Recurring Jøkulhlaups From Koppangsbreen, Norway

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Abstract

The aim of this study was to examine why the recurring jøkulhlaups occurred from Koppangsbreen glacier in northern Norway during the summer of 2013. Various data provided by NVE, available satellite and aerial ortho images and measurements of the glacier thickness were collected in order to calculate the stability of the ice damming the lake. Meteorological data was used to estimate the input to the lake between each event, in order to compare lake volumes against dam thickness so as to obtain better knowledge on the triggering mechanisms at each event.

The glacier dam failed although it had been stable before due to decreased thickness of the glacier dam. At the time of the recurring events in 2013, the glacier thickness must have reached a threshold limit that allowed for a number of re-occurring jøkulhlaups to occur. By the time the subglacial tunnel, that was formed by the first event, had closed, the lake volume had been refilled enough to trigger a new event. The rapidly recurring events could probably trigger drainage at lower lake volumes due to the already existing tunnels and fractures created by the preceding event. Additionally, warm water due to high temperatures over a long period might have contributed in this process. When the water drainage first was initiated, maintenance and expansion of the subglacial tunnel occurred due to the thermal energy of the flowing water. Due to further glacier decrease, there is currently no risk of another events.

Preface

4th June 2013, a jøkulhlaup released approximately 1.9 mill m^3 water which drained towards Koppangen village in northern Norway. The flood resulted in evacuation of people and several houses at Koppangen village downstream. At intervals of three to ten days, six more jøkulhlaups followed within the next 30 days, and another two before the summer season was over. No other known site has had as many recurring jøkulhlaup events during such a brief period, worldwide. The events at Koppangsbreen are therefore unusual and exceptional (Jackson and Ragulina, 2014).

CONTENT

ACKNOWLEDGEMENTS

First of all I would like to thank my supervisors Miriam Jackson at the Norwegian Water Resources and Energy Directorate (NVE) and Chris Nuth at University of Oslo (UiO) for being my supervisors and for giving me the opportunity of writing this thesis. Thank you for being encouraging, enthusiastic, inspiring, and helpful, for giving me reviews and feedbacks even when time was scarce. I would also like to thank NVE and UiO for the sponsor and funds I have been given that have brought me to inspiring field trips throughout of Norway. Additionally, I would also like to thank the employees at NVE for being so accommodating, helpful and inclusive during my stay at their open office landscape, for the cheerful lunches, dinners, excursions and tasteful cakes. Finally I would like to thank all my friends and family for always being supportive and understanding, whenever needed.

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1. INTRODUCTION AND BACKGROUND

A jøkulhlaup, or Glacier Lake Outburst Flood (GLOF), is any large and sudden release of accumulated water from a glacier (Jackson and Ragulina, 2014). Water bodies can accumulate in glacier- or proglacial moraine-dammed lakes, or under, within or on top of the glacier itself (subglacial, englacial, supraglacial) (Paterson, 2010) (Liestøl 1956, Nye 1976**).** In Norway, jøkulhlaups occur most commonly due to drainage of glacier-dammed lakes.

Jøkulhlaups are considered severe geomorphological hazards because of their abrupt and irregular occurrence, along with their high damage potential. (Carrivick and Tweed 2016) Entire lakes can drain in hours only, hence the normal glacier discharge can increase by more than one order of magnitude, within a short time. Sudden movement of such large volumes of rapid flowing water, possibly mixed with other material, can cause substantial damage far from the dam, having hazardous impact on humans and human structures, situated close to or near the outlines of the flood (Bjornsson 2002#) (Benn and Evans 2014) (Engeset, Schuler et al. 2005)

1.1. Jøkulhlaup Mechanism

The typical evolution of a jøkulhlaup begins with the formation and growth of a water body and it ends when the water drains. Formation is initiated by accumulation of water from precipitation, seasonal melting and/or geothermal heating and occurs wherever free drainage is prevented by some form of barrier. This could be bedrock, moraine or the glacier ice. In the latter case, water storage is located in cavities or channels within, under or on top of the glacier itself. Continuous supply of additional rain- and meltwater, maintains further growth, which increase the volume until a critical moment, the floatation pressure, is reached. Supplementary inflow will eventually result in dam breakage and the water drains in a jøkulhlaup, sometimes emptying the entire lake, sometimes not. (GLACIER HYDROLOGY CH3#)

Lakes can be formed along the glacier margins where they are being dammed by the ice itself, in depressions on the glacier surface on top of the glacier, or enclosed within or beneath the glacier. They can also be trapped between a retreating glacier front and a moraine, rock or sediment from previous glaciation. The lake types are thus generally divided into glacier-dammed, moraine-dammed and subglacial lakes (Benn, 2014).

Glacier-dammed lakes form along the glacier margins where the water is dammed by the glacier itself. Supraglacial lakes form in depressions on the glacier surface on top of the glacier, englacial lakes form when water is enclosed within and subglacial lakes form underneath the glacier. Proglacial moraine-dammed lakes form where water is trapped between a retreating glacier front and a moraine, rock or sediment from previous glaciation.

Marginal or glacier-dammed lakes are formed where the glacier itself blocks a stream or the escape route of meltwater. Moraine-dammed lakes are formed where water is trapped behind a moraine and, in front of a glacier, see figure 1. The moraine is typically from a retreating glacier. This dam type is very unstable and difficult to monitor in regard to hazard forecasting. Moraines consist of till which is unconsolidated and unsorted glacier debris with low shear strength, but the moraine may also enclose a core of ice. An ice core makes it difficult to calculate and predict how much accumulated water volume the moraine dam can resist before failure, as it depends on its size which consequently controls how much time will pass before melt. The presence of an enclosed ice core within the moraine is neither unusual nor evident, which magnifies the difficulties in outburst prediction because it demands detection of its presence, size and geometry within the moraine. This is often time-consuming and expensive {Benn, 2014 #14}

An englacial lake is formed where water accumulates within a glacier, enclosed by surrounding ice, for example during the formation of englacial drainage system where moulins, cavities and/or channels connects to form voids where water can accumulate.

A supraglacial lake is formed where water accumulated in a depression on top of the glacier surface, due to depletion in the bottom topography or space-releasing phenomenon causing the overlying ice to sink down, creating a trough on the glacier surface. This could be a collapse of a tunnel below, subglacial lake with melting of ice at the bottom due to geothermal heat etc. supraglacial meltwater may accumulate here and infiltrate downwards through the glacier towards the bottom {Benn, 2014 #14}.

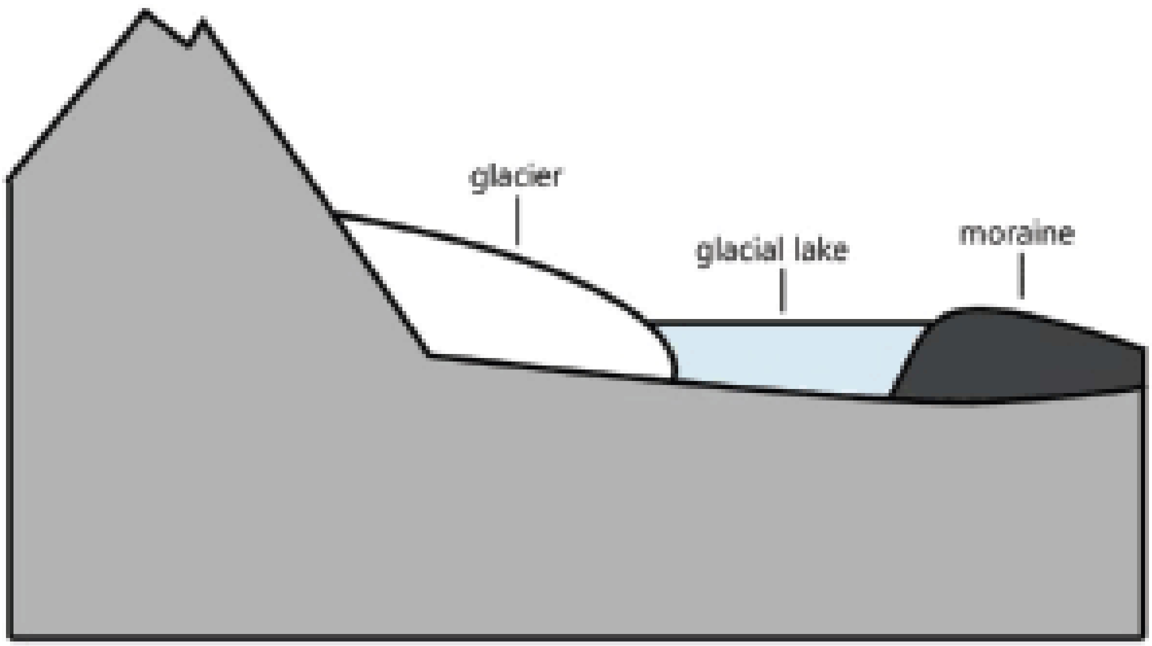


Figure 1: Illustration of moraine-dammed lake, trapped between the glacier and the moraine. source: https://www.nap.edu/openbook/13449/xhtml/images/p46.jpg

Subglacial lakes form due to bottom topography, see figure 2 a, or where the glacier is situated above an area of geothermal heat, figure 2 b. these are common in Iceland, where there is geothermal and volcanic activity due to the diverging mid-ocean ridge in the Atlantic Ocean. Some of the most famous jøkulhlaup events in the world, are the subglacial outburst floods from Grímsvötn underneath the glacier Vatnajökull, Iceland, which is well-known for its periodic events due to geothermal heating and sometimes subglacial volcanic eruptions (#islandsk referanse her). The heat melts ice at the bottom of the glacier, which accumulates and forms a lake. If not all the overlying ice melts, the water will eventually force its way out between the glacier and the bedrock or through it by forcing open a tunnel. Since water contracts when transitioning from frozen to liquid, the melting releases space and the overlying ice may sink down, creating a trough at the glacier surface, disclosing the subglacial lake at its bottom. This trough might eventually become the location of where a supraglacial lake will accumulate, due to precipitation and surface meltwater. {Benn, 2014 #14}

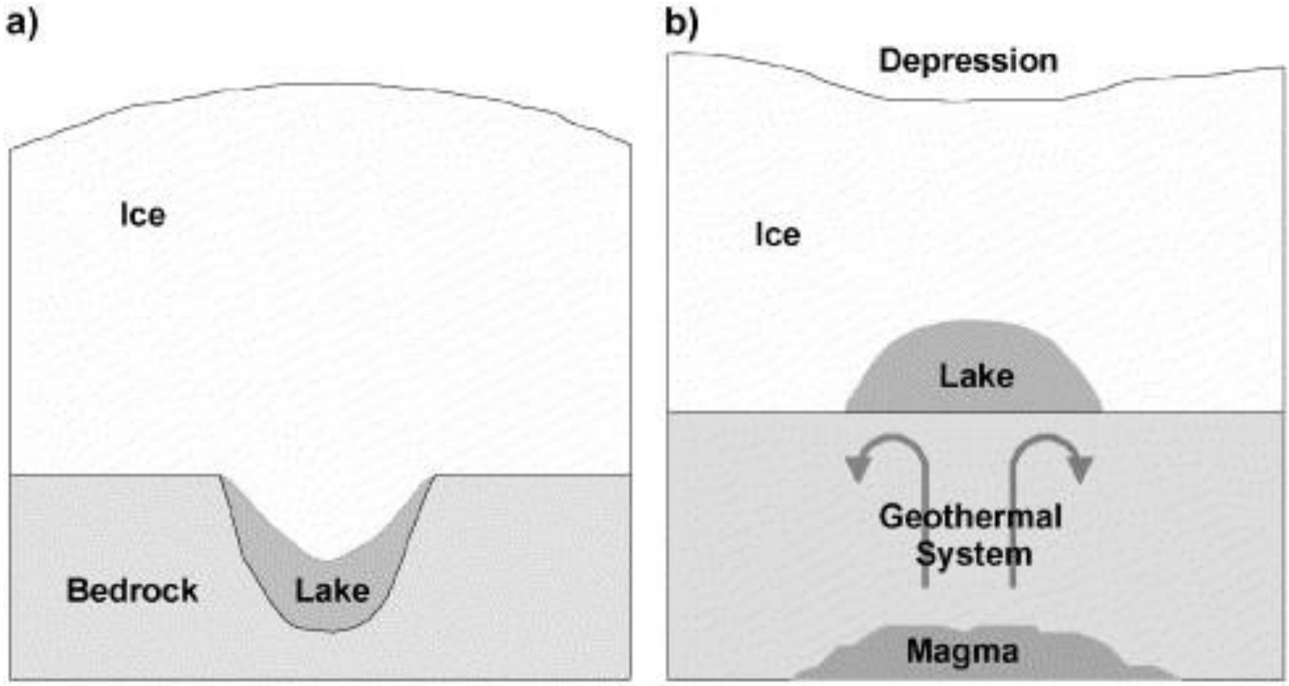


Figure 2: Schematic drawing of two types of sub-glacial lakes, the one in 2 a is formed due to the bottom topography while the one in 2 b is formed above a geothermal system.
Source: {Björnsson, 2003 #5} https://ars.els-cdn.com/content/image/1-s2.0-S0921818102001303-gr3.jpg

TRIGGERING MECHANISMS

Dam failure occurs when the retaining forces of the dam are exceeded by the potential energy of the water, given by the gravitational forces that constantly pull the water downwards. (GLACIER HYDROLOGY CH3#)

A jøkulhlaup outburst occurs when the resisting forces of the dam are exceeded and the dam no longer can hold the water in place. This occurs when the water pressure increases or the dam strength is reduced. Several triggering mechanisms might cause failure:

There are several mechanisms to sudden drainage of a lake, different trigger and drainage mechanisms operate for the different lake types. The mechanisms by which jøkulhlaups are triggered vary with position of water body and the nature of the dam holding back the waters. Triggering mechanisms might be water flowing below or through an ice-dam, water overflows and rapidly cuts through ice, sediment barrier or bedrock, or when a subglacial reservoir grows to such a volume that it collapses.

If the water is retained by glacier ice, initiation of water flow occurs due to the density differences between liquid water and solid ice. The physical property determines the floatation pressure of the glacier dam. In theory, this occurs when the pressure in the water volumes overcomes the floatation pressure of the ice, equalized by a level difference of 9/10. However, some dams are stable until this point is reached, while others release their water at a lower level, such as Blåmannsisen (e.g. Björnsson, 1992; Walder and Costa, 1996; Jóhannesson, 2002).

Ice melts faster with presence of water. When the water has first started flowing, melting and erosion will expand already existing cavities or channels and result in a positive feedback mechanism so that drainage out of the lake increases rapidly, resulting in a complete or partial draining of the lake in only a few hours or days [NVE]. During an event the discharge can therefore increase by several orders of magnitude (e.g. Björnsson, 2002). This can have catastrophic impact on people and society, causing considerable damage and destruction to people and society living close to or near the outlines of the floods.

Moraine-dammed lakes drain after failure (caused by rapid incision) of the sediment barrier by outpouring waters. Once incision begins water flow concentrated through the outlet can accelerate erosion and enlargement of it, starting a positive feedback process resulting in rapid release of large amounts of sediment-laden water. The onset of rapid incision of the barrier may be triggered by waves generated by glacier calving, ice avalanching, increase in water level associated with glacier advance. (Nepal august 1985, Zimmermann 1986) (Peru,Lliboutry 1977) (British Colombia, Clague 1985 & Evans and Clague 1993 1994). Installation of outflow pipes can dramatically reduce the risk as it allows the water to drain continually away, but may not be economically feasible in remote areas in poor countries.

Glacier-dammed lakes drain after failure of glacier dam or when water melts or erodes an overflow channel into dam surface (along over glacier surface, along ice margins, through neighbouring cols = lowest point between two peaks. Jøkulhlaup hydrographs or discharge curves of such lakes display rapid increases up to a maximum, followed by gradual decrease. Some glacier-dammed lakes drain slowly, for example Margin of Sydgletscher, south Greenland 1981, which emptied in 14 days, at an average rate of 200 m^3s^{-1}. (Dawson 1983)

The glacier-dammed and subglacial lakes may drain rapidly via channels at or near the bed as the result of changes in glacier hydrological system or changing relationships between pressure conditions in the water reservoir and surrounding ice, since liquid water is denser than solid state ice.

Theoretically, initiation of drainage occurs when the lake is filled to a water level higher than the potential barrier at the glacier bed, as this allows drainage underneath (Björnsson 1974). As jøkulhlaups occur in different settings, there are also different mechanisms for drainage initiation. Some lakes drain when the lake level is at the same height as the ice barrier, having reached its flotation pressure which is lower than the height of the glacier dam, while others drain when the lake level flows over the ice barrier, and others drain at lower levels (Jackson and Ragulina, 2014).

