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Controls on the sedimentology of an ice-contact jökulhlaup-dominated delta, Kangerlussuaq, west Greenland

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Abstract

This paper characterises the sedimentary impact of a glacial outburst flood or 'jökulhlaup' on an ice-contact delta topset at Russell Glacier, Kangerlussuaq, west Greenland. Rapid drainage of an ice-dammed lake in July 1987 generated a jökulhlaup with a peak discharge of $\sim 1300 \text{ m}^3 \text{ s}^{-1}$, which drained across a 500-m-wide, 200-m-long, delta top into a proglacial lake. The delta topset comprises boulder clusters, ice block obstacle marks with relief of up to 4 m, and is graded to lake levels up to 6 m higher than those during typical non-jökulhlaup conditions. The delta top was dissected by the 1987 jökulhlaup causing a fan-shaped extension of the delta front by 30 m. Surface grain size on the delta decreases rapidly away from the main flood flow direction, reflecting rapid downstream reduction in sediment transport capacity. The 1987 jökulhlaup was predominantly fluidal and turbulent and had peak stream powers of 2846 W m^{-2} proximally and $<400 \text{ W m}^{-2}$ distally. Delta topset sedimentation can be characterised by four lithofacies associations in order of decreasing flow energy: (A) coarse-grained deposits related to a flow expansion; (B) finer-grained peripheral deposits located at the margins of the main flow; (C) lobate bars and delta fronts deposited within distal locations and (D) fine-grained deposits at distance from the delta front associated with slackwater conditions. Jökulhlaup-dominated delta topsets are controlled by the geometry of the channel expansion into the proglacial lake, jökulhlaup hydrograph form, the sediment availability and character, proglacial lake basin depth and surface area, lake outflow spillway erodibility and cross-sectional area, and history of previous jökulhlaups.

Keywords: Greenland; Jökulhlaup; Delta; Kangerlussuaq; Rapid lake level change; Sedimentology

1. Introduction and aims

Glacial meltwater is important for glacier dynamics, sediment transfer, and the creation of distinctive landforms and sedimentary successions. Proglacial meltwater systems extend the influence of glaciers well beyond their margins. The hydrological system of a glacier can therefore exert considerable control on depositional process and form with proglacial meltwater systems. We can therefore use glacifluvial and glacilacustrine landforms and sedimentary successions to reconstruct ancient glacier hydrological systems. Modern proglacial fluvial systems provide valuable process-form analogues for meltwater events with a range of magnitudes and frequencies. Although there is growing recognition of the impact of glacier outburst floods or 'jökulhlaups' within modern proglacial river systems, there are relatively few studies of flood process and impact.

Numerous spectacular, coarse-grained, fluvio-deltaic deposits have been attributed to the action of glacier outburst floods or 'jökulhlaups', generated by the

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sudden drainage of ice-dammed lakes, in former ice-marginal locations in both Europe and North America (Melander, 1977; Ulfstedt, 1979; Johansson, 1988; Kehew, 1993). Reconstruction of palaeoflow magnitude and frequency is based either on delta morphology and sedimentology (Elfström, 1983; Williams, 1983; Rossbacher and Rhodes, 1984; Elfström and Rossbacher, 1985; Rossbacher and Rhodes, 1987; Elfström, 1987, 1988) or upon historical knowledge of jökulhlaups (Shakesby, 1985). Previous studies have therefore been unable to tell us about the magnitude and frequency of delta forming flows or provide information about delta development during individual jökulhlaups. This study examines the topset of an active coarse-grained jökulhlaup delta in order to improve our understanding of the relationship between jökulhlaup characteristics and delta topset morphology and sedimentology. This paper provides the first study of the sedimentological and morphological response of a coarse-grained, ice-contact, jökulhlaup-dominated delta to an observed jökulhlaup, July 17–19, 1987. Knowledge of modern jökulhlaup characteristics over the delta is crucial for interpretation of sedimentary structures as well as improving understanding of the controls on delta morphological change and internal sedimentary characteristics. Process observations allow patterns of both spatial and temporal variability of deposition across the delta surface to be distinguished.

2. The study area and methods

The field area is located at 67°00′N 50°43′W on the western margin of the Greenland ice sheet (Fig. 1). The jökulhlaup delta lies 30 km northeast of Kangerlussuaq (Søndre Strømfjord) on the northern margin of the Russell Glacier (Fig. 1). Repeated jökulhlaups at the Russell Glacier are well documented (Sugden et al., 1985; Gordon, 1986; Scholz et al., 1988; Russell, 1989, 1993). The jökulhlaup delta is located at the southern edge of a proglacial lake ‘Outlet Lake 1’ and is graded to a variety of levels up to 6 m above the normal, non-jökulhlaup, lake level (Fig. 1). The field area experiences a semi-arid climate with annual precipitation of only 150 mm per year. Maximum temperatures, recorded approximately 30 km from the field site, at Kangerlussuaq can reach 24 °C, compared with an absolute minimum of −43 °C. The surface of Outlet Lake 1 is typically frozen from late October to June. The lack of snow cover at the ice sheet margin during the winter months prevents the occurrence of a spring snow melt flood or ‘nival event’. Non-jökulhlaup ‘normal’ discharge to Outlet Lake 1 is derived solely from ice-surface ablation within a small (<2 km^2) supraglacial catchment. Normal discharge drains across the western and central portions of the delta via two channelised streams which have a combined maximum discharge of ~10 m^3 s^{-1}. These streams occupy less than 10% of the

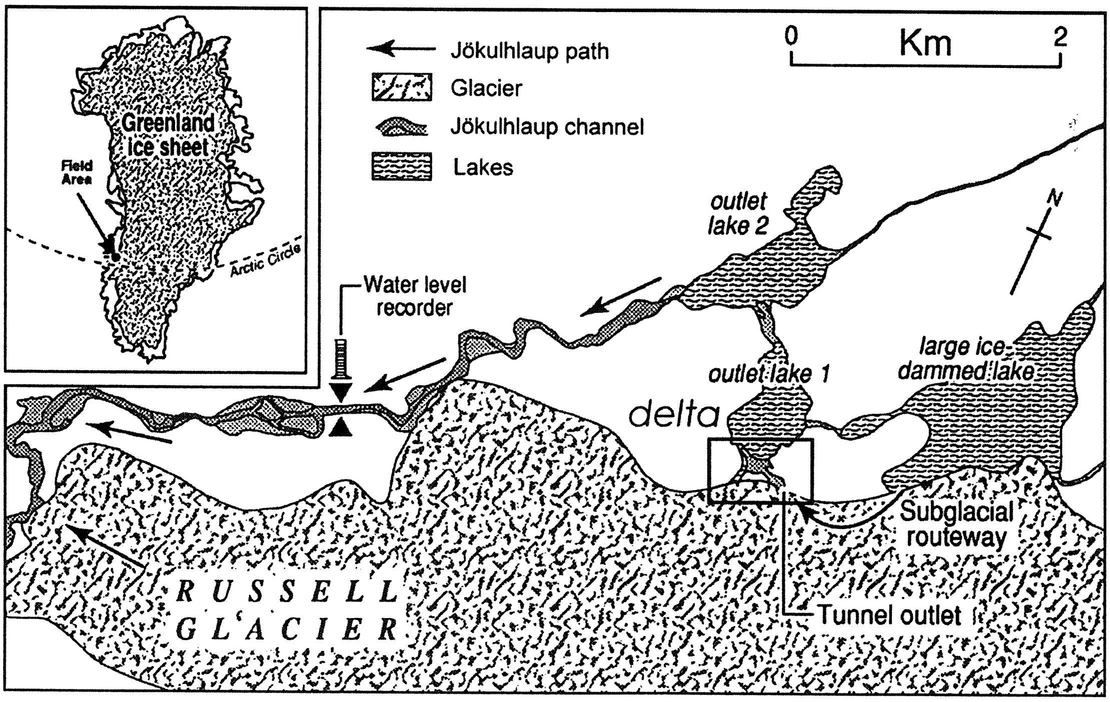


Fig. 1. Location of the field area within Greenland and the delta (boxed) in relation to Russell Glacier.

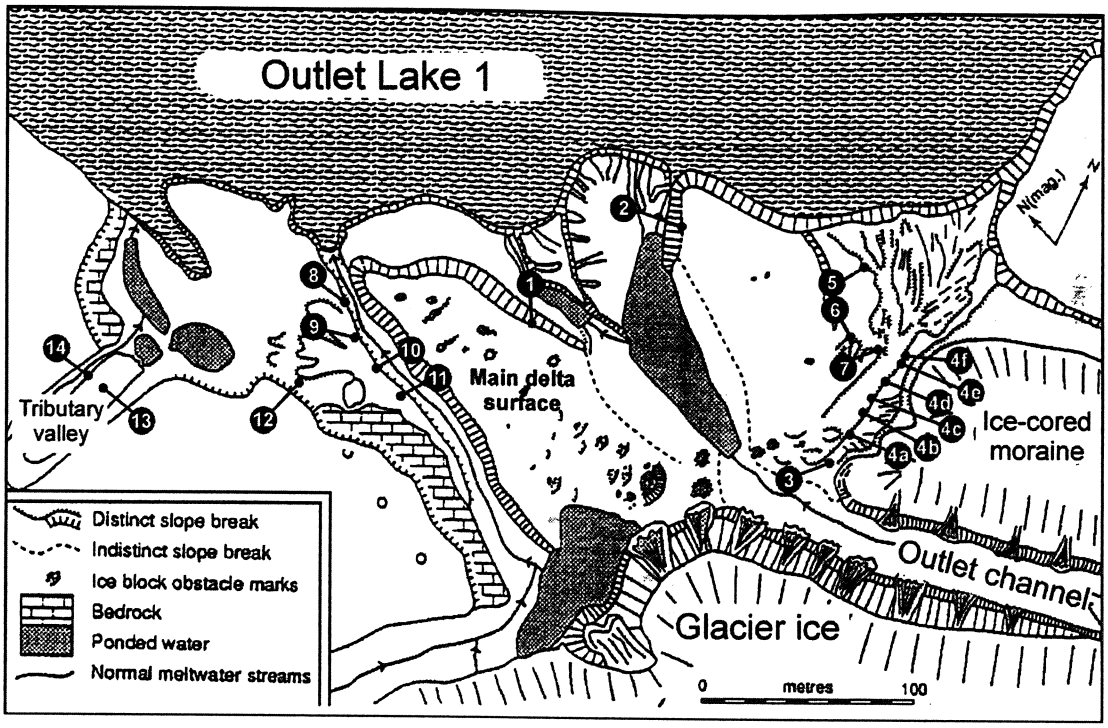
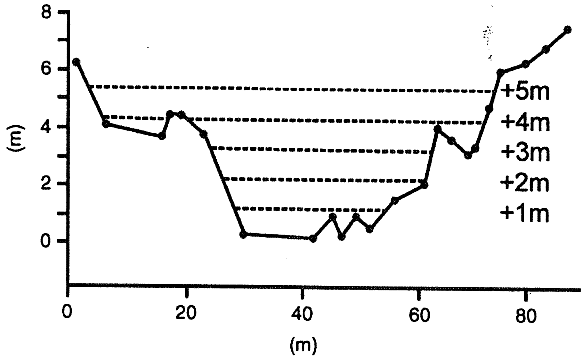


Fig. 2. Post-1987 geomorphological map of the delta showing the location of sections.

delta surface area and are incapable of reworking their boulder-lined beds.

