Jökulhlaups in Skaftá: A study of a jökulhlaup from the Western Skaftá cauldron in the Vatnajökull ice cap, Iceland

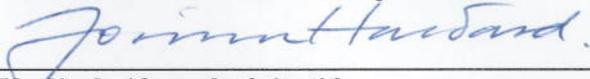Bergur Einarsson, Veðurstofu Íslands

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Jökulhlaups in Skaftá: A study of jökulhlaup from the Western Skaftá cauldron in the Vatnajökull ice cap, Iceland

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Abstract

Fast-rising jökulhlaups from the geothermal subglacial lakes below the Skaftá cauldrons in Vatnajökull emerge in the Skaftá river approximately every year with 45 jökulhlaups recorded since 1955.

The accumulated volume of flood water was used to estimate the average rate of water accumulation in the subglacial lakes during the last decade as $6\,\mathrm{Gl}$ ($6 \cdot 10^6\,\mathrm{m^3}$) per month for the lake below the western cauldron and $9\,\mathrm{Gl}$ per month for the eastern cauldron. Data on water accumulation and lake water composition in the western cauldron were used to estimate the power of the underlying geothermal area as $\sim 550\,\mathrm{MW}$.

For a jökulhlaup from the Western Skaftá cauldron in September 2006, the lowering of the ice cover overlying the subglacial lake, the discharge in Skaftá and the temperature of the flood water close to the glacier margin were measured. The discharge from the subglacial lake during the jökulhlaup was calculated using a hypsometric curve for the subglacial lake, estimated from the form of the surface cauldron after jökulhlaups. The maximum outflow from the lake during the jökulhlaup is estimated as $123\,\mathrm{m^3\,s^{-1}}$ while the maximum discharge of jökulhlaup water at the glacier terminus is estimated as $97\,\mathrm{m^3\,s^{-1}}$. This jökulhlaup was a fast-rising jökulhlaup as other jökulhlaups in Skaftá and cannot be described by the traditional Nye-theory of jökulhlaups. The total volume of flood water was estimated as $53\,\mathrm{Gl}$. The average propagation speed of the subglacial jökulhlaup flood front was found to be in the range 0.2–$0.4\,\mathrm{m\,s^{-1}}$. The volume of storage in the subglacial flood path, reached a maximum of $35\,\mathrm{Gl}$ which corresponds to two-thirds of the total flood volume. The volume of subglacial storage was an order of magnitude larger than could have been melted with the initial heat of the lake water and heat formed by friction in the flow along the flood path. The largest part of the space for subglacial storage was therefore formed by ice lifting and deformation induced by subglacial water pressures higher than ice overburden pressure.

The discharge data and the derived size of the subglacial flood path, as indicated by the volume of water stored subglacially, indicates a development towards more efficient subglacial flow over the course of the jökulhlaup. Thus, a discharge in the

range 80–90 $\mathrm{m^3\,s^{-1}}$ was flushed through the flood path near the end of the flood with only one-third of the flood path volume that transported a similar discharge a day or two after outflow started at the terminus. This may be interpreted as a development towards conduit flow and/or initial storage in subglacial reservoirs that do not contribute much to the transportation of flood water.

Measurements of the flood water temperature indicate that the jökulhlaup water is at or very close to the freezing point when it emerges at the glacier terminus. This indicates a very effective heat transport from the flood water to the ice walls of the subglacial flood path as all initial heat of the lake water and heat formed in the flood path by potential energy dissipation has been lost from the flood water.

A coupled sheet–conduit model for subglacial water flow was used to simulate the jökulhlaup. The model was forced with the estimated outflow from the subglacial lake. The simulations were not successful as a realistic subglacial pressure field could not be obtained for a reasonable fit of the jökulhlaup discharge at the glacier terminus. This indicates that the physical basis of the model is insufficient to provide a realistic description of fast-rising jökulhlaups.

Ágrip

Í Skaftá hafa mælst 45 jökulhlaup síðan 1955. Þessi hlaup vaxa hraðar en hefðbundin jökulhlaup frá Grímsvötnum og hafa því verið nefnd „hraðvaxandi jökulhlaup". Þau eru upprunnin úr lónum sem eru undir eystri og vestari Skaftárkatli í vestanverðum Vatnajökli. Kötlunum er viðhaldið af jarðhita.

Út frá gögnum um rúmmál hlaupvatns er meðalvatnssöfnun í vestari katlinum undanfarin áratug metin 6 Gl á mánuði og 9 Gl á mánuði í eystri katlinum. Afl jarðhitasvæðisins undir vestari katlinum má meta $\sim$550 MW út frá efnasamsetningu lónsvatnsins og meðalvatnssöfnun í katlinum.

Sig íshellunnar yfir lóninu, rennslið í Skaftá og hiti á hlaupvatninu nærri jökuljaðri var mælt í hlaupi úr vestari katlinum í september 2006. Útrennslið úr lóninu varð mest 123 $m^3\,s^{-1}$ en rennsli hlaupsins við jökuljaðar varð mest 97 $m^3\,s^{-1}$. Þetta hlaup er ekki hægt að skýra með viðteknum kenningum um jökulhlaup fremur en önnur Skaftárhlaup. Heildarrúmmál hlaupsins var 53 Gl. Geymsla vatns í hlaupfarveginum undir jöklinum var reiknuð út frá rennslisgögnunum en hún varð mest 35 Gl, sem samsvarar tveimur þriðju af heildarrúmmáli hlaupsins. Geymslan í flóðfarveginum er stærðarþrepi meiri en það rúmmál sem upphafsvarmi í vatninu í lóninu og varmi sem myndast vegna núnings í rennslinu geta brætt. Aflögun og lyfting íss, vegna vatnsþrýstings hærri en fargþrýstings, eiga því mestan þátt í að mynda flóðfarveginn. Framrásarhraði hlaupsins undir jöklinum var reiknaður út frá rennslisgögnunum og reyndist hann 0.2–0.4 $m\,s^{-1}$.

Rennsli hlaupsins og mat á stærð farvegarins undir jöklinum út frá rúmmáli vatns sem þar hefur safnast fyrir benda til þess að viðnám gegn vatnsrennsli við jökulbotn minnki eftir því sem líður á hlaupið. Undir lok hlaupsins runnu á bilinu 80–90 $m^3\,s^{-1}$ um farveg sem var einungis einn þriðji hluti af rúmmáli farvegar sem flutti svipað vatnsmagn á fyrsta eða öðrum degi eftir að hlaupið hófst við jökuljaðar. Þessi niðurstaða er vísbending um þróun farvegarins úr breiðu, óafmörkuðu rennsli yfir í skilvirkar rásir. Einnig kann að vera að í upphafi hlaupsins hafi töluvert vatnsmagn safnast upp í vatnsgeymum undir jöklinum sem ekki tóku mikinn þátt í því að flytja rennsli.

Mælingar á vatnshita í Skaftá benda til þess að hlaupvatnið sé við eða mjög nærri frostmarki þegar það kemur undan jöklinum. Þetta bendir til þess að undir jöklinum flæði varmi mjög ört úr flóðvatninu í umliggjandi ísveggi því að allur varmi í lónvatninu og sá varmi sem myndast vegna viðnáms í rennslinu hefur tapast þegar hlaupið kemur fram undan jökuljaðrinum.

Reynt var að herma rennsli hlaupsins með líkani sem lýsir víðáttumiklu, óafmörkuðu rennsli undir jöklinum sem tengist afmörkuðu rennsli í rásum. Ekki tókst að fá við-unandi niðurstöðu þar sem reiknaður þrýstingur í hlaupfarveginum var ekki trúverðugur. Þetta bendir til þess að fræðilegur grundvöllur líkansins sé ófullnægjandi til þess að skýra hraðvaxandi jökulhlaup.

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Chapter 1

Introduction

Jökulhlaups from subglacial lakes are an important part of the hydrology of Iceland. They are most often caused by an interaction between glaciers and geothermal/volcanic activity and jökulhlaups are the most frequent hazard related to volcanic and geothermal activity in Iceland (Guðmundsson *et al.*, 2008). Jökulhlaup may endanger people and livestock and they can damage infrastructure, power supply systems, communication lines, hydropower plants, vegetation and even alter landscape to great extent. A good understanding and knowledge of jökulhlaups is therefore needed.

Jökulhlaups from subglacial lakes under the so-called Skaftá cauldrons in western Vatnajökull have been well known since 1955 (Zóphóníasson and Pálsson, 1996). These jökulhlaups emerge in the river Skaftá and are characterized by a rapid increase in discharge, rising to maximum discharge in one to two days. Jökulhlaups in Skaftá are therefore classified as "fast-rising jökulhlaups" based on a classification of jökulhlaups into two main categories from the form of the discharge curve: fast-rising jökulhlaups and slowly-rising jökulhlaups. Slowly-rising jökulhlaups typically rise exponentially to maximum discharge in one to two weeks and recede in one to three days while fast-rising jökulhlaups generally rise to a maximum in one to three days and they recede in one to two weeks (Björnsson, 2002). Slowly-rising jökulhlaups are well understood by a theory developed by Nye (1976) but fast-rising jökulhlaups are not as well understood.

Fast-rising jökulhlaups are known from many different source lakes in Iceland and have been observed in many rivers. They range over many orders of magnitude in maximum discharge. Small jökulhlaups in Skaftá have maximum discharge on the order of $10^2 \, \mathrm{m^3 \, s^{-1}}$ while large jökulhlaups caused by eruptions in the subglacial volcano Katla have maximum discharge on the order of $10^5 \, \mathrm{m^3 \, s^{-1}}$ (Zóphóníasson and Pálsson, 1996; Tómasson, 1996). As fast-rising jökulhlaups have a rapid discharge

increase they may be extremely dangerous as time for response is short.

Various measurements on a jökulhlaup from the Western Skaftá cauldron in September 2006 are presented and interpreted in this thesis. The thesis forms a part of the Skaftá cauldrons research project which is an extensive campaign involving measurements within the Skaftá cauldrons, in the subglacial lakes and of the jökulhlaups originating in them. The project was initiated in June 2006 with a drilling through the 300 m thick ice shelf[1] covering the western cauldron. The temperature in the lake was measured and a water sample was obtained, a pressure and temperature sensor was deployed at the bottom of the lake and a GPS instrument was placed at the center of the ice shelf. Furthermore, the water temperature was measured in the Skaftá river 3 km downstream from the glacier terminus.

The thesis starts with an overview of jökulhlaups in Iceland and at other locations of the world. Chapter 3 describes the Skaftá cauldrons and jökulhlaups in Skaftá and brings together the main results of previous research of these jökulhlaups. The theory of jökulhlaups and of water flow in glaciers in general is described in Chapter 4 with an emphasis on the lack of understanding of fast-rising jökulhlaups. Chapter 5 is about the measurements that were carried out as a part of the preparation of this thesis and Chapter 6 describes the results and contains an interpretation of these results on the background of the theories presented in Chapter 4. Chapter 7 describes an application of a mathematical model of fast-rising jökulhlaups to the jökulhlaup from the western cauldron in 2006. The thesis ends with a chapter about the conclusions that may be drawn from this work about the nature of fast-rising jökulhlaups.

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[1]The phrase "ice shelf" is used in this thesis to describe the overlying ice cover of the cauldrons as the weight of the ice covers is mostly supported by floatation although they are fundamentally different from the large ice shelves of Antarctica, Greenland and Canada.

Chapter 2

Overview of jökulhlaups

A jökulhlaup is a sudden emptying of a water body that is formed by a glacier. The water bodies can be glacier surface lakes, glacier marginal lakes and subglacial lakes. Normally, these water bodies drain out through subglacial channels. Breaking of ice dams is also known to have caused jökulhlaups but is much less common. Glacier marginal lakes can also have a steady drainage on the surface. Jökulhlaups span an enormous range both in maximum discharge and flood water volume.

At the end of the last glaciation, great jökulhlaups from marginal lakes of the glacial period ice sheets are known to have occurred. Hundreds or thousands of cubic kilometers of water drained in outbursts with maximum discharge of millions of cubic meters per second (Clarke *et al.*, 2004; Baker and Bunker, 1985; Baker *et al.*, 1993). At the other end of the scale are small surface lakes in the ablation areas of glaciers that release few tens or hundreds of cubic meters every few hours or every few days (Björnsson, 1976).

Hundreds of jökulhlaups are known in Iceland. On the average 5–10 jökulhlaups occur annually in the country (Sigurðsson and Einarsson, 2005). The best known are repeated outbursts from Grímsvötn, the Skaftá cauldrons and Grænalón in Vatnajökull. Jökulhlaups in connection with volcanic eruptions are also well known. The Katla jökulhlaups from Mýrdalsjökull and the 1996 jökulhlaup from Grímsvötn are the best known jökulhlaups of this type. In addition to these main outbursts sources, many others are known.

2.1 Jökulhlaups from marginal lakes

Most of the Icelandic ice caps have marginal lakes. Many lakes are by the southern and eastern margin of Vatnajökull but marginal lakes are also known in other locations

(Björnsson, 1976). Generally, jökulhlaups from marginal lakes have decreased due to thinning of glaciers during the twentieth century because lower ice dams cause the lakes to burst at a lower water level. This leads to smaller but more frequent jökulhlaups.

Other types of floods can also originate in marginal glacier lakes. For example, a flood in the river Farið in 1999 was caused by a surging glacier that advanced into Hagavatn lake by Langjökull ice cap and forced water out of the lake (Sigurðsson and Einarsson, 2005). The landslide "Steinsholtshlaup" in 1967 can also be mentioned. Then a huge landslide fell down on the Steinsholtsjökull outlet glacier from Eyjafjallajökull ice cap. It travelled down the glacier and caused a flood wave in a marginal lake by the glacier terminus. The flood emerged in Krossá river and farther downstream in Markarfljót river (Kjartansson, 1967).

