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Repeated jökulhlaups at Russell Glacier, Kangerlussuaq, West Greenland: a reappraisal of the jökulhlaup cycle concept

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ABSTRACT

Two recent jökulhlaups from an ice-dammed lake on the northern margin of Russell Glacier in West Greenland mark the onset of a renewed jökulhlaup cycle after 20 years of stability. We present a novel record of successive ice-dammed lake (IDL) drainage events and associated ice margin dynamics spanning ~ 25 years, which is enabled via robust calculations of lake volumes and peak discharges based on intensive field surveys, and high spatial resolution orthophotographs of the lake basin and ice margin. These data enable the first field-based examination of controls on jökulhlaup magnitude and frequency for this system and provide a test of prevailing models relating fluctuations of glacier margin position to jökulhlaup cyclicity. We find that jökulhlaup magnitude is not entirely controlled by declining lake surface elevations due to a reduced threshold of flotation from vertically thinning ice, but also by horizontal ice margin fluctuations encroaching into, or withdrawing from the lake basin. Importantly, such ice margin dynamics not only are a response to, but also a cause of, ice-dammed lake formation and fluctuations. Russell Glacier jökulhlaups have a much higher peak discharge than predicted by the Clague-Mathews relationship, due to an unusually short englacial/subglacial routeway and the presence of a thin ice dam that permits incomplete sealing of jökulhlaup conduits between lake drainages. We demonstrate that the passage of jökulhlaups through an inter-linked system of proglacial bedrock basins results in a significant attenuation of peak discharge downstream. Improved understanding of jökulhlaup dynamics requires accurate information about ice-dammed lake volume and ice-proximal jökulhlaup discharge. The 2007 jökulhlaup is the largest documented from this site making this event consistent with the onset of a new cycle of drainage, as envisaged by Evans and Clague's model of jökulhlaup cycles. However, we present field data from a rather more stable ice-dammed lake configuration than that described by Evans and Clague. Finally, we consider that at our study site

the jökulhlaup cycle stopped even though the lake was still present. Therefore, whether lake drainage occurs at Russell Glacier depends upon internal glacier-hydrological mechanisms and leads us to suggest that the IDL at Russell Glacier represents a new end member of the Evans and Clague jökulhlaup cycle model. Correct identification of the volume of IDL's is crucial for providing accurate estimates of the total volume of meltwater runoff which goes into temporary storage at the Greenland Ice Sheet margin; even the relatively small IDL at Russell Glacier contributes 2 - 3% of the annual runoff from the entire Kangerlussuaq basin. It is clear that larger ice-dammed lakes may prevent the direct runoff of significant volumes of meltwater over annual to decadal timescales.

INTRODUCTION AND AIMS

Glacier-dammed lakes exist worldwide and often pose an extreme hazard when they drain rapidly generating glacial outburst floods or 'jökulhlaups'. Indeed mountain glacier margin recession and accompanying increases in meltwater production have resulted in the formation and destabilization of many ice-dammed lakes (e.g. Wiles and others, 2008). Whilst controls on the drainage of a number ice-dammed lakes have been described from intensive field campaigns (e.g. Walder and others, 2006; Huss and others, 2007), detailed investigations of cycles of repeated drainage of a single ice-dammed lake are relatively rare. Whilst large numbers of ice-dammed lakes at the margins of the Greenland Ice Sheet (GrIS) are known to drain suddenly, controls on longer-term drainage cycles have yet to be examined.

This paper examines the characteristics of and controls on the behaviour of an ice-dammed lake (IDL) at Russell Glacier, near Kangerlussuaq in West Greenland. The IDL on the northern flank of the Russell Glacier drained in July 1987 (Russell and de Jong, 1988; Russell, 1989, 1993a) and this event marked the termination of a cycle of jökulhlaups initiated in the late 1940s (Russell, 1992). A jökulhlaup on August 31, 2007 therefore marked renewed ice-dammed lake drainage after 20 years of stability (Mernild and others, 2008; Mernild and Hasholt, 2009). A subsequent jökulhlaup on August 31, 2008 demonstrated that the 2007 jökulhlaup indeed marked the onset of a new jökulhlaup cycle.

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The onset of a new jökulhlaup cycle at Russell Glacier is important because ice-dammed lake level, extent and stability are well-known to fluctuate in response to climate-driven fluctuations of ice thickness and glacier margin position (Thórarinsson, 1939; Evans and Clague, 1994; Anderson and others, 2003). However, most knowledge of ice-dammed lake drainage processes is derived from the analysis of lakes impounded by glaciers which have undergone progressive retreat since the

Little Ice Age (e.g. Clague-Mathews, 1973; Desloges and others, 1989; Walder and Costa, 1996; Ng and Björnsson, 2003; Ng and others, 2007). For example, Evans and Clague's (1994) model of the 'jökulhlaup cycle' is based upon the study of retreating glaciers and proposes that lake drainage occurs as a cycle of repeated floods that have peak discharges that successively decrease due to progressive reductions in ice-dammed lake volume.

This paper aims to elucidate the factors controlling jökulhlaup magnitude and frequency allowing existing models of jökulhlaup cyclicity to be tested. To do this we: (1) reconstruct the volume and peak discharge of the 2007 jökulhlaup from Russell Glacier; (2) recalculate the volume and peak discharge of jökulhlaups in 1984 and 1987 from the same ice-dammed lake utilizing new data; and (3) evaluate controls on the jökulhlaup cyclcity at Russell Glacier.