MAINTENANCE OF DRAINAGE TUNNELS

The maintenance of drainage paths is an important part of the mechanism, being necessary for such a large amount of water to flow over such a brief periode. Channel wall processes are melting, freezing and ice deformation. Melted open by frictional heat and closed by plastic deformation at the same time. A jøkulhlaup ends abruptly when it hits rock level for the lake, all the water is drained, when tunnels collapse due to reduced pressure.

Jøkulhlaups have two phases, the early phase when the lake grows and the potential hazard increases, then a late phase when the lake drains {Björnsson, 2010 #6}. The period of water accumulation before rapid drainage occurs can be several years but the duration of the jøkulhlaup outburst event itself is short, lasting only hours or days (Liestøl 1956). After an outburst, water can accumulate again and i**n some places the occurrence might appear to be cyclic, such as in Iceland.** The duration of the first phase where water accumulation builds up is not constant throughout the world as it varies with lake characteristics such as dam type, lake size, catchment area and local climate and/or the presence of geothermal activity. At Vatnajokull in Iceland, the duration of build-up before release is typically six years but in other places outbursts happen only occasionally. Hence precise prediction is not always possible and assumptions on how long it will take before a lake reaches its unstable level are complicated. The forecasting of future events and developing risk assessment methods and early warning systems is therefore difficult but important because the world's population grows, remote areas become more inhabited and rural settlements expands (Carrivick and Tweed 2016).

In Norway, jøkulhlaups mostly occur due to drainage of glacier-dammed lakes. {Benn, 2014 #14}. Drainage from glacier-dammed lakes occurs when the lake level is high enough to overcome a potential barrier at the glacier bed, enabling discharge underneath the glacier (e.g. Björnsson, 1974; Nye, 1976; Fowler, 1999). In some cases, the water flows over the glacier as occurred at Rembesdalskåka, a glacier outlet of Hardangerjøkulen (Liestøl, 1956) in southern Norway. Because of their far-reaching effect and their erratic nature, jøkulhlaups pose a significant hazard and can cause substantial loss of human life as well as damage to agricultural land and infrastructure. Reviews of the geographic distribution and drainage characteristics of glacier-dammed lakes are provided by Björnsson (2002).

JØKULHLAUP DISCHARGE

Estimates of the possible magnitude of future floods is important because they are so destructive to humans and human structures. Peak discharges are the most erosive and destructive phases of floods and methods to predict these are important. A relationship between the volume of water released from a glacier-dammed lake and peak flood discharge was calculated by Clague and Matthews (1973), and has been modified several times since (Costa, 1988; Desloges (1989). This method of discharge prediction – the Clague-Mathews formula – is not based on any physical mechanism but appears to give reasonable results. However, it is the only one that can be used for prediction as others require either flood duration or flood depth, which is impossible to know in advance of a jøkulhlaup outburst flood.

Observations in North America, Iceland and Scandinavia show that peak flood discharges are two to six times higher than mean discharge for the whole event thus if one knows the volume of water released, and the duration of the flood, mean and peak discharges can be calculated. Which means it is a method useless for determining magnitude of future floods (since one cannot know its duration in advance). Slope area method – a more physical based method of calculating peak discharges – which is based on measurements of dimensions and slope of channels during peak flow conditions – either from direct observations or geomorphological evidence.

EFFECTS AND IMPACTS OF JØKULHLAUPS

Jøkulhlaups represent a significant hazard to many glaciated regions of the world. Figure 4 gives an overview of proportion of glacier outburst flood records by major region and the number of outbursts per dam type or trigger. The term jøkulhlaup is Icelandic and referred originally to outburst floods related to geothermal heating and volcanic activity beneath glaciers on Iceland, exclusively. Figure 4 shows that jøkulhlaups occur in many other places all over the world and that this reference is not representative for the entire world. Therefore, the term now involves outbursts anywhere in the world, which originate from glaciated areas. It includes outbursts related to other triggers and other dam types, such as ice, moraine and bedrock as well. As seen on the map by Carrivick and Tweed (#2016), Iceland is the only place where outbursts are triggered by volcanic eruptions (red pie chart), but the most common dam type/trigger is ice (illustrated with blue pie chart) and the second most common is moraine (orange pie chart).

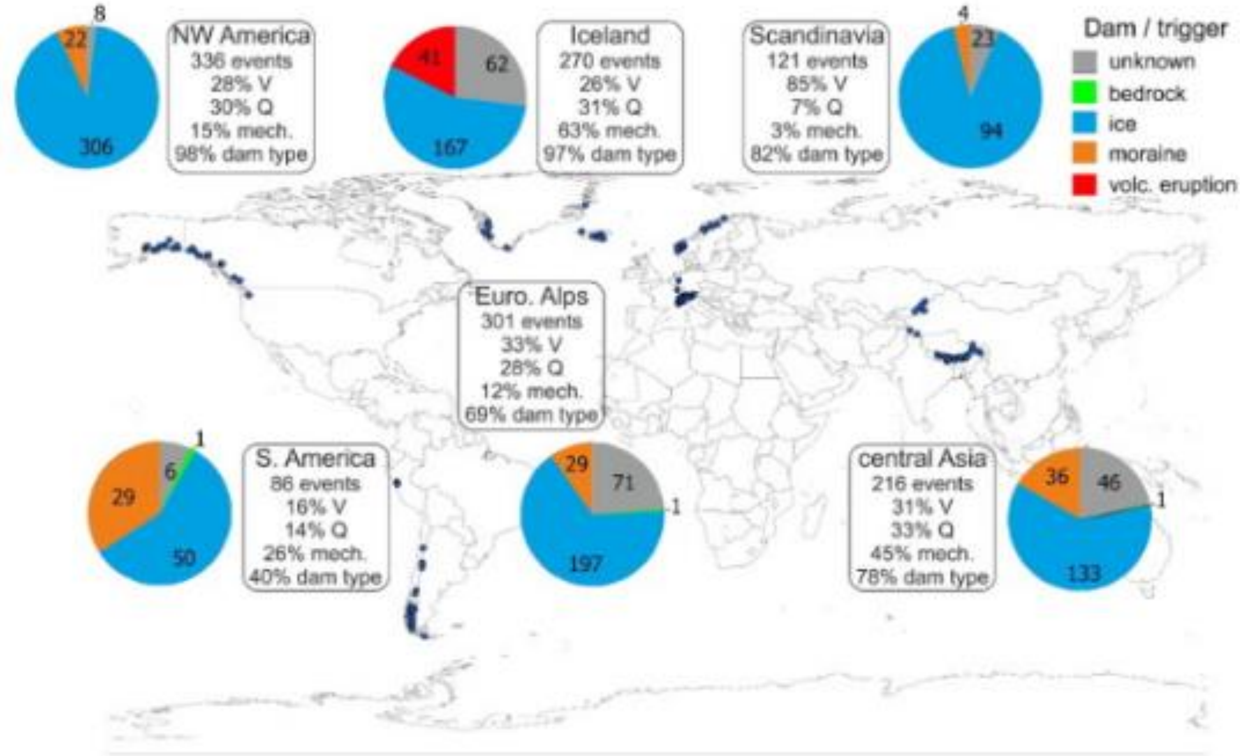


Figure 3: Worldwide location of jøkulhlaups. Numbers of pie charts are the number of floods per dam type/trigger.
Source: (Carrivick and Tweed 2016)

The map over the worldwide geographic distribution is only an estimate based on recorded jøkkulhlaups. It is not certain that the documented outbursts, represent every outburst ever happened because There is a lack of historical documentation. Until recent years there has not been an interest in documenting such events and previous events might have occurred without being documented. Because their occurrence is mainly restricted to poorly or not populated rural mountainous areas, events might also have occurred without anyone noticing. Reconstruction of previous events by geomorphological investigations is possible but reconstruction of historical events is difficult, if not impossible, considering how glaciated areas constantly change over time. When a glacier advances, it can easily erase and remove evidence of events prior to its change/growth/advancing, which can happen on timescales from less than one day to millions of years (ch.1.4 pdf glaciers). Consequently, the map of Carrivick and Tweed is not an absolute list of jøkulhlaups whereabouts but it shows how important it is to not only focus on Iceland but include other mountainous areas of the world.

Even though it shows that North-West America has had most events (336), followed by the European Alps (301), Iceland (270), Central Asia (216), Scandinavia (121) and South America (86), it does not mean these are the areas where outbursts have caused the most damage and destruction, in terms of relational, material and economic losses. (Björnsson 2003) (Jackson and Ragulina 2014) (Carrivick and Tweed 2016).

Figure 5 A, shows the number of glacier outburst floods per 25 years by major region, showing that the number of jøkulhlaupps has increased in all regions, Scandinavia, South America, North America, European Alps, Central Asia and Iceland. Figure 5 B shows the number of glacier outburst floods as a global cumulative total, also illustrating a global increase in events. This may be due to increased instability of dams, which can be linked to several reasons, for instance an increase in global temperature but it may also have other explanations. It can be due to the increased documentation of events (due to increased focus and interest in jøkulhlaups reseach). Also, the x-axis is limited to displaying records from the last 500 years only. Therefore, the increase in number of events is not necessarily due to the increased frequency of occurrence.

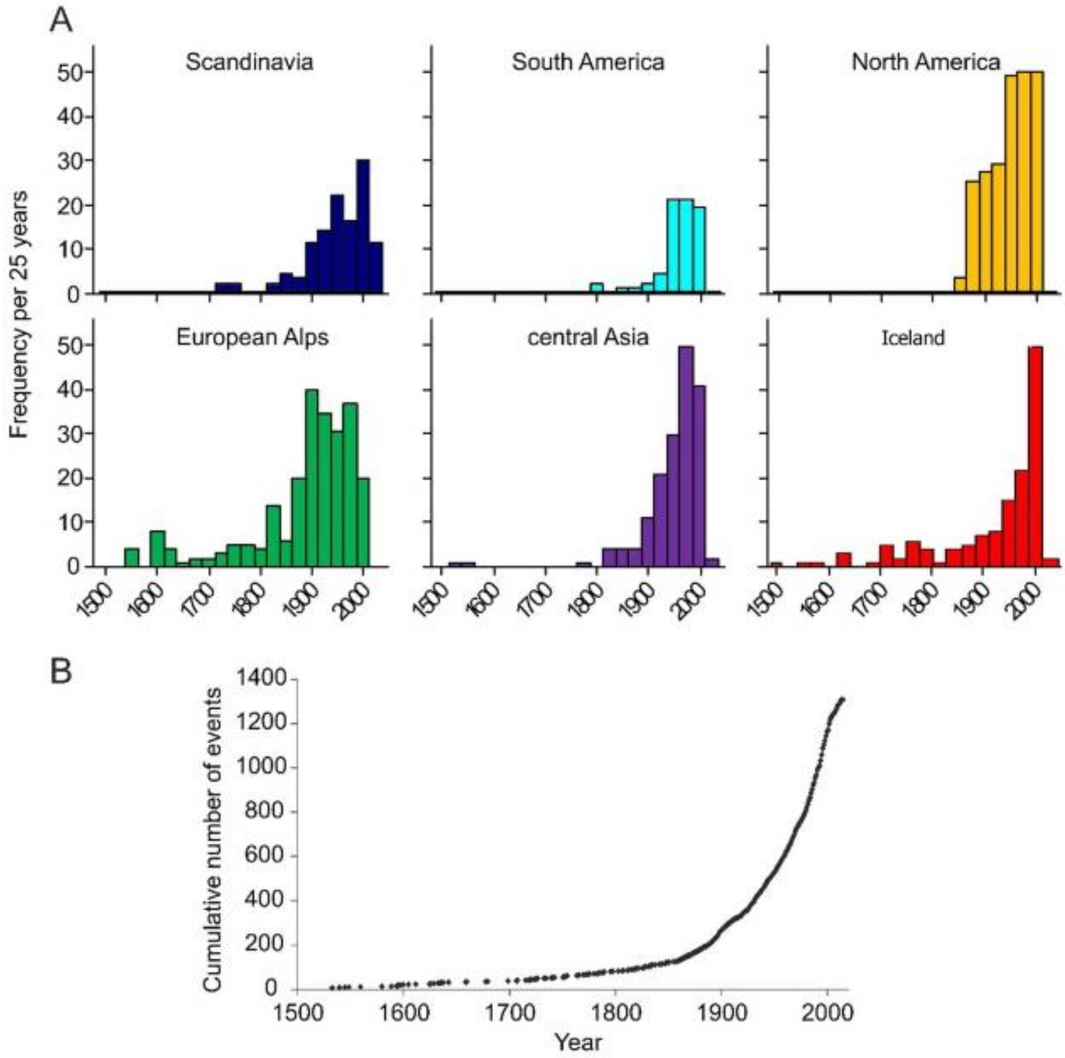
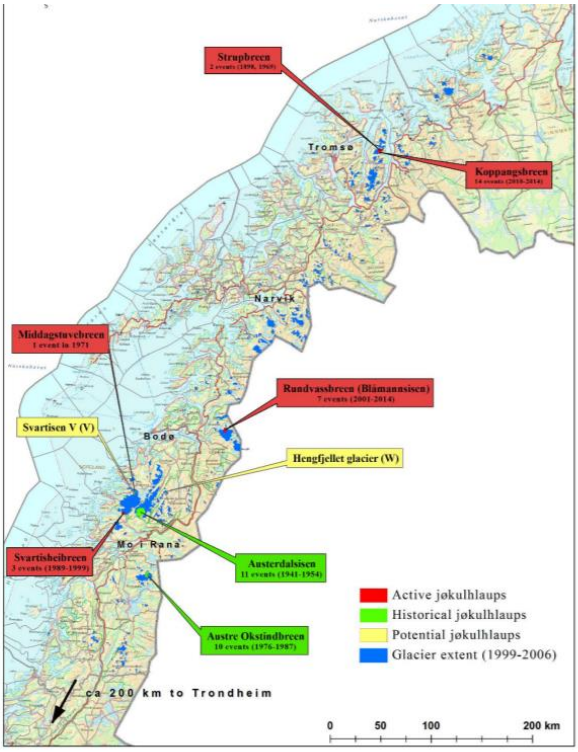
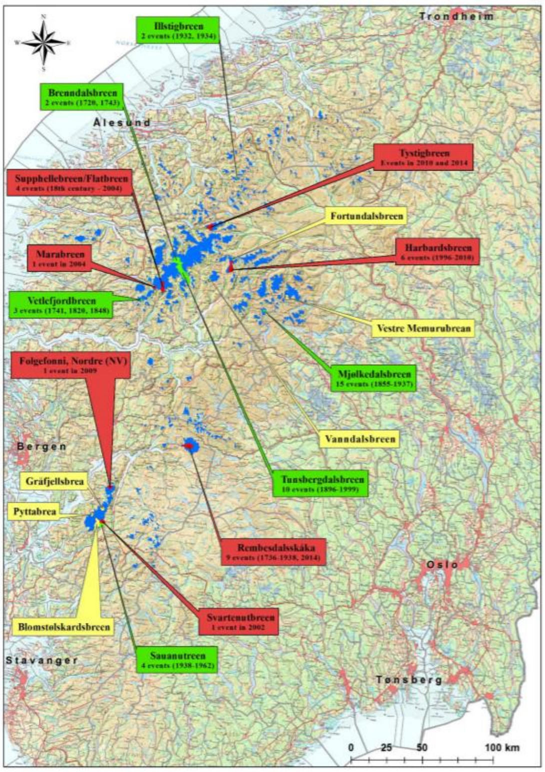


Figure 4: Number of glacier outburst floods per 25 years by major region (A) and as a global cumulative total (B). Note that the x-axis is limited to displaying records from the last 500 years.
Source: (Carrivick and Tweed 2016)

1.2. Former Jøkulhlaups in Norway



Map 1a – Glaciers with jøkulhlaup events in Northern Norway. Svartisheibreen also had an event in 2014.



Map 1b – Glaciers with jøkulhlaup events in Southern Norway.