The largest jökulhlaups from marginal lakes in Iceland in historical times are from Grænalón at the margin of Skeiðarárjökull and Vatnsdalslón at Heinabergsjökull, both in the Vatnajökull ice cap. The outbursts from Grænalón have diminished during the last decades because of thinning of glacier and changes in the outflow path (Björnsson, 1976). The situation is similar for Vatnsdalslón which now is emptied in one to three small outbursts each summer (Sigurðsson and Einarsson, 2005).

2.2   Jökulhlaups due to volcanic eruptions

In historical times, the most dangerous jökulhlaups in Iceland have been caused by subglacial volcanic activity. Katla and Öræfajökull are the most important in this context. Katla is a volcanic caldera under the Mýrdalsjökull ice cap. Eruptions in the caldera melt great amounts of ice and cause huge jökulhlaups with maximum discharge on the order of 1–$3{\cdot}10^5\,\mathrm{m}^3\,\mathrm{s}^{-1}$ and volumes of 1–$8\,\mathrm{km}^3$. Jökulhlaups from Katla rise very rapidly, reaching maximum discharge in a few hours but the duration is only 3–5 days (Björnsson, 2002). The situation is similar at Öræfajökull which also has an ice-filled volcanic caldera. Huge jökulhlaups were caused by eruptions in Öræfajökull in 1362 and 1727 (Björnsson, 1975).

Jökulhlaups caused by volcanic activity are also known from Eyjafjallajökull ice cap in 1821–1823 and seven jökulhlaups in the river Jökulsá á Fjöllum in the period 1655 to 1729 most likely originated from Vatnajökull near Kverkfjöll. In addition, jökulhlaups have probably been caused by volcanic activity in Snæfellsjökull, Tindfjallajökull, Torfajökull, Kerlingarfjöll and Tungnafellsjökull (Björnsson, 1975).

2.3 Grímsvötn and the Skaftá cauldrons

The best known subglacial lakes in Iceland are Grímsvötn and the Skaftá cauldrons. The Skaftá cauldrons are described in the next chapter of this thesis. Grímsvötn is a subglacial lake in a volcanic caldera in the interior of the Vatnajökull ice cap (Björnsson, 1988). Jökulhlaups from Grímsvötn have been known since at least the fourteenth century (Þórarinsson, 1939, 1974). In the beginning of the twentieth century there were about ten years between outbursts but the floods diminished with time and became more frequent. After a catastrophic, rapidly rising flood caused by the Gjálp eruption in 1996 (Björnsson 1997; Snorrason *et al.*, 1997) and with increased geothermal activity in Eastern Svíahnjúkur, Grímsvötn has drained continuously or in small and irregular jökulhlaups (Sigurðsson and Einarsson, 2005). In 1938, a similar large and rapidly rising jökulhlaup, also caused by a volcanic eruption, emerged from Grímsvötn (Þórarinsson, 1974).

Lake Grímsvötn and the associated jökulhlaups in the river Skeiðará are the cradle of jökulhlaup research. The anatomy of slowly-rising jökulhlaups from Grímsvötn was described by Björnsson (1975) and a physical model explaining the approximately exponential increase in the flood discharge was derived by Nye (1976). Slowly-rising jökulhlaups from Grímsvötn have since been analyzed and modelled by *e.g.* Spring and Hutter (1981), Björnsson (1992) and Clarke (2003). Jökulhlaups from Grímsvötn have also been studied in other contexts such as in connection with glaciohydraulic supercooling (Roberts *et al.*, 2002) and an interesting interplay between glacier surging and the hydraulics of a jökulhlaup was observed for a jökulhlaup from Grímsvötn in 1991 (Björnsson, 1998).

Grímsvötn focused increased attention to fast-rising jökulhlaups by the catastrophic flood in November 1996. Water melted by the subglacial volcanic eruption of Gjálp to the north of Grímsvötn, accumulated in the subglacial lake (Guðmundsson *et al.*, 1997) until the water level reached the flotation level of the ice dam and a jökulhlaup started (Björnsson, 1997). The jökulhlaup challenged conventional theories for jökulhlaups from Grímsvötn. The lake level rose unusually high before the flood was initiated but Grímsvötn had been previously known to release jökulhlaups at water levels lower than needed for floatation of the ice dam (Björnsson, 1988). In the beginning the flood, the water burst through the several hundred meters thick ice near the glacier terminus, indicating high subglacial water pressure, and the discharge increase during the rising phase of the jökulhlaup was extremely rapid compared with observed jökulhlaups during the previous several decades. The 1996 jökulhlaup has been described by Björnsson (1997; 2002) and Snorrason *et al.* (1997) and a conceptual model describing the

propagation of the flood front as subglacial pressure wave was presented by Jóhannesson (2002). Fowler (1999) analyzed the breaking of the subglacial seal at the beginning of the 1996 jökulhlaup and concluded that the high water level reached before the onset of the 1996 flood could be explained with the traditional theories of Röthlisberger (1972) and Nye (1976) but he did not address the rapid discharge increase following the breaking of the seal. Flowers *et al.* (2004) concluded that a model fundamentally different from the Nye-model involving coupled sheet–conduit subglacial water flow was needed to explain the 1996 jökulhlaup and presented a mathematical model based on this concept, which was able to reproduce the observed rapid rise in the discharge.

2.4   Jökulhlaups in Iceland during the Holocene

Great jökulhlaups occurred during the Holocene and at the end of the last glaciation in Iceland. A large glacier-dammed lake was formed in the central highland at Kjölur at the end of last glaciation. Initially, the lake drained in jökulhlaups down the river course of the river Blanda but then as the glacier thinned it started to drain in jökulhlaups to the south. First over Bláfellsháls, west of Bláfell but later east of Bláfell down the river course of the river Hvítá. The outbursts east of Bláfell were the largest with maximum discharge over $3 \cdot 10^5 \, \text{m}^3 \, \text{s}^{-1}$ and an estimated volume of $28 \, \text{km}^3$. Evidence for these jökulhlaups can be observed along the river courses of Blanda and Hvítá and at Kórinn, which is a gorge in the southern flanks of Bláfellsháls, eroded in jökulhlaups. Rows of paleo-coastlines of the lake can be seen on the mountainsides by Kjölur (Tómasson, 1993).

Evidences for large jökulhlaups north of Vatnajökull along the river course of the river Jökulsá á Fjöllum are also known. There were probably around 16 events in total with a one large outburst late in the Holocene (about 2900–2000 before present) (Kirkbride *et al.*, 2006). Possible locations of sources lakes for these floods are Bárðarbunga and Kverkfjöll but Grímsvötn has also been mentioned in this connection. At least two floods flowed through the Kverkfjallarani north of Kverkfjöll making Bárðarbunga and Grímsvötn unlikely source for at least these two floods (Carrivick *et al.*, 2004). It is also possible that a glacier-dammed lake was formed behind glaciers flowing down from Bárðarbunga and Kverkfjöll that reached together (Tómasson, 1973; Einarsson, 1976). The discharge in the largest events has been estimated from flood marks and hydraulic modelling to be $\sim 9 \cdot 10^5 \, \text{m}^3 \, \text{s}^{-1}$ (Alho, 2005). The volume has not been evaluated as the source lake is not known.

2.5 Jökulhlaups in America and Asia at the end of the last glaciation

The largest known jökulhlaups on Earth come from glacier-dammed marginal lakes formed at the end of last Ice Age. Of these, the Missoula jökulhlaups have gained the greatest attention (Baker and Bunker, 1985). The Missoula jökulhlaups originated in huge glacier-dammed lakes in the watershed of the Columbia River in western USA. The idea of Bretz (1923) that jökulhlaups had formed the various sedimentary and erosional features that are noticeable along the course of the Columbia River was highly debated when it was first proposed . "In the beginning it met great resistance as it was thought to break one of the principles of Geology, the principle of uniformitarianism, – that the present is the key to the past" (Tómasson, 1973). Today it is widely accepted that these jökulhlaups took place but the number of events is still not clear although it is known that there was more than one event. The discharge in the largest outburst is believed to have been at least $17 \cdot 10^6 \, \mathrm{m^3 \, s^{-1}}$ and it is possible that this was by far the largest of a series of jökulhlaups occurring along the Columbia river course towards the end of the last glaciation (O'Connor and Baker, 1992). The volume of water in Lake Missoula available for outburst has been calculated as $2184 \, \mathrm{km^3}$ (Clarke *et al.*, 1984). The Missoula jökulhlaups are believed to have taken place 17000 to 12000 years ago (Baker and Bunker, 1985).

A few thousand years later, a huge lake formed at the southern margin of the Laurentide Ice Sheet. It is called Lake Agassiz and formed $11700 \, \mathrm{C^{14}}$ years ago. The lake is believed to have emptied into Hudson Bay in a huge jökulhlaup causing a sudden input of freshwater into the North Atlantic that may have disturbed the ocean current system and caused a cold spell 8200 years ago. The maximum volume of outburst water has been estimated to have been $40,000$–$151,000 \, \mathrm{km^3}$. The discharge of the jökulhlaup has been modelled using the theory of Nye (1976) and Spring and Hutter (1981). The outburst was simulated to last approximately half a year and reach a maximum discharge of $5 \cdot 10^6 \, \mathrm{m^3 \, s^{-1}}$ (Clarke *et al.*, 2004). Geological evidence for the outburst path has been hard to find as it is on the bottom of Hudson Bay. The discharge estimates are therefore only built on modelling and might be an underestimate. Evidences of large jökulhlaups at the end of last glacial maximum are known from other places on Earth and formation of glacier-dammed lakes at the margin of the receding ice sheets was probably common. One of these locations is the Altay Mountain in Siberia. The height of water level marks there indicate a maximum discharge of up to $18 \cdot 10^6 \, \mathrm{m^3 \, s^{-1}}$ (Baker *et al.*, 1993).

2.6 Jökulhlaup sediment concentration and Lahars

Jökulhlaups have high sediment concentrations where significant erodible material is available in their course. This greatly increases their erosional power and affects the density of the flood water. Icebergs that break of the glacier or frazil ice formed by supercooling are sometimes also carried by the flow. Lahars are extreme examples of jökulhlaups with a high concentration of suspended sediments. They are landslides formed by a pyroclastic flow that melts snow or glacier ice or alternatively a jökulhlaup that flows down the flanks of a volcano where a lot of ash is available for erosion (Thouret *et al.*, 2007).

Chapter 3

The Skaftá cauldrons and jökulhlaups in Skaftá

3.1 The subglacial lakes

There are two powerful subglacial geothermal areas under the western part of the Vatnajökull ice cap forming surface cauldrons by melting the glacier from the base. The centers of the cauldrons are approximately at N 64°29.25' W 17°30.50' and N 64°29.75' W 17°37.00' and they are called the Eastern and Western Skaftá cauldrons, respectively (Figure 3.1). The eastern cauldron is a little less than 3 km in diameter while the western one is about 2 km in diameter. The cauldrons are shown on Figure 3.2. The lows in the glacier surface lead to local minima in the fluid potential at the base of the glacier and therefore lakes are formed under both cauldrons, sealed by the ice overburden pressure at the rim (Björnsson, 2002).

The temperature in the subglacial lakes has been measured, one profile in June 2006 through the lake below the western cauldron and two profiles in June 2007 through the lake below the eastern cauldron. The western subglacial lake was stably stratified at the location of the profile with two main water masses, a 90 m thick upper water mass with a temperature of 4.7±0.1°C underlain by a $\sim$10 m thick cooler tongue with a temperature of 3.5°C. Between the water masses there was a $\sim$10 m thick transition layer and at the bottom of the lake there was a thin layer of warmer water with temperature up to $\sim$5°C (Jóhannesson *et al.*, 2007). The vertical temperature distribution in the eastern subglacial lake was much more uniform at 4.0±0.1°C for both profiles except in a very thin layer by the bottom in one of the profiles where the temperature rose up to $\sim$7°C (data from the Skaftá cauldrons research project).