STUDY AREA AND METHODS

Russell Glacier is an outlet glacier on the western margin of the GrIS and is located 35 km east of Kangerlussuaq (Søndre Strømfjord) (Fig. 1). Within the Kangerlussuaq meltwater catchment, the Russell Glacier meltwater sub-catchment has an area of ~ 90 km^2 (van de Wal and Russell, 1994) with summer ablation controlled discharges of up to 70 m^3s^{-1} (Russell, 1989; van de Wal and Russell, 1994; Russell and others, 1995). Within the Kangerlussuaq meltwater catchment jökulhlaups are known to have resulted from the drainage of ice-marginal lakes (Russell and others, 1990) supraglacial lakes (Russell, 1990, 1993) and ice-debris dammed lakes (Russell and others, 1995). Repeated jökulhlaups from the largest ice-marginal ice-dammed lake (IDL) on the northern flank of the Russell Glacier are relatively well documented (e.g. Sugden and others, 1985; Gordon, 1986; Scholz and others, 1988; Russell & de Jong, 1988; Russell and Marren, 1999; Russell, 1989, 1993, 2007, 2009). The 1984, 1987, 2007 and 2008 jökulhlaups all routed through a 1 km long sub- or englacial tunnel at the south-western corner of the IDL (Figs. 2, 3A and 3B) before expanding out across a coarse-grained delta into Outlet Lake 1 (OL1) (Figs. 1, 2) (Russell, 1989, 2007; Russell and Marren, 1999). Jökulhlaups drain from OL1 northwards into Outlet Lake 2 (OL2) by a bedrock spillway whose cross-sectional shape controls the level of the upper lake (Figs. 1 and 2) (Russell, 2007). On exiting OL2, jökulhlaups flow though a series of bedrock-confined channels and outwash plains before flowing west along the glacier snout (Fig. 1).

Fieldwork conducted in February 2008 comprised bathymetric surveying (Figs. 3C and 3D) using a Trimble LS4600 Single frequency differential Global Positioning System (dGPS). This has accuracy ~ 10 cm due to use in Real Time Kinematic (RTK) mode. Fieldwork conducted in May 2008 included precision surveying of ice-dammed lake level, proglacial lakes and proglacial jökulhlaup channel cross-sectional areas, whilst 2007 jökulhlaup lake level and wash limit evidence was still intact. Maximum flood water levels were determined from wash limits comprising erosional trimlines, trash lines of fine woody debris (shrub twigs and leaves), silt cover on vegetation and the upper extent of ice block melt out deposits (e.g. Carrivick and Rushmer, 2009). All positions and heights surveyed in May 2008 were obtained using a Leica GPS500 differential Global Positioning System (dGPS) used in Real Time Kinematic (RTK) mode with occupation of temporary base-station points. Base stations were precisely located by post-processing relative to continuous dGPS data for Kellyville, located 45 km away. Our surveyed points (e.g. Fig. 3C) have a mean three-dimensional accuracy of < 10 cm, and commonly < 1 cm, which makes our measurements immediately comparable in accuracy to previous surveys (e.g. Russell, 1989). The relative accuracy between the Trimble and Leica dGPS locations is assessed with points common to both data sets; these points fall within 0.5 m of each other and the average elevation difference is 0.211 m.

Peak discharge was reconstructed using three techniques: (1) the Clague-Mathews (1973) relationship utilising ice-dammed lake volume; (2) the slope-area technique at three stable proglacial channel cross-sections; and (3) a spillway equation used to reconstruct flow exiting OL1 over a bedrock lip.

The Clague-Mathews relationship was used to estimate jökulhlaup peak discharge from ice-dammed lake volume (Clague and Mathews, 1973; Ng and Björnsson, 2003):

$$Q_{max} = 75V_{max}^{0.67} \tag{1}$$

Four variants of the slope-area technique were used to reconstruct peak flood discharge within proglacial jökulhlaup channels. A bedrock spillway between OL1 and OL2 provides a stable channel cross-sectional geometry, confirmed by pre- and post-2007 channel cross-sectional surveys and observations. Reconstructions downstream of OL2 were made at a stable reach used to gauge the 1987 jökulhlaup (Russell, 1989; 1993, 2009; Russell and Marren, 1999). Reconstruction of 1987 jökulhlaup peak discharge was undertaken for comparison with discharge measured directly during that event, thus providing validation of the hydraulic techniques used. Mean flow velocities for each channel cross-section were calculated using three variants of the standard Manning resistance equation as well as an adaptation of the Keulegan equation (Thompson and Campbell, 1979; Russell and others, 1999; 2007; Carrivick and others, 2004; Carrivick, 2007a). The Manning and Keulegan resistance equations required the following input data: (1) Energy gradient or water surface slope; (2) Channel hydraulic radius; and (3) Grain roughness (Maizels, 1983; Henderson, 1966; Chow, 1959). Four methods were used to characterise grain and form roughness in order to identify potential problems of incorporating the channel resistance. Water surface slope was derived from well-defined wash limits. Channel hydraulic radius was calculated from cross-sectional areas and wetted-perimeters derived from surveyed channel cross-sections. Channel roughness was characterised by the Darcy-Weisbach friction factor '*f*' and Manning's '*n*'. The modified Keulegan equation proposed by Thompson and Campbell (1979), being the least empirical and least site-specific, was used to calculate the friction factor '*f*' in the Darcy-Weisbach equation (Church and others, 1990). The Darcy-Weisbach equation was used to obtain mean velocity, v:

$$v = \sqrt{\frac{8gds}{f}} \qquad (2)$$

where s is the slope, d is the flow depth (m), g is the acceleration due to gravity (9.81 ms^{-2}) and f is the friction factor calculated using Thompson and Campbell's (1979) equation:

$$\frac{1}{\sqrt{f}} = \left(1 - 0.1\frac{k_s}{R}\right) \bullet 2\log_{10}\left(\frac{12R}{k_s}\right) \qquad (3)$$

where R is the hydraulic radius and k_s is the size of the roughness elements equal to 4.5 times the boulder size D_i. In this case, we calculated k_s as the proportion of the flow depth, d, occupied by flow resistance elements (ΔA):

$$k_s = 4.5d\Delta A \qquad (4)$$

The proportion of the flow depth occupied by flow resistance elements was estimated visually. A single '*f*' value was calculated for each cross-section. Calculated '*f*' values were converted to Manning's '*n*' using an equation presented by Richards (1982):