The 1987 jökulhlaup emanated from an englacial tunnel before spreading out over the jökulhlaup delta and Outlet Lake 1 (Figs. 1 and 2) (Russell, 1989, 1993). Analysis of air photographs (1943, 1953, 1957, 1968 and 1985) suggest that the outlet channel was created and progressively eroded by jökulhlaups since 1943 and that the delta extended into Outlet Lake 1 between 1953 and 1987. Water exits Outlet Lake 1 northwards into Outlet Lake 2 by a bedrock spillway whose cross-sectional shape controls the level of the upper lake (Figs. 1 and 3).

Jökulhlaup discharge was gauged 5 km downstream of the delta and both proglacial lakes between July 17 and 19, 1987 (Russell, 1989, 1993) (Fig. 1). Post-jökulhlaup measurements obtained in 1987 and 1988 allowed reconstruction of peak flood flow conditions. Post-1987 jökulhlaup delta surface contours were interpolated from 690 points surveyed on the delta and surrounding deposits and 50 surveyed and plumbed lake depths (Fig. 4). Delta long- and cross-profiles were surveyed using a theodolite relative to local datum (Fig. 4). Peak jökulhlaup water surface gradients were obtained from surveys of wash limits identified immediately after the event and from comparison of photographs taken during and after the jökulhlaup. Morphological change as a result of the 1987 jökulhlaup was identified from the comparison of pre- and post-jökulhlaup survey data and photographs. Natural and hand excavated sediment sections were logged after the jökulhlaup. The axis of the 10 largest clasts was measured at ~100 locations



Outlet lake 1 level above normal (m)	Cross-sectional area of spillway (m^2)	Surface breadth (m)	Estimated discharge (m^3s^{-1})
1	19	25	52
2	50	35	188
3	88	39	413
4	142	63	668
5	211	70	1150
5.4	225	72	1250

Fig. 3. Surveyed cross-section of spillway between Outlet Lakes 1 and 2. Discharge figures were calculated for different spillway cross-sectional areas using Eq. (1).

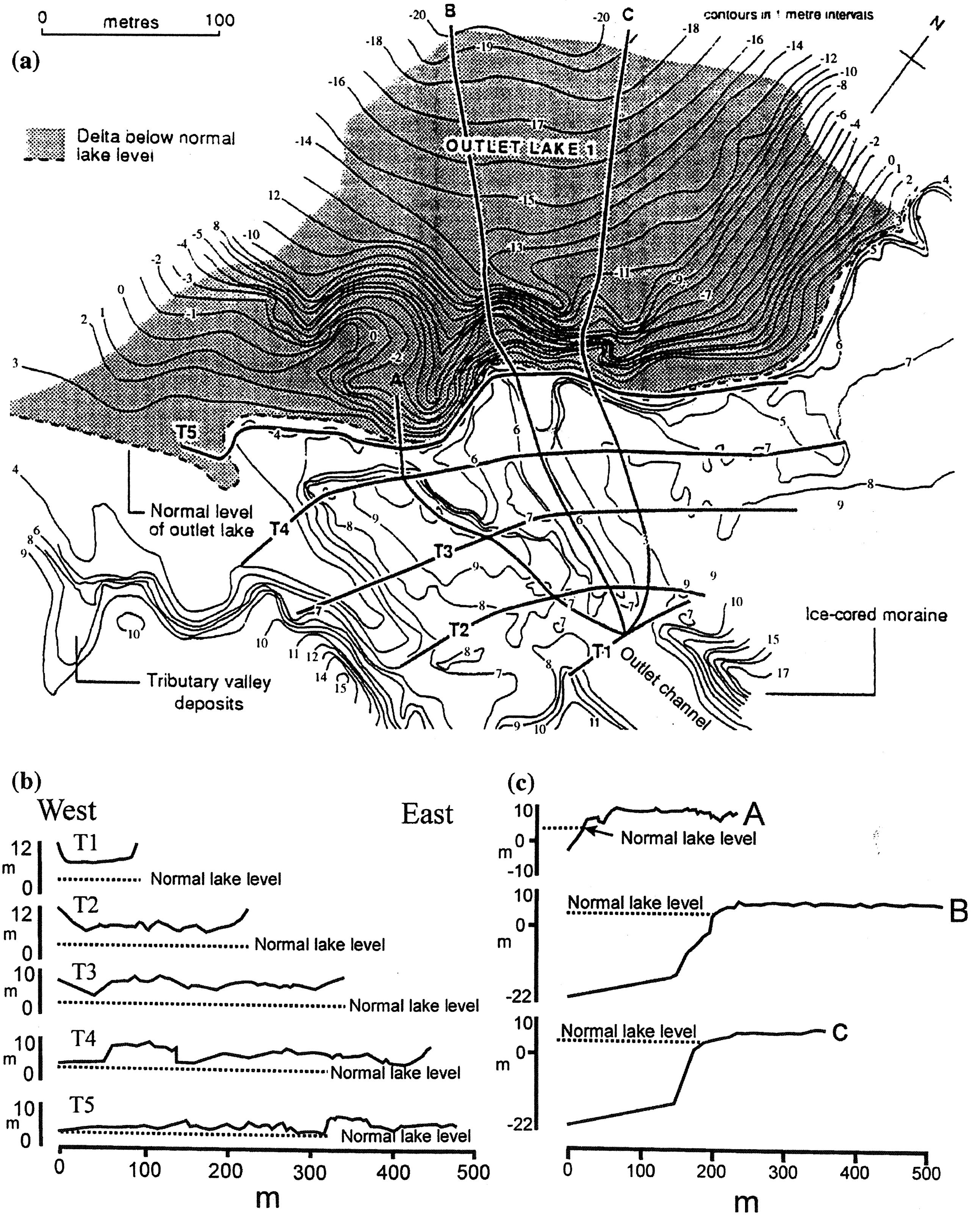


Fig. 4. (a) Post-jökulhlaup contour map of the delta and Outlet Lake 1 showing locations of delta cross-sections and long profiles. (b) Cross-sections of the post-jökulhlaup delta used to reconstruct hydraulic parameters in Table 2. (c) Long profiles of the post-jökulhlaup delta.

across the delta flow directions on the delta surface were measured using the orientation of ice block obstacle marks and clast imbrication (Fig. 5a). Sedimentary units were classified according to the lithofacies scheme adapted from Miall (1985) and Postma (1990) (Table 1).

3. 1987 jökulhlaup

The jökulhlaup was observed by the author between 1600 h and 1700 h on July 18, 1987 near flood peak. Flows were observed emanating from the outlet channel