The composition of inflow into the cauldrons has been estimated for the west-
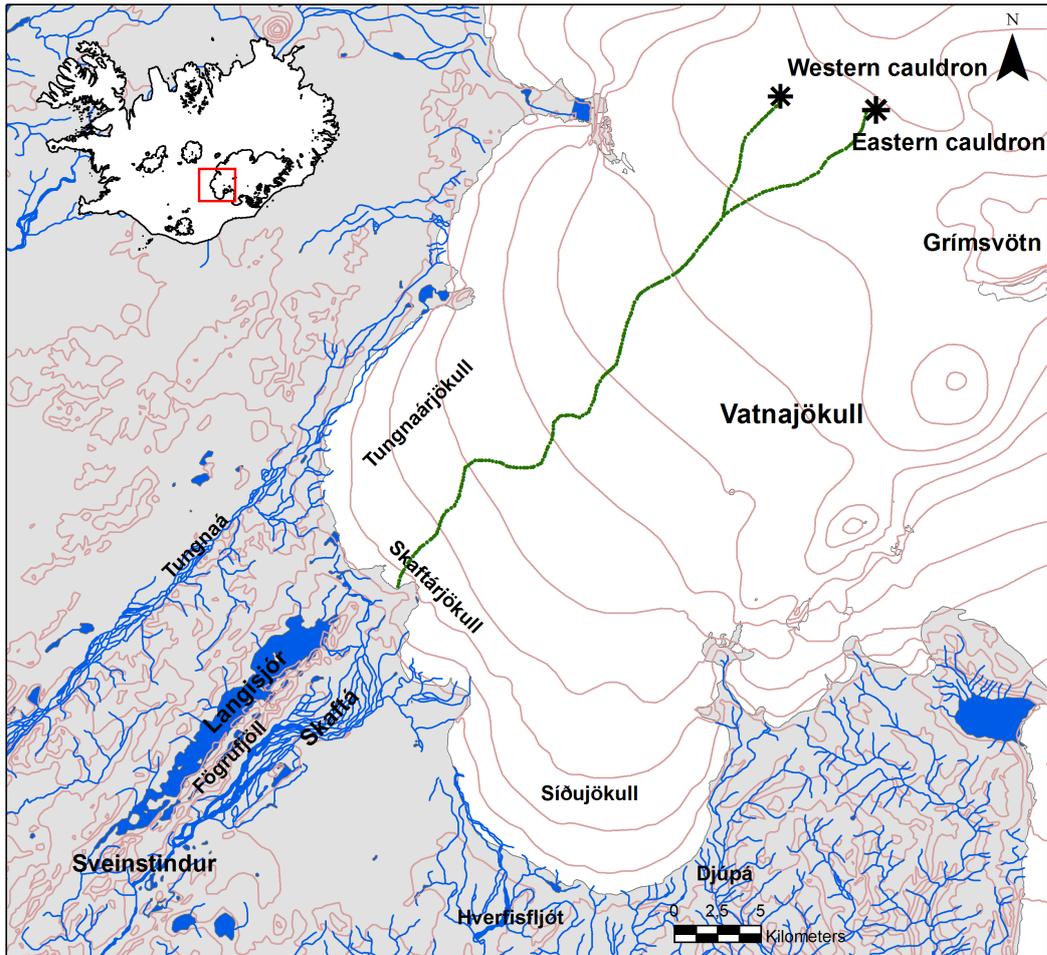
Figure 3.1: Overview map of western Vatnajökull showing the Skaftá cauldrons. The subglacial flood paths and river Skaftá are shown. Data on the location of the flood path are provided by the Institute of Earth Sciences at the University of Iceland.

ern cauldron from the chemical composition of the water and the lake temperature by Jóhannesson *et al.* (2007). The results indicate that geothermal melting of ice is the largest (71%) component of inflow into the lake, followed by geothermal fluid (19%) and surface meltwater (10%). The power of the geothermal area below the cauldron can be estimated from these ratios and the average water accumulation in the subglacial lake. The average water accumulation during the last decade is $6\,\mathrm{Gl}$ per month, or $2.3\,\mathrm{m}^3\,\mathrm{s}^{-1}$, giving $1.6\,\mathrm{m}^3\,\mathrm{s}^{-1}$, or $1600\,\mathrm{kg}\,\mathrm{s}^{-1}$, melting of ice which requires $550{\cdot}10^3\,\mathrm{kJ}\,\mathrm{s}^{-1}$. The power of the geothermal area can therefore be estimated as $\sim$550 MW. This estimation is crude and many factors are poorly known or not accounted for, the main uncertainty being the source area for groundwater inflow into the geothermal circulation. If the feeding of the geothermal circulation is from within the watershed of the cauldron, the power to melt the corresponding amount of ice must come from the geothermal area. If on the other hand the feeding is from other areas

Figure 3.2: Top: The Skaftá cauldrons on 25 February 1986. View towards east, the western cauldron in the lower left corner, the eastern cauldron near the center and Grímsvötn is distinctly visible in the right side of the photograph. A depression over the subglacial flood path from the eastern cauldron can be seen stretching southwest from the cauldron. The next jökulhlaup from the western cauldron was the following summer so it can be assumed close to full on the photograph. Bottom: The western cauldron on 19 August 2000, shortly after a jökulhlaup in Skaftá that originated in the cauldron, view towards southeast. Photographs: Oddur Sigurðsson.

than the watershed of the cauldron, as assumed in these calculations, this power is not needed to melt ice.

It may be mentioned that there is a possibility that a considerable part of meltwater from the Skaftá cauldrons does not accumulate in the subglacial lakes but infiltrates into the porous bedrock below the ice and emerges as spring water far from the glacier. An indication of this is sulfate-rich groundwater that originates from the glacier (Sigurðsson, 1997). The amount of sulfate-rich groundwater coming from the glacier has been estimated to be 15–$20\,\mathrm{m^3\,s^{-1}}$ (Vatnaskil, 2005; Hreinsson, 2004). This is much more than can be expected to originate from the cauldrons, three to four times the water equivalent of the accumulation of snow over the watershed of the cauldrons. It has been estimated that flow from the cauldrons, in addition to the jökulhlaups, could be 2–$5\,\mathrm{m^3\,s^{-1}}$ at maximum (Vatnaskil, 2005). It is possible that part of the sulfate-rich groundwater from the glacier comes from the cauldrons, the origin of the rest remains unclear. In the calculation on the power of the geothermal area, this possible extra water is omitted but taking it into account would give higher results.

3.2 Jökulhlaups in Skaftá

As the water level in the subglacial lakes rises, the lakes expand and the surface lows in the glacier flatten out. Finally the subglacial seal is lifted and the water escapes the subglacial lakes in jökulhlaups. The flood water flows about $40\,\mathrm{km}$ subglacially until it emerges as a flood in the river Skaftá (Figure 3.1), which gives the cauldrons their name. When the water is expelled from the subglacial lake, the overlying ice collapses and a deep cauldron is reformed. The maximum depth of the cauldron after a jökulhlaup is $\sim\!150\,\mathrm{m}$ for the eastern cauldron and $\sim\!100\,\mathrm{m}$ for the western cauldron. At their filling point, just before a jökulhlaup, they are only tens of meters deep (Guðmundsson and Högnadóttir, 2002). The western cauldron is close to full on Figure 3.2 top, while the lower photograph shows the cauldron shortly after a jökulhlaup. Jökulhlaups in Skaftá occurs almost every year and 45 jökulhlaups are known in the river since 1955. The timing, discharge and volume of jökulhlaups from the cauldrons are shown in Tables 3.1 and 3.2. Data for jökulhlaups prior to 1997 are from Zóphóníasson and Pálsson (1996) while data for more recent events are from the database of the Hydrological Service. The origin of the small jökulhlaup in October 1995 was unclear until Magnússon *et al.* (2007) used interferograms based on InSAR data to show that this event originated in the eastern cauldron.

Jökulhlaups from each cauldron display different characteristics. The jökulhlaups

Table 3.1: Jökulhlaups from the Eastern Skaftá cauldron. Q_m is measured total discharge in Skaftá during each jökulhlaup and Q_b is estimated base flow of other water than the flood water. Likewise V_m is the volume of all measured discharge and V_b is the volume of the base flow during the flood.

Year, month	max Q_m $(m^3\,s^{-1})$	max $Q_m - Q_b$ $(m^3\,s^{-1})$	V_m (Gl)	$V_m - V_b$ (Gl)
1955 09	755			213
1957 05	650			110
1960 09				205
1964 03	885			257
1966 10	1040			230
1970 01	1380	1226	493	394
1972 07	1090	922	407	295
1974 12	1270	1224	306	274
1977 02	1160	1108	320	285
1979 09	1220	1140	346	297
1982 01	1140	1077	384	341
1984 08	1560	1426	411	336
1986 11	1308	1290	247	239
1989 07	1288	1159	354	279
1991 08	1120	928	356	236
1992 09	167	124	167	98
1995 07	1994	1759	599	368
1995 10	96	62	73	37
1997 07	921	728	330	184
2000 08	1240	1083	365	221
2002 09	689	582	267	160
2003 11	241	207	139	98
2006 04	1370	1340	300	270
2008 10	1350	1290	300	265

The origin of the 1957, 1960, 1964 and 1966 jökulhlaups is not certain but is most likely the eastern cauldron. The discharge and volume for the 1995 jökulhlaup are a sum from for Skaftá, Hverfisfljót and Djúpá (see discussion in Section 3.4). Data for the 2006 and 2008 jökulhlaup are preliminary. Discharge measurements are from different gauges in Skaftá, see a description of hydrological measurements in the main text. Note that discharge measurements at Sveinstindur underestimate the jökulhlaup discharge at the glacier margin at high discharge levels due to water that is lost from the main river course as further discussed in the main text.

from the eastern cauldron are normally larger. They most often rise in one to two days reaching maximum discharge in excess of $1000\,m^3\,s^{-1}$. Maximum discharge of nearly $2000\,m^3\,s^{-1}$ has been measured. This discharge peak, which is much higher than the normal discharge of Skaftá, lasts for three to four days (Zóphóníasson and Pálsson, 1996). The flood then recedes quite fast. The volume of the floodwater is on

Table 3.2: Jökulhlaups from the Western Skaftá cauldron. Q_m is measured total discharge in Skaftá during each jökulhlaup and Q_b is estimated base flow of other water than the flood water. Likewise V_m is the volume of all measured discharge and V_b is the volume of the base flow during the flood.

Year, month	max Q_m (m³ s⁻¹)	max $Q_m - Q_b$ (m³ s⁻¹)	V_m (Gl)	$V_m - V_b$ (Gl)
1968 09				110
1971 07	658	532	450	230
1973 12	238	191	151	72
1975 09	307	210	172	53
1977 08	545	390	277	134
1980 01	444	351	214	121
1981 08	472	250	360	110
1983 09	222	138	177	99
1986 07	418	275	242	102
1988 08	394	301	188	106
1990 10	129	101	129	75
1994 08	968	726	330	162
1996 08	672	521	249	147
1997 07	335	169	197	60
1998 09	292	149	323	116
2000 08	699	561	223	150
2002 07	720	620	287	161
2003 09	436	211	290	87
2005 08	723	523		174
2006 09	194	100	145	53
2008 08	390	240	212	102

The origin of the 1968 jökulhlaup is not certain but is most likely the western cauldron. The discharge and volume for the 1994 jökulhlaup are a sum for both Skaftá and Hverfisfljót (see discussion in Section 3.4). Data for the 2008 jökulhlaup are preliminary. Discharge measurements are from different gauges in Skaftá, see a description of hydrological measurements in the main text. Note that discharge measurements at Sveinstindur underestimate the jökulhlaup discharge at the glacier margin at high discharge levels due to water that is lost from the main river course as further discussed in the main text.

average about 250 Gl, with a range from 100 Gl to 400 Gl. A total of 24 events are known since 1955 so the average time between jökulhlaups from the eastern cauldron is approximately 28 months. Discharge curves for the last two jökulhlaups from the eastern cauldron are shown on Figure 3.3.

The cumulative volume of outflow from the eastern cauldron is shown, as a function of time, on Figure 3.4. In total, approximately 5500 Gl of water have been released from the eastern cauldron since the first recorded jökulhlaup in 1955, giving an average rate of water accumulation in the subglacial lake of 9 Gl per month. The rate of water
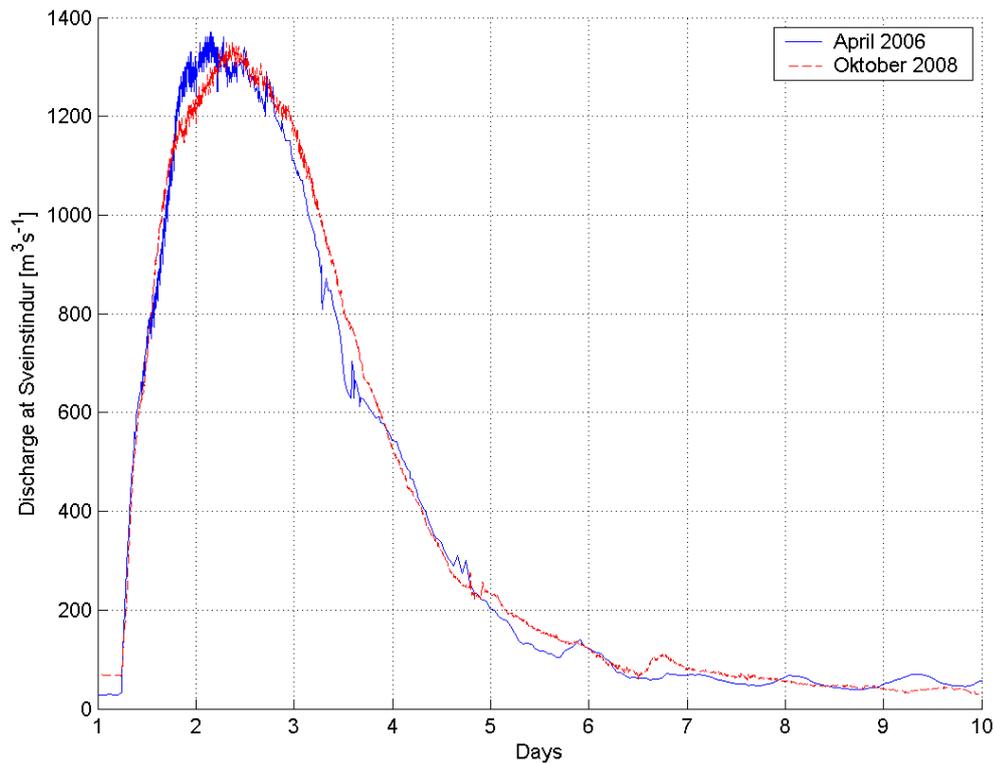
Figure 3.3: Discharge in the 2006 and 2008 jökulhlaups from the Eastern Skaftá cauldron.

accumulation corresponding to the jökulhlaup volume and the numbers of months that elapsed from the preceding jökulhlaup is variable and has a range from 5 Gl to 12 Gl per month. No clear trends can be noticed except a rather low rate of water accumulation between the first four jökulhlaups and between the last six. The low rate of water accumulation between the first four jökulhlaups might be caused by an underestimate of the flood discharge (Zóphóníasson and Pálsson, 1996).