$$n = \sqrt{\frac{fR^{1/3}}{8g}} \qquad (5)$$

Manning's n values were calculated from grain size characteristics using the Manning-Limerinos and Manning-Strickler equations:

$$n = 0.038 D_{90}^{1/6} \tag{6}$$

$$\frac{n}{d^{1/6}} = \frac{0.113}{1.16 + 2.0 \log \dfrac{d}{D_{84}}} \tag{7}$$

where d = flow depth and D_{84} and D_{90} are the 84th and 90th percentile grain-size values, respectively. D_{84} and D_{90} percentiles were derived for each cross-section from estimates of surface grain sizes (Maizels, 1983; Ryder and Church, 1986; Russell, 1994). However, *n* values calculated from grain-size distribution were compared with those derived solely from water surface slope and the hydraulic radius using Jarrett's equation (Equation 8) (Jarrett, 1984):

$$n = 0.32 S^{0.38} R^{-0.16} \tag{8}$$

where R = hydraulic radius (m) and S is the energy slope (-). Mean flow velocities were derived for each proglacial channel cross-section using each of the four variants of the slope-area technique presented above. Peak flood discharges were derived by multiplying mean flow velocities by channel cross-sectional areas.

Instantaneous outflow discharge from the maximum level of OL1 was calculated using the following spillway equation:

$$Q_w = \left(\frac{A^3 g}{b} \right)^{0.5} \tag{9}$$

where A = the cross-sectional area of the spillway (m^2), g = gravitational constant (9.81 ms^{-2}), and b = water surface width (m).

RESULTS

Ice-dammed lake (IDL) volume: 1984-2008

Drainage of the IDL on the northern margin of Russell Glacier occurred on 31 August, 2007 and lasted for approximately 17 hours (Mernild and others, 2008; Mernild and Hasholt, 2009) (Fig. 2). Our direct measurements of the IDL basin in May 2008 show that it did not drain completely and lowered in level by 49.2 m ± 0.2 m (Table 1), differing significantly with Mernild and others (2008) estimate of ~ 25 m. Lake volume loss during the 2007 jökulhlaup was 39.1 ± 0.8 x 10^6 m^3 (Table 1). Our lake volume loss figure is far greater than the 11.3 x 10^6 m^3 reported by Mernild and others (2008) and Mernild and Hasholt (2009). Mernild and others (2008) tenuously ascribed a further 17.5 x 10^6 m^3 to 'frictional melting of ice as the flood travelled in contact with the glacier together with an input from base flow'.

Glacier surface elevation in the vicinity of the ice-dammed lake increased by ~ 22 m ± 5 m between 1985 and 2006 (Figs. 3A and 3B). As the latest pre-jökulhlaup orthophotograph was only available for 2006, our glacier surface elevation is based upon comparisons of oblique field photos and corresponding natural markers in the field such as those indicated in Figure 2. Besides an ice thickness change, an ice margin advance is evidenced by field observations of a newly-formed push moraine within the IDL basin as well as along the rest of the local ice margin. Whilst Russell Glacier is known to have advanced between 1968 and 1999 (Knight and others, 2000), it is interesting to note that glacier extent into the ice-dammed lake basin has decreased by 100 - 180 m (Figs. 3A and 3B). In months immediately following the 2007 drainage, we observed that the Russell Glacier ice margin advanced into the lake basin by approximately 50 m. However, no change in ice margin position was observed following the August 2008 lake drainage. The position of the ice margin is important because it controls the exact relationship between ice-dammed lake level and volume (Fig. 4). Due to ice margin fluctuations, IDL volume for the maximum lake level increased by 8 x 10^6 m^3 from 1985 to 2006 (Fig. 4).

Lake elevation was not measured immediately before the 2008 drainage. We therefore calculated a lake water recharge in order to estimate lake elevation and volume immediately before the August 2008 drainage. A mean refill rate of 0.6 - 1 m^3 s^{-1} is based on discharge measurements of two major proglacial streams between July 15 and August 3, 2008, made by Bazeley (pers. comm.). A second approach using our bathymetric model of the ice-dammed lake basin (Fig. 3D), and a measured lake surface rise between August 2 and 4, 2008 of 0.32 cm ± 0.2 m (from 423.71 m) produces an average refill rate of 1.29 m^3 s^{-1}. We consider this latter recharge rate is consistent with the proglacial stream discharge since it includes additional meltwater contributions from subglacial melting and ice calving. Photographic evidence on September 2, 2007, September 17, 2007 and October 22, 2008 confirm that the post-drainage lake surface level was similar in 2007 and 2008, and that post-drainage refill to the lake basin was negligible.

IDL volume calculated from our new lake bathymetric data (Figs. 3C and 3D) has enabled re-calculation of ice-dammed lake volumes for the 1984 and 1987 jökulhlaups as reported by Sugden and others (1985) and Russell (1989), respectively. Sugden and others' (1985) volume estimate of 22.3 x 10^6 m^3 for the 1984 drainage has been recalculated as 25.2 x 10^6 m^3, whilst Russell's (1989) estimate of 32-36 x 10^6 m^3 was recalculated as 31.3 x 10^6 m^3 (Table 1).

Jökulhlaup peak discharge: 1984-2008

The Clague-Mathews (1973) relationship provides peak discharge values of 652, 753, 875 and 416 m^3 s^{-1} for the 1984, 1987, 2007 and 2008 jökulhlaups, respectively (Table 1). However, peak discharges calculated using slope-area techniques utilising proglacial flood wash limits and channel characteristics (Table 2) provide consistently higher estimates of peak discharge. The spillway equation used at the outlet of OL1 (Fig. 1) yielded peak discharges of 2030 and 2774 m^3 s^{-1} for the 1987 and 2007 jökulhlaups, respectively (Table 3). The spillway-derived peak 2007 jökulhlaup discharge agrees well with the average peak discharge calculated from the results of four slope-area techniques, for the same reach, which ranged from 1887 to 2806 m^3 s^{-1} (Table 3). Using the same slope area techniques, peak discharge at the gauged reach was 1647 m^3 s^{-1} (Table 3). The dramatic attenuation of peak discharge between OL1 spillway and the gauged reach, of ~ 40% over just ~ 5 km, is attributed to temporary storage within the OL2 basin (Fig. 1).