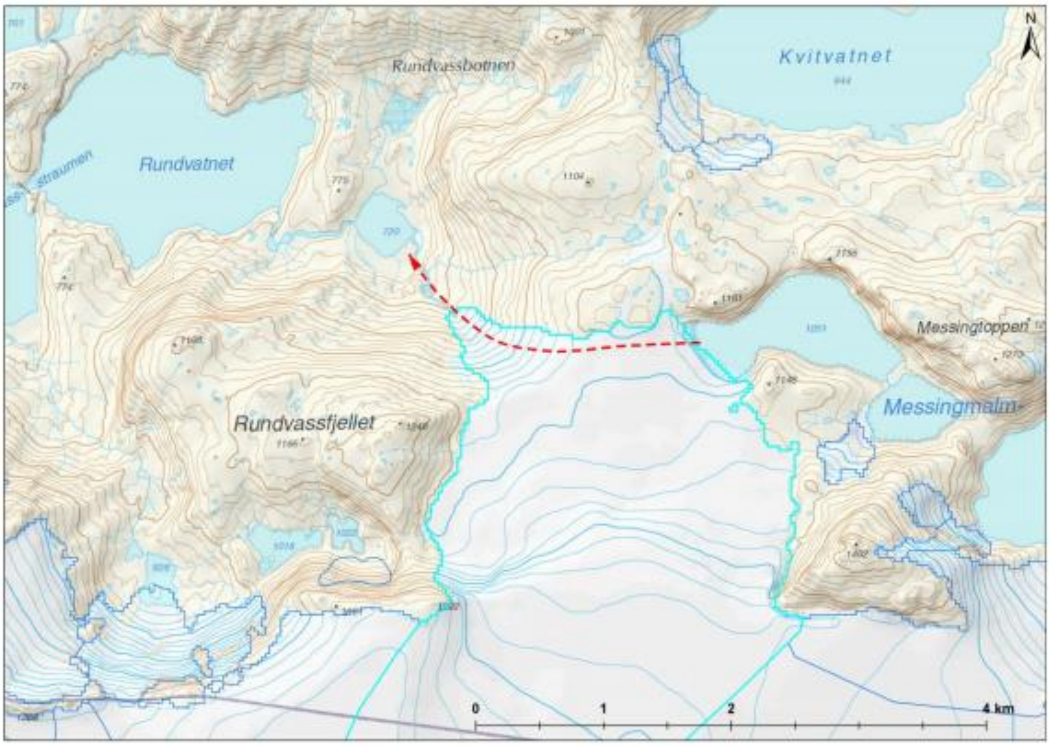
Figure 5

In Norway there are currently at least 20 glacier-dammed or supraglacial lakes. Previously several others have existed but many of these have disappeared due to retreating glacier fronts which no longer prevents water from flowing freely and dams are not formed. Figure 6 and 7 above illustrates the distribution of active jøkulhlaups glacier (red), historical (green) and potential (yellow) in northern, and southern Norway, where an active jøkulhlaup glacier is one where a previous event has occurred and it is thought that an event could occur in the future; a historical classification means that previous events have occurred but the glacier has changed so much that no future events can occur and potential means that there is a glacier-dammed lake or other potential glacier flood hazard but no previous events have been recorded. . An example of a glaciers where a historical jøkulhlaup have occurred, is Brenndalsbreen, an outlet of Jostedalsbreen in Sogn & Fjordane, which had an outburst as far back as in 1720 and another in 1934. Another example is Mjølkedalsbreen in Oppland, southern Jotunheimen, which had 15 jøkulhlaups between 1855 and 1937 [Map 1a,b]. Examples of active jøkulhlaups are Supphellebreen/Flatbreen in Sogndal, Sogn & Fjordane, Engabreen in Meløy, Nordland, and Rundvassbreen/Blåmanssisen in Fauske, Nordland.

The most dangerous jo¨kulhlaups in Norway were from lake Demmevatn at Rembesdalskåka, western outlet of Hardangerjøkulen ice cap in southern Norway, which drains to the populated valley Simadalen. In total, 9 events from this lake has been documented since the 18th century. “After a catastrophic jo¨kulhlaup in 1893, a rock tunnel was constructed to drain the lake artificially. However, a further decrease in glacier volume inflicted another two serious floods in 1937 and 1938 (Elvehøy and others, 2002)“. “A new tunnel completed in 1938 lowered the lake an additional 50 m and a diversion tunnel was constructed in the 1970s for hydropower purposes. However, extensive thinning of the lowermost part of the glacier led to a new flood in summer 2014, the first since 1938.”

Two other sites of special interest are the jøkulhlaups from Supphellebreen/Flatbreen and Blåmannsisen. Supphellebreen/Flatbreen because it is the only moraine-dammed lake in Norway (?), and because an outburst in 2004 triggered a 240 000 m3 debris flow, which caused extensive property damage but there were no injuries or loss of life. Blåmannsisen is of interest because the glacier-dammed lake previously drained the opposite direction. Blåmannsisen in Nordland is of special interest because the glacier-dammed lake upper Messingmalmvatn (usually called Vatn 1051) previously drained the opposite direction, eastwards over the border into Sweden. On 6th September 2001, the first known jøkulhlaup from Blåmannsisen occurred when 40 million cubic metres of water were suddenly released from an glacier-dammed lake over the course of 35 hours, draining westwards. Almost the entire lake drained through a 4.5 km long tunnel under the glacier Rundvassbreen, which caused the lake level to fall by up to 50 m (Engeset, 2002). The outburst did not cause any injuries or loss of life or caused any material damage, but was on the contrary economically beneficial because it drained into Elkem ASA’s hydropower reservoir, Sisovatn, which is where the jøkulhlaup first was detected since the water level inexplicably rose 2.4 m over a short period. Since then, several more outbursts have occurred in successive years and the glacier have glacier-dammed lake have been closely monitored since the initial event. Measurements of the glacier surface near the glacier-dammed lake showed a thinning of about 26 m between 1961 and 2001 (Engeset et al, 2005)

After the initial jøkulhlaup in 2001, the lake took three years to refill to its previous level, and water once again drained eastwards. It was not until late August 2005 (27th – 29th) that another jøkulhlaup occurred. There was heavy rainfall in the days preceding the jøkulhlaup, which may have triggered the event. During this second jøkulhlaup, 35 million cubic metres of water were released over a period of 36 hours. The jøkulhlaups in 2001 and 2005 occured when the lake was completely full. However, the jøkulhlaups in 2007 (29th August) and in 2009 (6th – 7th September) drained from a lake that was only half-full. The time between subsequent events decreased from 4 years to 2 years. The next jøkulhlaup occurred just a year later in September 2010 when the lake was less than half-full. Water drained under glacier here 189 of 11 m between 2002 and 2009, and this continued presumably in 2010, allowing water to escape at a lower water level than previously (Kjøllmoen et al., 2011). A new jøkulhlaup occurred one year later in September 2011, and again the lake was less than half-full. The water level in Vatn 1051 was about 12-13 m higher in September 2013 than it was just before the event in 2011. The glacier surface adjacent to the lake was 1.4 m lower compared with measurements from October 2011.Thus, ice pressure decreased and water pressure increased. Subsequent events occurred on 10th-12th August 2014 when for the first time since 2005 the glacier-dammed lake was more than half-full before the initiation of the jøkulhlaup, and then on 28th -29th September 2016 with a volume of about 26 million cubic metres (Kjøllmoen et al, 2017). The most recent event wason 25th to 26th August 2018, with a volume of approximately 2 million cubic metres, calculated from the increase in lake level of Sisovatnet, downstream from Rundvassbreen. The pattern of events shows that the mechanism of the initiation of a flood event is complex and that Blåmannsisen is still able to surprise us.



Map: Norge digitalt. The red dashed line shows the probable drainage path during a jøkulhlaup.

Figure 6: The red dashed line shows the probable drainage path during a jøkulhlaup from Blåmannsisen

The first known jo¨kulhlaup from the glacier-dammed lake Øvre Messingmalmvatn occurred in September 2001, releasing 4.0 107 m3 of water. The lake emptied in about 35 hours, starting at noon on 6 September 2001. Rundvassbreen, an outlet glacier from Bla˚mannsisen, dammed the lake that used to drain via a superficial spillway towards the east, on its western shore. The retreat of Rundvassbreen since 1961, revealed by map comparison, and results from mass-balance modelling indicate that the glacier decreased in volume. This is supported by the observed decrease of ice thickness in the ablation zone. The rate of thinning has increased since 1998 and presumably affected the stability of the ice dam so that the water impounded in the lake can now break the seal at the glacier bed and drain underneath the glacier. In September 2001, the lake water level was about 40 m below that required to float the ice dam. In line with the subflotation triggering, the asymmetrical shape of the flood hydrograph suggests that drainage occurred through a subglacial channel, the size of which was progressively enlarged by melting due to the dissipation of energy with the flowing water.

Table 1: All known events from koppangsbreen, which has occurred from 2010 to 2014. not included, is the latest jøkulhlaup that occurred 25.8.2018.
Source: (Jackson and Ragulina, 2014)

Table 14-2
Dates and approximate volumes of jøkulhlaups from Blåmannsisen. The volume in 2001 was estimated from the change in volume of water in the reservoir Sisovatnet. The water level after each event differed, and also accounts for different volumes calculated at the same pre-event water level.

Year	Date	Water volume	Water level before event
2001	5^{th} – 7^{th} September	~35 mill. m^3	≈ full (1053 m a.s.l.)
2005	27^{th} – 29^{th} August	35 mill. m^3	≈ full (1053 m a.s.l.)
2007	29^{th} August	20 mill. m^3	~ half-full
2009	6^{th} – 7^{th} September	20 mill. m^3	~ half-full
2010	8^{th} – 17^{th} September	11 mill. m^3	less than half-full
2011	22^{nd} September	12 mill. m^3	less than half-full (1029 m a.s.l.)
2014	10^{th} – 12^{th} August	35 mill. m^3	≈ full (probably 1053 m a.s.l.)
2016	28^{th} – 29^{th} September	26 mill. m^3	> half-full (estimated 1040.7 m a.s.l.)

1.3. Study Area

Koppangsbreen glacier (69.69° N 20.14° E) dams the glacier-dammed lake that drained 9 times during summer 2013. It is located on Lyngen peninsula in the Region of Troms (figure 9), where it is surrounded by several peaks of 1000 m.a.s.l., covers 4.14 km^2 that ranges from 509 to 1203 m.a.s.l. The glacier-dammed lake is located where rain- and meltwater is blocked between Store Koppangstinden (1224 m.a.s.l.) to the north and the glacier front to the south. Previously, the water had drained steadily eastwards through a small stream. During the jøkulhlaups, the water drained southwards, presumably through a subglacial tunnel which lead the water to an outlet close to the location of the glacial stream. From there, the water followed the same flow path as the glacial stream to Koppangsvatnet lake, and further down towards Koppangen village in Koppangselva river (figure 9).

Map: Norge digitalt

Figure 7: *Location of Koppangsbreen, Koppangen village and the glacier dammed lake, the latter illustrated with a red dot. The outlines of Koppangsbreen are highlighted with the blue line. Location of the study site within Norway is shown on the upper map to the right.*
Source: http://www.norgeskart.no

The first known jøkulhlaup from Koppangsbreen occurred 6th September 2010. Between then and 2014, a total of 14 events have occurred, at which 9 occurred between 4th June and 3rd September 2013. Normal repeat interval is one or several years, thus the fact that 7 events occurred in 30 days only is exceptional and extraordinary. Jøkulhlaups recurring at only a few days' intervals, has never previously been documented. In 2017 the thickness of the glacier dam had decreased enough for water to constantly flow freely underneath and the glacier tongue appeared to no longer be damming the lake.

Norge digitalt (27.07.2006). The blue arrows show where the water flowed during the first 2013 event.

Figure 8: Map of study area. Numbers 1 and 2 show the drainage outlet at the glacier snout during the jøkulhlaup and the drainage path before the jøkulhlaup respectively

The Lyngen area does not have a long history of glacier measurements, but it is assumed that glaciers in Troms were at their greatest extent around 1900s (Bakke m.fl. 2005). Since then, Norwegian glaciers have, in general, decreased although some have had shorter periods of increase (Andreassen m.fl. 2005). The first measurements of Koppangsbreen began in 1998, when NVE began registering its glacier length. Previous mapping of glaciers in this area is based on aerial photographs and maps from 1952-1971 (Østrem m.fl. 1973) which later has been improved using satellite photos from 1988 and 2001, as well as orthophotos and laser scanning in 2010 (Hausberg og Andreassen 2009). Altogether, these studies show that glaciers in this area shrink at a faster rate than glaciers further south in Norway, such as Engabreen in Nordland. (Jackson and Ragulina, 2014).

Since the Lyngen area does not have a long history of glacier measurements, knowledge of the climatic and glacial history of Koppangsbreen is limited. However, it is assumed that glaciers in Troms were at their greatest extent around 1900s (Bakke m.fl. 2005) and that Norwegian glaciers in general have decreased since then, apart from some shorter periods of increase (Andreassen m.fl. 2005).

Table 2: file:///C:/Users/Guro/Downloads/2010_NVE_rapport_Lyngen_1juni_2010.pdf

Period	Glacier area reduction (km^2)	Total decrease (%)	Annual decrease (%)
1955-1988	4,06 - 3,77	7,1	0,22
1988-2001	3,77 - 3,68	2,4	0,19
2001-2010	3,68 - 3,26	7,7	1.34

The first measurements of Koppangsbreen began in 1998, when NVE began registering its glacier length. Between 1978 and 2006, the glacier front retreated more than 300 m. Height models between 1955 and 2010 show that glacier thickness been lowered by 34 m. Previous mapping of glaciers in this area is based on aerial photographs and maps from 1952-1971 (Østrem m.fl. 1973) which later has been improved using satellite photos from 1988 and 2001, as well as orthophotos and laser scanning in 2010 (Hausberg og Andreassen 2009) [Jackson, Ragulina, 2014]. Altogether, these studies showed that glaciers in this area have shrunkat a faster rate than glaciers further south in Norway, such as Engabreen in Nordland. In 2011, a study on three specific glacier areas in Lyngen, Koppangsbreen included, was conducted. This analysis concluded that all three glaciers have decreased in extent and thickness since 1956. In table 4, the glacier area and changes for three periods are presented. In total, the glacier area decreased by 13 % between 1955-2010, from 4.06 km2 to 3.26 km2. The decrease in annual reduction rate for the second period, between 1988 and 2001, corresponds with observed climate conditions in the 1990s when there was a period of colder weather with more precipitation. In this period, many of the more mairitime glaciers in Norway grew. Altogether, this means that the area of Koppangsbreen gradually decreased between 1955 and 2010, but that the rate of decrease has increased since 2001.

1.4. The Jøkulhlaups Of 2013

The duration and volumes of the events prior to 2013 are not known, but observations by local population and local media informs that the first event of 2010 was so violent it changed the original river course and almost cut off a house in Koppangen village (Miriam Jackson, personal communication). The second outburst of 2011 was smaller than the first and the following two in summer and late autumn 2012, were both minor.

The first event of 2013 occurred on 4th June and lasted for about 11 hours. Estimations by NVE gave an approximate flood volume release of 1.9 million m^3 water, based on GNSS measurements of the lake level prior to and after drainage on 5th June, the calculated lake area on the ortho image from 16 August 2011. The water level prior to drainage was based on stranded blocks of snow on the ground, indicating initial shoreline at 525 m.a.s.l.

Figure 9: Koppangsbreen glacier-dammed lake after drainage 5th June 2013. Initial shorelines can be seen by stranded snow blocks.
Photo: Miriam Jackson, NVE

The following 6 events occurred at intervals of 3-10 days and lasted for 1-2 hours, see table 6. The two last events occurred at intervals of more than one month and drained in 20 hours and 8 hours, respectively.

Table 3: Date, time, and period length of All known events from Koppangsbreen, which has occurred from 2010 to 2014.

Date of event	Time	Duration
6 September 2010	-	-
2011	-	-
Summer 2012	-	-
Automn 2012	-	-
04.06.2013	kl. 15-02	11 h
09.06.2013	kl. 09-11	2 h
19.06.2013	kl 13-15	~2 h
23.06.2013	kl 17:30-19:30	~2 h
26.06.2013	kl 08-09:30	~1,5 h
30.06.2013	kl 15-16:20	~1 h 20 min
03.07.2013	kl 19:20-20:25	~1 h 5 min
12.08.2013	20 t	20 h
03.09.2013	kl 21-05	8 h

GNSS measurements of the water level after drainage on 27th June 2013, showed that lake level change prior to and after drainage was 509 m.a.s.l. and 506 m.a.s.l., respectively. The lake level prior to the events was based on field observations of initial shoreline. The flood volumes released during the jøkulhlaup on 3rd July were smaller and due to the protective actions conducted after 30th of June, the flood protections were not damaged. Another two events followed before the summer season of 2013 was over. The last jøkulhlaup from Koppangsbreen occurred the 17th June 2014 and had a duration of about 20 hours

1.5. Objectives

The objectives of this study are to analyse the recurring jøkulhlaups, or Glacier Lake Outburst Floods (GLOF), from Koppangsbreen in northern Norway during the summer of 2013. The aim of this study is to examine why these events occurred and why they occurred in such a short period in 2013. In order to accomplish this, we will first use various data collected by NVE and published in NVE reports, complemented by available satellite and aerial ortho images during the past decades. In addition, we will collect measurements of the glacier thickness in order to calculate the stability of the ice damming the lake. Finally, we will combine meteorological data to estimate the water production and input into the lake.

The basic questions that govern this analysis is: Why did the glacier dam fail although it had been stable before? What changes of the glacier or the lake might explain the sudden instability? Did the lake level increase enough for water to escape underneath the glacier between each event?

2. DATA

Since there are no persistent measurements from/of Koppangsbreen, the amount of data available is limited. This thesis is therefore based on data collected by a variety of methods.

