from 250 ka (MIS 8) to 150 ka (Sandfell, MIS 6). The MIS 8 glaciation was limited in size (Liesecki & Raymo, 2005), but MIS 6 was a glaciation with massive storage of ice, favouring volcanism during deglaciation.

The traditional viewpoint with respect to the flow pattern of the ice stream is indicated in Figure 1B and 2B (route 1); our new data suggest the pattern that is shown in Figure 1A. These changes after the building of the Snæfell volcano from approx. 250 ka (MIS 8) are geomorphologically self-evident on the Jökuldalheidi plateau (Fig. 1C). The recent 'abnormal' entrenching both in the upper Jökuldalur valley and downstream of Brù also suggests more intense drainage during and following the last deglaciation than during the Saalian, and is thus an indirect indication of the northern routing (Fig. 2B, route 3) of the ice stream during MIS 6 (Van Vliet-Lanoë et al., 2007).

The regional ice-flow pattern was diverted from the NE to the N from the Middle Pleistocene onwards, due to the development of the Snæfell volcano after MIS 8 and probably also to the reactivation of the Kárahnjúkur ridge. As a consequence, the Héraðsfloí ice stream was partly starved, and the Vopnafjördur ice stream developed. The changing pattern is clearly related to subglacial volcanism during the large glaciations such as MIS 6 (see Section 'Morphology and ice-flow lines'), but the northern elongation of the Mulalón palaeolake indicates that a similar situation also existed about 760 ka ago. All these data suggest that a northern routing of ice streams occurred more than once after large glaciations during which subglacial volcanism was active. This explains the large submarine platform fan in front of the Vopnafjördur (Fig. 1A) despite the limited size of the catchment area upstream. These fans were presumably formed by successive lake-drainage events during the Middle Pleistocene. The building of the Snæfell and Kárahnjúkur led to the required major change in the Middle Pleistocene sediment supply.

The present-day spectacular incision of the Jökuldalur valley was facilitated by the existence of previous entrenchings. The Jökuldalur canyon incised stepwise in a glacially abraded, wide valley formed after the Brù-Háls lava flow (ending approx. 480 ka). The upper Jökuldalur was filled locally by the Kárahnjúkur hyaloclastites during MIS 8, but was re-exploited by rivers during MIS 7 and MIS 5e, and by glaciers during MIS 6. No evidence of volcanic damming has been observed that may be related to the infillings formed by Hálslon units C, D and E. Potential damming objects are terminal moraines, lodgement-till accumulations (but the till bodies are small in this region), landslides and even ice from a downstream glacial centre, e.g., the Fjällgarðar ridge. Thus, they are easily destroyed, but not necessarily by a brutal breaching.

Downstream of the concrete dam, the Eemian (units C and D) is preserved at 560 m a.s.l., with evidence of an early incision down to 530 m in the North. During the mid-Eemian, the Jökulsa stream changed its course, flowing temporarily to the East between the Sandfell and the Kárahnjúkur, probably due to the landslide from which the scar is still visible on the western face of the Kárahnjúkur. The temporary palaeovalley was 40 m deep (Guðmundsson & Helgason, 2004), reaching a level of 540 m a.s.l. Several terminal moraines certainly formed during the mid-Eemian cooling, stacked between the Kárahnjúkur and the Sandfell, as shown by the glaciotectonic deformation and compaction of unit D before the deposition of unit E and the onset of the main glaciation (Glacial C). These accumulations are responsible for the lacustrine episodes during MIS 5e. Jökulhlaup activity is recorded from deglaciation pulses (base of units C and E).

Late Glacial terminal moraines exist downstream in side valleys of the Jökuldalur valley; they were occupied by the Younger Dryas glacier (Fig. 2C) during the Bùrfell stage (Fig. 8), although the Sauðar valley was already ice-free during the Preboreal advance. During deglaciations, sediments accumulated in the form of damming fans, and the river then partly flowed over the surface of older fills. When the rate of deglaciation diminished, the fairly narrow incision of the canyon was renewed especially by jökulhlaup erosion (Rudoy, 2002). This can hardly be attributed to the sudden failure of the volcanic dam from palaeolake II during the last deglaciation (Knudsen & Marren 2002) as upstream, at Trolagill, the postglacial incision is about 60 m deep in basalts. Palaeo-erosional features have been partly erased by recent jökulhlaup activity, especially at the level of the main terrace (dam workplace). Some traces of these floods, surging following Björnsson (1992) from the Kverkjökull volcano through the Kringilsa valley, are preserved both upstream in the Jökuldalur and downstream in the Jökulsa á Brù valleys. Due to the erosion of the loess cover, they can be ascribed mainly to the approx. 4200 BP jökulhlaup events in the rift zone described by Maizels (1997) and Carrivick et al. (2004). Thus, it may be deduced that mid-Holocene surge flood events accentuated a Late Glacial to Early Holocene valley (cf. Van Vliet-Lanoë et al., 2007).

Conclusions

Middle Pleistocene volcanism in NE Iceland has controlled changes in glacier and river-drainage patterns. The MIS 6 glaciation was a highly significant event, followed by the deposition of a thick deglaciation succession (Termination II). Giant lakes, such as the Halslón palaeolake, formed close to the end of the MIS 6 deglaciation and further during the mid-Eemian cooling deglaciation events.

During the phases of maximum ice extent, the ice flow was directed from the Brùarjökul glacier to the Vopnafjördur area, which has certainly occurred since the formation of the Snæfell and Kárahnjúkur volcanoes. The change of the ice-flow direction has been forced since the MIS 8 glaciation, modifying the offshore sedimentary budgets off the Vopnafjördur coast during major deglaciations. Compared with Termination II, the deposits of the Weichselian deglaciation (Termination I), which occurred very early, are limited in thickness. The Jökuldalur canyon that is incised into the Early Pleistocene hyaloclastites downstream of the Halslón lake, probably developed over a number of stages; the incision must have started during the Middle Pleistocene (approx. 400 ka); it diminished during deglaciations and was re-activated again by recent jökulhlaups.

The main lacustrine accumulation (palaeolake succession I) was formed and preserved primarily thanks to intra-Eemian glaciations, the so-called mid-Eemian cooling (MEEC). The best evidence of this controversial event is well preserved thanks to its vicinity to the Vatna ice cap. The local signature of the MEEC evidences an intermediate answer to the progressive cooling in North America and the temporary return to an interglacial on the western coast of Europe. In contrast to previous interpretations, the long glacial sedimentary development in NE Iceland, as described here, is consistent with the data from other Arctic regions, from Eurasia and from Greenland (Funder et al., 1991; Hjort, 1981; Svendsen et al., 2004), and thus increases the insight into the history of the northern Atlantic Middle Pleistocene.

Acknowledgements

This research was funded by Program no. 316 IPROCI of the Institut Paul-Emile Victor (IPEV).

Spot images were purchased thanks to the French PNTS and ISIS programmes, and the IPEV.

We also greatly thank Professor L. Marks and an anonymous reviewer for their suggestions to improve the content and the clarity of this contribution.

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Manuscript received 25 May 2010
revision accepted 25 November 2010