Observations of the ice dam and tunnel entrance

1985 and 2006 high spatial resolution digital orthophotographs illustrate that the position of this tunnel entrance remains fixed despite changes in the ice margin position and elevation (Figs. 3A and 3B). Visual comparison of photos in 1987 and May 2008 indicate maximum glacier margin advance of 50-60m and surface elevation increase of ~ 10m.

DISCUSSION

We have presented a record of drainage of Russell Glacier IDL since the 1950's, providing one of the longest records of ice-dammed lake drainage in Greenland. We have documented considerable variation in: (1) the levels to which the IDL drains from and to; (2) IDL volume (3) drainage frequency; and (4) peak discharge with distance downstream.

Controls on jökulhlaup magnitude

Jökulhlaup peak discharge is controlled by the volume of ice-dammed lake water released and lake drainage mechanisms (Tweed and Russell, 1999; Roberts, 2005). In recognition of the first of these controls the Clague-Mathews relationship has been used to predict the jökulhlaup peak discharge from the volume of water drained from the ice-dammed lake basin (e.g. Clague and Mathews 1973; Desloges and others, 1989). The Clague-Mathews relationship systematically under predicts jökulhlaup peak discharge at Russell Glacier. The volume to peak discharge relationship for the IDL lies between the regression lines for ice tunnel and dam-failure data sets (Fig. 4B). Jökulhlaups within the empirical data set used to define the Clague-Matthews relationship exhibit a more sedate rising stage, in line with predicted exponential tunnel expansion, as represented by Nye-Clarke equations (Roberts, 2005). The rising stage of the 1987 Russell Glacier jökulhlaup hydrograph is clearly linear (Russell, 1989, 1993), reflecting more rapid water release and higher peak discharge than predicted by the Nye-Clarke and Clague-Matthews equations. Furthermore, jökulhlaups with a linear rise to peak discharge will be subject to a greater degree of downstream discharge attenuation than those with an exponential hydrograph shape (Roberts, 2005). We argue that the Russell Glacier jökulhlaups are better characterised by a linear rise to peak discharge hydrograph.

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The unusually short englacial/subglacial tunnel length (~ 1 km) is an important factor in explaining higher than predicted jökulhlaup peak discharges. If the IDL drained subglacially for 8 km to the Russell Glacier snout, a lower peak discharge would result due to lower conduit enlargement rates. A relatively thin ice-dam would also be more prone to flotation allowing water to exit the lake more rapidly. In 2007 and 2008 the IDL drained to a level of 404.9 m, approximately 10 m above the conduit outlet where the jökulhlaups entered OL1. This suggests the presence of a bedrock lip as a control on minimum lake level. Observations of jökulhlaup inlet tunnels in the same place during successive drainage events implies that bedrock topography provides both conduit location and form at this location. Jökulhlaup flow through an incised bedrock-walled 'Nye channel' would also favour a quicker rise to peak discharge due to the concentration of hydro-mechanical erosion processes allowing the rapid evacuation of ice from the bedrock channel (Roberts, 2005).

Hydraulic resistance exerted by proglacial jökulhlaup routeway will modify a flood hydrograph with distance from source (e.g. Carrivick 2007a, 2007b) and therefore complex proglacial topography, such as the inter-linked bedrock basins typical of western Greenland causes a dramatic attenuation of peak discharge downstream (Sugden and others, 1985). This may partially explain why our slope-area and spillway crest calculations of jökulhlaup peak discharge are so much larger than the peak discharge estimates provided by Mernild and others (2008) and Mernild and Hasholt (2009), who utilise stage records ~ 35 km downstream at Kangerlussuaq. Our calculations of peak discharge for the 1987 jökulhlaup display a similar rate of downstream attenuation from the IDL, to OL1 and OL2. The gross underestimation of 2007 and 2008 jökulhlaup volumes presented by Mernild and others (2008) and Mernild and Hasholt (2009) highlight the perils of extrapolating

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stage-discharge relationships to include higher stage events. Calculations of jökulhlaup volume are best based upon knowledge of the ice-dammed lake basin rather than gauging 35 km downstream.

Jökulhlaup timing and frequency

Russell (1989) and Russell and de Jong (1988) suggested that the Russell Glacier IDL would preferentially drain later in the melt season when sub- and englacial water pressures are more likely to be lower. Thus ingress of IDL lake water to the glacier plumbing system would be facilitated. Known drainage dates for the Russell Glacier IDL of July 18 (1987), August 19 (1984) and August 31 (2007 and 2008) (Sugden and others, 1985; Russell, 1989; Mernild and Hasholt, 2009) lend some support to this argument. Ice-dammed lake drainage can be triggered by sudden reductions in subglacial water pressure resulting from reduced meltwater production (Tweed and Russell, 1999). Interestingly, Mernild and Hasholt (2009) pointed out that the 2007 and 2008 jökulhlaups occurred 4 days after a cooling of mean air temperature. The 1987 jökulhlaup occurred 4-5 days after a reduction in air temperatures recorded in Kangerlussuaq (Fig 5). The 1987 stage record for the gauged reach indicates a steady reduction in river base flow in week preceding the jökulhlaup pointing to a *meltwater reduction – pressure trigger* for lake drainage (Fig. 5). Precipitation experienced by the lead author at Russell Glacier during the 1987 jökulhlaup and recorded by Mernild and Hasholt (2009) in Kangerlussuaq before the 2007 jökulhlaup is not thought to be significant in triggering the ice-dammed lake drainage as the amount of precipitation is not considered sufficient to offset the reduction in ablation resulting from the reduction in incoming solar radiation. Although van de Wal and Russell (1994) were able to register a non-convective rainfall event in the 1991 discharge hydrograph record from Russell Glacier, ablation-controlled discharge variations provide the greatest overall impact on meltwater runoff rate. Although it may be possible to invoke a *meltwater reduction – pressure trigger* for the IDL, the specific timing or triggering threshold for a jökulhlaup during the melt season is likely to reflect the year to year variations in seasonal evolution of the subglacial drainage system which is coupled to the intensity of the melt rate (Tedesco, 2007; Mernild and Hasholt, 2009).