2.1. Glacier Geometry

Changes in glacier geometry was studied by several means, including satellite photos and topographic maps, which were compared to a study of glacier change in northern Norway (Winswold, Andreassen, 2010). The survey of glacier changes in northern Norway by Andreassen and Winswold is based on topographic maps from the 1950s, satellite photos from 1988 and 2001, and by laser scanning and ortho image of 2010. Moreover, three orthoimages taken by Terratec AS (norgeibilder.no) are available on 5.8.2006, 19.9.2011 and 16.9.2016, which help calculate the lake area. ASTER orthophotos and DEMs are available in 2001, 2008, 2012, 2013 and 2017 (Nuth, personal communication). These are generated by using the procedures of Girod et. al. (2017), which provide a DEM accuracy of up to 5 meters.

2.2. Glacier thickness

To map the glacier thickness and the glacier and lake changes between the most recent jøkulhlaup and present day, field measurements from the ground and from a helicopter were collected on 26th September and 24th October 2017, in cooperation with the University of Oslo (UiO) and the Norwegian Water Resources and Energy Directorate (NVE). Additionally, ground-penetrating radar (GPR) was used to detect depth to bedrock, map the bedrock topography and to find the current thickness of the glacier ice, on 24th October 2017. A total of 11 GPR profiles were collected across the glaciers tongue (Figure 11), close to the glacier-dammed lake. The GPR profiling was collected with a MALÅ operational system, a 50 MHz antenna and furthermore processed in ReflexW Software.

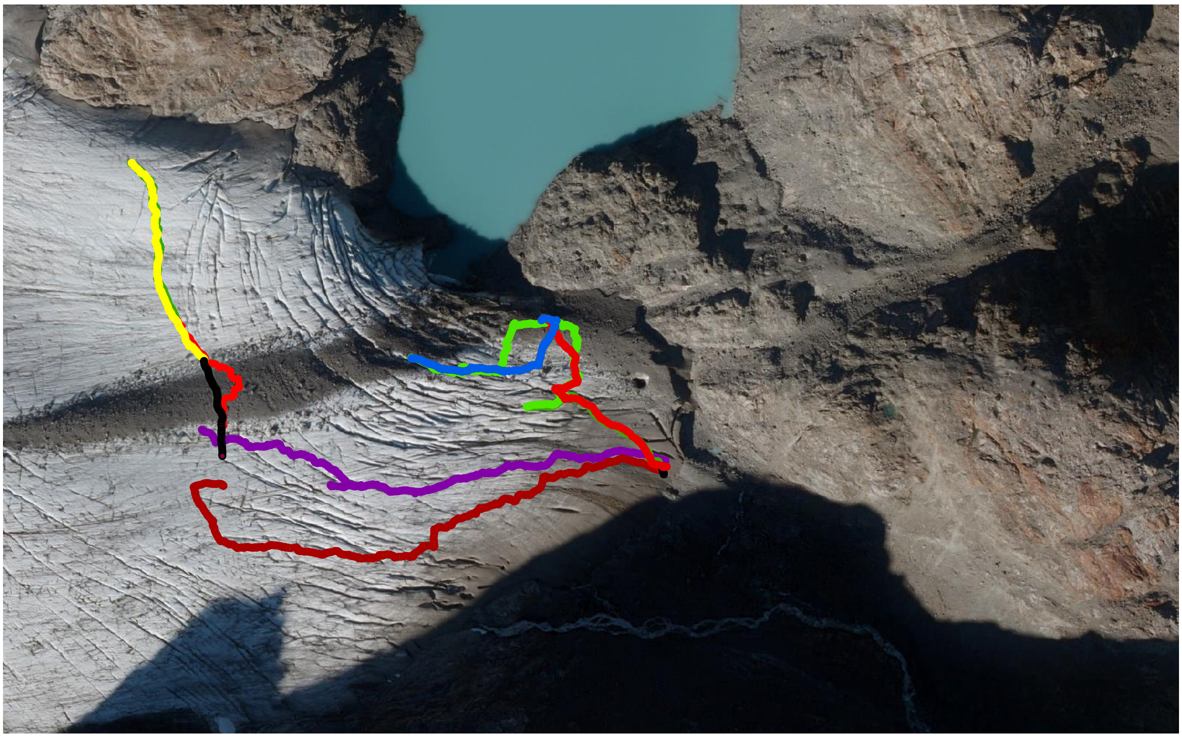


Figure 10: Map showing location of the collected GPR profiles, illustrated with coloured lines.

2.3. Lake levels

Global Navigation Satellite System (GNSS) was used by NVE to survey the level of the lake on 5.6.2013 and 27.6.2013 (Jackson and Ragulina, 2014). On 26.9.2017 and 24.10.2017 additional GNSS surveys of the lake were made in field, in cooperation with NVE and UiO (NVE internal note, December 2017)

2.4. Meteorological data

Meteorological data was used to calculate inflow to the glacier-dammed lake between each event. Since there is no meteorological station situated by Koppangsbreen glacier, meteorological data was downloaded from the open portal site seNorge.no. The site provides diurnal time series of interpolated meteorological data for every square kilometer of Norway from 1957 and up to present day. It is a cooperation between NVE (nve.no), the Norwegian Meteorological Institute (met.no) and the Norwegian Mapping Authority (statkart.no). The meteorological and other relevant data are presented on the senorge.no website under four main themes; snow, water, weather and climate. The interpolated meteorological data is used as input to a snow model to model other relevant parameters such as runoff. The meteorological and hydrological data used in this thesis, are air temperature, precipitation and snowmelt (given as metres water equivalent). The two latter were used to calculate runoff from the glacier, and air tempreuatre was used to calculate melting glacier ice by using a degree-day model for the selected area.

The snow model used in SeNorge.no interpolates data from the closest situated meteorological stations. For Koppangsbreen, there are only two nearby meteorological stations that measured parameters of interest over the investigated period. These are

Nordnesfjellet and Gjerdvassbu meteorological stations, illustrated with red dots close to Lyngseidet on the map in figure 12. Nordnesfjellet is furthest to the east, whereas Gjerdvassbu is furthest to the west. The meteorological station between these two, was installed after the events in 2013. The location of other nearby meteorological stations can also be seen but data from them were not used in this thesis.

Nordnesfjellet was installed in January 2010 and is situated 13 km southeast, across the fjord. It measures precipitation and temperature. Gjerdvassbu is situated 11 km to the south and provides records of precipitation, wind, temperature and snow depth since November 2011. This means that the interpolated precipitation and snow depth values are based on only one nearby source, whilst the interpolated temperature values are based on only two. In a coastal climate, such as Lyngen, variations in local weather is common, hence it is possible that the interpolated data for Koppangsbreen provide values of some uncertainty, depending on the sources of error associated with the snow model. This influences the reliability of each calculation used to estimate inflow to the glacier-dammed lake later in this thesis.

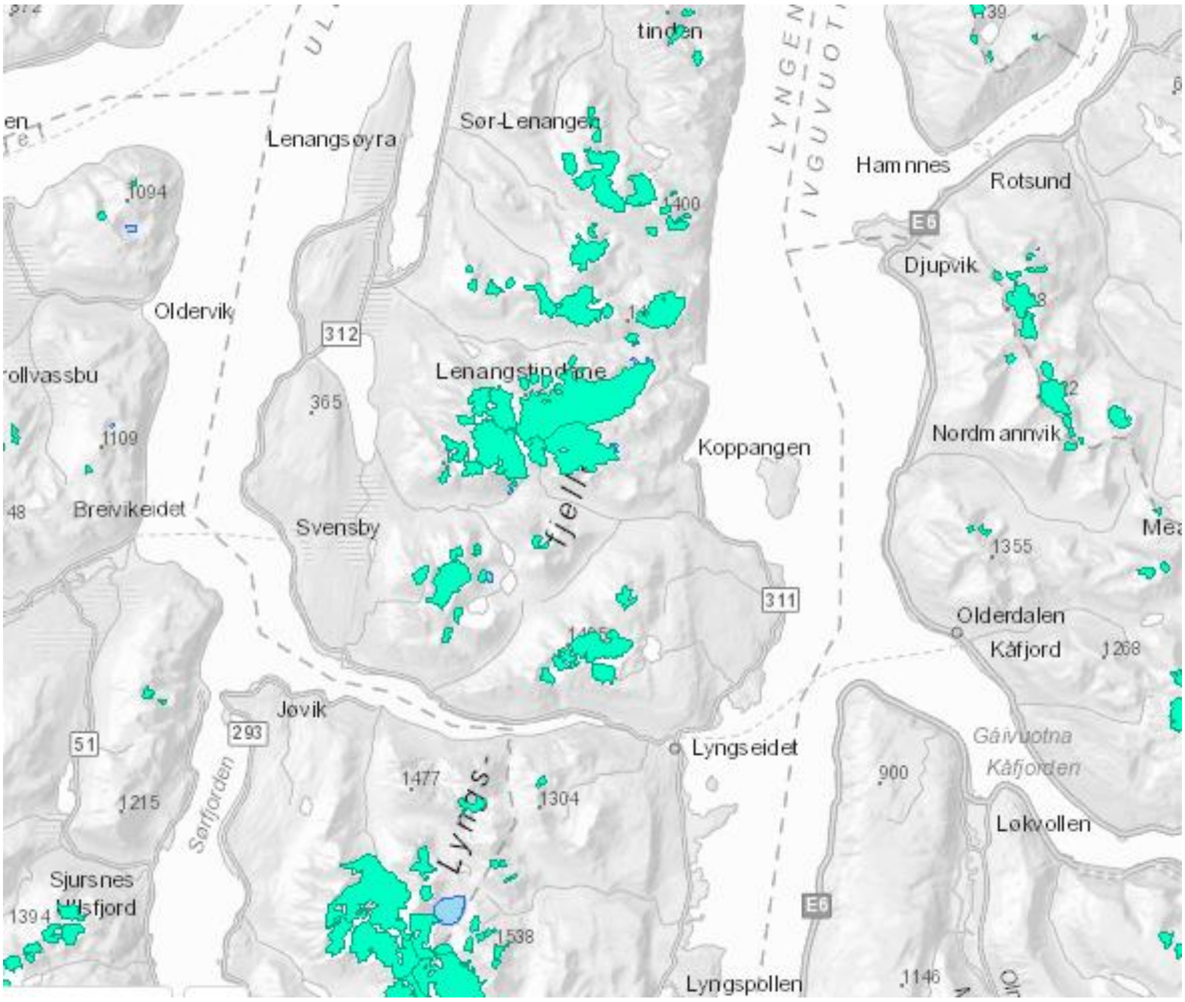


Figure 11: Location of the closest meteorological stations, situated near Lyngseidet. Nordnesfjellet is the red dot furthest to the east, whereas Gjerdvassbu is furthest to the west.
Source: Miriam Jackson, NVE

3. METHOD

3.1. Glacier Geometry

Glacier geometry was studied in order to analyse/investigate the stability of the ice damming the lake, determined by the relationship between thickness of the glacier dam and lake volume.

GLACIER-DAMMED LAKE AREA

Glacier and lake area was calculated for 2006, 2011 and 2016. The lake areas were digitized from the ortho images of the respective dates, in the geospatial processing program ArcMap 10.4. The areas were then calculated by the use of tools in the same program. The glacier and lake area for 2016 was assumed to represent the glacier and lake area in 2017. Likewise, glacier and lake area, subglacial tunnel length and distance from the glacier-dammed lake to the sea in 2013, has been based assumed to be as in September 2011, two years prior to the events. Considering how lake area change with changing lake volumes, and that a jøkulhlaup occurred at an unknown date in 2011, inflow estimations based on this area might be inaccurate.

CATCHMENT AREA OF THE KOPPANGSBREEN GLACIER-DAMMED LAKE

The catchment area of the Koppangsbreen glacier-dammed lake was generated in NEVINA, NVEs hydrological application containing national catchment areas, which altogether constitute the database of the national hydrographic division of catchment areas in Norway, REGINE. NEVINA is meant for larger areas hence the boundaries of the catchment area was therefore adjusted manually by digitizing in NEVINA, to more accurately follow the contour lines of topographic maps from hoydedata.no. The catchment area was furthermore imported as a polygon to ArcMap and calculated by using the program tools.

JØKULHLAUP/FLOOD VOLUMES

To estimate the total volumes released during the jøkulhlaups, the change in water level prior to and after an event was multiplied with the original lake area before drainage (m^2).

Equation 1:

$$\Delta\, lake\ volume = \Delta\, lake\ level \cdot lake\ area\ before\ drainage$$

The same formula was used to calculate what the estimated inflows between each event in 2013 would mean for the rise in water level. The change in water level prior to and after an event from June 2013, was based on NVEs GNSS measurements of the water level after drainage and observed shorelines from stranded snow blocks. The GNSS measurements from September and October 2017, was used to calculate lake volume change between 2013 and present-day.

AVERAGE DISCHARGE

Average discharge was calculated by dividing released lake volume by the duration of the flood, to get insight in the intensity of each flood.

Equation 2:

$$\frac{flood\ volume\ (m^3)}{duration\ (s)} = average\ discharge\ (\frac{m^3}{s})$$

The flood durations are based on observations from the local population and workers in the field.

3.2. Glacier Thickness

Ground Penetrating Radar (GPR) profiles were collected to detect depth to bedrock thus the current thickness of the glacier dam. Basically, the radar transmitter sends electromagnetic waves down into and through the surface, while the receiver records the time it takes for the waves to be reflected and returned to the surface (REFERENCE). This is called the travel time or two-way travel time, since the waves that travel down return up back to the surface again. The travel time can later be converted into depth assuming the electromagnetic permissivity remains constant.

Electromagnetic (EM) waves are reflected at the transition zones between layers of different densities, chemistry, and general characteristics. For glaciers, cold and pure ice will give the best images as it is relatively transparent to electromagnetic (EM) waves, while water-filled voids will cause diffractions and create scattered profiles (Bamber, 1987#). The steep area surrounding Koppangsbreen, could also cause diffractions and scattered profiles due to reflections from nearby bedrock, which could absorb the signal energy before it reaches the bottom of the glacier and the transition zone of interest. Ideally, the transition zone between ice and bedrock would be detected, and in theory it would also be possible to record the transition between ice, air or water in a subglacial tunnel, and then the underlying bedrock if the profiles were taken just above the tunnel location.

The formula used to convert two-way travel time (TWT) to depth in meters [m], is as follows:

Equation 3:

$$\frac{V(TWT \cdot 10^{-9})}{2} = \frac{1{,}67 \cdot 10^8\ \cdot \frac{m}{s}\ (TWT \cdot 10^{-9}\ s)}{2} = \frac{1{,}67 \cdot 10^8\ (TWT \cdot 10^{-9})}{2}$$

Where V is the electromagnetic permissivity, the main parameter used to calculate radar wave velocity in ice, which is a constant of 1.67 * 10^8 [m/s] (Sevestre, Cuffey and Patterson) and TWT is the two-way travel time in nanoseconds [ns].

Figure 11 under data shows the location of the profiles on the glacier, illustrated by the coloured lines. Figure 27 shows for which path of the profiles the two-way travel time has been converted to depth, illustrated with round points. The ortho image from 2016 was used as background layer, to give the most precise and accurate presentation relative to rocks, moraine and distance to mountainsides along the glacier edge and front.

3.3. Discharge Calculations

Discharge calculations were made in order to obtain further insight in the possible flood mechanism, investigate the potential flotation pressure of the glacier dam and to see whether there was significant input to the glacier-dammed lake between each event.

CATCHMENT AREA DETERMINATION

To calculate input to the glacier-dammed lake, the catchment area first had to be determined. This was done in NEVINA, an NVE web-based application used for delineating catchments areas. The catchment area for Koppangsbreen and the glacier-dammed lake was then calculated automatically in ArcMap.

GRID CELL DIVISION

Estimates of input were based on runoff from each square kilometre within the catchment area. The runoff from each square kilometre, was calculated based on meteorological data interpolated for each grid cell. Since interpolated meteorological data could differ between two adjacent grid cells, the catchment area had to be divided into the same grid cells as used in the snow model in SeNorge.no.