The linear rise at the beginning of the two jökulhlaups on Figure 3.3 is notable. It is also interesting how similar these two events are. Although the discharge curves of jökulhlaups from the eastern cauldron are normally similar these two are remarkably alike. Discharge curves of other jökulhlaups from the eastern cauldron can be seen in Zóphóníasson and Pálsson (1996). It is also worth noting that these two jökulhlaups hit the subglacial hydraulic system in a totally different state as the 2006 jökulhlaup occurred in the spring while the 2008 flood occurred in the autumn. The seasonal timing does not seem to affect the form of the discharge curve. A break in the discharge rise can be seen at approximately $650 \, \mathrm{m^3 \, s^{-1}}$. It is likely that this effect is not caused by a decrease in the rate of change in the jökulhlaup discharge at the glacier snout but is rather caused by flow of water out of the river course of Skaftá and onto porus lava fields between Sveinstindur, where the discharge of Skaftá is measured, and the

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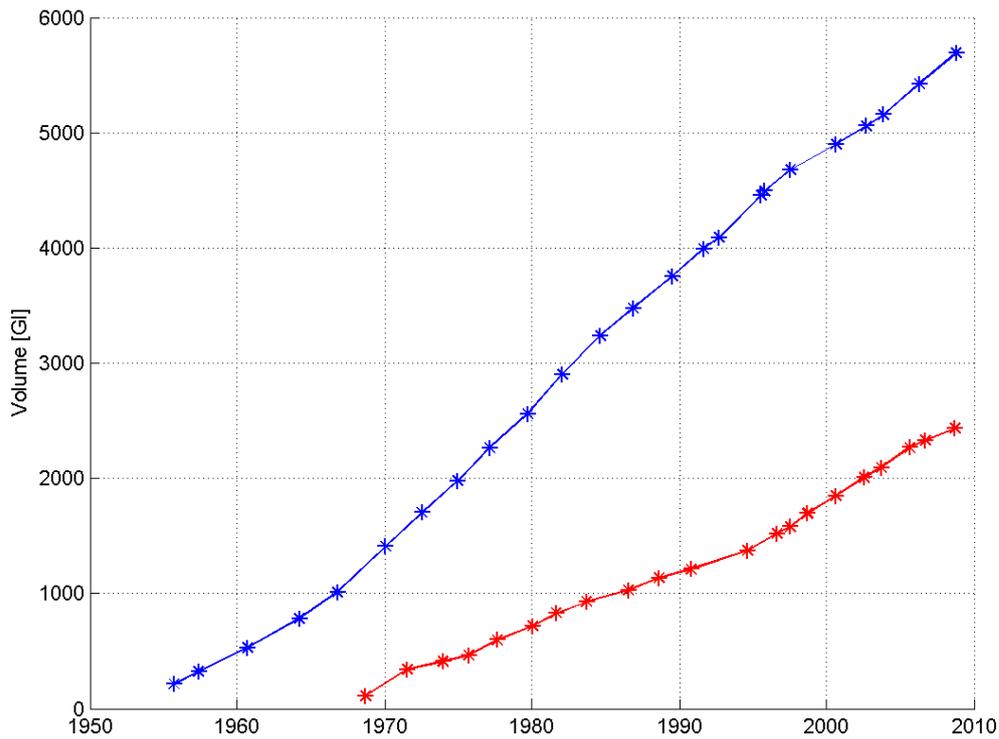
Figure 3.4: Cumulative volume of floodwater from the Skaftá cauldrons, shown with blue curve for the eastern cauldron and with red curve for the western cauldron. Jökulhlaup events are indicated with stars.

glacier. Marks of distributaries flowing out of the river course have been photographed and measured at the same locations for the 1995 jökulhlaup. It is thus likely that discharge measurements at Sveinstindur underestimate the jökulhlaup discharge due to water that is lost from the river course between the ice margin and the gauging station. The deficiency is hard to estimate quantitatively but measurements in the 1995 jökulhlaup indicate that ungauged discharge was $300 \, \mathrm{m}^3 \, \mathrm{s}^{-1}$ when gauged discharge at Sveinstindur was $1310 \, \mathrm{m}^3 \, \mathrm{s}^{-1}$. Part of this discharge infiltrates into the lava fields while most of the rest merges back to the river below the gauging station.

Jökulhlaups from the Western Skaftá cauldron are considerably smaller than from the eastern cauldron but they have been growing in volume and maximum discharge during the last decades, with some exceptions. The maximum discharge of jökulhlaups from the western cauldron is in the range 60–$230 \, \mathrm{m}^3 \, \mathrm{s}^{-1}$ with an average of $120 \, \mathrm{m}^3 \, \mathrm{s}^{-1}$. The volume of the floodwater from the western cauldron ranges from $50 \, \mathrm{Gl}$ to $230 \, \mathrm{Gl}$, with an average of little more than $100 \, \mathrm{Gl}$. Discharge curves for the last two jökulhlaups from the western cauldron are shown on Figure 3.5. The character of jökulhlaups from the western cauldron is different from the outbursts from the east-

ern cauldron. In contrast to the jökulhlaups from the eastern cauldron, the jökulhlaups from the western cauldron have a less distinct discharge peak. They maintain a higher discharge than the normal discharge of Skaftá for a longer time and recede much more slowly with a tail of about one week. During the last decades, with increasing volume and maximum discharge, the jökulhlaups from the western cauldron have become more similar in shape to the jökulhlaups from the eastern cauldron (Zóphóníasson and Pálsson, 1996 and hydrographs since 1996). Since the first recorded jökulhlaup from the western cauldron in 1968, in total 21 events have been recorded, corresponding to an average time between jökulhlaups from the western cauldron of 24 months.

The cumulative volume of outflow from the western cauldron is shown, as a function of time, on Figure 3.4. In total, approximately 2325 Gl of water have been released from the western cauldron since the first recorded jökulhlaup in 1968, giving an average rate of water accumulation in the subglacial lake of 5 Gl per month. A sharp break in water accumulation can be noted in 1994. Before 1994 the average rate of water accumulation was 4 Gl per month while after 1994 the average was 6 Gl per month. This difference indicates an increase in the power of the subglacial geothermal area, and thereby the melting, below the western cauldron. It is also possible that small jökulhlaups such as the jökulhlaup in September 2006 were not noticed before 1994. Missing jökulhlaups would cause an underestimate of the true rate of water accumulation.

3.3 Routing of jökulhlaups from the ice margin to the discharge station at Sveinstindur

In addition to some effect on the form of the jökulhlaup discharge wave, there is also a delay of several hours between the flood wave peaks at the glacier margin and at Sveinstindur, as the flow path between these places is 25 km long. The progress of the flood wave was simulated by Jónsson (2007), using the HEC-RAS hydraulic model. He found a delay of approximately 4 hours for the small jökulhlaup in September 2006. This is shorter than the normal travel time along this stretch of the river during summertime which has been estimated as 7 hours (Kristinsson, 2005). The jökulhlaup flood wave will travel somewhat faster than the actual flow speed of the water because flood waves propagate faster than the water flow as kinematic waves (Viessman and Lewis, 2002). However, for large jökulhlaups, the flood wave is much larger than the normal discharge in Skaftá, so that the flood peak at Sveinstindur can only to a small extent be built up from water that is present in the river course before the flood.
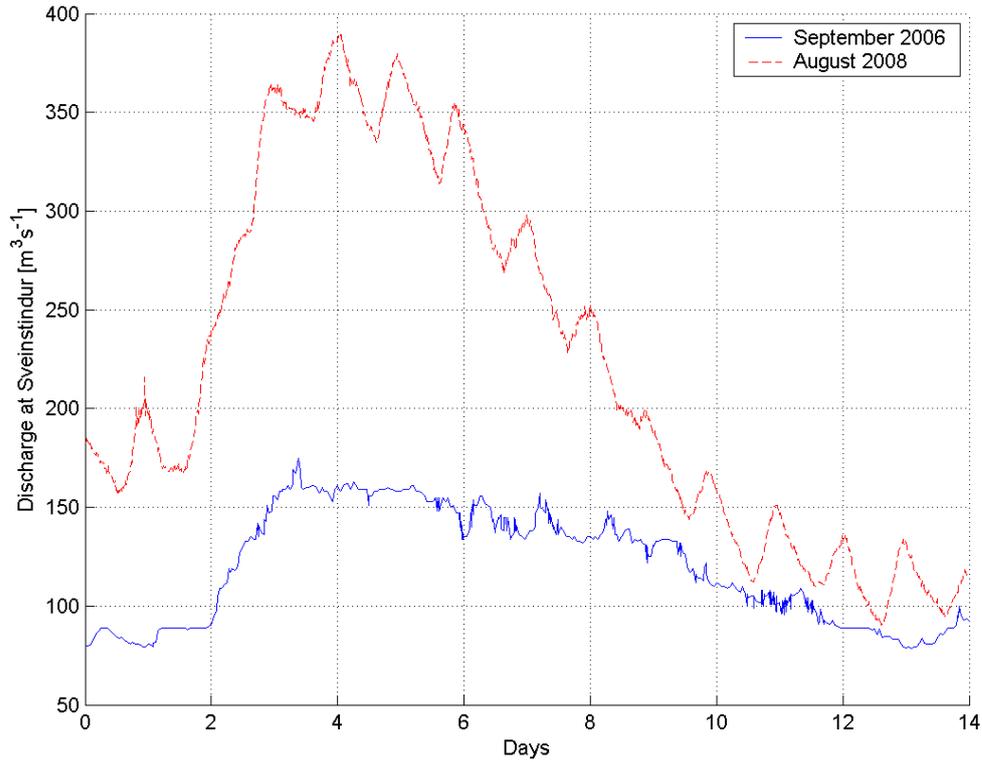
Figure 3.5: Discharge in the 2006 and 2008 jökulhlaups from the Western Skaftá cauldron.

Therefore, the flow speed of the water in a larger flood wave will be a larger ratio of the wave propagation speed than for smaller flood waves.

To calculate a rough approximation for the delay between the glacier margin and Sveinstindur for different jökulhlaups, simple calculation for flow in open channels and the results of Jónsson's (2007) HEC-RAS simulation can be used. Let c denote the propagation speed of a flood wave in an approximately rectangular channel, u_1 and u_2 the flow speeds behind and in front of a flood wave, respectively, and h_1 and h_2 water depth behind and in front of the flood wave. Then

$$h_1(u_1 - c) = h_2(u_2 - c), \tag{3.1}$$

because of continuity. Solving for c leads to

$$c = \frac{h_1 u_1 - h_2 u_2}{h_1 - h_2}. \tag{3.2}$$

Using the Manning equation for discharge per unit width, q,

$$q = \frac{1}{n'} h R_H^{2/3} S_0^{1/2}, \tag{3.3}$$

where n' is the Manning coefficient, h is the flow depth, R_H is the hydraulic radius and S_0 is mean channel slope; the relationships: $u \sim h^{2/3}$, $q \sim h^{5/3}$ and $u \sim q^{2/5}$ can

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be derived for a wide rectangular channel for which $R_H \approx h$ and $u = q/h$. Now using $u_1 h_1 = q_1$, $u_2 h_2 = q_2$, the abovementioned relationships and Equation (3.2) imply

$$c = \frac{q_1 \left(1 - \frac{q_2}{q_1}\right)}{h_1 \left(1 - \left(\frac{q_2}{q_1}\right)^{3/5}\right)}. \tag{3.4}$$

Since the delay is inversely proportional to c, this equation for c can be used to estimate roughly how the delay varies with the flood discharge magnitude q_2. The estimate provided by Equation (3.4) is of course rather rough as it is based on viewing the flood front as a discontinuous step function whereas the real flood wave has a more complex shape.

Equation (3.4) shows that as the ratio q_2/q_1 decreases, that is as the flood becomes larger compared with the base flow, c becomes closer to u_1 as already noted. For the jökulhlaup in September 2006, an average propagation speed $c = 1.7 \, \mathrm{m \, s^{-1}}$ was simulated with the HEC-RAS model for a flood wave with a discharge $q_1 = 150 \, \mathrm{m^3 \, s^{-1}}$, leading to travel time of 4 hours along the 25 km between Sveinstindur and the glacier margin. Assuming a base flow of $q_2 = 75 \, \mathrm{m^3 \, s^{-1}}$, Equation (3.4) gives flow speed $u_1 = 1.2 \, \mathrm{m \, s^{-1}}$.

Using this result for average flow speed and the Manning's equation, a representative mean channel width between the glacier margin and Sveinstindur can be calculated for the jökulhlaup in September 2006. With a higher discharge, the flow will of course become wider but assuming that this represents the average width of the main channel where most of the flow takes place in jökulhlaups, flow speed and travel time can be calculated for larger jökulhlaups. Calculations for a jökulhlaup with an initial flood wave of $Q = 700 \, \mathrm{m^3 \, s^{-1}}$, as can be assumed for large jökulhlaups from the eastern cauldron, results in propagation speed of $2.6 \, \mathrm{m \, s^{-1}}$ and a delay time of little less than three hours.