Drainage of the Russell Glacier IDL between the 1950's and 1987 is considered to have occurred every 2-3 years based upon evidence from aerial photographs, sedimentary data and lake re-fill time (Gordon, 1986; Russell, 1989; 1992). Mernild and Hasholt (2009) suggested that the IDL drained in consecutive years in 1983 and 1984 mirroring the 2007 and 2008 events. Although Sugden et al. (1985) suggested that the IDL may drain annually; their argument for a jökulhlaup in 1983 was based upon wash limit evidence of a jökulhlaup of equivalent or greater magnitude to their observed

1984 flood. The 1984 jökulhlaup had a volume of 25.2 x 10^6 m^3, twice the volume of the 2008 jökulhlaup, making it impossible for the lake to have drained in 1983.

We consider the 2007 and 2008 jökulhlaups to be the first recorded consecutive annual drainage events from the Russell Glacier IDL. The 2007 jökulhlaup was the largest recorded from this lake whilst the 2008 jökulhlaup was just one third of the volume of water that drained in 2007. To explain occurrence of two such different jökulhlaups from the same IDL in close succession we have to invoke different drainage and triggering mechanisms.

We suggest that the subglacial passage of the 2007 jökulhlaup weakened the ice dam sufficiently to allow the 2008 drainage to occur at a much lower lake level. It is feasible that much of the subglacial conduit remained open between the two events. It is also possible that the lake did not re-seal after the 2007 drainage. A substantial accumulation of naled (aufeis) ice was observed in May 2008 directly in front of the 2007 jökulhlaup conduit outlet indicating a steady feed of water from the conduit through the winter months. Discharge of water from the tunnel outlet in late May 2008 may also have been from leakage of the IDL. Observations of the 2007 jökulhlaup tunnel inlets in late May 2008 show that they were covered by ice cliff collapse breccia. Similarly, the 2008 tunnel inlets were observed to collapse within two months of IDL drainage. It is feasible that rising IDL levels during the 2008 melt season allowed the flotation of ice cliff collapse debris from the conduit entrances.

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Ice-dammed lake – glacier interactions

Although observations of ice dynamic responses to lake drainage have been made (e.g. Sugden and others, 1985; Walder and others, 2006; Mayer and others, 2008), detailed consideration of lake volume, lake levels and ice margin position have hitherto not been reported for multiple jökulhlaups from the same system. We have demonstrated that following the 2007 drainage, glacier advance into the IDL basin reduced the lake basin volume. Our data are compatible with observations made by Sugden and others (1985) for the same IDL who reported glacier margin advance following the 1984 jökulhlaup. Jökulhlaup volume and peak discharge therefore decreased. Such feedbacks between ice-dammed lakes and ice margin position would produce progressive decreases in jökulhlaup magnitude within a jökulhlaup cycle.

Controls on jökulhlaup cycle at Russell Glacier

The fact that the 2007 jökulhlaup is the largest documented from this site is consistent with the onset of a new cycle of drainage, as envisaged by Evans and Clague (1994). However, we present field data from a rather more stable ice-dammed lake configuration than that described by Evans and Clague (1994); their hypothetical examples concentrate on a piedmont setting where glacier margin position, lake extent and lake volume fluctuate relatively quickly over time during glacier advance and retreat. Consequently, at our study site, the jökulhlaup cycle stopped even though the lake was still present. Therefore, whether lake drainage occurs at Russell Glacier depends upon internal glacier-hydrological mechanisms. We therefore suggest that the IDL at Russell Glacier represents a new end member of the Evans and Clague (1994) jökulhlaup cycle model. Whilst the renewed cycle of drainage at Russell Glacier might be regarded as being subordinate to the major cycles reported by Evans and Clague, it is driven by glacier margin fluctuations. This is important because it is an ice-dammed lake system where the jökulhlaup cycle continues during localised ice margin advance and because a complete glacier advance-retreat cycle has not yet been studied.

CONCLUSIONS

Ice-dammed lake drainage at Russell Glacier has renewed after twenty years of dormancy. Our analysis reveals that the key factor controlling the occurrence and character of jökulhlaups at this site is the pattern of ice margin advance and retreat. This control of ice margin dynamics has been identified for alpine glaciers where overdeepenings promote subglacial storage of meltwater (e.g. Thórarinsson, 1939; Carrivick and Rushmer, 2009) but has not been identified on the edge of ice sheets. Specifically, at Russell Glacier there was a period of ice margin advance from 1987 to ~ 1999 and this was subsequently followed by recession (~50 m) and ice surface lowering (~ 10m); i.e. ice thinning. The drainage of the same lake in 2008 suggests that the 2007 jökulhlaup was the first in a renewed phase of ice-dammed lake drainage (Evans and Clague, 1994; Clague and Evans 1997). The 2007 jökulhlaup is the largest documented from this site and this fact is consistent with the model of Evans and Clague (1994). We draw attention to note that Evans and Clague's model is entirely driven by ice margin retreat producing a reduction in ice-dammed lake level, jökulhlaup volume and peak discharge. In contrast, we find that reduced lake volumes and thus lower jökulhlaup peak discharges can also be produced by localised ice margin advance and/or glacier thinning. Our robust calculations of lake volume drained and peak discharge argue for a non-exponential model for the drainage of IDL, and thus a mechanism that is determined by a relatively thin ice dam, a short bedrock controlled subglacial tunnel and hydro mechanical erosion of glacier. We reconstruct a 40% attenuation in peak discharge over just ~ 5 km due to temporary storage in bedrock basins. Correct identification of the volume of IDL's is crucial for providing accurate estimates of the total volume of meltwater runoff which goes into temporary storage at the ice-sheet margin. Using Mernild and Hasholt's (2009) discharge figures, drainage of the relatively small IDL at Russell Glacier contributes 2 - 3% of the annual runoff from the entire Kangerlussuaq basin. It is clear that larger ice-dammed lakes, of which there are many in western Greenland, may prevent the direct runoff of significant volumes of meltwater over annual to decadal timescales.