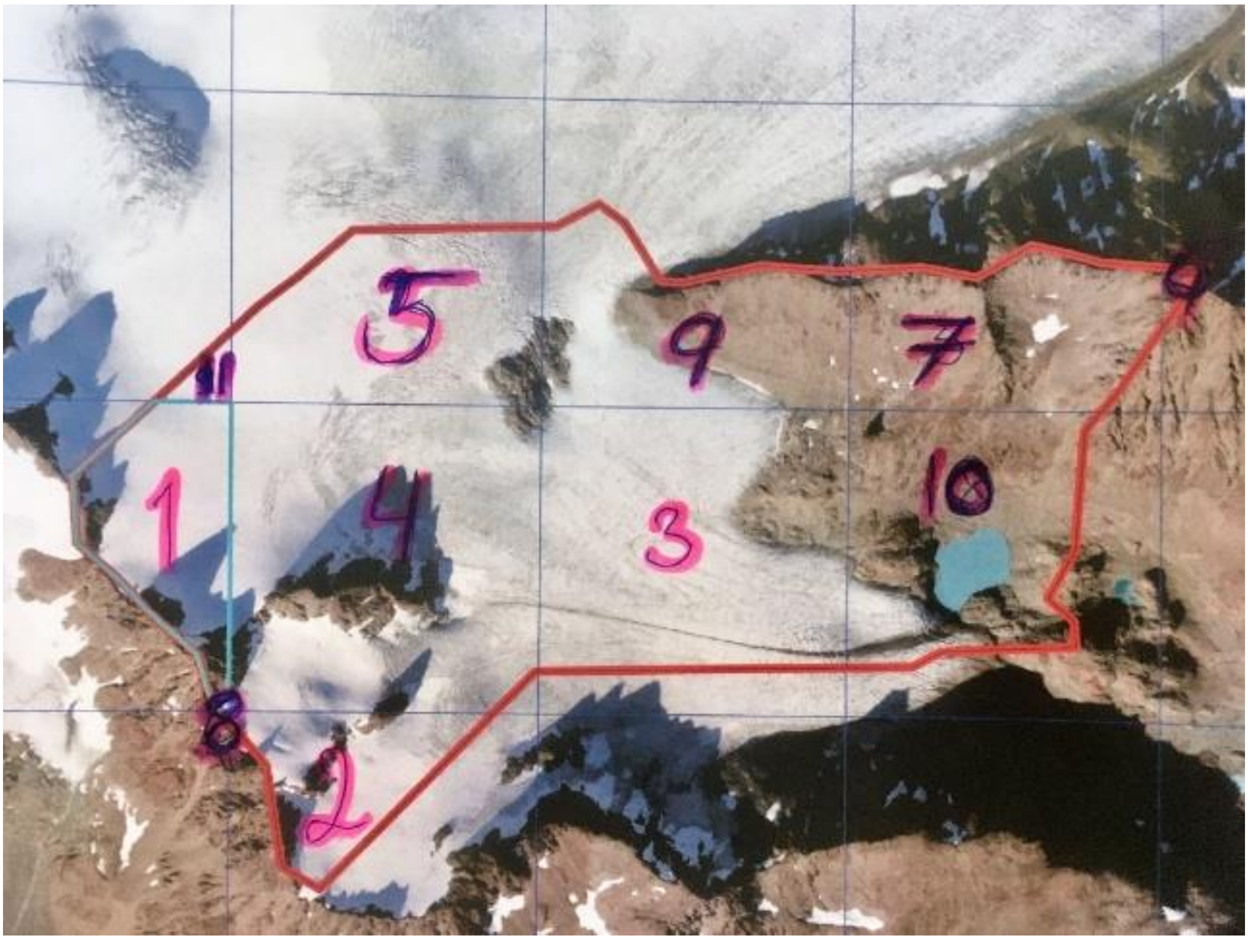


Figure 12: *Grid cells used in the calculation in SeNorge, with the generated catchment area and an aerial photo of Koppangsbreen from 2016 as background layer*

In total, the catchment area of the glacier-dammed lake consists of portions of 11 grid cells. These are numbered 1-11 in figure 13. Furthermore, the size of the catchment area within each specific grid cell was determined. The area was calculated by using geospatial program tools in ArcMap (table 3). Some of the grid cells cover very small areas of the catchment area, but these were included in the first estimate of inflow for the sake of completeness

Table 4: *Size of catchment area, percentage catchment area and percentage glacier within the 11 cells that constitute the catchment area*

GRID CELL	Catchment area (m^2) to desimaler	Percentage of grid cell in catchment area	Glacier area 2011 (m^2)	Percentage glacier cover in the area of each grid cell used
1	309 980	0,3	250 288	0,8
2	256 659	0,3	238 908	0,9
3	852 209	0,9	793 216	0,9
4	989 315	1,0	907 160	0,9
5	506 181	0,5	506 176	1,0
6	12 142	0,0	0	0,0
7	456 226	0,5	0	0,0
8	447	0,0	0	0,0
9	499 133	0,5	300 505	0,6
10	586 535	0,6	235 975	0,4
11	27 458	0,0	27 458	1,0
TOTAL	4 496 285		3 259 685	

To calculate runoff from glacial melt within each grid cell, the percentage of glacier ice within each grid cell, if present, was calculated. These are also shown in table 3. The percentage glacier area was based on calculated glacier area from the ortho image of 2011, by tools and functions in ArcMap.

GLACIER MELT ESTIMATION

Calculations of inflow between each event were based on the interpolated meteorological data from SeNorge.no. The calculations were based on three sources, precipitation, snow melt and glacier ice melt. senorge.no was used to get the sum of precipitation and snow melt. The amount of glacier-ice melt was calculated by using the terrain and degree-day model, based on water equivalents and daily mean temperatures.

For snow and ice to melt, their temp must first increase to melting point 0°C before melting of snow can occur (Hock, 2003). Ice does not start to melt before the overlying snow has melted (because the snow will then be insulating and will prevent the ice from melting). In addition, ice cannot melt before its temperature exceeds 0°C, which means there might still not be melting even if the air temperature > 0°C because heat will then be used to heat the snow and ice pack from below zero to 0°C (Hock, 2003). The water equivalent given in senorge is a rough estimate of when ice melting begins, as this occurs when the water equivalent is shown as zero. Then there is no more snow on the ground or on the glacier and melting can occur.

The meteorological data for snowpack (swe>0) says when there is snow covering the ground and also gives an estimate of the potential metres water equivalent of the remaining snow based on the calculated density.. The amount of snow is given in water equivalents (s.w.e.) The snow model used in senorge takes into account that the snow pack can hold up to 10 % free water and the values for rain- and snowmelt are therefore based only on runoff from the snow pack. This is not the case for the amount of inflow to the lake coming from melting glacier ice, because the degree-day model is based only on daily mean temperatures and the degree-day factor.

However, the snow model sets the water equivalent equal to zero when less than 50 % of the grid cell is covered by snow, which means that in senorge the grid cell is snow free whilst in reality the ground could still be covered by 49 % snow. This is seen in a picture taken on 5th June 2013, where snow is clearly visible on the ground but in senorge, the water equivalent for this day was zero. This shows that the inflow calculated in this thesis is an approximation, since it are based on volumetric runoff from glacial melt found by using this model when it is assumed the snow has melted.

The amount of glacier ice melt is estimated by multiplying the temperature values for the days where swe=0 with the degree day factor (6.4 mm/day/degree). This gives an approximate w.e. of ice melt for each cell. This applies only for the cells where there is glacier ice, which excludes gridcells number 6, 7 and 8. The values for cell 1-5 and 9-11 must further be multiplied with the percentage glacier ice constituting each cell.

When the swe=0 it is estimated that there was no more snow on the ground (which is not always true because as mentioned previously, pictures taken on 5th June, the day after the first outburst, shows that there is still some snow covering the ground around the glacier lake - an area that is covered by cell number 10, which in senorge shows a swe-value equal to zero on this date. So the estimated values are not always 100% true to what was the reality).

When water equivalent = 0, the amount of melting glacier ice can be calculated from the degree day-method: the temperature for that date (in respective grid cell) is multiplied by 6.4 mm/day/degree to get amount of mm melt per day per grid cell

INFLOW ESTIMATIONS

Interpolated meteorological data from senorge model is based on the closest meteorological stations, so the actual values locally are often quite different than the interpolated values. This is particularly relevant for the values of rain- and snowmelt because there is significant topography in this area.

The Diurnal time series of interpolated meteorological data for the three parameters daily mean temperature (°C), rain and snowmelt (mm) were downloaded from 1.5.2013-23.9.2013, for each of the 11 grid cells. Inflow from rain and snowmelt was calculated by summarizing daily volumetric values for each square meter between each event. Runoff from glacial melt was calculated based on a terrain and degree-day model. The values for rain- and snowmelt were converted to volumetric units by multiplying with the size of the catchment area within each cell. This gave daily volumetric values of runoff from rain and snowmelt from each square meter of the catchment area.

Equation 4:

Equation showing the conversion from mm to m^3:

$$\left(\frac{mm}{\frac{1000\,mm}{m}}\right) \cdot m^2 = \left(mm \cdot \frac{m}{1000\,mm}\right) \cdot m^2 = m^3$$

Inflow from glacial melt has been calculated by using the terrain and degree-day model. When there is no more snow covering the ground, the amount of glacial melt can be estimated by multiplying daily mean temperature for that day with a degree-day constant. In this thesis, the constant used was 6.4 mm/°C/d (Hock, 2003). Since glacial melt is not initiated until all of the overlying snow has melted, the daily mean temperature for the dates with water equivalent (s.w.e.) equal to zero was multiplied with the degree-day factor for these days, to get daily amount of glacial melt. This was done for each grid cell and later converted to volumetric units by multiplying with the percentage glacier area within each grid cell. Temperature values are regardless of area and remained the same.

Estimations of inflow to the glacier-dammed lake between each jøkulhlaup was made for the entire catchment area, 11 grid cells, and for a smaller area, the four grid cells closest to the glacier-dammed lake. The inflow volumes were converted to meters lake level rise by using equation 1, basing the lake area on the 2011 orthoimage.

DURATION OF EACH PERIOD BETWEEN EVENTS

Each value of the meteorological data provided by senorge.no is given at 08:00 every morning. This value is the mean value of the preceding 24 hours, thus the diurnal measurements ends at 07:59 the following morning. Each period has therefore been set to begin the following hour after the preceding jøkulhlaup ended, and it restarts at zero after each event. This was done because the events began and ended at such varying times of the day. Since the period between the events were so brief, many at which only lasting for 3 days, it was presumed that each hour influenced inflow between the events. However, this assumes that the duration of each event are precise, but this documentation is based on observations by local population and since it is a less populated area and the events were so many, there is a large probability that these duration times are not perfectly true.

Inflow prior to the first event on 4th June has been neglected due to the chances of errors related to inflow calculations for this period by the method used in this thesis. Inflow is calculated by summarizing all theoretical runoff to the lake non-stop, without restrictions concerning time, maximum lake volume, evaporation or restraint within the snow or ice, thus it would have been necessary to know when the filling of the lake began and what volume the lake had on this day. Both at which are unachievable given the limited information available

The first period therefore begins the day after the first event, on 5th June.period therefore begins after the preceding 4th June 2013 ended, thus at 02:01 on 4th June, and lasts until the preceding jøkulhlaup occurred, which for the first period means at 08:59 on 9th June. Since the first event ended at 02:00 on the 5th June, 6 hours of the mean value for the 4th is included in the period between the first and second event. These 6 hours are therefore added to the sum of the following 4 days of the period. Additionally, 1 hour of the 9th June mean values should be added to the sum, since the second event occurred at 09:00 on the 9th June, 1 hour after the mean values of the 9th had begun.

4. RESULTS

4.1. Glacier Area and Elevation Change Prior To the Jøkulhlaup

Comparisons of the three orthoimages from 5.8.2006, 19.9.2011 and 16.9.2016 show how the glacier and lake outlines had evolved prior to the events in recent years (figure 13). The decrease in glacier area has resulted in retreat of the glacier front and decrease in lake area. Lake area measurements from the three orthoimages clearly show what impact the decreasing lake area has had on the glacial lake, as they show a reduction from about 99 000 m2 to 49 000 m2 between 2011 and 2016. In only 5 years, the lake area has been reduced by 50 000 m2, to about half its size. This is further illustrated in figure 14 [Jackson, Ragulina, 2014#].

Table 5: Lake area change between 2011 and 2016s

Date	Year	Lake area
19.9.	2011	0,1 x 10^6 m^2
16.9.	2016	0,5 x 10^6 m^2

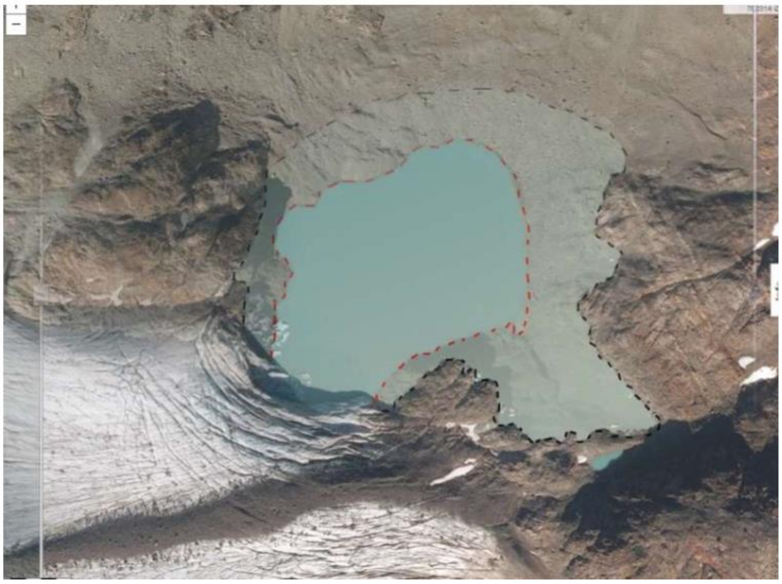


Figure 13: Figure 14 shows the glacier dammed lake in 2011 (black line) and 2016 (red line). Source: Miriam Jackson, NVE

Lake area measurements from the three ortho images clearly show what impact the decreasing lake area has had on the glacial lake, as they show a reduction from about 99 000 m^2 to 49 000 m^2 between 2011 and 2016. In only 5 years, the lake area has been reduced by about half its size. This is further illustrated in figure 14 [Jackson, Ragulina, 2014#].

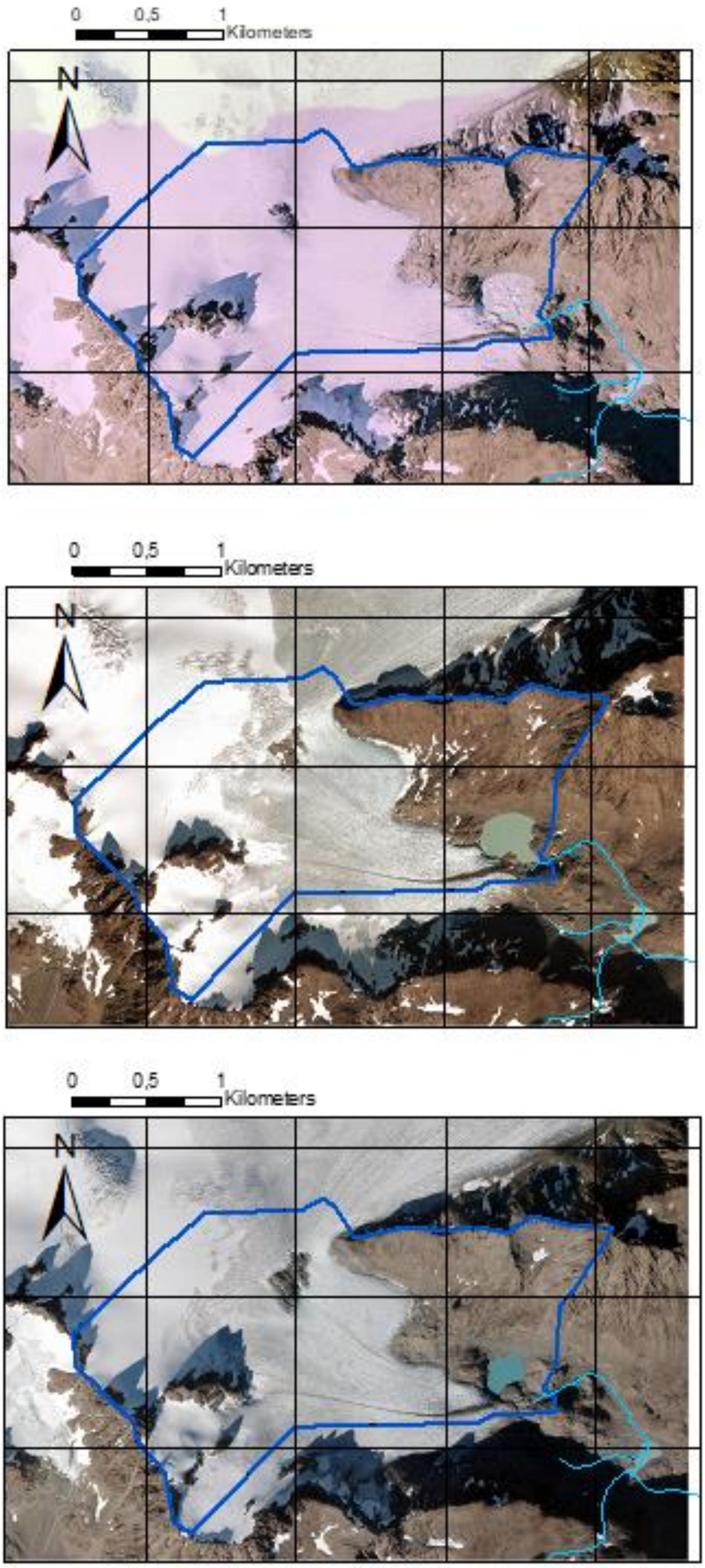


Figure 15: ***Catchment area, glacial lake and streams in 2006, 2011 and 2016, from top to bottom***