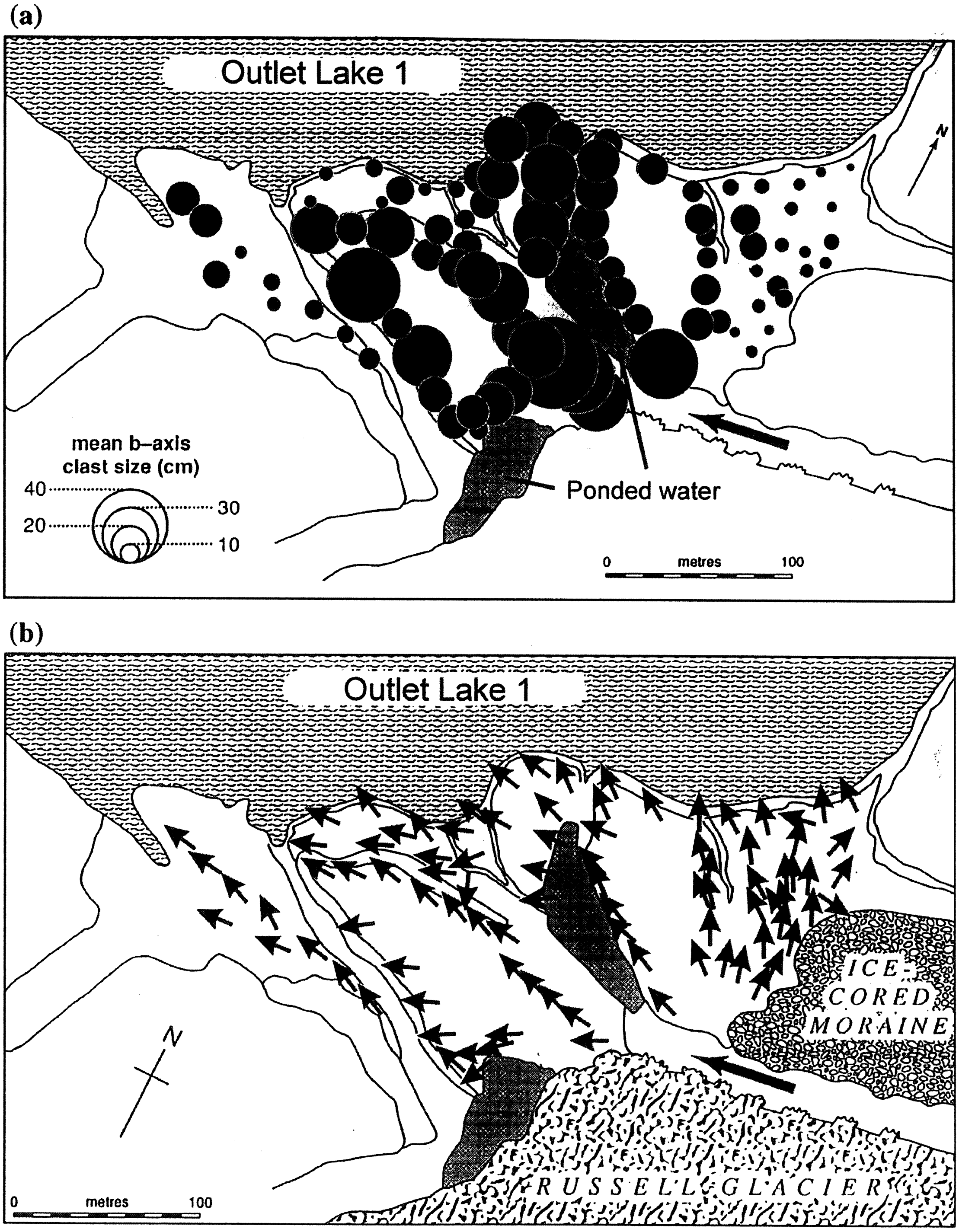


Fig. 5. (a) Delta surface maximum grain size based on the average of the 10 largest clasts sampled at each location. (b) Former flow directions over the jökulhlaup delta based on surface imbrication, clustering, ice block obstacle marks and channelisation.

Table 1
Lithofacies codes adapted from Miall (1985) and Postma (1990)

Lithofacies code	Sedimentary structure	Depositional process
Bm	Clast-supported, polymodal, boulders containing cluster bedforms and occasional imbricate structures	Deposition of bedload within a turbulent flow.
Bms	Massive, structures polymodal, matrix-supported cobble gravel	Rapid deposition of a range of grain sizes from a turbulent fluidal flow
Gm	Clast-supported, imbricated, cobble gravel	Deposition from bedload within a turbulent fluidal flow
Gms	Massive clast-supported, polymodal, clustered cobble-gravel	Rapid, deposition of a turbulent flows; clast transport by traction, saltation and suspension
Gh	Horizontally-bedded gravel	Deposition from fluidal flows; transport in turbulent suspension prior to late-stage traction transport
Gp	Planar cross-stratified gravel	Small-scale bar or delta units; migration of large bedforms
Sp	Planar cross-stratified sand	Bar front deposition; upper or lower flow regime sand waves
Sr	Ripple sand	Lower flow regime condition
Sh	Horizontally-bedded sand	Upper and lower flow regime plane bed condition

and spreading out over most of the delta surface (Fig. 6a). Only small clusters of coarse material could be seen above the peak water level (Fig. 6a). The flow separated into two main channels around an area of high ground immediately opposite the tunnel mouth (Fig. 6a). Numerous ice blocks, up to 5 m in diameter, released from the area around the tunnel outlet, travelled down the channel, grounding audibly (Russell, 1993, Fig. 4a and b). A jet of high velocity, turbulent, water observed entering the lake basin extended 100–150 m beyond the delta edge in front of each of the main channels (Fig. 6a and b). Small tributary embayments on the western margin of the main delta were submerged by relatively slackwater characterised by large-scale, eddy circulation within which numerous ice blocks were caught. The maximum level of Outlet Lake 1 during the jökulhlaup was 5.4 m above its normal level (Fig. 6b and c).

Comparison of pre- and post-flood survey data revealed that an estimated 4700 m^3 of sediment were supplied to the 1987 jökulhlaup from moraine talus slopes flanking the outlet channel, accounting for the bulk of sediment supplied to the delta. The sediment acquired represents a minimum jökulhlaup sediment concentration by volume of only 0.00015%. Smaller amounts of additional sediment may also have been supplied from subglacial erosion of active glacier ice, entrainment of supraglacial debris, and erosion of older, non-ice-cored, moraine ridges. Limited sediment availability is consistent with observations of turbulent flows in suggesting that the 1987 jökulhlaup was entirely fluidal in nature, dominated by turbulent flow processes.

Peak jökulhlaup discharge over the delta was determined by calculating the outflow discharge associated with different levels of Outlet Lake 1 (Russell, 1989, 1994). Instantaneous outflow discharge from the maximum level of Outlet Lake 1 was calculated from the following empirical relation:

$$Qw = (A^3 g/b)^{0.5} \tag{1}$$

(accurate to c. ±20%, Featherstone and Nalluri, 1984; Russell, 1994), where A = the cross-sectional area of the spillway (m^2), g = gravitational constant (9.81 m s^{-1}), and b = water surface width (m). This equation relies on the principle that water flows freely over the spillway crest as in the case of a weir. A discharge of 1300 m^3 s^{-1} obtained from the above method compared with 1080 m^3 s^{-1} gauged downstream (Russell, 1989, 1993). Discharge estimates provided here represent minimum values as inflow to Outlet Lake 1 from the jökulhlaup tunnel outlet must have been higher than outflow.

Stream powers and Froude numbers were calculated to quantify spatial variations in the flow regime and sediment transport capacity of peak jökulhlaup flows over the delta surface and to allow better comparison with other flood studies (e.g. Baker and Costa, 1987). Cross-sectionally averaged mean peak flow velocities were calculated from pre-1987 jökulhlaup channel cross-sectional area and discharge. Discharges of 1080 m^3 s^{-1} and 1300 m^3 s^{-1} were taken as representing minimum and maximum values for peak flood flows over the delta respectively (Table 2). The specific weight of water was assumed to be 9800 N m^3. Stream powers and Froude numbers were then calculated using equations presented by Baker and Costa (1987).

The results show a marked reduction in stream power, mean velocity and consequent transport capacity with distance from the tunnel mouth (Table 2). Flows in the outlet channel (TI) were capable of transporting material up to 10 cm in diameter as suspension load (Fig. 9 in Komar, 1988). However at the delta edge flows were capable only of transporting material of up to 1 cm diameter in suspension (Komar, 1988). The marked decline in sediment transport capacity suggests a large transfer of material from

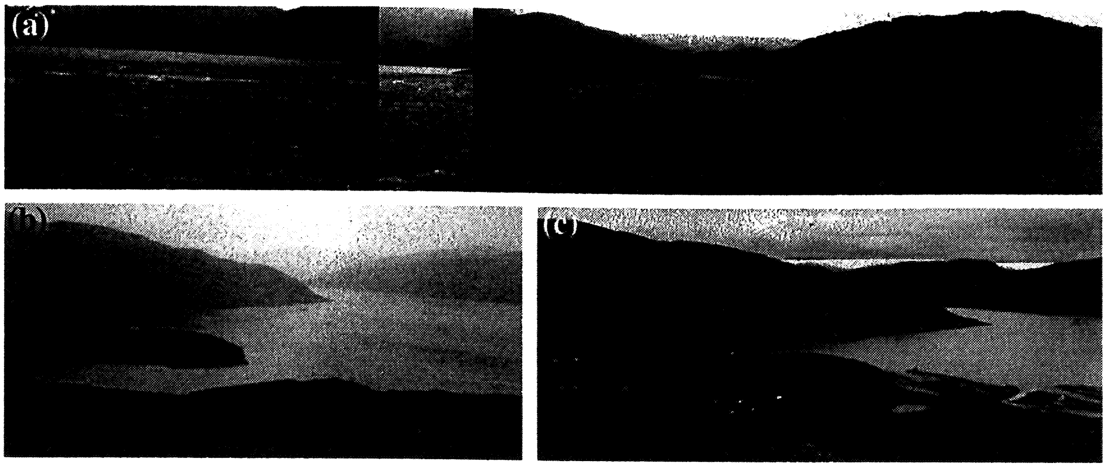


Fig. 6. (a) Peak of 1987 jökulhlaup over the delta and into temporarily raised Outlet Lake 1. The mouth of the outlet channel is ~60 m wide. Note the zone of turbulence extending beyond the delta front (middle-left). (b) View to the north of inundated tributary valley at the jökulhlaup peak. The jet of water extending from the western delta channel can be seen at the right of the picture. (c) Post-jökulhlaup view of the tributary valley showing stranded ice blocks and the light coloured slackwater deposit.

suspended load to bedload between TI and T5 over a distance of 150–200 m (Fig. 4b).

4. Delta topset morphology and change

Surface grain size decreased rapidly away from the tunnel mouth, both down-delta and laterally (Fig. 5a). The delta surface was visibly modified by the 1987 jökulhlaup (Fig. 7). The pre-1987 jökulhlaup delta top lay at an elevation of 7.9–8.8 m. The delta surface was in places lowered to 6.2–7.5 m by the 1987 jökulhlaup (Figs. 7 and 8). A shallow channel between the eastern and western delta segments was deepened and enlarged, supplying sediment to the delta front (Figs. 7 and 8). Areas of net deposition on the delta are located downstream of the main areas of erosion (Fig. 8). The delta front advanced 30 m into Outlet Lake 1 along a 150 m front (Fig. 7a–d). Deposition took the form of a higher-gradient, heavily-channelised, boulder lobe graded to lower lake levels of between 4 and 6 m (Fig. 7c and d). Adjacent parts of the high level delta surface were extended lakeward 5 to 20 m (Figs. 7c, d and 8). Lower lying parts of the delta were infilled by finer sand and gravel, changing post-jökulhlaup meltwater flow routing.

Marked grain size reduction with distance from the outlet channel corresponds closely with reductions in stream power and sediment transport capacity at peak 1987 jökulhlaup discharge. A general increase of delta-surface grain size with elevation suggests the highest delta surfaces are composed of high-stage deposits and that waning-stage deposition of finer-grained material on the jökulhlaup occurred in lower lying, peripheral channels. Coarse sediment found at low elevations constitutes a lag deposit produced by winnowing of finer-grained sediment. Delta morphological change during the 1987 jökulhlaup consisted mainly of modification and extension of pre-existing form. Although some delta front extension may have occurred during the jökulhlaup rising stage due to a greater availability of sediment, major waning stage jökulhlaup incision produced the greatest delta extension and morphological change during the 1987 jökulhlaup. Waning stage incision occurred as a response to the rapid fall in level of Outlet Lake 1.