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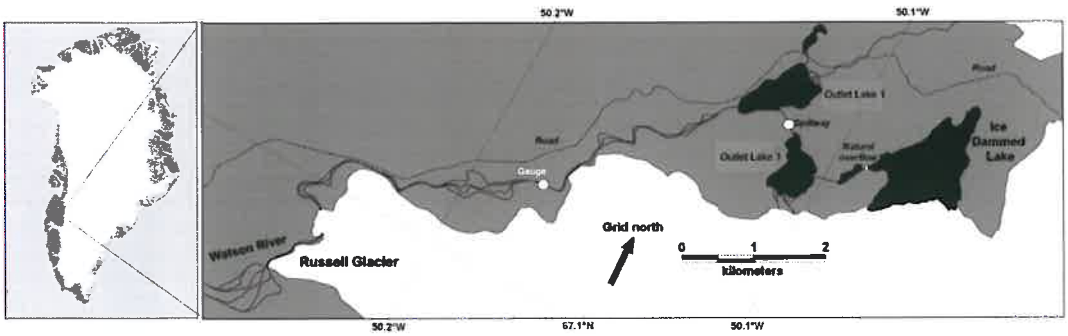

Figure 1. Location of the ice-dammed lake (IDL), Outlet Lake 1 (OL1) and Outlet Lake 2 (OL2) in relation to the northern margin of Russell Glacier, West Greenland. The ice-dammed lake drains through an englacial-subglacial tunnel for a distance of ~ 1 km routing jökulhlaups into OL1. Proglacial jökulhlaup peak discharge reconstructions were undertaken at the spillway between OL1 and OL2 and at the 1987 gauged reach.

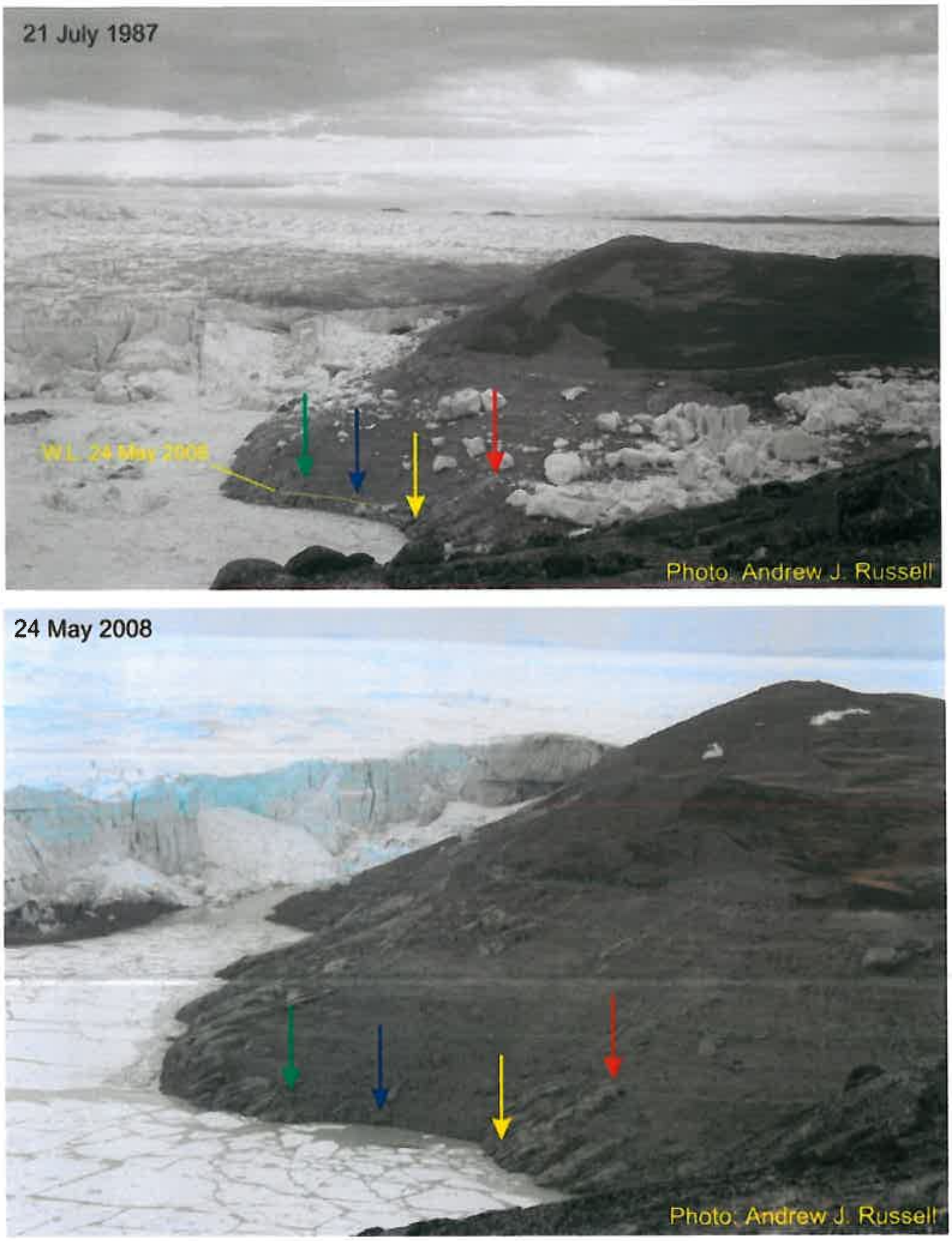

Figure 2. Ice dammed lake (IDL) on the northern margin of the Russell Glacier, on July 21, 1987 and May 24, 2008. Both photographs were taken immediately after a jökulhlaup and permit comparisons between lake level drop; see the coloured arrows for distinguishing common points, and for examining ice margin position and thickness.