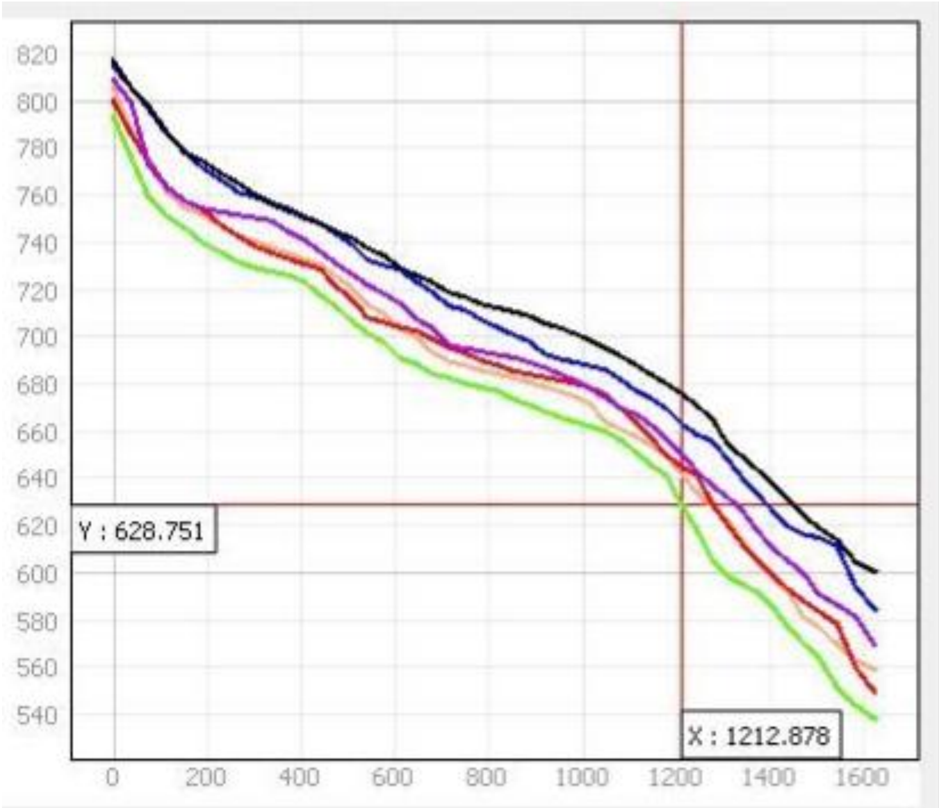
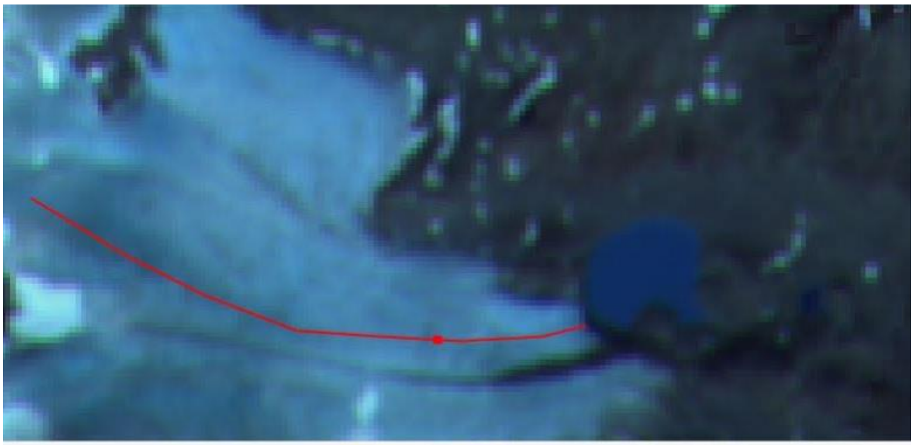


Figure 16: DEM showing elevation progression of the front from the mid 90s to 2017
Source: Chris Nuth, personal communication



Figur 17: Location of DEM son the glacoer.

DEMs showing the evolution of glacier thickness in recent years.

4.2. Lake level evolution

The estimated inflow to the glacier-dammed lake prior to each event, is presented in table 7. The potential rise in lake level based on these volumes between each jøkulhlaup is also shown. The difference between the two methods using 11 grid cells and 4 grid cells is further illustrated by the difference in lake level rise, in figure 15.

The estimated inflow to the glacier-dammed lake prior to the events, and the influence that these volumes had on lake level rise between each jøkulhlaup, are presented in table 7. There is a clear difference between the inflows based on 11 cells and 4 cells. The difference in lake level rise is further illustrated in figure 15. Inflow estimates based on 11 cells gave considerably more varying and extreme volumes for the different periods, than those estimated with 4, which gave more even inflow estimates throughout the season. The highest estimated inflow based on 11 cells, is half the size of that estimated based on 4. The inflow estimates based on 4 cells will therefore be used in further analyses, since it seems to give more realistic inflow.

Tabell 6: Estimated inflows to the glacier-dammed lake between each jøkulhlaup, based on the entire catchment area (grid cell 1-11) and on a smaller area surrounding the lake (grid cell 3, 7, 9 and 10). The respective influence on lake level rise, is also included.

Date of jøkulhlaup	INFLOW [m^3] (1-11)	INFLOW [m^3] (3, 7, 9, 10)	Lake Level rise [m^2] (1-11)	Lake Level rise [m^2] (3, 7, 9, 10)
9 June 2013	400 000	309 000	4	3
19 June 2013	521 000	364 000	5	4
23 June 2013	847 000	401 000	9	4
26 June 2013	597 000	275 000	6	3
30 June 2013	960 000	415 000	10	4
3 July 2013	878 000	403 000	9	4
12 August 2013	7 558 000	3 404 000	77	34
3 September 2013	3 608 000	1 695 000	37	17

Inflow to the glacier-dammed lake between the first 6 events, varied from 0.25 mill m^3 to 0.42 mill m^3 thus they volumes were all quite similar. This can explain the similar duration times, which varied from about 1 to 2 hours (table 3). Inflow volumes during the longer periods, such as prior to 12th August and 3rd September, were two or three times as high but this can be explained by the fact that these periods were about two or three times as long as those prior to the 6 first events, thus this is legit. Although a lake level rise of 34 m prior to the event on 12th August seems unlikely, one must take into account that the method only summarize inflow throughout the period regardless of the eastbound stream through where the water would drain if it rose to water levels higher than its inlet.

Daily discharge of 0.85 m^3 per day.

506 + 34 = 540 m.a.s.l.

506 + 19 = 525 m.a.s.l.

If glacier thickness was in fact 50 m thick in 2013, and the lake level after drainage at 506 m.a.s.l., this mean that the first, 8^{th} and 9^{th} event drained when the water levels were close to the floatation pressure of the ice (9/10).

The events that drained at the highest water levels, were below the floatation pressure of the ice. 34/50=0.7.

Inflow estimates based on runoff to the lake from the entire catchment area (A) are naturally larger than for those based on runoff from the four closest grid cells only (B). This is enhanced by the fact that a large section of the glaciated area lies within the area covered by method A, which means that a large portion of runoff from glacial melt would be included in the inflow volumes of method A. The difference is seen clearly when looking at inflow estimates over longer periods, such as the periods prior to 12th August and 3rd September. Estimated inflow based on method A is about twice the volume hence twice the lake level rise as that based on method B for the same periods. Estimated inflow based on A prior to the event on 12th August is 7.6 mill m3 whereas it is 3.4 mill m3 using B. These volume equals a rise in lake level of 77 m and 34 m, respectively. As the elevation of the outlet from the lake when it doesn't drain under the glacier is x m asl, the lake is not able to rise more than x m. Hence, increases in lake level above this value are not realistic, but are used for illustrative purposes only.

Although an inflow volume of 7.6 mill m3 between 3rd July and 12th August seems high, it is not so unrealistic considering the long duration of the period. The inflow estimate is the summary of theoretical runoff from the entire catchment area over40 days, which is equivalent to a daily inflow of 190 000 m3. Figure 16 shows that several days between May and September 2013 had daily precipitation and/or glacial melt values of half this volume, even though it shows volumes for method B. The fact that estimated inflow is the sum of these two parameters, a daily inflow of 190 000 m3 is acceptable and could have occurred on a warm day with a lot of precipitation and glacial melt. The average discharge values for the jøkulhlaups on 12th August and 3rd September are not given, as it wasn't possible for the volumes calculated to be retained in the lake, hence average values calculated for discharge would be meaningless.

Inflow estimates based on A gave considerably more varying and extreme volumes for the different periods, than those estimated with method B, which gave more even inflow estimates throughout the season. Method A show inflow volumes that range from 400 000 m3 to 7,6 mill m3 throughout the season, while method B gave inflow volumes ranging from 275

000 to 3,4 mill m hence the highest estimated inflow based on method B, is half the size of that estimated by method A. The inflow estimates based on method B will therefore be used in further analyses.

Inflow based on method A gave a potential lake level rise of 4 m in the 4 days between the 19th and 23rd June and between 26th and 30th June. Inflow based on method B for the same days, give a lake level rise of 0 and 1 m, respectively. Method B gives more realistic inflow estimates and furthermore method A shows theoretical inflow to the lake from the entire catchment area, without taking into account that not all of this will reach the lake.

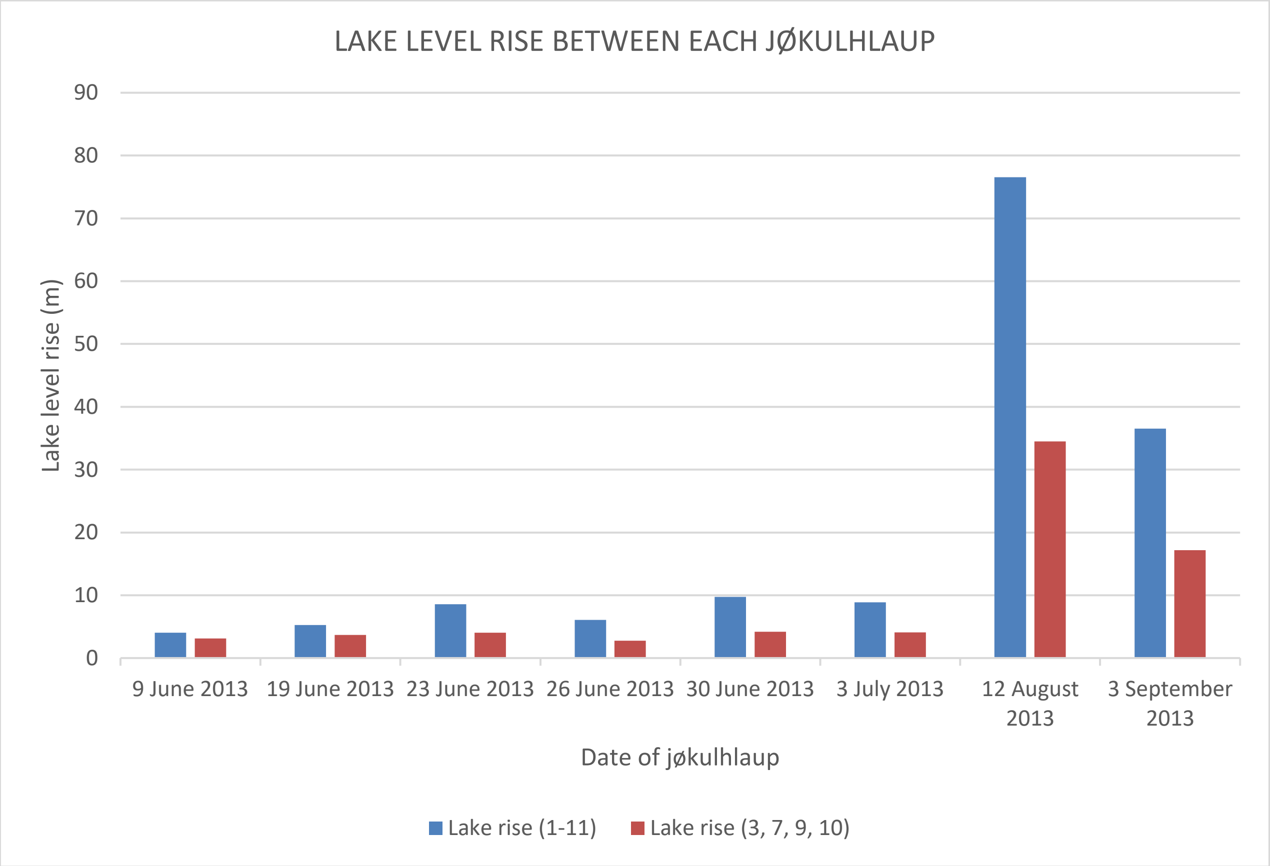


Figure 18: Lake level rise between each jøkulhlaup, based on calculated theoretical inflow to the glacier-dammed lake from 11 gridcells (dark blue) and 4 gridcells (light blue).

In order to obtain better insight in whether the jøkulhlaups drained when the glacier-dammed lake was full or not, the elevation of minimum and maximum water level was sought. The maximum water level was defined by the eastbound stream through where the glacier-dammed lake drained prior to the jøkulhlaups, in times when the glacier dam was thicker and allowed the lake volume to be larger. Whenever the lake level reached the elevation of the eastbound stream inlet, it drained eastwards. According to DEMs from 2013 the stream inlet is situated between 530 and 540 m.a.s.l.

Although the lake has never been observed completely emptied, observations of the lake basin topography after drainage has revealed a rock threshold underneath the glacier dam, presumably where the subglacial tunnel inlet is located. GNSS measurements on 5th and 27th June 2013 both show that lake level after drainage was at 506 m.a.s.l. which give reason to believe that this also was the height of the rock threshold. If that was the case, this would also define the level at which the glacier-dammed lake no longer would be dammed by the glacier dam but by a physical barrier, the rock threshold. Lake volumes lower than the threshold would then no longer be able to drain in a jøkulhlaup, since the lake would no longer be glacier-dammed, and there would no longer be any risk of danger. However, observations of the subglacial tunnel inlet from 2017 revealed that the thickness of the glacier dam had decreased enough for water to constantly drain underneath. Since GNSS measurements from 26th September and 24th October 2017, show that the lake level was at about 497 m.a.s.l. and the water was in fact draining freely in a subglacial tunnel situated on top of the rock threshold, it is consequently impossible that the lake level drainage was prevented by the rock level at 506 m.a.s.l. in 2013. Alternatively, the water flow was prevented due to closure or collapse of the subglacial tunnel, or that the subglacial tunnel inlet was shaped in such a way that the water did not drain at the very base/bottom of the glacier, in contact with the bedrock.

Having roughly estimated volumes for when the lake is full, it is clear that the inflow estimations contains some weaknesses. It does not take into account that whenever the lake is full, water drains eastwards until the glacier dam breaks and water drains underneath. Naturally, inflow estimations based on the longest periods will be much higher, which could explain why the inflow prior to the jøkulhlaup on 12th August and 3rd September are so much higher than for the rest of events. For this reason, average discharge calculations for these events are not possible to calculate since they would not be realistic or credible.