Table 2
Hydrolic variables for the five delta cross-section as shown on Fig. 4 for the peak of the 1987 jökulhlaup

Cross-section	A (m^2)	Slope	Discharge ($m^3\ s^{-1}$)	Velocity ($m\ s^{-1}$)	Power ($W\ m^{-2}$)	Froude number
T1	250	0.0181	1300	5.2	2846	1.1
T2	579	0.0181	1300	2.2	1119	0.3
T3	685	0.0181	1300	1.9	705	0.4
T4	1075	0.0181	1300	1.2	541	0.1
T5	1613	0.0181	1300	0.8	490	0.04
T1	250	0.0181	1080	4.3	2365	0.8
T2	579	0.0181	1080	1.9	929	0.2
T3	685	0.0181	1080	1.6	585	0.3
T4	1075	0.0181	1080	1.0	449	0.07
T5	1612	0.0181	1080	0.7	407	0.03

A=Channel cross-sectional area, Slope=water surface gradient, Discharge, mean velocity for channel cross-section, Power=unit stream power, Froude number.

5. Delta-top sedimentary characteristics

Four lithofacies associations were identified on the delta surface. Lithofacies association A is composed

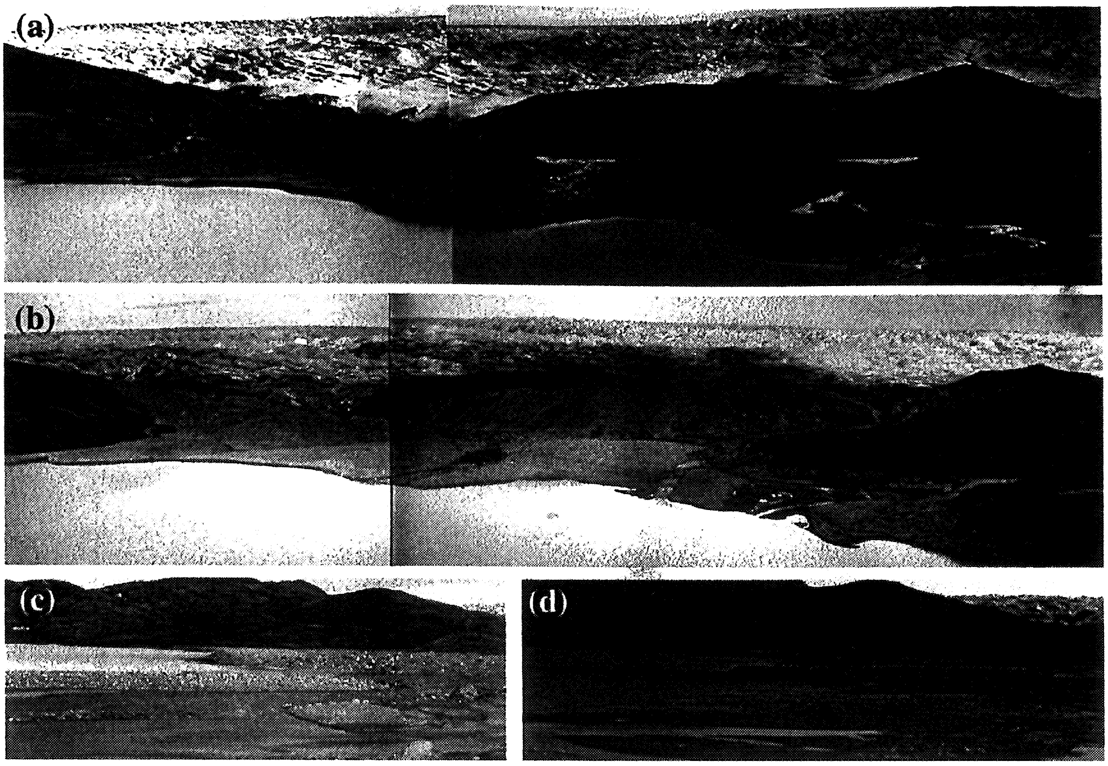


Fig. 7. (a) View to the south of jökulhlaup delta in October 1986. Note the extent of the delta front and the course of the small stream at the right. (b) Delta in July 1988. The delta front has advanced and locations of the small streams have changed. (c) and (d) Pre- and post-jökulhlaup views of delta front showing the lakeward advance of a low delta lobe and lowering of the high level delta surface.

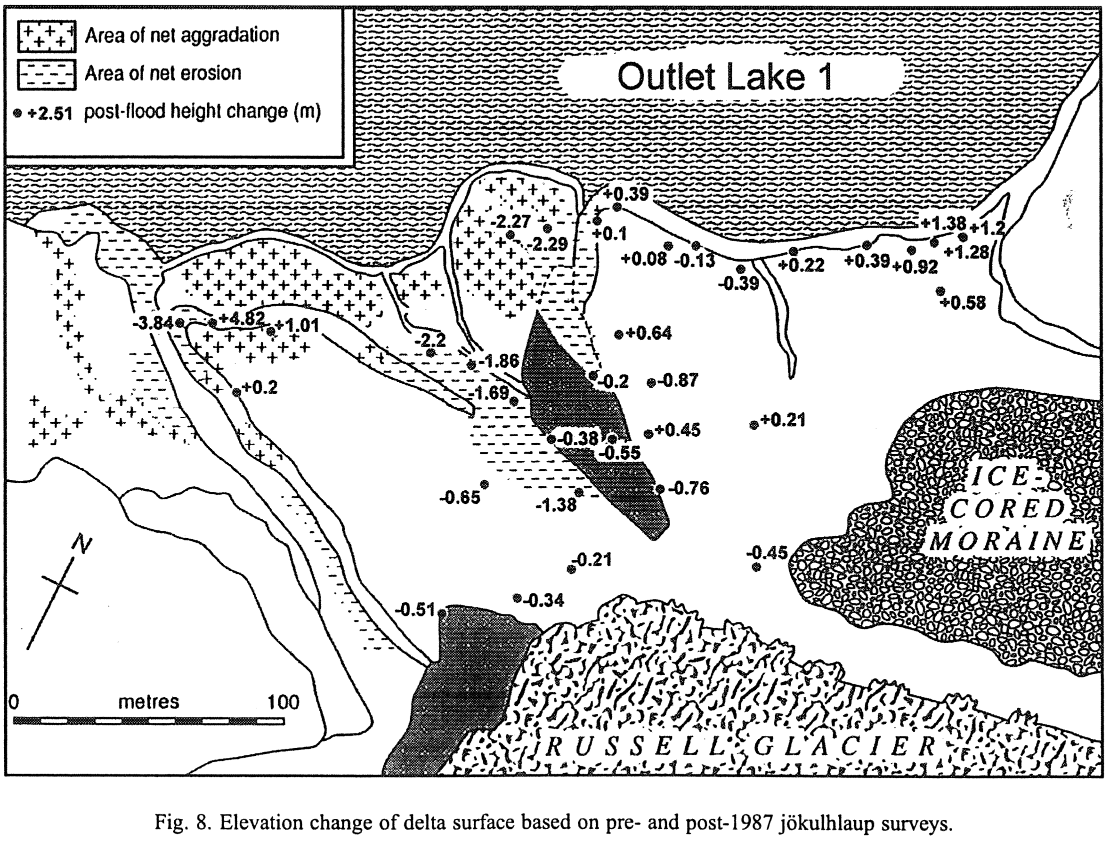


Fig. 8. Elevation change of delta surface based on pre- and post-1987 jökulhlaup surveys.

mainly of boulder-sized sediment and is located, at a distance of 150 m, directly in front of the jökulhlaup outlet. This association is found from, and immediately opposite, the main tunnel outlet and forms the core of the delta topset succession lithofacies association B comprises gravel-sized sediment and despite its proximity (<30 m) from the jökulhlaup outlet, occupies a relatively sheltered location behind an ice-cored moraine ridge. Lithofacies association C is located on the distal delta margins of the delta within at a distance of ~100 m from the jökulhlaup outlet. Lithofacies association D is located in a small tributary valley 300 m from the jökulhlaup outlet.

5.1. Lithofacies association A (Bm, Bms, Gm, Gms, Sh)

5.1.1. Description

Sections 1 and 2 (Fig. 2) contain boulders up to 2 m in diameter within a cohesionless matrix of poorly sorted coarse sand and gravel (Bm and Bms) (Fig. 9a and b). Occasional clusters of open-framework gravel lie between boulders (Fig. 9a and b). Some of the largest clasts show high-angle a-axis imbrication (Bm). However the predominance of the sub-angular to sub-rounded clasts makes imbrication much less well defined giving the deposit a structureless appearance (Bms). Bm and Bms facies are capped by a veneer of surface sand and gravel (Sh and Gm) (Fig. 9c).

5.1.2. Interpretation

Evidence of clustering and localised high-angled imbrication in the Bm, Bms, Gm and Gms facies of Sections 1 and 2 are consistent with sheet-like traction deposition from a highly turbulent flow (Brayshaw, 1985). Structureless, poorly sorted sediment between clusters was deposited in zones of proximal and distal flow separation associated with large boulders and cluster bedforms during a general phase of aggradation. Poor sorting of these deposits indicates rapid deposition from a flow with high sediment transport capacity. Most poorly imbricated, clast-supported gravel facies have a similar origin to the boulder facies described above.

The delta surface above Sections 1 and 2 lies at elevations of 7 to 10 m. Individual boulders and cluster bedforms on the main delta surface above Sections 1 and 2 were not moved during the 1987 jökulhlaup. A nearby pre-jökulhlaup wooden survey peg remained after the flood, further illustrating the lack of disturbance of the main delta surface. Massive boulder and gravel units are interpreted to have been deposited during the rising limb of an earlier jökulhlaup, with surface sand and gravels representing high stage sediment transport across the main delta surface during the 1987 jökulhlaup.