3.4   The subglacial flood paths

Although the ice drainage basins of the Western and Eastern Skaftá cauldrons are adjacent to and a continuation of the ice drainage basins of Tungnaárjökull and Sylgjujökull (Björnsson, 1988), the flood water from the cauldrons emerges from Skaftárjökull which is to the east of Tungnaárjökull (Figure 3.1). The water follows a subglacial valley to the east of a subglacial ridge that is a continuation of Fögrufjöll, lying in the NE-SW direction of fissures in the area. The flood water crosses a pass in the ridge and emerges in the western branch of Skaftá to the west of the Fögrufjöll Ridge (Björns-
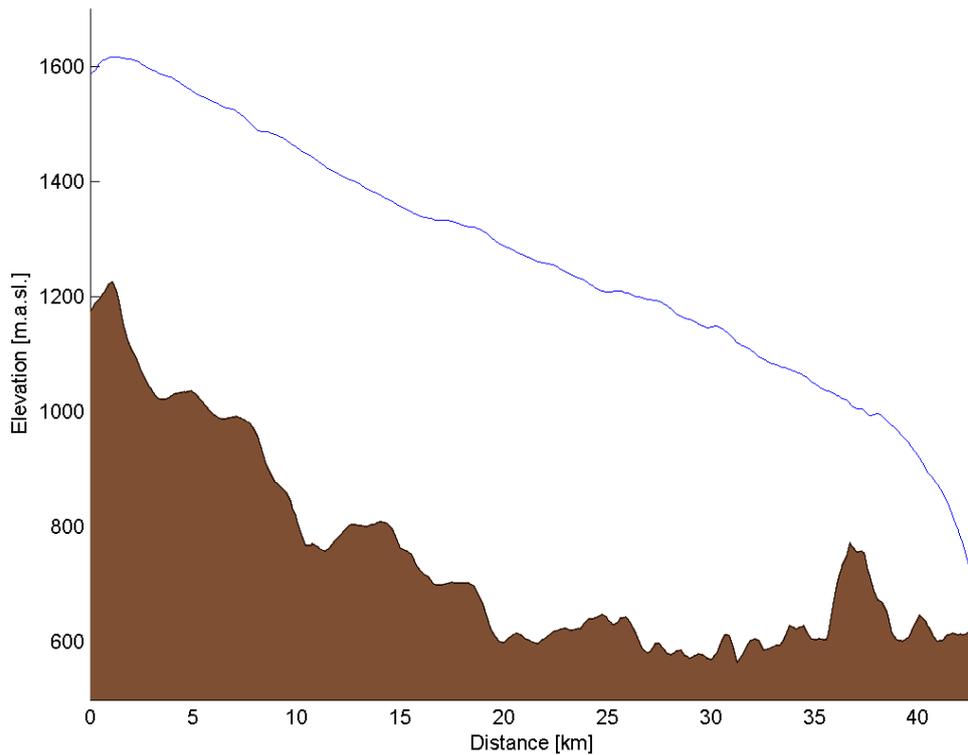
Figure 3.6: A transect of the subglacial path of jökulhlaups from the Eastern Skaftá cauldron and the glacier surface. Data are provided by the Institute of Earth Sciences at the University of Iceland.

son, 1988). Transects of the flood paths from the eastern and the western cauldrons are shown on Figures 3.6 and 3.7 respectively (data provided by the Institute of Earth Sciences at the University of Iceland, 2008).

The theoretical water potential calculated by Magnússon (2003) fails to capture the westward flow across the pass in the subglacial ridge. The water potential indicates that the jökulhlaups should emerge in the eastern branch of Skaftá east of Fögrufjöll. The reason for this discrepancy might be: a possible small-scale basal feature such as a canyon that is not resolved in the subglacial bedrock measurements, that the theoretical potential is too simplistic a view of the flow of subglacial water, or that the water pressure at the glacier bed might be different in jökulhlaups than under normal conditions (Magnússon, 2003).

There is a steep rise at the end of the subglacial flood path because the glacier has considerable overdeepening near the snout (Figures 3.6 and 3.7). The water flows down the subglacial hydraulic water potential, which is largely controlled by the slope of the glacier surface, but up the bottom slope at the downstream end of the overdeepening. The adverse bottom slope is 1.9 times the glacier surface slope and therefore

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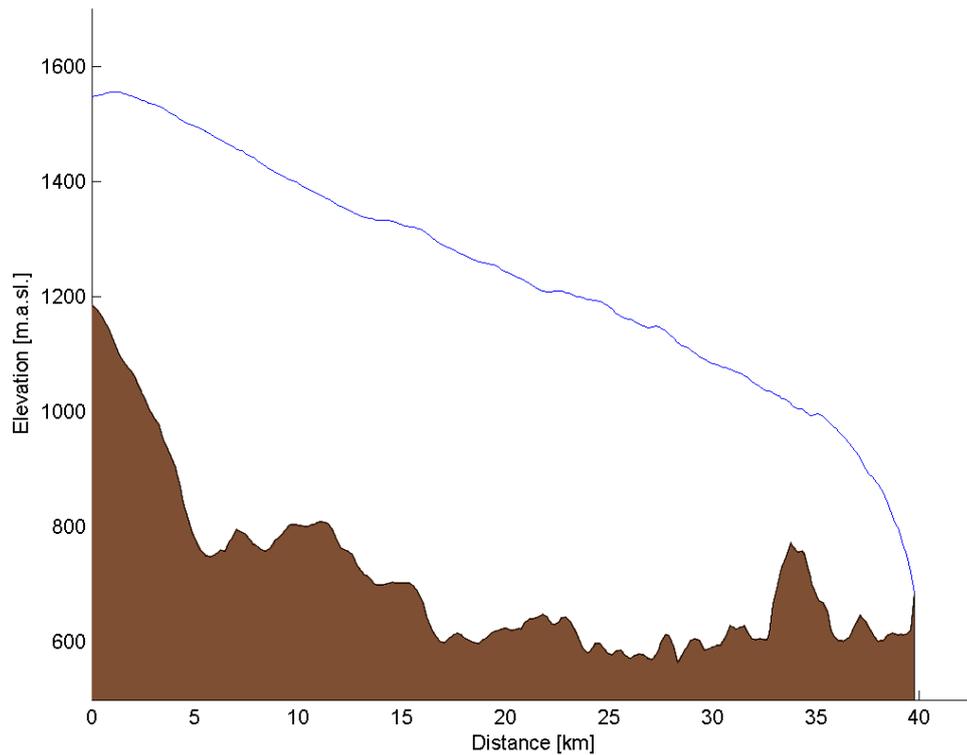
Figure 3.7: A transect of the subglacial path of jökulhlaups from the Western Skaftá cauldron and the glacier surface. Data are provided by the Institute of Earth Sciences at the University of Iceland.

glaciohydraulic supercooling of jökulhlaup water near the snout is possible. Glacio-hydraulic supercooling will theoretically happen if the adverse bottom slope is more than 1.7 times the surface slope (Alley *et al.*, 1998). This value corresponds to water that is unsaturated with air because the water is flowing fast and inside an ice tunnel and therefore changes in dissolved air content are unlikely. The temperature of the flood water in jökulhlaups from both Skaftá cauldrons has been measured to be within a few hundredths of a degree from the freezing point (data from the Skaftá cauldrons research project).

Under normal circumstances, the jökulhlaups emerge only in the western branch of Skaftá, coming from the Skaftárjökull glacier, but incidents are known when the flood water has also emerged in other rivers. In 1994, the jökulhlaup from the western cauldron emerged in both Skaftá and Hverfisfljót, which is the next glacial river east of Skaftá (Figure 3.1). There are some indications from measurements of sediment load that a part of this flood water reached Djúpá (Figure 3.1), the next glacial river east of Hverfisfljót, even though the flood water was not noticeable in discharge measurements in that river. The 1995 jökulhlaup from the eastern cauldron emerged in Hverfisfljót

and as a clearly distinguishable discharge peak in Djúpá (Zóphóníasson and Pálsson, 1996). The reasons for this eastward flow of these jökulhlaups were surges in Skaftár- jökull and its adjacent glaciers, Síðujökull to the east and Tungnaárjökull to the west.

The surge of the snout of Síðujökull took place in the beginning of 1994 but first indications of the surge were noted already in 1990 as an unusual crevasse formation on upper Síðujökull and anomalous low frequency seismic activity (Björnsson *et al.*, 2003). The surges in Tungnaárjökull and Skaftárjökull reached the terminus in October 1994 (Sigurðsson, 1996) and continued for 10 months but an increase in ice velocity had been noted already in 1992 and 1993 (Björnsson *et al.*, 2003). These surges had a major influence on the subglacial hydraulic system. In the 1994 surge of Síðujökull, the surge bulge was observed to restrict the subglacial water flow and, as the bulge reached the snout, silty water emerged from behind it (Sigurðsson, 1995). The surges also affect the surface geometry of the glacier and thereby the water potential as it is largely controlled by the surface shape.

Surges have been observed to have a restricting effect on jökulhlaups. A jökul- hlaup from Lake Grímsvötn in 1991 ceased suddenly before the lake drained out. This jökulhlaup occurred during a surge of Skeiðarárjökull, the glacier that the jökulhlaup from Grímsvötn emerge from. The flow speed of the flood water under the glacier was also abnormally low in this case. After the surge was over, a full jökulhlaup from Grímsvötn developed (Björnsson, 1998). This was not the case for the 1994 and 1995 jökulhlaups from the Skaftá cauldrons but as noted above their course was altered.

Jökulhlaups from the Skaftá cauldrons have also spread to other rivers in non- surging conditions as large jökulhlaups are known to have affected Tungnaá river to the west of Skaftá. This is known to have happened in the April 2006 jökulhlaup from the eastern cauldron (Oddur Sigurðsson, personal communication, 2008).

3.5   The river Skaftá

Skaftá is one of the larger rivers in Iceland. It has a glacial component and a consider- able groundwater component (Rist, 1990). The mean discharge at Skaftárdalur valley is a little over $100 \, \mathrm{m^3 \, s^{-1}}$. The glacial component in the river causes the summer discharge to be higher than the winter discharge because of summer melting but the groundwater component is much more stable within the year. At the glacier margin, the river is formed from several small branches that combine in two main branches, the eastern and the western branch. The western branch comes from the western part of Skaftárjökull while the eastern branch flows from the eastern part of Skaftárjökull

and the western part of Síðujökull. The branches join and the river runs mainly in one course, but is braided into two or more branches at some locations, until it enters the lowland. There it is divided into three distributaries, one flowing onto the lava field from the 1783 Laki eruption, another flowing eastwards between the lava field and the highland edge and a third branch flowing by the west side of the lava field. The eastwards flowing distributary keeps the name Skaftá but the westwards flowing distributary is called Eldvatn by Ásar. Eldvatn by Ásar joins other tributary rivers (Hólmsá and Tungufljót) as it reaches Mýrdalssandur and is called Kúðafljót after that. Before 1783, Skaftá fell in great gorges from the highland and then mainly in similar course as the eastward flowing distributary but with some smaller branches flowing southwards over old lava fields, now overlain by the 1783 lava (Sigurðsson, 1997). In the 1783 eruption, the lava flow followed the Skaftá gorges and then spread over the lowlands. The gorges were filled with lava so now the middle part of Skaftá, that formerly followed the gorges, flows on top of a recent lava field.

Hydrological measurements have been made in Skaftá since 1951. In the beginning the water level was read every third day but in 1966 a continuous recording instrument was installed. These measurements were all done in the eastern branch of the river at Skaftárdalur, about 50 km downstream from the glacier. To connect discharge to water level measurements, discrete discharge measurements have been done for a wide range of water levels. These measurements are used to calculate a rating-curve describing the relationship between water level and discharge. In 1972, water level gauges were constructed at Sveinstindur (closer to the glacier than before, 25 km down river) and at Kirkjubæjarklaustur (a small village by the most easterly distributary of the lower part of Skaftá). Finally, a gauge was built in Eldvatn by Ásar in 1993 (Zóphóníasson and Pálsson, 1996). Cableways for discharge measurements have been constructed at all these locations except at Skaftárdalur were a boat is used.

In 2002, a water level gauge was constructed only 3 km from the glacier in the western branch of Skaftá but the river course was so unstable in that place that a useful relationship between water level and discharge could not be made. Therefore the gauge was taken down in 2004. There is still a cableway at this location and discrete discharge and chemical concentration measurements are sometimes made. In the summer of 2006, a thermometer was placed at this measurement site in the western branch especially to capture the temperature of the jökulhlaup waters as they emerge from the glacier.

At present, Sveinstindur is the main measuring station for jökulhlaups. It is the station closest to the glacier where discharge can be calculated and almost the entire river flows in one branch by the measurement station. In large jökulhlaups, a branch

however, forms to the east of the main branch. Water conductivity, water temperature, light absorption in the water, air temperature and air humidity are measured at Sveinstindur in addition to water level, which is measured with two independent sensors. The conductivity and light absorption sensors have given valuable information about the initiation of jökulhlaups. Peaks are observed in both conductivity and light absorption in connection with jökulhlaups and rising conductivity is often the first sign of an impending jökulhlaup (Gunnar Sigurðsson, personal communication, 2008).

As mentioned before, jökulhlaups from the Skaftá cauldrons have been known since 1955 and descriptions of cauldrons in this area exist since 1945 (Þórarinsson and Rist, 1955). Jökulhlaups in Skaftá about every year at least since 1910 are described in available accounts. These floods were small in magnitude before 1955 and the river could be crossed on horse within one day from the start of the flood (Björnsson, 1977). The reason for this difference in magnitude down in the lowlands might be that Skaftá fell into the lake Langisjór, with an area of $26\,km^2$, before 1967 (Tómasson and Vilmundardóttir, 1967). Part of the jökulhlaup water or all of it probably followed the same route so the flood wave was dampened in the lake and part of the sediment load was deposited in the lake. In addition, the outflow from Langisjór is through a narrow gap with a width of 23 m and steep walls which restricts the outflow and increases the dampening further (Zóphóníasson and Pálsson, 1996). It is also possible that the geothermal activity under the glacier was different but this is unknown.

3.6 Earthquake tremors

Earthquake tremors have been noticed in connection with jökulhlaups from the Skaftá cauldrons. Both small-scale events in connection with the initiation of the jökulhlaups, most likely caused by icequakes (Matthew Roberts, personal communication, 2008), and larger events in later stages. The larger events take place after the main discharge peak is over. The larger events are low-frequency tremors and have been associated with either boiling in the geothermal system because of the pressure release, as the water is drained from the cauldron, or a small eruption, which would likewise be caused by the pressure drop (Björnsson and Einarsson, 1990). Such events have been observed for both cauldrons. About a day after the tremor, a peak in chemical concentrations in the river has been observed at Sveinstindur for some events, for other events a similar peak has been observed later farther down the river. There is also a distinct change in the sediment load of the river, and for one incident (the July 1995 jökulhlaup from the eastern cauldron), an increase in discharge seems to have been associated with the tremor (Zóphóníasson and Pálsson, 1996). The changes in sediment load and chem-

ical concentrations are so distinct that the river has been observed to become much more light-colored than the normal color for jökulhlaup flood waters far down the river course in the lowland.