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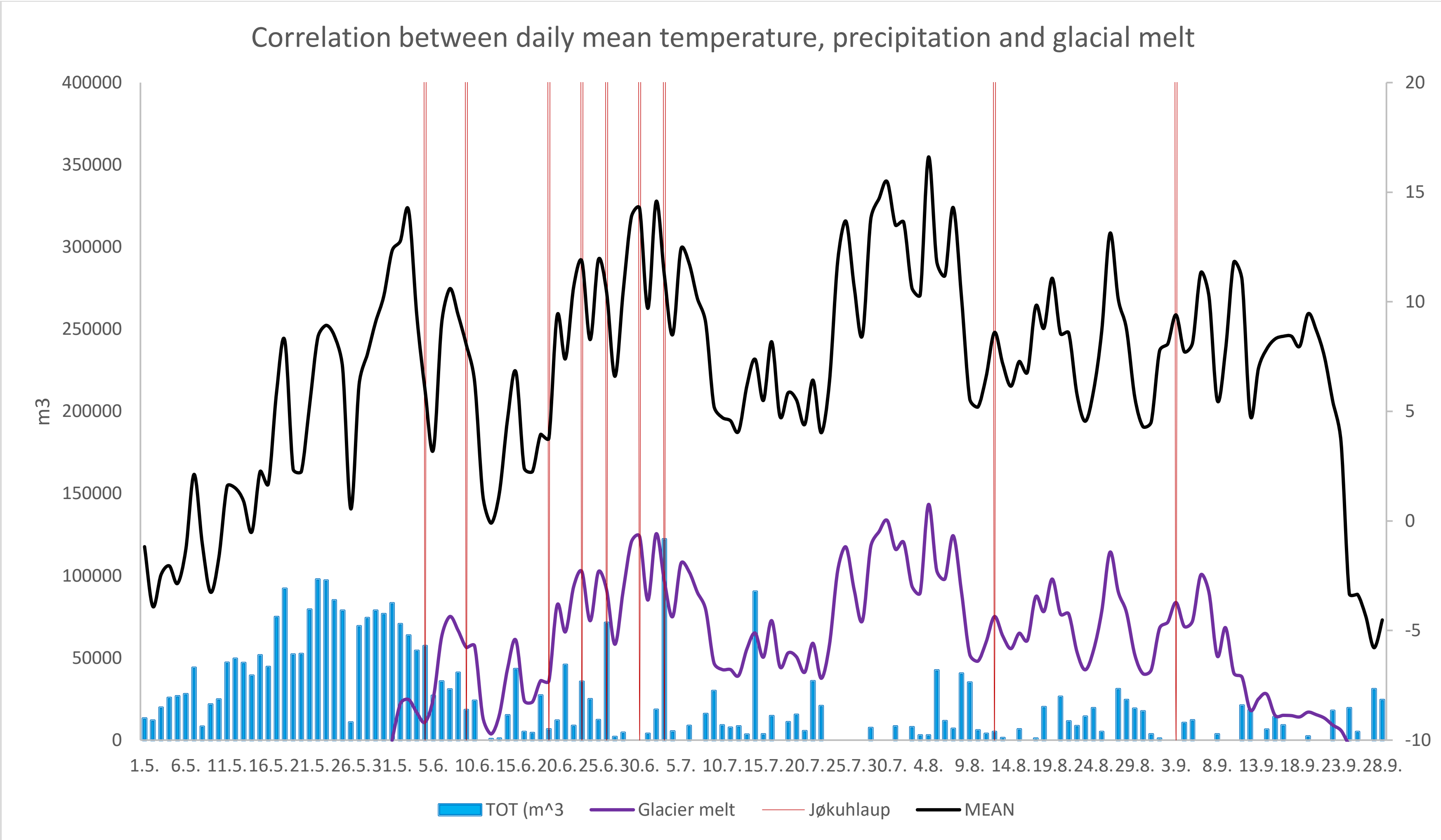

Figure 19: Illustration of daily mean temperatures(MEAN) and diurnal inflow to the glacier-dammed lake from precipitation (rain and snowmelt) (TOT (m^3) and glacier melt between 1.5.-29.9.2013, based on inflow calculations from the four grid cells situated closest to the glacier-dammed lake (3, 7, 9 and 10). The red lines represent times of jøkulhlaup occurrence.

The correlation between daily mean temperatures and diurnal inflow from precipitation and glacier melt between 1.5.2013 and 29.9.2013, based on 4 cells, is plotted in figure 19. The red vertical lines represent times of jøkulhlaup occurrences. The first event on 4th June is included. Volumes of precipitation and glacial melt are given in m^3 (lefthand side), while temperatures are given in degrees celsius (righthand side). Diurnal variations in precipitation (rain or snowmelt) are represented by blue charts, diurnal variations in glacial melt are represented by the purple line and daily mean temperatures with the black line.

The graph show a clear correlation between temperature and glacial melt, which is logic since glacial melt is dependent on temperature. The temperatures varied from -8 to +18 and the graph show that glacial melt has was not initiated before 1st June, indicating that there was still snow on the ground until this day.

Inflow to the glacier-dammed lake was mostly based on precipitation in the period prior to the first event 4th June. During this first period, precipitation correlates with temperature. In order for glacial melt to begin, all the overlying snow must melt, which did not occur until 1st June.

Which means that according to the SeNorge model there was still snow on the ground up until this day. This explains the large amounts of precipitation in May, as this most likely was due to snow melt, which would be shown under this parameter. If the graph had only shown the correlation between glacial melt and snowmelt in one grid cell, the charts for snow melt wold have been equal to zero the same day that the glacial melt would begin. However, since the method summarize inflow from four grid cells, these overlap because there could still be some snow left to melt in for instance grid cells number 10, while glacial melt had begun in grid cell number 3. In addition, the charts show rain or snowmelt values, thus it is still possible that rain and glacial melt occurs at the same time. In fact, presence of water enhance the rate of glacial melt, this is somehow visible from the results but not always. The overall correlation is that glacial melt increase when temperatures increase, while rain- and snowmelt increase with temperature to some extent but not as certain.

The correlation between rain- and snowmelt, glacier melt and temperature between 1.5.2013 and 29.9.2013 is shown in figure 19. The red lines represent day of jøkulhlaup. The graph does not show any distinct correlation between event and amount of days between each event but it does show a clear correlation between temperature rise and increased precipitation the days prior to a jøkulhlaup event. According to timeseries from senorge.no for other years, the temperatures during the summer season 2013 does not differ considerably from other years but the rapid temperature increase during the month of May might have had a large influence on the triggering of the first event. The temperature quickly increased from -2°C to +12°C during this period prior to the first jøkulhlaup. In the first half of May, from 1st to 15th, daily mean temperatures varied from -4°C to +1°C. The second half of May, had no temperatures below zero. At the end of May and at the start of June, daily mean temperatures were above 10°C. This would naturally have influenced snow- and glaciermelt in the area.

The eight outburst occurred 12th August, 40 days later. During these 40 days, theoretical inflow was 7 557 797,7 m^3 for the entire catchment area and 3 403 800,9 m^3 when only counting inflow from the 4 grid cells closest to the lake. This corresponds to a theoretical lake level rise of 76,6 m and 34,5 m, respectively. However, the lake level cannot rise more than 22 m, due to the eastbound stream at 527 m.a.s.l. through where the lake water flows whenever this maximum lake volume is exceeded. In addition, 4 x 10^2 m^3 in about 6 weeks (= 38 days) equals to ~1 m^3 inflow per day. The lake drained in 20 hours.

Calculated lake volumes released during the jøkulhlaup on 4th June based on GNSS measurements of the changes in water level from the day after the events, show a lake volume release of similar quantities like the inflow based on only the four closest grid cells for the same event. The same applies for the 26th June 2013, this strengthens the reliability of the calculated inflow based on the model only including the four closest grid cells, surrounding the glacier-dammed lake.

GNSS measurements made on 26th September and 24th October 2017, show that lake level is below the level at which the subglacial tunnel inlet is situated, on the rock threshold at 506 m.a.s.l. This means the lake was not dammed by the glacier per 2017. The lake level would have to increase to the more than 506 m.a.s.l. to pose any danger of another jøkulhlaup.

Since the surrounding area is very steep, the water inflow can be large early in the summer when there is a lot of snowmelt and at the time of the events, the glacier dam was just about thick enough to dam the water. The ice in the threshold area has now decreased to a thickness thin enough for water to flow freely underneath and the water currently seems to flow freely through the tunnel and is no longer glacier dammed by the glacier ice.

		INFLOW m^3	**m lake rise**	
DATE	**DURATION EVENT**	Grid 3, 7, 9, 10	Grid 3, 7, 9, 10	**Average discharge (based on 4 closest cells) [m^3/s]**
4 June 2013	11 h	220 080,1	2,2	48,0
9 June 2013	2 h	308 643,9	3,1	42,9
19 June 2013	~2 h	363 953,8	3,7	50,5
23 June 2013	~2 h	400 601,5	4,1	55,6
26 June 2013	~1,5 h	274 943,6	2,8	50,9
30 June 2013	~1 h 20 min	414 613,3	4,2	86,4
3 July 2013	~1 h 5 min	402 864,1	4,1	103,3
12 August 2013	20 h	3 403 800,9	34,5	-
3 September 2013	8 h	1 695 017,0	17,2	-

Average discharge show that duration of the events decrease while their intensity and damage potential increase throughout the year. The average discharge for 17 august and 3 september could not be calculated as they woul not give any credible numbers, due tot the errors of the inflow estimation methods.

Based on the duration the jøkulhlaups lasted for 11 hours, the average discharge of the first event was 48 m^3/s (equation 2).

$$\frac{1{,}9\ x\ 10^6\ m^3}{11h\ x\ 3600\ s/h} = 48\ m^3/s$$

These calculations are not very accurate but they illustrate the magnitude of the flood, which caused substantial damage down by Koppangen.

By multiplying the lake level decrease between 2011 and 2016 with the lake area decrease for the same period, a rough estimate of the volume change during recent years is given. It is rough because the formula does not take into account that the bottom topography of the lake narrows to a crescent shape but assumes the lake is shaped like a box with straight walls. GNSS measurements of the lake level prior to the first event in 2011 and on 26th September and 24th October 2017, show that the lake level decreased from 525 m.a.s.l. in 2011 to 497 m.a.s.l. in 2017, which is a water level reduction of 28 m. Since the lake area decreased by

50 000 m^3 during the same period, this gives a volume change of approximately 1.4 million m^3.

$$(525 - 497)m \cdot (99\,000 - 49\,000)m^2 = 28\,m \cdot 50\,000\,m^2 \approx 1{,}4 \cdot 10^6\,m^3$$

As the lake area decreased, so did the potential hazard of a jøkulhlaup since a potential flood consequently would release smaller volumes of water and the floods will be smaller.

4.3. Change in ice-barrier and ice-dam stability prior to the jøkulhlaup

In this study, GPR profiles of glacier ice were collected to detect depth to bedrock thus the current thickness of the glacier dam in total, 11 GPR-profiles could be further processed and these 11 have been used in this thesis.

The GPR profiles show that the glacier dam in 2017 had decreased to a thickness of about 40 m. By multiplying the lake level change between 2011 and 2016 with the lake area of 2016, the lake volume of 2017 is estimated to be

506-497=9m

9m * 0.5 x 10^6 m^2 = 4.5 x 10^6 m^3

The floatation pressure is determined by the thickness of the glacier dam and the lake volume. In theory, a glacier dam is stable until the water level has reached 9/10 of its height. Since the density of liquid water is lower than that of solid ice, the relationship between a stable or unstable glacier dams depends on glacier geometry and lake volume at the time of drainage. The dam remains stable until the water level equals the overburden pressure of the glacier dam. When the resisting forces of the glacier ice no longer can withstand the pressure from the water volume in the lake, the glacier dam breaks. Lake volume is dependent on glacier thickness to overcome the floatation pressure. This means that the water level required to equalize the overburden pressure of the glacier dam, is determined by glacier geometry at the time of drainage. Increased glacier thickness requires larger lake volumes, and vice versa. To some extent, the amount of englacial and subglacial fractures influence the stability of the glacier dam, along with slope gradient of the underlying topography because it influences the potential energy of the water. The gravitational forces pulling the water downwards increases with increasing slope gradient and elevation range. The lake volumes is only dependant on the surrounding topography, local weather and season/time of year.

If the lake volumes show the lake drained at a water level below that required to equalize the overburden pressure, the theoretical explanation is not sufficient enough and other relationships must be taken into account. For instance, the degree and development of fractures and tunnels within the ice.

Floatation pressure is reached when the water level is at 9/10 the height of the glacier ice. Based on the DEMs of 2013 and the glacier thickness in 2017, the glacier thickness in 2013 is assumed to be approximately 10 m higher than that of 2017, meaning 50 m. The elevation of the glacier thickness at 50 m, is 485 m.a.s.l. + 50 m = 535 m.a.s.l. If the triggering mechanism that initiated the floods was that the water volumes overcame the floatation pressure of the glacier ice, this mean the water level would have to be at 9/10 of this, which is 45 m. This is equal to an elevation of 530 m.a.sl. Which furthermore means that the water volumes would have to increase by (530-506=) 24 m between each event. Table based on method B, show that this was not the case, since the lake levels only rose about 3-10 m during the periods between the following event. The triggering mechanism could therefore not be due to overcoming of the floatation pressure.

DEPTH TO BEDROCK / GLACIER THICKNESS

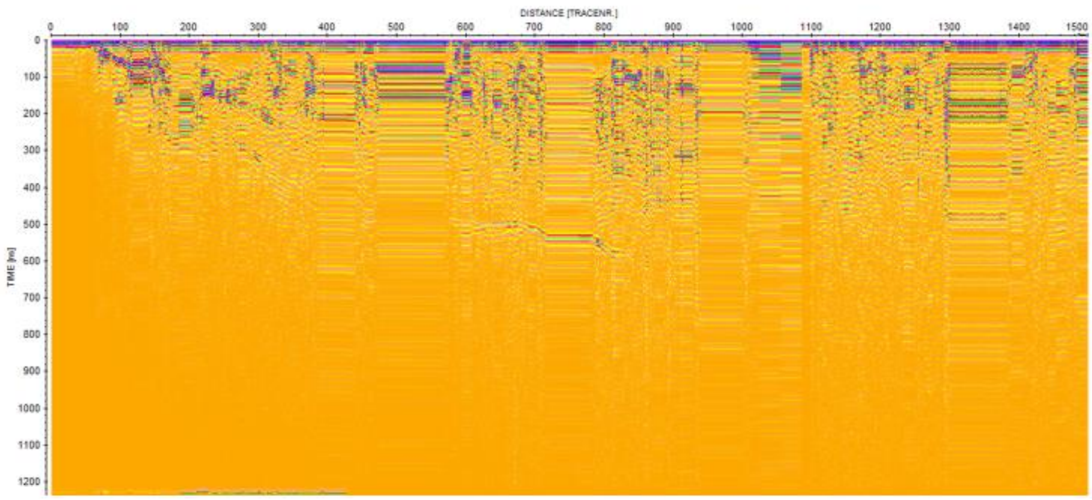

Figure 20: GPR-profile of the end of the glacier. The transition between glacier ice and bedrock is most distinct on the left part of the profile, showing how the glacier thickness increases as the GPR is moved further onto it.

GPR-profile of the end of the glacier. The transition between glacier ice and bedrock is most distinct on the left part of the profile, showing how the glacier thickness increases as the GPR is moved further onto it. The image has not been turned to upraised position.

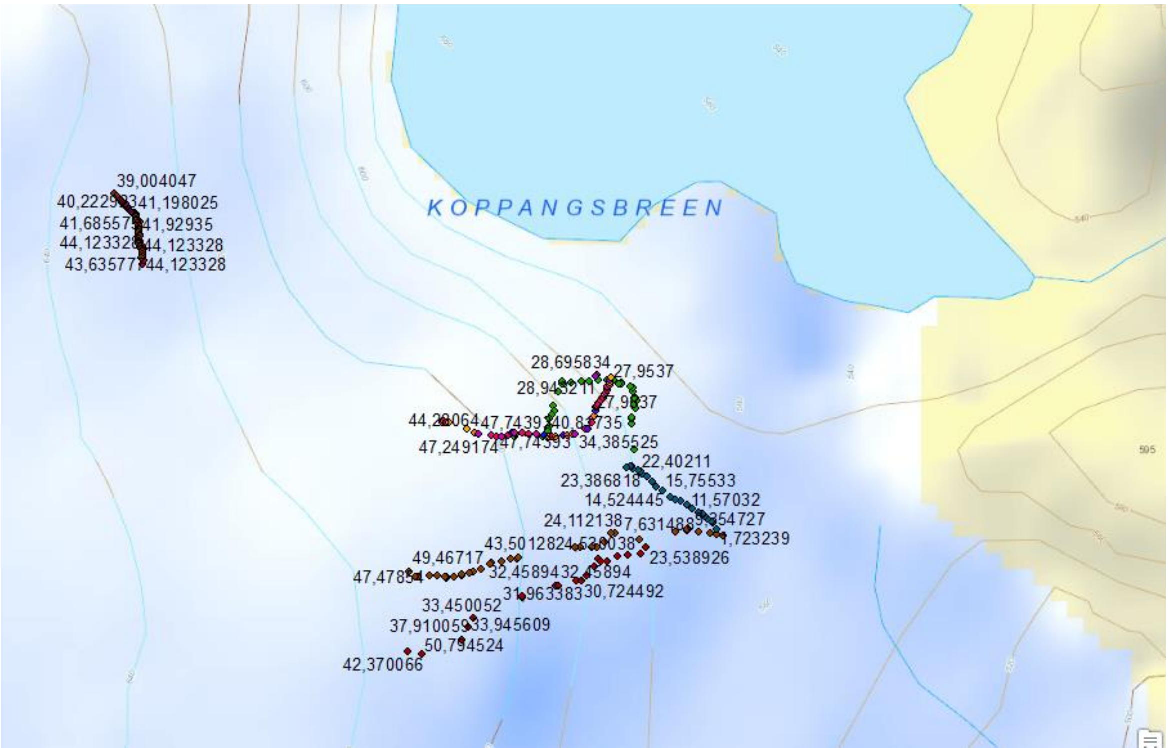

Figure 21: valid depths are acquired to actually show the glacier thickness values. The thickness was not that accurately measured, thus the decimals should be neglected.

The GPR profiles show that the thickness of the glacier closest to the location of the presumed subglacial tunnel inlet, in 2017, was about 40 m.