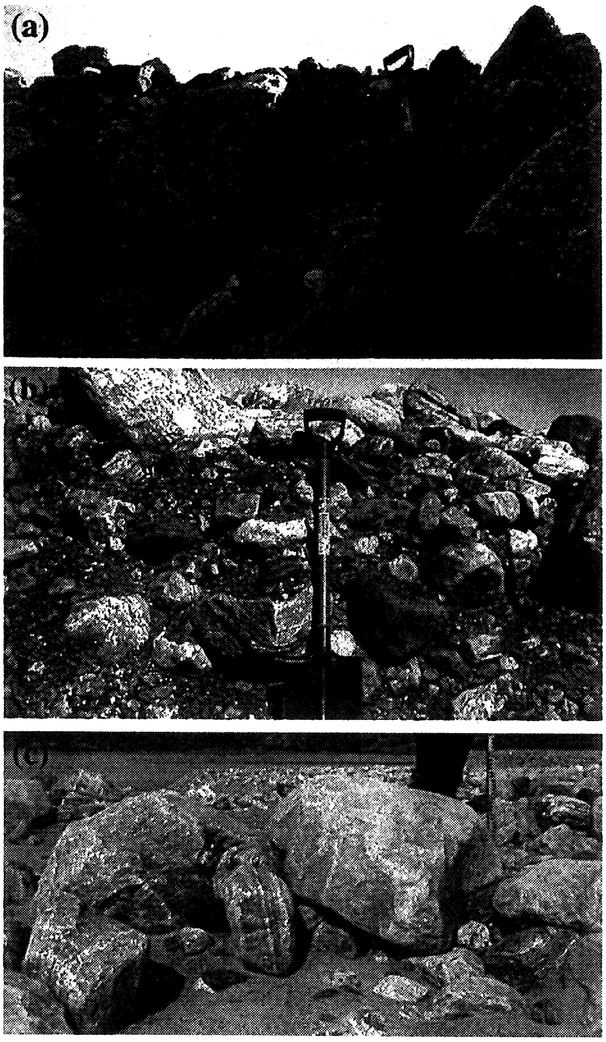


Fig. 9. (a) Section 1 on the west-central portion of the delta (shovel is 1 m in length). The coarsest sediment is clast-supported but is filled with pockets of loosely packed, finer, poorly sorted sediment. (b) Section 2 on the eastern main delta surface. The largest material is dominantly clast-supported, but contains pockets of finer, more poorly sorted, matrix material. (c) Boulder cluster bedform on the main delta surface above Section 1.

5.2. Lithofacies association B (upward-coarsening Gh)

5.2.1. Description

Lithofacies association B consists primarily of upward-coarsening, horizontally-bedded gravels (Gh) underlain by clast-supported cobble-gravel (Gm). Lithofacies association B is relatively fine-grained despite being only 20–80 m from the outlet channel. Most of these successions are located 90° from the main flow direction in the lee of an ice-cored moraine ridge and at

Fig. 10. Delta surface in the lee of ice-cored moraine ridges. Note the lobate form of the bar. Sections 3 and 4 (Fig. 2) show the internal characteristics of this deposit. Location of Section 4 marked by a broken line. Numerous perched delta fronts and chute and lobe structures can be seen in the background.

an elevation of 8–9 m (Figs. 2 and 10). Sections 3 and 4 (Fig. 2) are dominated by crude horizontally-bedded, poorly sorted, gravel (Gh), forming 0.6–1.0 m-thick upward-coarsening successions (Figs. 10, 11a, b, c, and 12a). Thinner (<0.1 m thick) open-framework and sandy-matrix-supported couplets occur within thicker upward-coarsening beds. The largest clasts within these exposures are located near the surface of the deposit, except at Section 3(a) where occasional out-sized clasts are present at depth and Section 4 where cobble-sized sediment occurs at a depth of 1.6–2.0 m (Figs. 11 and 12a). A single upward-coarsening unit of horizontally-bedded gravel is present at the base of Section 12 (Fig. 12).

5.2.2. Interpretation

The crude stratification and poor sorting of the Gh facies suggest rapid deposition from a turbulent flow. The presence of clast imbrication and cluster bedforms indicates sheet-like deposition of both large clasts (10 cm+) and coarse sand (Fig. 11b and c). The Gh facies at Sections 3 and 4 are very similar to horizontally-bedded gravel units attributed by Brennand (1994) to traction deposition within fluidal flows. Rare angular clasts at Section 4 were probably split from larger clasts during traction transport (Fig. 12a and b).

Lithofacies association B is interpreted to have been deposited during the rising stage of the 1987 jökulhlaup. Boulder-sized sediment travelling as bedload at flood peak would have had sufficient momentum to be carried directly from the outlet channel onto the main delta surface. All grain sizes present at Sections 3 and 4 (<~10 cm diameter) were however capable of having been transported in suspension within the nearby main outlet channel at flood peak (Table 2). Sudden stream power reduction of flow exiting the main outlet channel would have generated suspension rain-out in the lee of the ice-cored moraine ridge in the vicinity of Sections 3 and 4. Following rain-out, sediment was transported briefly as traction load before depositing a series of progressively coarser horizontal gravel sheets to form a lobate bar (Fig. 10). Cobble-sized material at the base of Section 4 is thought to represent the pre-1987 jökulhlaup surface (Fig. 12a and b). Waning-stage jökulhlaup flow incised this part of the delta, feeding reworked sediment lakeward, as evidenced by lobate deposits at the foot of successive chute channels (Fig. 10).

5.3. Lithofacies association C: delta topset lobate bar deposits (Gp, Sp, Sh, Gm)

5.3.1. Description

Planar cross-stratified gravel (Gp) is the most common lithofacies found at Sections 5–12 (Figs. 2 and 12a, c and d). It was identified in sections excavated into lobate deposits located in channelised and more distal parts of the delta surface. In some cases cobble- and boulder-sized clasts occur at the base (Section 10) or surface (Section 6) of the facies, obscuring cross-strata (Fig. 12d). Cross-stratified units at Sections 5 and 7–12 are commonly bounded by erosional contacts (Fig. 12a). A round silt ball, about 10 cm in diameter was noted in the lower cross-stratified unit of Section 7 (Fig. 12a). Cross-strata dip directions were generally towards Outlet Lake 1 at all sections. Cross-stratified sand (Sp) was only found at Sections 7, 8 and 11, grading into cross-stratified gravel units.

5.3.2. Interpretation

Planar cross-stratified sand and gravel units were deposited by downstream migration of sediment waves or bedforms. Near-surface, cross-strata dip directions at

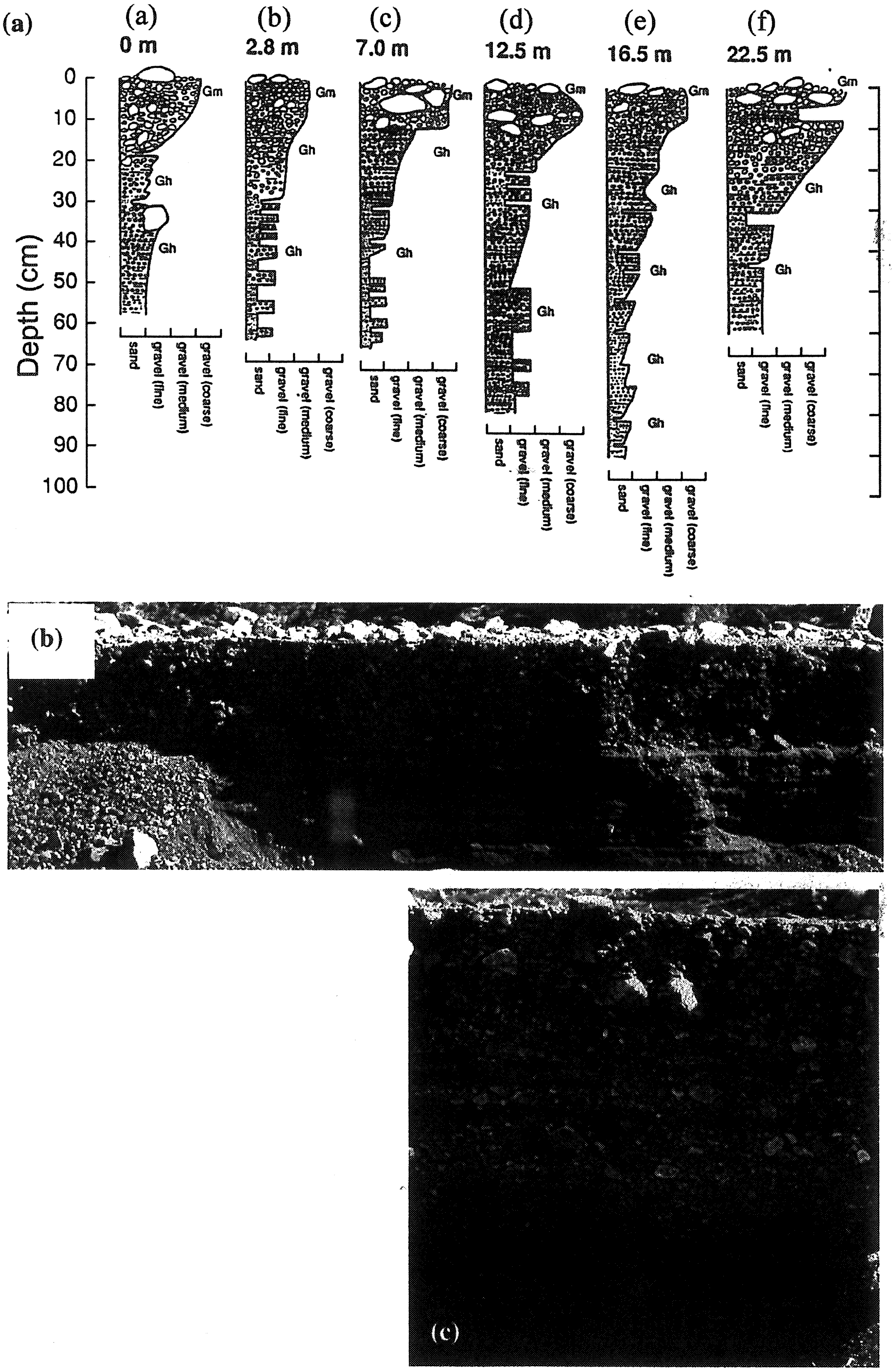


Fig. 11. (a) Log of Section 4. Note upward-coarsening succession of horizontally-bedded gravel. (b) Section 4c. (c) Section 4f. Note clast clustering and weak imbrication within the coarsening-upward succession. Largest clasts are 8–10 cm in diameter.