Chapter 4

Theory of jökulhlaups

The flow of water in glaciers is heterogeneous in space and time and affected by many factors, for example: meltwater and precipitation input, glacier geometry and morphology, snow and firn pack, and amount and properties of sediments at the glacier bed. Different modes of water flow in glaciers have been described. Here a short overview of the hydrology of temperate glaciers will be given.

Water in glaciers mainly originates from ablation of ice and snow or from precipitation but basal melting may also contribute, though typically in smaller amounts. The water draining from a temperate glacier is normally observed to emerge at the glacier snout in many outlets cut into the glacier underside. The discharge is normally at maximum in late summer and with a marked diurnal variation. The diurnal variation is superimposed on a much more slowly varying base flow. During wintertime, the discharge from the glacier is much smaller and then without a diurnal variation, or not observable at all.

Water can flow on the glacier surface, within the glacier or subglacially. Good overviews of glacier hydrology are given by Paterson (1994) and Fountain and Walder (1998).

4.1   Surface flow on glaciers

The glacier surface is normally snow covered during winter. As spring comes and melting starts, water initially accumulates in the snowpack and quickly eliminates the sub-zero temperatures near the surface of the glacier. Some of the meltwater is retained in the snowpack but most of it percolates down until it meets impermeable ice in the ablation zone or firn in the accumulation zone. In the accumulation zone, percolation continues through the firn until the water reaches the horizon of glacier ice. Above the

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firn–ice interface, the firn becomes water-saturated and downslope water flow, similar to groundwater flow, is formed. At crevasses, the water is drained from the firn, in a similar way as water drains from a perched unconfined aquifer. These processes are slow and smooth out diurnal and longer variations in meltwater and precipitation input in the accumulation zone. Water flow through firn is therefore believed to be the main source of the base flow of glacial runoff (Fountain and Walder, 1998). These processes continue during the entire summer. In the ablation zone, similar processes take place at the snow–ice interface in early summer but as the snow melts and bare ice is exposed, surface channels start to form. Later in the summer, these channels resemble ordinary river systems except that much of the water is captured by crevasses and moulins (Paterson, 1994). The supraglacial part on the ablation zone is similar to the runoff system in karst regions. As winter returns, input from melting and precipitation decreases and these surface forms (channels, crevasses and moulins) become snow filled or snow covered.

4.2 Englacial flow

As the water flows into crevasses and moulins it becomes englacial and flows to the bottom of the glacier through moulins and in englacial conduits. Whereas moulins are nearly vertical shafts formed by overdeepening of water-filled crevasses, englacial conduits are formed by water flowing at the bottom of crevasses that melts its way down into the glacier by frictional heat (Fountain and Walder, 1998). An arborescent network of englacial passages formed from water flow through small englacial veins or microfractures is also part of the system. Inter-granular veins at crystal boundaries in the ice are on the other hand believed to drain a negligible part of the total water flow. The reason for this is that inter-granular veins are observed to be quite small and may often be blocked by air bubbles. The presence of surface pools and streams on glaciers, furthermore indicates that ice is relative impermeable (Fountain and Walder, 1998). Theoretical calculations have been carried out for the formation of englacial conduits and evidence of them has been seen in boreholes (Fountain and Walder, 1998; Shreve, 1972).

4.3 Subglacial flow in Röthlisberger channels

This system is sometimes referred to as the fast system or the arborescent subglacial hydraulic system. As water is often observed to exit glaciers in conduits at the inter-

face between the glacier and the bed and conduits at this interface have been followed upglacier during times of low discharge, it has been concluded that during the melt season a large part of the discharge from the glacier takes place in conduits at the glacier–bed interface. Röthlisberger (1972) described this type of channel and introduced equations that describe how heat formed by frictional energy dissipation in the water flow melts the conduit walls and enlarges the conduits. Heat for melting can also originate from the adjustment of the water temperature to a colder pressure melting point, in case the water flows towards a higher pressure, or in heat that the water carries with it as it flows into the glacier if it originates from rain or from a glacial lake (Björnsson, 1988). Röthlisberger also described how ice flow caused by overburden pressure counteracts enlargement of the channels by melting and tends to close the conduits and how a conduit can be maintained in a steady state by a balance between these two effects. He furthermore showed that under a steady state the system will be arborescent because the water pressure in two adjacent conduits will be lower in the one with a higher discharge. This causes large conduits to grow at the expense of the smaller ones and the system will form a tree-like pattern where small conduits merge into larger and larger conduits.

Röthlisberger's reasoning has been debated and the assumption of steady state is probably unrealistic in the ablation area of a glacier were the water input is constantly changing with daily variations in melting (Fountain and Walder, 1998). Other mechanisms for the growing of larger conduits rather than small ones have been suggested, such as a larger generation of heat per unit of surface area in a large conduit compared with a small one, or that larger and deeper drawdown cones around inlets of larger conduits compared with smaller ones cause lower pressures in the larger conduit. The effect of conduit flow dominating the subglacial hydraulic system, as expected in the ablation zone in summertime when inflow is high, is a large transportation capacity and therefore fast travel speeds. The water pressure at the glacier bed will also be comparatively low and the effect of basal water on sliding speed and other dynamic processes will be localized in the small area covered by the conduits.

4.4   Subglacial flow in linked cavities

This system is sometimes referred to as the slow system or the non-arborescent subglacial hydraulic system. As the glacier slides over irregularities in the glacier bottom, cavities are formed on the lee sides. The cavities are full of meltwater and connected through small orifices so that water can flow through the system. The water pressure is high and flow velocities are slow. Furthermore, the water pressure increases with

increased flux so there is no tendency for the larger channels to grow more rapidly than the smaller ones (Paterson, 1994). Therefore, this system can be maintained distributed over large areas at the glacier bed and the water has multiple different flow paths that do not tend to merge with one another as the water flows downglacier. This system is believed to be dominant during wintertime when meltwater input is low and it has been suggested to be connected to glacier surges (Kamb, 1987; Raymond, 1987).

4.5   Other modes of subglacial flow

Other modes of subglacial flow are also known. Where the bed is made up of soft or highly soluble material, water can flow in channels incised into the bed rather than channels melted into the ice that are called Nye channels (Paterson, 1994). An idea of water being mainly drained from a glacier in a thin sheet under more or less the whole glacier was suggested by Weertman (1962) but it was rejected by Nye (1973) that such a film could play a significant part in draining the glacier. It is though likely that a much thinner film of water, connected to the regelation process, is present at the glacier–bed interface (Nye, 1973). Porus sediments and bedrock under glaciers may also in some places drain considerable part of water from the glacier as groundwater flow.

4.6   Water flow in jökulhlaups

Many jökulhlaups are observed to rise in a regular fashion such that the discharge increases approximately exponentially by many orders of magnitude over a period from several days up to a couple of weeks. A theory for such slowly-rising jökulhlaups is well established and has been used successfully to describe many jökulhlaups, in particular jökulhlaups from Grímsvötn in Vatnajökull (Nye, 1976; Spring and Hutter, 1981; Björnsson, 1992). Many jökulhlaups are however observed to rise much faster than can be explained by this theory and they are not as well understood. The 1996 jökulhlaup from Grímsvötn (Snorrason *et al.*, 1997; Björnsson, 1997; Jóhannesson, 2002) is a well known example of a fast-rising jökulhlaup as well as a jökulhlaup from Grímsvötn in 1938 (Þórarinsson, 1974) and most jökulhlaups in Skaftá (Björnsson, 2002). Here the classic theory of slowly-rising jökulhlaups will be described.

The mechanism of slowly-rising jökulhlaups was described by Nye (1976) with jökulhlaups from Grímsvötn in Vatnajökull Iceland as an example. Similarly to flow in Röthlisberger channels, the flow is assumed to be in a conduit in the ice, either at

the glacier bottom or englacial. This conduit is assumed to be straight with uniform geometry and to connect the source lake to the glacier snout. Frictional heat released in the flow by potential energy dissipation and the initial heat of the source water melts the walls of the tunnel so the tunnel grows. This leads to a positive feedback because the larger tunnel carries more flow causing more frictional heat to be released which causes further enlargement of the tunnel and so forth. Ice deformation caused by ice overburden pressure tends to close the tunnel and as the water level in the source lake falls, the jökulhlaup is terminated by tunnel closure or by the exhaustion of water in the reservoir.

In contrast to normal flow in Röthlisberger conduits where the water source is limited and the system may find a stable equilibrium for a constant discharge by changing the pressure and tunnel size, the pressure head is almost constant for a jökulhlaup of this type, determined by the pressure drop from the source lake to the glacier snout. This causes the tunnel size and discharge to be unstable and this fundamental instability causes subglacial lakes to drain in outbursts and not by steady flow (Nye, 1976).

Unstable flow in conduits can be described by differential equations for the balance of mass, the balance of momentum, the balance of energy and for changes in tunnel geometry and the flow of the surrounding ice. During the rising phase of the jökulhlaup, this set of differential equations, with simplifications, has the following solution for discharge as a function of time (Nye, 1976)

$$Q = \left(\frac{4}{k_1(t_0 - t)}\right)^4 , \tag{4.1}$$

where t_0 is the time of a hypothetical vertical asymptote. The coefficient k_1 is given by

$$k_1 = \frac{(4/3)(\rho_w/\rho_i)g}{(S/R_H^2)^{1/4}Ln'^{3/4}}(\Delta z/\Delta x)^{11/8} , \tag{4.2}$$

where ρ_w is the density of water, ρ_i is the density of ice, g is the acceleration of gravity, Δz is the vertical height difference from the water level in the reservoir to the glacier terminus, Δx is the horizontal length of the tunnel, S is the tunnel cross-sectional area, R_H is the hydraulic radius of the tunnel, L is the latent heat of melting and n' is the Manning's roughness coefficient.

Equation (4.1) describes an approximately exponentially rising discharge. Some days before the vertical asymptote, the discharge starts to diminish and the simplifications behind Equation (4.1) break down as tunnel closure by deformation becomes more important than melting by frictional heating.

Many simplifications are made in Nye's theory, the most important of which are that the potential gradient driving the flow is equal to the mean potential gradient averaged from source lake to glacier snout and that the initial heat of the water is ignored.

Calculations of heat transfer from the water to the conduit walls are also omitted and spontaneous heat transfer is assumed. Somewhat unexpectedly, this assumption seems to be closer to reality for many jökulhlaups than a more physically-based formulation based on empirical expressions for heat transfer in pipe flow which seems to result in too slow heat transfer (Jóhannesson, 2002).

Some of these simplifications were relaxed by Spring and Hutter (1981) and Clarke (2003) who used numerical methods to solve a system of equations similar to the full equations derived by Nye (1976) without any of the simplifying assumptions used to derive Equation (4.1). This work shows that Nye's theory provides a good physical description of slowly-rising jökulhlaups but at the same time demonstrates that even the full theory without simplifications and with some additional features proposed by Spring and Hutter and other authors is unable to explain fast-rising jökulhlaups such as the jökulhlaups from the Skaftá cauldrons.

Chapter 5

Data

An extensive campaign involving measurements within the Skaftá cauldrons, in the subglacial lakes and of the jökulhlaups originating in them, was initiated in 2006. The 300 m thick ice shelf covering the western cauldron was penetrated with a hot water drill in June 2006 (Thorsteinsson *et al.*, 2007; Jóhannesson *et al.*, 2007). Temperature profiles in the lake were measured and a water sample was taken for geochemical and microbiological studies. A pressure and temperature sensor was deployed at the bottom of the lake and connected with a cable to a continuously recording datalogger at the ice shelf surface. A continuously recording differential GPS instrument was placed at the center of the ice shelf and a water temperature logger was placed in the Skaftá river some distance downstream from the port where the river emerges from beneath the ice. In addition, data from the Sveinstindur hydrological station in Skaftá have been used in this study.

The discharge in the jökulhlaup from the Western Skaftá cauldron in the autumn 2006 was measured at the hydrological station at Sveinstindur, as for other recent jökulhlaups from the Skaftá cauldrons. The discharge at the outlet of Skaftá at the glacier snout was back-calculated using flood routing with the HEC-RAS hydraulic model. Furthermore, the outflow from the cauldron during the jökulhlaup was calculated from GPS measurements of the elevation of the ice shelf.

5.1   Discharge in Skaftá and the September 2006 jökulhlaup

The hydrological measurement station in the Skaftá river closest to the glacier margin is at Sveinstindur, 25 km downstream from the glacier. Water level measurements and a rating curve, derived from point measurements of discharge and water level, were

used to calculate the discharge. The water level is measured with a *Druck* pressure sensor and the data are recorded with a *Campbell CR-10X* data logger.

In addition to the jökulhlaup component, the discharge measured at Sveinstindur includes the normal discharge from the glacier and from tributary rivers. Both of these components depend on weather.

Two major tributary rivers enter Skaftá between the glacier snout where the jökulhlaups emerge and Sveinstindur, where the discharge of the jökulhlaup is measured. These rivers are the outlet from lake Langisjór and the eastern branch of Skaftá coming from the eastern part of Skaftárjökull and the western part of Síðujökull. A number of smaller creeks from Fögrufjöll also join the river upstream of Sveinstindur. During the autumn, the discharge in these tributary rivers is mainly controlled by precipitation. During the jökulhlaup in the autumn 2006 (Figure 3.5) there is only one considerable precipitation event measured at the nearby precipitation stations, Laufbali and Kirkju-bæjarklaustur. This event occurred on the 5 October, with a cumulative precipitation of about 40 mm during 3 hours in Laufbali. This precipitation caused a small discharge peak in Djúpá and Hverfisfljót, which are nearby rivers that have similar discharge characteristics as Skaftá, but this peak is small compared with the discharge of the jökulhlaup. Therefore, the discharge of tributary rivers entering the western branch of Skaftá between the glacier snout and Sveinstindur was assumed to be constant during the jökulhlaup. This discharge has been measured by discrete discharge measurements to be between 20–40% of measured discharge at Sveinstindur for similar total discharge during autumn, see Figures 2 and 3 in Kristinsson (2005). The mean discharge during the days around the jökulhlaup was around $75 \, \text{m}^3 \, \text{s}^{-1}$ and therefore the value of this constant tributary discharge may be assumed to be approximately $20 \, \text{m}^3 \, \text{s}^{-1}$.