4. DISCUSSION

4.1. Scenarios of glacier change and future jøkulhlaups

Glacier geometry is one of the factors that sets the appropriate conditions for jøkulhlaup release thus glacier geometry was studied to examine changes in glacier and lake area prior to, and after, the jøkulhlaups of 2013. The aim was to get a better insight in why so many jøkulhlaups occurred in 2013, and to investigate if there is a present or future risk. Changes in glacier area would affect thickness and stability of the glacier dam, which determines how much water the dam can withstand before breakage, hence the maximum volume of the glacier-dammed lake. Changes in glacier area would also affect the lake volume by its influence on inflow. A retreating glacier could enlarge the glacier-dammed lake due to increased melt runoff, but it could also decrease the lake volume, due to increased temperatures and increased evaporation. Since lake volume is dependent on glacier dam and inflow, which furthermore are dependent on glacier area, changes in glacier geometry would have had great impact on the events of 2013.

Changes in glacier and lake area prior to, and after, the jøkulhlaups of 2013 was therefore studied. Prior to the first jøkulhlaup in 2010, the glacier area, and consequently its thickness, was larger. This allowed for the lake to grow large enough to spill over at a point at about 530-540 m.a.s.l., depending on the accuracy of the provided DEMs. Based on previous surveys of the glacier and the orthoimages between 2006 and 2011 it is evident that Koppangsbreen glacier area is retreating, thus its thickness decreasing. When the thickness of the glacier decrease, so does the stability of the glacier dam and the volumes of water that it can withstand. The glacier continued thinning (DEMs) from the early 2000s through to 2013. Simultaneously, as the glacier retreat, the lake area decreased, as observed on the three orthoimages of 2006, 2011 and 2016. The lake area decreased presumably due to lower inflow volumes. According to GNSS measurements based on shoreline prior to the event on 4th June 2013, the lake level was 526 m.a.s.l. at this point. This measurement is a very rough estimate, but it shows that the lake level is below that of the glacier-dammed lake spill. This means that the lake water was no longer draining out of the lake but remained dammed between the glacier front and the surrounding bedrock. As the glacier retreated, the lake volumes grew to reach the floatation pressure of the glacier dam in the jøkulhlaups prior to the 2013 events. In 2013, the glacier thickness must have reached a threshold limit that allowed a number of re-occurring jøkulhlaups to occur. The events seem to correlate with estimated daily input from rain, snow- and glacial melt that refilled the lake enough to trigger new flood initiations 3-10 days after each event. According to estimated inflow between each of these events, they seem to have triggered dam failure when the lake had refilled to about the same volumes. Inflow based on four grid cells show that this was at a lake level rise of about 4 m. Assuming that the GNSS measurements after drainage on 5th and 27th June 2013 are accurate, this means the lake level was at about 510 m.a.s.l. at the time of the 6 events recurring after periods of less than one month. Since there are no available data for glacier thickness in 2013, it is difficult to calculate what was the floatation pressure of the glacier dam at the time. However, a rough assumption based on the rate of glacier thickness decrease according to the DEMs, and the calculated thickness from the GPR profiles of 2017, the glacier thickness is assumed to have been about 50 m thick in 2013. Since field observations from the days after drainage showed that the glacier dam seem to lie above a rock threshold, it is assumed that this threshold contributed in dammin the water and preventing the lake from draining completely. If this is the case, the elevation of the rock threshold is likely to have about the same elevation level. Hence, if the rock threshold at the subglacial tunnel inlet at the bottom of the glacier is 506 m.a.s.l. and the glacier was 50 m thick in 2013, this means at 556 m.a.s.l. If the lake volume had drained completely at 506 m.a.s.l., one would have to count the overlying water in order to compare against glacier thickness. If the lake level prior to the first event was 20 m and the glacier thickness at the time was 50 m, this relationship is far below that required to trigger glacial uplift by overcoming the floatation pressure.

The aim was to get a better insight in why so many jøkulhlaups occurred in 2013, and to investigate if there is a present or future risk. Changes in glacier area would affect thickness and stability of the glacier dam, which determines how much water the dam can withstand before breakage, hence the maximum volume of the glacier-dammed lake. Changes in glacier area would also affect the lake volume by its influence on inflow. A retreating glacier could enlarge the glacier-dammed lake due to increased melt runoff, but it could also decrease the lake volume, due to increased temperatures and increased evaporation. Since lake volume is dependent on glacier dam and inflow, which furthermore are dependent on glacier area, changes in glacier geometry would have had great impact on the events of 2013.

By studying the correlation between lake volumes and glacier dam thickness at the time of the events, the floatation pressure determining when the glacier dam is stable and not, can be found. To be able to determine at what level the glacier dam is stable and not, the lake volumes at each event must be compared with the respective glacier dam thickness at the time of each drainage. Since Koppangsbreen has not been regularly monitored, the reduction in glacier dam thickness is estimated from previous measurements of glacier-dam thinning…..

Floatation pressure is reached when the water level is at 9/10 the height of the glacier ice. Based on the DEMs of 2013 and the glacier thickness in 2017, the glacier thickness in 2013 is assumed to be approximately 10 m higher than that of 2017, meaning 50 m. The elevation of the glacier thickness at 50 m, is 485 m.a.s.l. + 50 m = 535 m.a.s.l. If the triggering mechanism that initiated the floods was that the water volumes overcame the floatation pressure of the glacier ice, this mean the water level would have to be at 9/10 of this, which is 45 m. This is equal to an elevation of 530 m.a.sl. which furthermore means that the water volumes would have to increase by (530-506=)24 m between each event. Table 6 based on method B, show that this was not the case, since the lake levels only rose about 3-10 m during the periods between the following event. The triggering mechanism could therefore not be due to overcoming of the floatation pressure.

The GPR profiles show that the glacier dam in 2017 had decreased to a thickness of about 40 m. By multiplying the lake level change between 2011 and 2016 with the lake area of 2016, the lake volume of 2017 is estimated to be

506-497=9m

9m * 0.5 x 10^6 m^2 = 4,5 x 10^6 m^3

Most likely, the subglacial tunnel had begun its closure after each event but since this is a process that normally takes a couple of days to complete, the lake was able to refill to some extent due to the large inflow. By the time the subglacial tunnel managed to close 100 %, the lake volume had grown enough to release another jøkulhlaup, which could take advantage of the already existing weaknesses in the ice, such as cracks and voids. According to figure and table 6 the jøkulhlaup that has occurred at the approximate same intervals, has also had the approximate same volume inflow between the preceding evets. According to their duration and average discharge, these also correspond to events that has had about the same inflow volume.

4.2. Sensitivity analysis

Inflow estimates based on runoff from the entire catchment area (11 cells) are naturally larger than for those based on runoff from 4 only. This is enhanced by the fact that a large section of the glaciered area, thus a large portion of runoff from glacial melt, lies outside the area that is included in the 4 cells. The difference is clearly visible when looking at inflow estimates over longer periods, such as the periods prior to 12th August and 3rd September. Estimated inflow based on 11 cells is about twice the volume, thus twice the lake level rise, as those based on only 4, for the same periods.

The method used to estimate inflow is not considering that some runoff is not reach the lake but evaporate or is restrained within the snow or glacier ice along the way. It is assumed that all the runoff from glacial melt and rain and snowmelt directly flows into the glacier-dammed lake. The inflow estimates based on 11 grid cells may therefore not be appropriate since the values are unrealistically high. The estimates based on four cells, is therefore giving more credible and realistic estimates and these values are chosen to be used. However, they are still based on theoretical maximum inflow.

Between 3rd July and 12th August, inflow volumes of 7.6 mill m^3 or 3.4 m^3 was estimated for method A and B, respectively. These volumes would increase the lake level rise by 77 m and 37 m, respectively. Although such a large volume increase seem unlikely, it is not unrealistic considering the long duration of the period they were estimated for. The estimated inflow of 7.6 m^3 is the summary of theoretical runoff from the entire catchment area during 40 days, which is equivalent to a daily inflow of 190 000 m^3. Figure 16 show that several days between May and September 2013 had daily precipitation and/or glacial melt values of half this volume, even though it shows volumes based on method B. The fact that estimated inflow is the sum of these two parameters, a daily inflow of 190 000 m^3 could likely have occurred on a warm day with a lot of precipitation and glacial melt. However, it is not geophysical possible that the lake level rose 77 m, since it would have drained either eastwards through the east river spill of the glacier-dammed lake, or southwards above the glacier dam before reaching this elevation. In addition, the formula used to calculate lake level change, does not take into account that lake area greatly change with changing lake volumes. For this reason, it was not reasonable to calculate discharge values for the jøkulhlaups on 12th August and 3rd September, since it would not have given trustworthy results.

4.3. Comparison Blåmannsisen

Similar to Blåmannsisen, the lake downstream of the glacier-dammed lake is the first to receive the water from an outburst. This is Koppangsvatnet which is located at 411 m.a.s.l., 110 m below and approximately 233 m south-east of the glacier-dammed lake. This lake contributes in lowering the potential energy of the water and prevents the flood water from being more destructive. If additional actions to prevent damage is needed, one option is to strengthen the lake threshold so that the lake becomes big enough to withhold the water supply released in an outburst, like Sisovatnet below Blåmannsisen in Nordland.

The events at Blåmannsisen occurred at intervals of not 3 days but 3 years. Similar to Koppangsbreen, the events occurred when the lake was only half full or less, and at random intervals and random volumes. It is assumed that the same drainage mechanism operated for Blåmannsisen and Koppangsbreen, meaning that the subglacial tunnel did not get to close properly before the lake had been build-up to such a level that it could trigger new openings in the already existing cracks and voids in the glacier.

The events seem to correlate with estimated daily input from rain, snow- and glacial melt than refilled the lake enough to trigger new flood initiations 3-10 days after each event. Warm weather in May 2013, resulted in a lot of snowmelt around the glacier-dammed lake just before the first jøkulhlaup. The increased water inflow also increased the water pressure in the lake against the glacier ice dam which eventually succumbed and the water drained underneath the glacier. This could be the triggering mechanism to all the events because they have all been related to some precipitation the day prior to the outburst, apart from the outburst on the 23 June which followed two days after a day of heavy rain. Most likely the jøkulhlaups has ended due to the collapse of the ice dam in contribution with the rock threshold situated under the ice dam, which prevents complete drainage of the lake.

The DEM showing elevation progression of the front from the mid-90s to 2017, shows that in 2001 the lake indeed drained over the eastwards lake spill. Since then, glacier shrinkage and retreat eventually exposed a larger area for the lake to grow (possibly allowing for a larger liquid water volume). The lake then stopped flowing over the spill by 2008 due to unknown reasons, but after continued glacier thinning, the thickness probably reached some sort of threshold in 2012/2013 that allowed for the repetitive jøkulhlaups.Glaciers in northern Norway has been decreasing during the last 73 years and so is the case for the Koppangsbreen glacier. When the glacier area decrease, so does the thickness of the glacier dam, but also the inflow to the glacier-dammed lake from glacial melt. Reduction of glacier dam thickness and glacier-dammed lake therefore decrease simultaneously, thus in theory, the floatation pressure would remain unaffected because the ratio between glacier thickness and lake volume remain the same, as both sides of the equation decrease. Nevertheless, in 2013 the thickness of the Koppangsbreen glacier dam had decreased to such an extent that it was at the very limit of being able to withstand the lake and in the preceding years it has been too thin to dam the water anymore and the lake is currently constantly flowing through the subglacial tunnel. Between 2010 and 2014, the glacier dam has probably been in a transition zone between being stable and not, and in 2013 its thickness was at a very vulnerable level. If the lake were to grow in the future, the glacier dam could return to this unstable transition zone but if the glacier reduction continues the same way, the glacier will soon have retreated so much that it no longer goes as far down the mountain slope as to where the lake is situated today.

There is only a future risk if the glacier is to grow and the thickness of the glacier dam is to increase to a level at which it can dam the lake water again. Until then or as the current development indicates, the glacier dam will only continue to decrease until the glacier tongue has retreated so much it will no longer reach as far as to where the glacier-dammed lake is situated today. A scenario is of course that another glacier-dammed lake can form further up, but since the Koppangsbreen is so small, this is not very likely. Perhaps from Strupbreen or in the case of new growth.

What triggered the jøkulhlaups was that the lake volume increased to such a level to overcome the resisting forces of the glacier dam. The water forced open a subglacial tunnel, which was enlarged by the release of thermal energy produced by the flowing water, a process that was enhanced by the warm water due to the long-term high temperatures in the summer of 2013.

Currently, there is no risk of future jøkulhlaups since the glacier no longer is thick enough to withstand the lake volumes and the subglacial tunnel still is open so the lake water can flow constantly through. However, since glaciers are very sensitive to climate change and can rapidly respond by changing its geometry, a period of lower temperatures can result in glacier advance and the glacier dam can become thick enough to withstand the lake water again. If this occurs, the risk of jøkulhlaup will again be present, until the point where the glacier increases enough to remain stable no matter the lake volume dammed. Like it had prior to the first event in 2010.

5. CONCLUSION

Changes in glacier geometry are the reason why the jøkulhlaups occurred. The retreat of Koppangsbreen since 1961, revealed by map comparison indicate that the glacier has decreased in volume. The rate of thinning has increased since 1998 and presumably affected the stability of the glacier dam so that the water accumulated in the lake at the time of the events could overcome the floatation pressure of the ice at the glacier bed and drain underneath the glacier. The final trigger for all jøkulhlaups is that the lake volumes reaches the floatation pressure, which is 9/10 of the height of the glacier thickness, this could explain some of the large events at Koppangsbreen but for the rapidly repetitive ones alternate processes, such as melting from the warm water which may quickly help re-open up a subglacial channels to discharge the water could have been the trigger, for example from those events that occurred 3 days apart without a 25 meter rise in lake level. Previous jøkulhlaup events in Norway, such as those from Blåmannsisen, show that events are often triggered by precipitation (Jackson and Ragulina, 2014)

The graph show that there was quite large inflow from both glacial melt and precipitation, as well as high temperatures, during the periods of intervals at only 3-10 days. And that they all occurred at the very beginning of the season. There is no obvious difference between the first 7 events and the two final ones after periods of about one month.

The glacier dam failed although it had been stable before due to decreased thickness of the glacier dam. At the time of the recurring events in 2013, the glacier thickness must have reached a threshold limit that allowed for a number of re-occurring jøkulhlaups to occur. By the time the subglacial tunnel, that was formed by the first event, had closed, the lake volume had been refilled enough to trigger a new event. The rapidly recurring event could probably trigger drainage at lower lake volumes due to the already existing tunnels and fractures created by the preceding event. Additionally, warm water due to high temperatures over a long period might have contributed in this process. When the water drainage first was initiated, maintenance and expansion of the subglacial tunnel occurred due to the thermal energy of the flowing water.

FUTURE JØKULHLAUPS

Based on changes in the glacier geometry over the past years, especially since 2001, it is less likely that future jøkulhlaups will occur. Because the lake area already has decreased, the amount of water that could possibly be released, is smaller than before and the damage potential of future outbursts is reduced. On the other hand, when the glacier dam thickness also decreases, the volume required to overcome the resisting forces of the dam is reduced simultaneously, which was the case when the first jøkulhlaup occurred 4th June 2013. However, if the glacier area continues to decrease, the glacier tongue will retreat to such an extent that it does no longer dam the water and the glacier-dammed lake will not be formed.

In addition, the artificial flood embankment constructed by Koppangen village will most likely force the potential flood water to flow outside the riverbanks but stay in the river course and not change the river course and current drainage pathway and find new river courses.

However, glaciers have the ability to change very fast (Koppangsbreen glacier front has retreated by more than 300 m since 1998) and because there were so many recurring outbursts in 2013, Koppangsbreen glacier is still a source of concern. If the glacier grows the water can become glacier-dammed again. A scenario likely to occur if there are one or several precipitation rich winter with a lot of snow along with summer temperatures of today. During spring the inflow to the lake can increase in only 3-4 days which can cause rapid increase in water level due to inflow from precipitation, snowmelt from the surrounding mountains and snow and ice from the glacier which feeds/flows into the lake. This was observed several times in June 2013.

This could happen and its glacier tongue moves forward, closes the tunnel and other cavities and reinforces the ice dam. The lake can then grow and if the glacier decreases a subsequent jøkulhlaup might release even larger volumes of floodwater.

FURTHER WORK

Further research should involve additional GPR measurements of the bottom topography in the area, especially at the threshold, to obtain better insight in the drainage path underneath the glacier. Further research should also involve more accurately estimates of the glacier dam thickness at the time of, in order to compare to ratio of lake volume at each event.

ERRORS:

- Lake area used in all calculations for 2013 is based on satellite image from September 2011, it is uncertain whether this is the actual area of a full lake in 2013. Especially considering that the date of the jøkulhlaup vs date of image in 2011 is not known
- Interpolated meteorological data from senorge model is based on the closest meteorological stations
- Uncertainties connected to accuracy of catchment area since the program where it was generated (NEVINA) actually is meant for much larger areas

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