An upward-coarsening, cross-stratified unit occurs at an elevation of ~7.5 m at Section 6 within 100 m of the outlet channel (Fig. 12a). Nearby Section 7 is also dominated by an upward-coarsening succession of cross-stratified units. Rip-up clasts such as those in Section 7 are commonly found within rising-stage flood deposits. Rising-stage jökulhlaup deposition is also suggested by successions graded to the same surface. Progressively more vigorous flows are thought to have deposited increasingly coarse sediment into relatively deep water, creating the distinctive lobes prograding into rising proglacial lake levels. Exhaustion of available sediment may explain why the lobes did not prograde as far as the delta edge. The erosional contact between Gp units in Section 7 may represent a reactivation surface, suggesting that this lobate deposit was only extended by the 1987 jökulhlaup.

Upward-coarsening successions are found at Sections 8, 9, 11 and 12, whereas only Section 10 exhibits an upward-fining trend. Sections 8, 9, 11 and 12 were all excavated into a lobate bar unit at a low elevation (<4 m) within the main western channel. Despite the close proximity of these sections to each other, no obvious correlations can be made among them. The upward coarsening basal units of Sections 8–12 are consistent with deposition during a high-energy rising-stage jökulhlaup flow (Fig. 12a). Fining-upward successions at Section 10 may represent the progressive reduction of flow strength either on the falling flood limb or on initiation of backwater conditions at the jökulhlaup peak. Deposition was followed by waning-stage incision of the lobate bar unit.

Finer-grained Gm units less than 0.4 m thickness are found at Sections 4–6 and 8–11 (Fig. 12a). Gm units at Sections 6, 8, 9 and 11 cap distinctive coarsening-upward successions, whereas those at Sections 4, 5 and 10 are capped by fining-upward successions (Fig. 12a). Gm facies at Sections 6, 8 and 10 are associated with cross-stratified gravel, suggesting deposition on a bedform or sediment wave.

5.4. Lithofacies association D: distal sediments (upward coarsening Sh, Sr)

5.4.1. Description

Sections 13 and 14 are located in a small tributary valley (Fig. 2) and consist of coarsening-upward medium-coarse sand. Section 13 displays crude horizontal laminae (Sh), which become better defined upwards. Small-scale ripple trough cross-stratification (Sr) is locally present at this section. Section 14 contains horizontally-bedded sand (Sh) and local cross-stratified sand (Sr). Laminations became coarser 0.35 m from the top of the pit, with occasional clasts of -1 ϕ. Very angular gravel platy clasts lie on the deposit surface, with clast size decreasing with distance up the tributary channel.

5.4.2. Interpretation

Lithofacies association D occurs in a small tributary valley 300 m from the outlet channel at an elevation of 6 to 7.7 m (Fig. 6b and c). The deposit is thickest at the tributary mouth, and has the general morphology of slackwater deposits (Baker and Kochel, 1988a,b). Horizontally-bedded and ripple-laminatedsands (Sh and Sr) at Sections 13 and 14 indicate deposition within a relatively low energy flow. Very angular bedrock flakes on the surface of the slackwater deposit are likely to have been chipped from larger clasts transpofied in bedload across the delta. The bedrock flakes are platy in shape and would therefore have travelled relatively easily in suspension. At its peak, the 1987 jökulhlaup submerged the slackwater succession to a minimum depth of 1.36 m (Fig. 6b). Sparse vegetation cover on the slackwater sediment surface suggests that they were deposited by several jökulhlaups. The upwards-coarsening trend of lithofacies association D reflects the increasingly proximal location of the tributary valley, caused by progressive lakeward extension of the delta.

6. Discussion

6.1. jökulhlaup impact on delta

The 1987 jökulhlaup had relatively little impact on the morphology of the delta surface with the most significant changes consisting of delta front extension and subsequent incision of the main delta surface. It is clear that most delta topset deposits predate the 1987 jökulhlaup. Only discharges of $\sim 10^3$ m^3 s^{-1} can account for sufficient proglacial lake level rise to account for delta topset deposition. Within this catchment jökulhlaups provide the only means of attaining such high discharges. Delta surface incision was accompanied the deposition of a new central delta lobe (Figs. 7 and 8). The jökulhlaup caused net aggradation, mainly on the delta margins where relatively slackwater conditions prevailed. A staircase of incised delta lobes reflects rising-stage deposition and subsequent waning-stage erosion. Chutes and lobes at different heights on abandoned jökulhlaup deltas in Norway and Sweden have similarly been attributed to waning-stage erosion with simultaneous, short-lived lobe deposition (Shakesby, 1985; Elfström, 1987). Sedimentary evidence indicates that lobate forms on

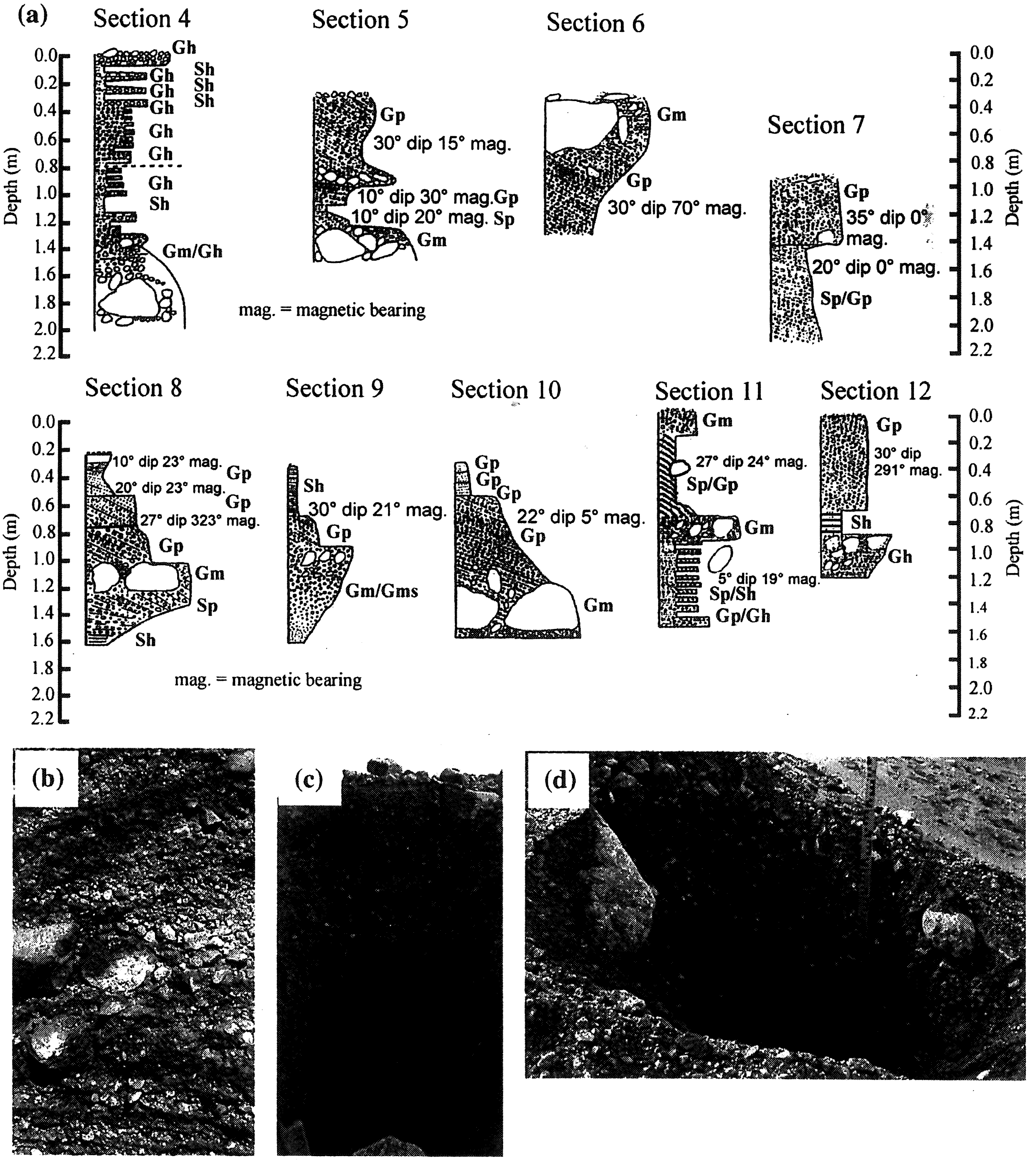


Fig. 12. (a) Logs of Sections 4–7 (top) and 8–12 (bottom). Poorly-sorted, crudely-bedded sediment at Section 4. (b) Clast-supported, open-framework layers alternate with finer, matrix-supported layers. The 8 cm wide, round clast in the centre of the photo was fractured during transport and prior to deposition. (c) Cross-stratified gravel at Section 7. (d) Cross-stratified gravel at Section 6 containing boulders.

Sections 5, 6, 7 and 12 are consistent with former flow directions inferred from the morphology of lobate sediment waves. Stacked cross-stratified units at Sections 8 and 10 indicate the passage of several sediment waves at these sites as part of an aggrading succession. Deposition of poorly sorted, coarse sand and gravel took place simultaneously with cobbles and boulders within foreset beds at Sections 5, 6, 8 and 9 (Fig. 12a).

the Greenland delta are associated with both the rising and falling stages of the jökulhlaup (Figs. 2 and 10). Incision of the main delta surface is attributed to headward erosion of channels as they adjusted to rapid base-level lowering. Zones of net delta surface aggradation coincide with pre-flood kettles related to the in situ meltout of isolated ice blocks (Fay, 2002).

The development of hummocky terrain on the Greenland delta during the 1987 jökulhlaup has been attributed to flows around stranded ice blocks (Russell, 1993). Flow acceleration and vorticity on the stoss-side of ice blocks produced erosional hollows on the delta surface (Russell, 1993). Russell (1993) further suggested that flow around stranded ice blocks could initiate waning stage bar incision. However, only evidence of stoss-side scour was observed around ice blocks, indicating that the ice blocks were not responsible for incision of the delta surface. The post-1987 delta surface displays more ice block obstacle marks than the pre-flood delta, suggesting the release of a greater number of large blocks of ice during the 1987 jökulhlaup than during the 1984 jökulhlaup.