The distance from the glacier snout to the gauge at Sveinstindur affects the timing and the shape of the flood wave at the gauge location. This needs to be taken into account in estimating the speed of the flood front under the glacier, the total storage of water under the glacier and other characteristics of the subglacial flood wave. The one-dimensional hydraulic model HEC-RAS was used by Jónsson (2007) to calculate the shape and timing of the jökulhlaup at the glacier snout, as mentioned in Section 3.3.

To obtain an estimate of the jökulhlaup discharge at the glacier snout, the normal glacial discharge component must be subtracted from the discharge calculated with the hydraulic model. The normal glacial discharge is mainly due to daily melting of the glacier but inflow from ground water and englacial storage, precipitation and other factors do also contribute but in smaller amounts. The daily melting is controlled by air temperature, wind, radiation and precipitation. Discharge of other glacial rivers in

the southern part of Iceland indicates that glacial runoff can be assumed to be fairly constant during the jökulhlaup, except for the daily variation and a distinct peak in the discharge from 27 September until 30 September. This discharge peak is connected to unusually warm temperatures during these days; daily average temperature was up to 5°C warmer than on other days before, during and after the jökulhlaup.

Based on these considerations, the normal glacial discharge was assumed to be constant except during the days of the temperature peak. The diurnal variation in the glacial discharge was not considered and therefore affects the jökulhlaup discharge but these variations are comparatively small compared with the changes in the jökulhlaup discharge. Furthermore, if the constant base discharge is estimated as the average of the diurnal variation during the days around the jökulhlaup the diurnal variation will average out in volume calculations.

As the discharge peak caused by the temperature peak is superimposed on and masks the early part of the jökulhlaup, it needs to be determined and subtracted. As a first approach, simple degree-day melt calculations were carried out for the ablation area of Skaftárjökull but it can be assumed that most of the discharge appearing from the glacier on a short time-scale originates as melt from the ablation zone. Temperatures from the weather station at Laufbali were used. Laufbali is located about 30 km southeast of the glacier at an elevation of 560 m a.s.l. A degree-day factor of $0.007 \, \mathrm{m_{w.e.}} \, °\mathrm{C}^{-1}\mathrm{d}^{-1}$ was used based on earlier studies of Icelandic glaciers (Jóhannesson, 1997). Assuming that all this melt appears as discharge with a delay of only a few hours gives an unreasonably large discharge peak. The reason for this failure is most likely a combination of several factors: part of the melt may filtrate into groundwater under and in front of the glacier, a part may become retained englacially within the glacier or in an early autumn snow layer in the upper part of the ablation zone. If the upper part of the ablation zone was covered with snow, melting there would have been much smaller than for bare ice.

To account for at least the retaining factors, another approach was used in preference to the degree-day calculations. A similar peak in temperature without any precipitation was found from another period and the corresponding discharge peak was used for comparison. This temperature peak was in the autumn of 2008. The 2008 temperature peak lasted only for two days while the temperature peak in 2006 was four days long. The ratio of the total number of degree-days for the duration of the temperature peaks was used to scale the volume of the discharge peak caused by the 2008 temperature peak to obtain the volume of the 2006 discharge peak. The time-scale for the 2008 discharge peak was then stretched according to the ratio of the durations of the temperature peaks. This scaled and stretched discharge peak was then adopted as an

estimate for the discharge contribution caused by the 2006 temperature peak. A constant base flow of $55\,\mathrm{m^3\,s^{-1}}$ due to groundwater flow and other discharge components that may be assumed to be constant for the duration of the jökulhlaup was furthermore assumed.

By this method a plausible time-series of the discharge from the glacier due to other reasons than the jökulhlaup was constructed. This discharge was then subtracted from the estimated discharge at the glacier snout to give the jökulhlaup discharge at the glacier margin (Figure 5.1).
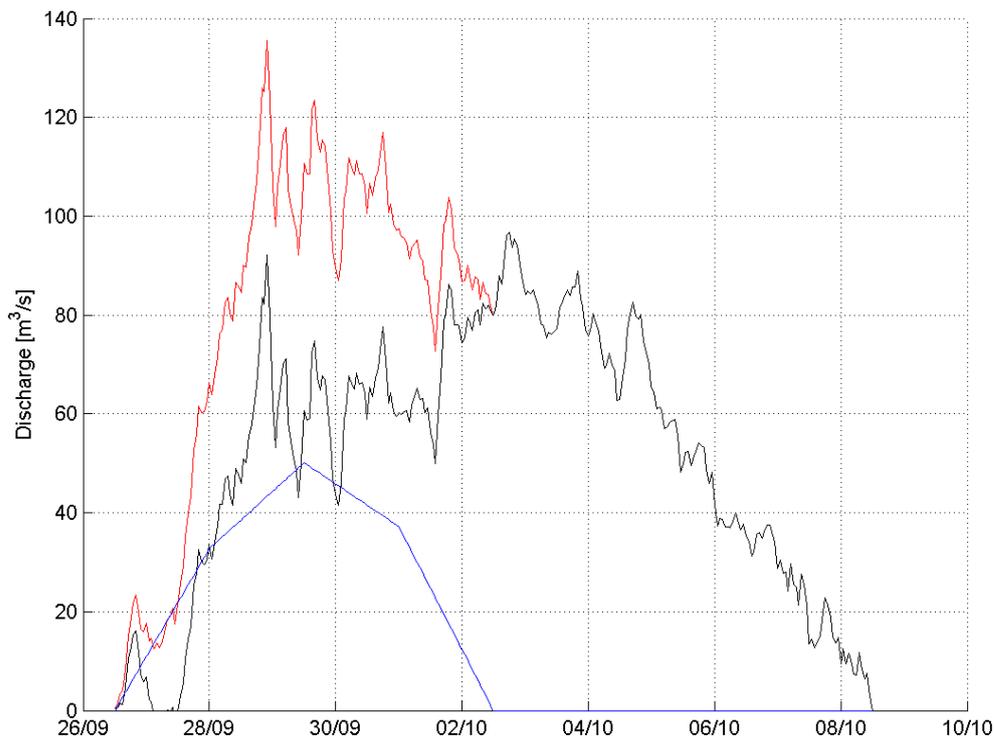
Figure 5.1: Jökulhlaup discharge from the Western Skaftá cauldron in September 2006 as calculated at the glacier snout. The discharge at the glacier margin back-calculated with the HEC-RAS model with a $55\,\mathrm{m^3\,s^{-1}}$ base flow subtracted is shown as a red curve. The estimated discharge peak because of the warm spell is shown with a blue curve and the resulting jökulhlaup discharge is shown as black curve.

5.2   Discharge from the cauldron

The outflow from the cauldron during a jökulhlaup can be calculated from continuous measurements of the water level of the subglacial lake by using a volume–elevation

relationship curve (hypsometry curve). During the summer 2006, the lake level was measured with a pressure sensor at the lake bottom and with a GPS receiver on the surface of the ice shelf. Unfortunately, the cable to the pressure sensor broke on the 16 September 2006, eight days before the jökulhlaup, and therefore only the GPS measurements are used here to calculate the discharge. The reason that the cable broke is probably the intense internal deformation in the ice shelf, which may be assumed to be in a state of horizontal compression combined with vertical straining.

The GPS instrument was a *Trimble 4000SE* GPS receiver with a *Trimble 4000ST L1 Geodetic antenna*. Due to the limited memory capacity, the receiver was set to log 15 second data for 5 minutes at 08:00 o'clock every morning, providing one GPS-location per day, which was found as an average of the measurements during the 5 minute logging interval. The antenna was mounted approximately 4 m above the snow surface on a mast that consisted of 3 stakes drilled approximately 2 m into the snow near the centre of the small hill in the middle of the ice shelf. The data were processed with the *Trimble Geomatics Office* software using base station data from Grímsfjall (GFUM, length of baseline ∼20 km) and Skrokkalda (SKRO, length of baseline ∼37 km) kindly provided by Halldór Geirsson at the Icelandic Meteorological Office. Based on fluctuations in the measurements between adjacent days during time periods of slow movements before and after the jökulhlaup, the accuracy of the measured z coordinates may be estimated to be better than 1 m.

The antenna height above the snow surface and the local geoid height of 67.7 m above the GRS80 reference ellipsoid (Moritz, 1980) were subtracted from the GPS measurements, giving elevation of snow surface above sea level. Variations in the height due to snow accumulation and melting are not accounted for, but the resulting height should provide an estimate of the elevation of the snow surface within 1 m. Small-scale variations in the snow surface elevation near the middle of the cauldron are much larger than this uncertainty. As relative elevation changes of the whole ice shelf are sought, this uncertainty does not affect the quality of the data for estimating outflow from the subglacial lake.

As mentioned, the GPS instrument had only limited memory capacity and could therefore not record for more than about 40 consecutive days. This is the cause for the gap in the GPS data from 25 July to 14 September (Figure 5.2). During a few days near the middle of September there are some fluctuations in the data. It is not clear whether these variations are due to measurement errors of 1–2 m or due to fluctuations in the ice shelf elevation when the seal preventing outflow from the lake was about to break. The rise of the ice shelf during the period 25 July to 14 September, when the GPS measurements are not available, is not fully consistent with the rise in the

water level during the same period. This difference could be due to a slow variation of the amplitude of undulations in the surface geometry of the ice shelf during this time period. This would require the central hill in the ice shelf geometry to have been lowered by 2–3 m in this time period with respect to the surrounding depression in the ice shelf.

The pressure transducer that was deployed on the lake bottom was a *Geokon 4500-SHSR*-7.5 MPa vibrating wire sensor connected to a *Campbell CR-10X* data logger on the surface through a steel reinforced cable. The accuracy of the sensor is ±0.1% of its 7.5 MPa measurement range corresponding to 0.75 m in water depth. The measurements are not corrected for variations in atmospheric pressure. As a change of 100 Pa in atmospheric pressure equals a water level changes of 1 cm, expected fluctuations in atmospheric pressure can cause errors of similar magnitude as the inaccuracy of the sensor. It was deployed in June 2006 after the drilling through the ice shelf. Sitting at the bottom of the lake, the sensor measured overburden pressure due to the weight of the ice shelf and the water column in the lake. This equals the water level in a borehole drilled through the ice shelf down into the lake. During the time period when both measurements are available, the water level measurements indicate that the elevation of the ice shelf gives a good indication of the lake level (Figure 5.2).

Two methods were tried to estimate the shape of the lake underneath the cauldron and variations in this shape during the emptying of the lake as this must be known to construct the volume–elevation relationship. For both methods, a good surface map is needed. The newest and by far the best map of the cauldron is the map published in the MS thesis of E. Magnússon (Magnússon, 2003). It is available on a 5x5 m digital grid and covers a large part of western Vatnajökull and incorporates both the Western and Eastern Skaftá cauldrons. The map is based on aerial SAR measurements from August 1998.

The last jökulhlaup from the western cauldron, before the measurements for the map were done, occurred in July 1997 so the cauldron had been filling for little more than one year. The next jökulhlaup from the cauldron was in September 1998 or shortly after the measurements. In terms of the jökulhlaup in September 1998, the cauldron could be assumed to be almost full in August 1998 but as the normal time between jökulhlaups is two years it could just as well be assumed to be half full. According to GPS measurements of the lowest point in the cauldron by Institute of Earth Sciences at the University of Iceland, the bottom position before the 1998 jökulhlaup was not lower than in the 1997 and 2003 jökulhlaups, for example, but it was approximately 15 m lower than in the 2000 and 2002 jökulhlaups (Guðmundsson and Högnadóttir, 2002). With regard to the volume of outburst water, the 1998 jökulhlaup was of full size for
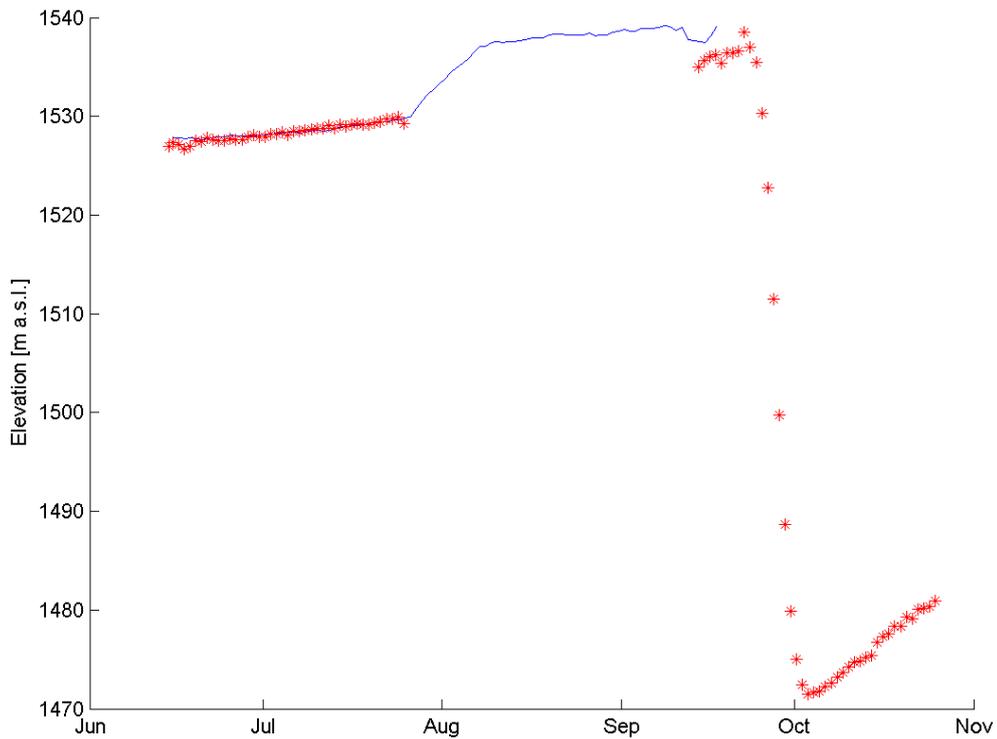
Figure 5.2: Ice shelf elevation in the Western Skaftá cauldron and measurements of bottom pressure in the underlying subglacial lake. The pressure measurements have been shifted so that they represent the elevation of the ice shelf surface. Pressure measurements are shown as a blue line while GPS elevations are marked as red stars.

the western cauldron while the 1997 jökulhlaup was small. The 2006 jökulhlaup was also small. Taking all this into account, the cauldron should be assumed close to full in August 1998. This has both advantages and disadvantages because one of the two methods to estimate the outflow needs to be based on a map of a full cauldron while the other should be based on a map of an empty cauldron. As the map of Magnússon is by far the best available, it is used for both methods but with some adjustments when a map of an empty cauldron is needed.