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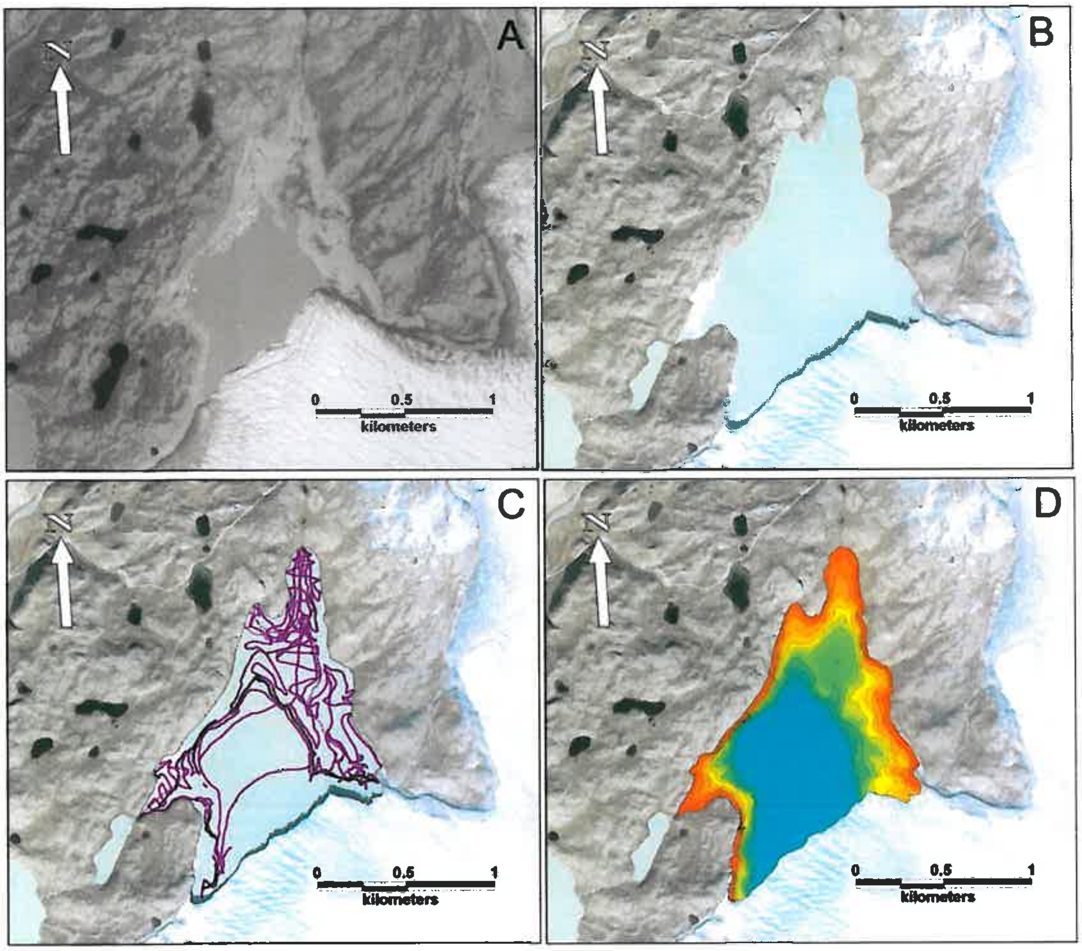

Figure 3. Recent visual history of ice-dammed lake (IDL) dynamics. (A) Aerial photograph from 1985 by Danish National Survey and Cadastre illustrates low lake level, smaller lake surface area and thinner ice. (B) 2006 aerial photograph by ARTEK illustrates maximum lake level, larger lake volume and thicker ice. Maximum lake level is controlled by a proglacial bedrock spillway which drains into OL1. (C) Tracks of a kinematic dGPS survey undertaken in February 2008 in order to interpolate lake basin bathymetry. (D) Lake bathymetric model with contour lines at 5 m intervals. Maximum lake level illustrated is 453.7 m. Lake levels were measured in May 2008.

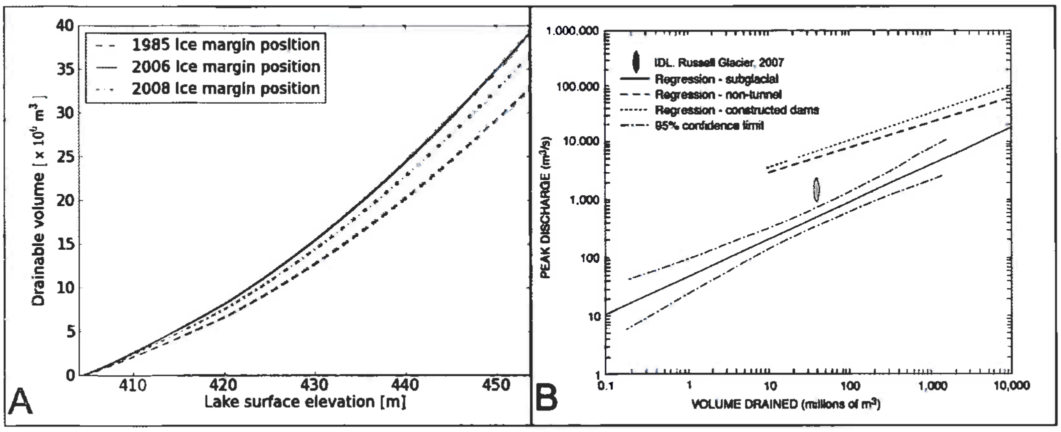

Figure 4. (A) Stage-volume graph for the ice-dammed lake (IDL). The three curves describe the stage-volume relationship based on the changes in ice margin position. The 1985 and 2006 curves are based on digitized ice margin positions from air photos, whereas the 2008 curve is based on GPS measurements. The grey zones indicate the change in drainable volume based on a 10 m advance or retreat of the entire ice margin. 1987 and 1984 volumes were recalculated based on the 1985 curve (Table 1). (B) The volume-peak discharge relationship between that of floods that route through englacial/subglacial tunnels and those that are triggered by dam failures. The grey ellipse in (B) is defined by the error of our calculations, particularly those of peak discharge in which we employ several different methods. Figure 4B is adapted from Walder and Costa (1996) and Costa and Schuster (1988).