The sedimentary signature of the 1987 jökulhlaup differs across the delta surface. Bar aggradation occurred by means of accretion of a series of gravel sheets (Gh) within the proximal part of the delta surface subject to slackwater conditions. These deposits are up to 1.5 m thick. Finer grained sediments (Sr, Sh) occurred to thicknesses of $<\sim 10$ cm in distal slackwater locations. Small-scale, <1 m high, delta lobes (Gp, Sp) formed in pre-existing channels and proximal slackwater locations.

6.2. Distribution of lithofacies associations on delta

The four lithofacies associations identified in the delta topset are the depositional product of multiple jökulhlaups, and each records a distinctive depositional environment.

Lithofacies association A (Bm, Bms, Gm, Gms, Sh) represents deposition from the highest energy flow conditions. It corresponds with the area of coarsest sediment directly in front of the main outlet channel. The boulder-sized sediment may have been deposited during several jökulhlaups, but sedimentary evidence of this is lacking. The first jökulhlaup to enter the lake basin would have had the greatest depositional impact due to the greater availability of sediment. It is therefore possible that the bulk of lithofacies association A was deposited during a single jökulhlaup and that subsequent jökulhlaups only reworked the delta surface. Deposition of a large volume of coarse-grained sediment at a channel expansion produces an expansion bar, which is commonly found within bedrock fluvial systems (Baker, 1984). Lithofacies association A is the core of the delta topset around which the other lithofacies associations are arranged (Fig. 13a and b).

Proximal slackwater facies (lithofacies association B) are located on the flanks of lithofacies association A and are associated with lower energy deposition. They are, however still intimately linked to sediment transport processes operating within the nearby outlet channel. Sediments of lithofacies association B are likely to be transported as suspended load from the outlet channel, and thus probably record multiple jökulhlaups. Although there is evidence of some waning stage erosion, the bulk of the lithofacies association B is preserved. Rapid reduction in waning-stage flow shut-off required for sediment preservation is facilitated by the location of the lithofacies in the lee of a moraine ridge and high above the main outlet channel (Fig. 13a and b).

Lithofacies association C consists of lobate bars or delta front deposits at distal locations, typically channels at the margins of lithofacies association A and at the distal margins of lithofacies association B. These deposits represent prograding bedforms: first; deposited diachronously from distal to proximal delta top locations as lake level rose; and second, as a series of lobes in association with waning-stage incision of rising-stage deposits as lake levels fell. Lithofacies association C is the product of complex interactions between base-level change and sediment deposition and reworking during a jökulhlaup (Fig. 13a and b).

Lithofacies association D is located beyond the main delta surface within a small tributary valley and is a product of low-energy fluvial processes consistent with its slackwater origin. Sediment was transported to this location as suspended load during repeated jökulhlaups. The coarsening-upward nature of the deposit is explained by the increasingly proximal location of the site as the delta front advanced during successive jökulhlaups. The lithofacies are likely to record all jökulhlaups at this site, although identification of individual jökulhlaups was not possible (Fig. 13a).

The arrangement of each lithofacies association and its relative abundance depend on local factors such as the orientation and dimensions of the outlet channel, geometry of the proglacial lake basin and the relationship between outlet channel geometry and proglacial lake basin morphology. An ice-walled outlet channel may also change its geometry over time, widening and lengthening during successive jökulhlaups due to the processes of thermal erosion and tunnel collapse.

6.3. Controls on jökulhlaup impact on delta morphology and sedimentology

Numerous factors influence the morphology and sedimentology of deltas subject to jökulhlaups. Jökulhlaup magnitude and form affect sediment entrainment and the timing of erosion and deposition (Maizels, 1989a,b). Within a confined proglacial lake basin, peak discharge and the cross-sectional area of the outlet spillway control the amount and rate of base level rise

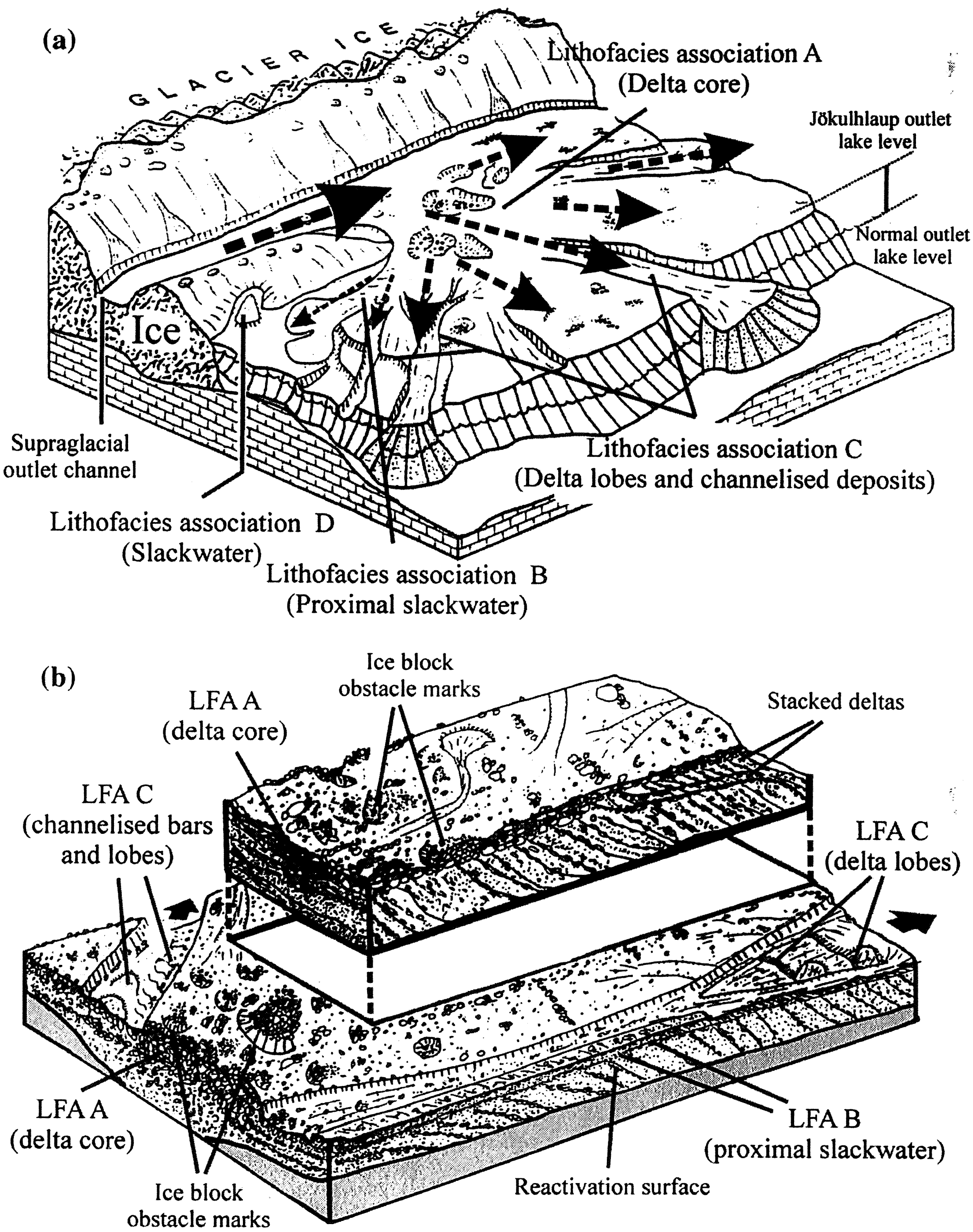


Fig. 13. (a) Schematic diagram showing the morphology of an ice-contact jökulhlaup delta. Locations of the lithofacies associations identified in this study (A–D) are indicated. (b) Block diagram showing idealised jökulhlaup delta surface and sub-surface sedimentology together with the lithofacies associations (A–C) identified in this study. Thickness of the foreset beds (diagram showing idealised jökulhlaup delta surface and sub-surface sedimentology together) is dependent on the depth of the proglacial lake. Topset thickness is determined by the amount of jökulhlaup-induced lake-level rise.

during a jökulhlaup (Shakesby, 1985; Kehew, 1993; Russell, 1994) (Fig. 14). Backwater effects during the rising flow stage progressively reduce stream power on the delta top and initiate complex spatial and temporal patterns of erosion and deposition. Variations in the cross-sectional area of the spillway or rate of discharge may alter the magnitude and duration of proglacial lake level rise (Fig. 14). The occurrence of temporarily raised lake levels during jökulhlaups increases the potential of waning stage flows to incise. The depth and surface area of the proglacial lake basin control rates of delta front advance and delta surface area (Fig. 14).

Delta morphological change influences the routing of sediment across the delta. The central portion of the pre-1987 jökulhlaup delta, enlarged by the 1987 event, is a major sediment pathway. A subsequent jökulhlaup of equivalent magnitude will not be able to rework or deposit sediment on the higher delta surfaces, as flows will be more channelised. Future fluctuations in sediment supply and routing mean that repeated jökulhlaups will have different erosional and depositional capacities. Differences in peak discharge between successive jökulhlaups will result in differences in temporary lake levels, which will increase the potential for jökulhlaup impact. Delta surface erosion will allow more sediment to be delivered to the delta front via channels.

The Greenland jökulhlaup delta has high relief amplitude of ~28 m compared with jökulhlaup deltas reported elsewhere (Shakesby, 1985; Elfström, 1987). In Greenland the presence of a relatively deep lake basin combined with low sediment supply rates have resulted in low rates of delta progradation.

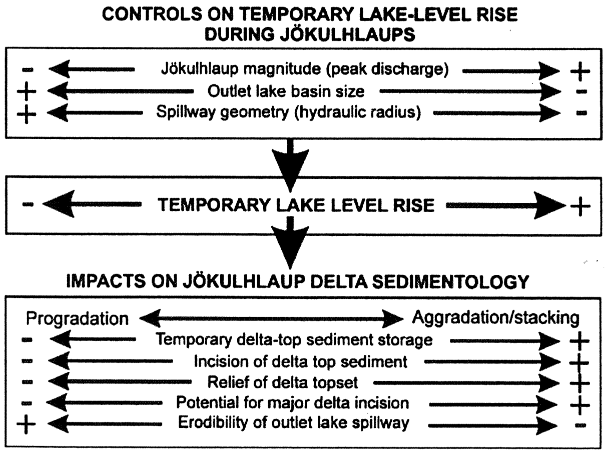


Fig. 14. Model of the controls on temporary lake-level rise during jökulhlaup (top box) and the consequent impacts on delta sedimentology (lower box).

7. Conclusions

The Greenland jökulhlaup delta has been formed solely by jökulhlaups spanning only ~0.4% of the melt season and representing 130 times the discharges of normal, ablation-generated flows. This study demonstrates that one or more high-magnitude events of short duration may dominate the sedimentary record of an ice-contact delta. A paucity of available sediment at the western margin of the Greenland ice sheet limits the impact of each jökulhlaup and facilitates high rates of delta-surface reworking.

Temporary lake level rise during jökulhlaups increases the amount of delta top sediment reworking and increases the thickness of the topset compared to non-jökulhlaup deltas. Slackwater conditions during temporary proglacial lake high stands result in stacking of lobate bars and small delta fronts in large erosional channels. The sedimentary expression of jökulhlaup-induced lake level variations may be confused with that of base-level change over longer timescales (e.g. Bowman, 1990; Martini, 1990; Mastalerz, 1990; Nemec, 1990; Postma, 1990). This study has demonstrated that numerous cycles of deposition and erosion can be recorded in the topsets of jökulhlaup deltas. The record of the largest jökulhlaups will be recorded within the delta core, which may have the form of an expansion bar. Multiple jökulhlaups will be recorded in lower energy deposition on the flanks of the delta core, and within slackwater environments within the proglacial lake basin.

Jökulhlaup deltas have distinctive landforms and deposits that aid in their identification within the sedimentary record. Jökulhlaup deltas form where flood waters enter ice-dammed lakes, proglacial lakes, and lakes within the downstream river system. Ice-contact jökulhlaup deltas are likely to be found in association with preferred subglacial drainage routes such as rock-cut meltwater channels (Shakesby, 1985). Linear tunnel deposits or subaerial channel deposits may lead to the apex of jökulhlaup deltas (Russell, 1994). A delta may form wholly or partially on ice, and melt of buried ice will produce large kettle-holes. A steep ice-contact face may be left during glacier retreat. Unless there is a glacier readvance, the delta will remain intact and subject to subaerial denudation. Delta surfaces composed of large clast size provide the greatest resistance to subaerial reworking processes and thereby stand the greatest chance of having their morphology preserved. Bedforms such as ice block grounding structures may provide flow direction indicators even when clast imbrication has been destroyed.

Examination of the relation between jökulhlaup flow processes and delta sedimentology has provided insights into the controls on jökulhlaup delta formation. This study contributes towards a better interpretation of jökulhlaup deposits in the sedimentary record of former ice-marginal areas. Improved reconstruction of palaeoflow magnitude and frequency is important for understanding the dynamics of former ice masses.

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References

Baker, V.R., 1984. Flood sedimentation in bedrock fluvial systems. In: Koster, E.H., Steel, R.J. (Eds.), Sedimentology of Gravels and Conglomerates. Canadian Society of Petroleum Geologists Memoirs, vol. 10, pp. 87–98.

Baker, V.R., Costa, J.E., 1987. Flood power. In: Mayer, L., Nash, D. (Eds.), Catastrophic Flooding. Allen & Unwin, pp. 1–25.

Baker, V.R., Kochel, R.C., 1988a. Paleoflood analysis using slackwater deposits. In: Baker, V.R., Kochel, R.C., Patton, P.C. (Eds.), Flood Geomorphology. John Wiley & Sons, Ltd, pp. 357–376.

Baker, V.R., Kochel, R.C., 1988b. Flood sedimentation in bedrock fluvial systems. In: Baker, V.R., Kochel, R.C., Patton, P.C. (Eds.), Flood Geomorphology. John Wiley & Sons, Ltd, pp. 123–137.

Bowman, D., 1990. Climatically triggered Gilbert-type lacustrine fan deltas. The Dead Sea area, Israel. In: Colella, A., Prior, D.B. (Eds.), Coarse-Grained Deltas. Blackwell, pp. 273–280.

Brayshaw, A.C., 1985. Bed microtopography and entrainment thresholds in gravel bed rivers. Geological Society of America Bulletin 96, 218–223.

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

Elfström, A., 1983. The Baldakatj boulder delta, Lapland, Northern Sweden. Geografiska Annaler 65A, 201–225.

Elfström, A., 1987. Large boulder deposits and catastrophic floods. Geografiska Annaler 69A, 101–121.

Elfström, A., 1988. Late Glacial hydrology of the upper Pite River valley, Swedish Lapland. Geografiska Annaler 70A, 99–123.

Elfström, A., Rossbacher, L., 1985. Erosional remnants in the Baldakatj area, Lapland, Northern Sweden. Geografiska Annaler 67A, 167–176.

Fay, H., 2002. Formation of kettle holes following a glacial outburst flood (jökulhlaup), Skeiðarársandur, southern Iceland. In: Snorasson, A., Finnsdottir, H.P., Moss, M. (Eds.), The Extremes of the Extremes: Extraordinary Floods, Proceedings of a Symposium, Reykjavik, Iceland, July 2000. International Association of Hydrological Sciences Publication, vol. 271, pp. 205–210.

Featherstone, R.E., Nalluri, C., 1984. Civil Engineering Hydraulics. Billings & Son Ltd., Worcester.

Gordon, J.E., 1986. Correspondence concerning glacial lake drainage near Søndre Strømfjord, West Greenland. Journal of Glaciology 32, 304.

Johansson, P., 1988. Deglaciation pattern and ice-dammed lakes along the Saariselkä mountain range in northeastern Finland. Boreas 17, 541–552.

Kehew, A.E., 1993. Glacial-lake outburst erosion of the Grand Valley, Michigan, and impacts on glacial lakes in the Michigan basin. Quaternary Research 39, 36–44.

Komar, P.D., 1988. Sediment transport by floods. In: Baker, V.R., Kochel, R.C., Patton, P.C. (Eds.), Flood Geomorphology. John Wiley & Sons, Ltd, pp. 97–111.

Maizels, J.K., 1989a. Sedimentology, paleoflow dynamics and flood history of jökulhlaup deposits: paleohydrology of Holocene sediment sequences in southern Iceland sandur deposits. Journal of Sedimentary Petrology 59, 204–223.

Maizels, J.K., 1989b. Sedimentology and palaeohydrology of Holocene flood deposits in front of a jökulhlaup glacier, South Iceland. In: Bevan, K., Carling, P. (Eds.), Floods, Hydrological, Sedimentological and Geomorphological Implications: an Overview. John Wiley & Sons, Ltd, pp. 239–253.

Martini, I.P., 1990. Pleistocene glacial deltas in southern Ontario, Canada. In: Colella, A., Prior, D.B. (Eds.), Coarse-Grained Deltas. Special Publication of the International Association of Sedimentologists, vol. 10, pp. 281–295.

Mastalerz, K., 1990. Diurnally and seasonally controlled sedimentation on a Glaciolacustrine foreset slope: an example from the Pleistocene of eastern Poland. In: Colella, A., Prior, D.B. (Eds.), Coarse-Grained Deltas. Special Publication of the International Association of Sedimentologists, vol. 10, pp. 297–309.

Melander, O., 1977. Geomorfologiska kartbladet 30J Rensjon. Statens Naturvårdsverk PM 858.

Miall, A.D., 1985. Giaciofluvial transport and sedimentation. In: Eyles, N. (Ed.), Glacial Geology: an Introduction for Engineers and Earth Scientists. Pergammon, Oxford, pp. 168–184.

Nemec, W., 1990. Aspects of sediment movement on steep delta slopes. In: Colella, A., Prior, D.B. (Eds.), Coarse-Grained Deltas. Special Publication of the International Association of Sedimentologists, vol. 10, pp. 29–73.

Postma, G., 1990. Depositional architecture and facies of river and fan deltas: a synthesis. In: Colella, A., Prior, D.B. (Eds.), Coarse-Grained Deltas. Special Publication of the International Association of Sedimentologists, vol. 10, pp. 13–27.

Rossbacher, L.A., Rhodes, D.D., 1984. Catastrophic flood features in Swedish Lapland as a terrestrial analog for Martian channel fans. NASA Technical Memorandum 87563, 319–321.

Rossbacher, L.A., Rhodes, D.D., 1987. Planetary analysis for geomorphic features produced by catastrophic flooding. In: Mayer, L., Nash, D. (Eds.), Catastrophic Flooding. Allen & Unwin, pp. 289–305.

Russell, A.J., 1989. A comparison of two recent jökulhlaups from an ice-dammed lakem, Søndre Strømfjord, West Greenland. Journal of Glaciology 35, 157–162.

Russell, A.J., 1993. Obstacle marks produced by flows around stranded ice blocks during a jökulhlaup in West Greenland. Sedimentology 40, 1091–1113.

Russell, A.J., 1994. Subglacial jökulhlaup deposition, Jotunheimen, Norway. Sedimentary Geology 91, 1–14.

Scholz, H., Schreiner, B., Funk, H., 1988. Der einfluss von Gletscherlaufen auf die schmelzwasserablagerungen des Russell-Gletschers bei Sondre Stromfjord (West Gronland). Zeitschrift für Gletscherkunde und Glazialgeologie 24, 55–74.

Shakesby, R.A., 1985. Geomorphological effects of jökulhlaups and ice-dammed lakes, Jotunheimen, Norway. Norsk Geografiska Tiddskrift 39, 1–16.

Sugden, D.E., Clapperton, C.M., Knight, P.G., 1985. A jökulhlaup near Søndre Strømfjord, West Greenland, and some effects on the Ice sheet margin. Journal of Glaciology 31, 366–368.

Ulfstedt, A.C., 1979. Geomorfologiska kartbladet 27G Sulitelma. Statens Naturvårdsverk PM 1230.

Williams, G.P., 1983. Paleohydrological methods and some examples from Swedish fluvial environments, I – cobble and boulder deposits. Geografiska Annaler 65A, 227–243.