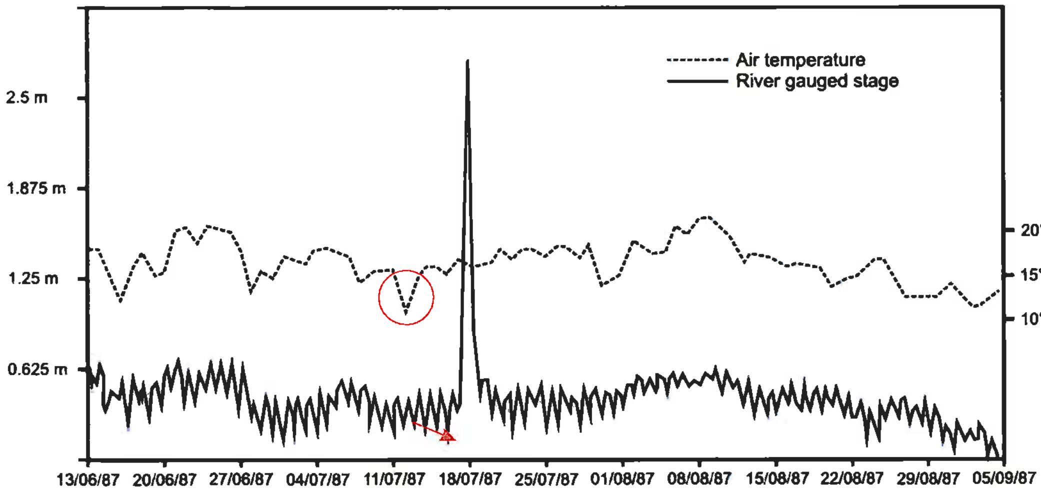

Figure 5. River stage from gauged reach and daily maximum air temperatures from Kangerlussuaq 13 June, 1987 to 05 September, 1987 (see Figure 1 for location). A dip in temperature (circled) precedes the jökulhlaup by 4 days reducing meltwater production rate as characterised by the reduced base flow component in the hydrograph (arrow).

Jökulhlaup (year)	Lake basin	Pre-drainage lake level (height above ellipsoid, m)	Post-drainage lake level (height above ellipsoid, m)	Lake level surface level change (m)	Volume change (x 10^6 m^3)	Peak discharge (Clague-Mathews) ($m^3 s^{-1}$)
2008	IDL	428.3 (1)	404.9	-20.0	12.9 ± 0.3	416
2007		453.7	404.5	-49.2	39.1 ± 0.8	875
1987		450.8 (2)	402.5 (2)	-48.3	31.3	753
1984		446.7 (2)	406.9 (2)	-39.8	25.2	652
2007	OL1	388.1	394.9	+6.8	0.84	-
1987		-	-	+6.5	0.81	-
1984		-	-	-	-	-
2007	OL2	359.9	363.9	+3.9	0.79	-
1987		-	-	+3.44	0.70	-
1984		-	-	+3.11	0.63	-

Table 1. Lake levels measurements and calculations used to produce jökulhlaup volumes. Error associated with all measurements is ± 0.2 m vertically, and 0.1 m horizontally. (1)Value derived from a dGPS measurement on August 2, 2008 and a recharge rate based on measurement of lake surface elevation rise between August 2 and 4, 2008 (1.29 m^3s^{-1}). (2)Water surface elevations for 1984 and 1987 were measured relative to an arbitrary datum (Sugden and others, 1985; Russell, 1989), but are presented relative to the spillway-controlled maximum lake level.

	Spillway cross-section 1	Spillway cross-section 2	Gauged reach 1987	Gauged reach 2007
Cross-sectional area (m^2)	311	382	174	224
Hydraulic Radius (m)	1.7	1.4	3	3
Slope (mm^{-1})	0.200	0.037	0.018	0.018
Width (m)	91	132	56	57
Average depth (m)	5.0	5.0	3.5	4.5
Grain size Min.	2.0	2.0	0.2	0.2
D_{84} (m) Max.	3.0	3.0	0.5	0.5
Grain size D_{90} (m)	3.0	3.0	1.0	0.5
% reduction in cross-section area due to roughness	25	25	5	5

Table 2. Input data for jökulhlaup peak discharge reconstruction using four variations of the slope-area technique.

(m^3s^{-1})	Spillway crest (1987)	Spillway crest (2007)	Spillway cross-Section 1 (2007)	Spillway cross-section 2 (2007)	Gauged reach (1987)	Gauged reach (2007)
Darcy-Weisbach	-	-	4070	3285	1207	1747
Manning-Limerinos	-	-	2113	1127	1063	1625
Manning-Strickler	-	-	3845	2051	1347	2011
Jarrett	-	-	1195	1085	778	1204
Weir Crest Equation	2030	2774	-	-	-	-
Average	2030	2774	2806	1887	1099	1647
1987 Gauged	-	-	-	-	1080	-

Table 3. Reconstructed jökulhlaup peak discharge for each of the slope-area variants and the weir crest equation. Average reconstructed peak discharge for the 1987 jökulhlaup is very similar to the peak discharge gauged by Russell (1989).