Before the map was used account had to been taken of the general lowering of the glacier between 1998 and 2006. GPS measurements were available at ice velocity stakes in the area around the cauldron but only stakes at some distance away from it were used in order that they are not affected by the elevation changes of the ice shelf of the cauldron. Elevation at the same locations was found from the map and compared to the GPS elevations in autumn 2006. The average difference between the elevations of the map and from the GPS measurements was 7 m. The map was always higher than the GPS measurements. The greatest difference was 10.0 m and the smallest was 3.2

m. Five stakes were used for this comparison. As Magnússon's map has a 5x5 m grid size and is a little noisy, the elevation from the map was determined as the mean of gridpoints within a 20 m x 20 m (4x4 points) square around the location of each stake.

Lake shape estimated by assuming floating balance. The first approach was to assume that the ice shelf over the cauldron is in a local floating balance. If the ice shelf floats on the water surface represented by the water level in the borehole, which was drilled through the ice shelf in June 2006, then the roof of the water body below it would represent the shape of the surface cauldron mirrored over the water surface and magnified with factor: $\rho_i/(\rho_w - \rho_i)$ where ρ_i is density of ice and ρ_w is density of water. This factor is close to ten for normal values for ρ_i and ρ_w. The idea is to think of the glacier as a giant iceberg submerged in water and calculate a fictitious bottom profile of that iceberg. At the location of the lake, this fictitious bottom surface will be higher than the glacier bottom and represent the roof of the lake. The factor $\rho_i/(\rho_w - \rho_i)$ is derived from equation

$$(h_u + h_l)\rho_i = \rho_w h_l \Leftrightarrow h_l = h_u \rho_i/(\rho_w - \rho_i), \tag{5.1}$$

were h_u is height of the part of the "iceberg" that is above the water and h_l is the height of the part that is below the water surface. In terms of the surface elevation of the ice shelf, z_i, the elevation of the water level, z_w, and the elevation of the bottom of the ice shelf (the lake roof), z_{ib}, Equation (5.1) can be written as

$$(z_i - z_{ib})\rho_i = (z_w - z_{ib})\rho_w \tag{5.2}$$

and the unknown z_{ib} can be calculated by

$$z_{ib} = (z_w \rho_w - z_i \rho_i)/(\rho_w - \rho_i) \tag{5.3}$$

Note that at the edge of the lake (at the grounding line) z_{ib} will become equal to the elevation of the bedrock bottom below the glacier and the lake, z_b, and Equation (5.3) is not valid outside this line.

Before the surface can be reflected by Equation (5.3), account has to be taken of the low density of the surface firn layer. The layer is few tens of meters thick and with density rising from some hundred kilograms per cubic meter at the surface to $830\,\mathrm{kg\,m^{-3}}$ at the transition of firn to ice and then up to $\sim 910\,\mathrm{kg\,m^{-3}}$ by compression of air bubbles because of overburden pressure. It is assumed in the derivation of Equation (5.3) that the whole glacier has the same density of $910\,\mathrm{kg\,m^{-3}}$ and to account for this assumption, the calculations are done in ice thickness equivalents but not in the actual thickness.

Density profiles from ice cores are known for few places in Iceland. The largest core projects until now are the 415 m long core from Bárðarbunga in 1972, a 100 m long core from Hofsjökull in 2001 and a 115 m long core from Grímsvötn in 2002 (Steinthórsson, 1977; Thorsteinsson *et al.*, 2002; Thorsteinsson *et al.*, 2003). The density profiles from Hofsjökull and Grímsvötn were used to calculate the difference between the actual thickness of the firn layer and its ice equivalent thickness, to account for the lower density of the firn layer. For Grímsvötn, the reduction was 3.5 m but 8.3 m for Hofsjökull. The reason for this difference is the different rate of transformation from snow to ice, mainly because of the different elevation of the drill locations. The transformation is faster as the elevation is lower and the reduction is then smaller. The elevation of the top of Hofsjökull is 1800 m a.s.l. and 1450 m a.s.l. at Grímsvötn compared to ∼1550 m a.s.l. at the bottom of the Western Skaftá cauldron. Linear interpolation in elevation between the reduction for Hofsjökull and Grímsvötn was used to assess the reduction value of 4.9 m for the cauldron. This value was subtracted from the elevation of the ice shelf as the firn layer is all expected to be above the water table.

Before the jökulhlaup in September 2006, the elevation of the central hump in the cauldron was 1538 m a.s.l. but on Magnússon's map it is 1542 m a.s.l. Therefore, the cauldron is at a 4 m higher stage on the map than before the jökulhlaup in September 2006. To account for this difference, 4 m were subtracted from the map elevations of the ice shelf.

The western cauldron is bowl shaped except for the hump in the center. As the cauldron fills, this hump becomes less pronounced but it reforms during lowering of the ice shelf in jökulhlaups and during the initial rise of the ice shelf. It is most likely that this hump is an ice dynamic thrust phenomenon connected to ice flow into the cauldron. The hump does thus not represent the shape of the subglacial lake and was therefore flattened before the use of elevations from Magnússon's map for lake shape estimations. The whole area of the hump was given an elevation of the low around the hump which is at 1525 m a.s.l.

As a test of the assumption that the shelf is locally floating, it can be examined whether the ice shelf is in local floating balance at the location of the drill hole from June 2006. The elevation of the water level and ice shelf thickness are known at that location. Using Equation (5.1) to calculate the expected thickness of the ice shelf below the water surface results in ∼50 m smaller calculated ice thickness than measured. Because of the buoyancy of this extra ice, an upward pressure of approximately a half bar will act on the ice shelf. The ice shelf is therefore not in local floating balance at the location of the drill hole. This situation is probably similar for the whole ice shelf

and this force creates the shear stress at the ice shelf edge, that is needed to cause the strain associated with the lifting of the ice shelf.

Estimating the subglacial lake volume by mirroring was attempted although the discussion above indicates that it is not fully valid. Equation (5.3) was applied at each point in the grid of the surface from Magnússon's map, after the abovementioned adjustments. In June 2006, the water level was at an elevation of 1488 m a.s.l. The ice shelf rose 12 m before the jökulhlaup in September and as it is assumed to be floating the water level is expected to have undergone the same rise. The water level at the start of the jökulhlaup is therefore set equal to 1500 m a.s.l.

To form a subglacial lake from the resulting surface, the surface needs to be cut with a surface representing the bedrock. A map based on radio echo-sounding provided by the Institute of Earth Sciences at the University of Iceland shows that the center of the western cauldron is on top of a relatively flat ridge but the southern side lies above southwards facing slopes with an elevation drop of a few tens of meters. The map is only based on measurements done outside of the cauldron because of attenuation of the radio signal in the high conductivity geothermal water in the subglacial lake. This affects the accuracy of the map in the area of the cauldron. The average bedrock elevation under the cauldron can roughly be estimated as 1170 m a.s.l. A point measurement of bedrock elevation is available at the location of the June 2006 borehole, giving 1109 m a.s.l. whereas the bedrock map indicates an elevation of 1180 m a.s.l. at that location. This indicates considerable small-scale elevation variations under the cauldron that are not resolved on the bedrock map. It might be a weak assumption to approximate the bedrock as a flat surface at an elevation of 1170 m a.s.l. but it is used anyway due to lack of more detailed data.

The resulting lake resembles an upside-down elliptical bowl with a short axis of ∼150 m and a long axis of ∼300 m. The sides of the overlying ice shelf are convex and the depth of the lake is about 200 m at its deepest location. The total volume of the estimated subglacial lake is 87 Gl.

The form of the hypsometric curve for the lake is dependent on the inner strength of the overlying ice shelf. If the ice shelf has no strength, the grounding line will move inwards during outflow from the lake but the remnant part of the lake will keep its original shape. This is equivalent to excising the lowest part of the lake corresponding to the elevation drop and the hypsometry curve will be as for a reversed normal lake with the same shape. If the ice shelf has high strength and is able to carry a large part of its weight, it will deform more as a cantilever beam. The grounding line will then remain in its original location but the shape of the lake deforms during outflow. For this case, the hypsometry curve will be close to a straight line. The actual hypsometric

curve is probably somewhere in between these two extreme scenarios.

For the scenario of a lake underneath an ice shelf with no inner strength, the outflow volume during ice shelf lowering of 67 m, as in the September 2006 jökulhlaup, will be the volume of the lowest 67 m of the lake. This volume is 48 Gl which is somewhat too low compared with the volume of flood water at the glacier terminus of 53 Gl. For a lake underneath an ice shelf with more load bearing capacity, deforming as a cantilever beam gives an even lower volume for the same elevation drop.

Because of the uncertainty in the estimation of the bedrock shape, the lake shape estimated by this method did not give enough outflow volume and that the sides of the obtained ice shelf are convex counter to what is expected, a different method was also tried. In addition, there are clear indications, as mentioned above, that the ice shelf is not in floating balance. During the rapid lowering of the shelf in a jökulhlaup, the deviations from local floating balance are likely to be larger than for the slow lifting of the shelf between jökulhlaups.

Lake shape estimated by the form of an empty cauldron.   If the bottom under the cauldron can be assumed to be relatively smooth and flat compared with the surface changes of the ice shelf during a jökulhlaup and the lake assumes the form of a half dome extending upwards from the glacier bed into the ice, the cauldron will take the shape of an inverted lake when the lake is emptied.

As mentioned above, the central hump in the cauldron is most likely not connected to the shape of the lake and thus the map with the hill flattened out was also used in this analysis. As the GPS instrument measuring the lowering of the ice shelf during the jökulhlaup was on the central hump the elevation measurements were lowered according to the elevation difference between its location and the elevation used for that location after the hump was flattened out. This difference was 16 m.

The cauldron may be assumed close to full on Magnússon's map so the map needs to be adjusted to represent an empty cauldron as required by this method. To do so, a GPS profile measured shortly after the jökulhlaup in September 2006, was used to estimate the elevation difference between the state of the cauldron on Magnússon's map and the state of the cauldron after the 2006 jökulhlaup, as a function of elevation. For each 2.5 m elevation band in the cauldron the drop between the map and the GPS profile was calculated. The difference increases towards the cauldron center and resembles a parabola, therefore a quadratic function in elevation was fitted to the data.

The greatest lowering according to the fitted function is 32 m at the center but the largest difference between the elevation at the location of the permanent GPS instru-

ment taken from Magnússon's map (assuming a flattened hump) and the measurements from the permanent GPS instrument is 69 m. The reason for this difference is that the GPS profile was taken on the 26 November, two months after the jökulhlaup and the ice shelf had then risen considerably since the end of the jökulhlaup. To account for this difference, the quadratic function was scaled with a factor of 69/32.

The quadratic function is assumed to represent the elevation difference between the cauldron as it is on Magnússon's map and its stage after the September 2006 jökulhlaup. The shape of an empty cauldron is then obtained by applying the function to the map. The hypsometric curve may then be built for the subglacial lake from the shape of an empty cauldron as it is assumed to represent the inverted shape of the subglacial lake.

As for mirroring, the hypsometric curve depends on the inner strength of the ice shelf. The hypsometric curves for a lake underneath an ice shelf with no inner strength (the same as is obtained by filling the surface cauldron as a lake) and the hypsometric curves for a lake underneath an ice shelf with high strength will form an envelope for the true hypsometric curve. As the ice shelf overlaying the subglacial lake formed with this method is concave, the hypsometric curve for high strength will give more volume than the hypsometric curve for low strength for the same drop in lake level, counter to the case of mirroring.

Before the September 2006 jökulhlaup, the ice shelf in the cauldron center was at 1523 m a.s.l. Using the hypsometric curve for a subglacial lake underneath an ice shelf with low strength this level gives a volume of 32 Gl while the hypsometric curve for a subglacial lake underneath an ice shelf with high strength gives 78 Gl. The volume of jökulhlaup water at the glacier margin was 53 Gl which falls approximately midway between the two curves. Thus a weighted average of the two curves for ice shelf with low and high strength, giving the volume of 53 Gl for an ice shelf level at 1523 m a.s.l., was used to estimate the hypsometric curve for the subglacial lake (Figure 5.3). This hypsometric curve was used with the GPS measurements of lowering of the ice shelf to calculate the outflow from the subglacial lake during the jökulhlaup.

## 5.3 Temperature measurements

Temperature recorders were deployed in the subglacial lake in the western cauldron and in Skaftá at Fagrihvammur 3 km from the glacier snout in order to obtain temperature measurements from flood water at the beginning and the end of the subglacial flood path. The temperature sensor in the subglacial lake was built into the *Geokon*
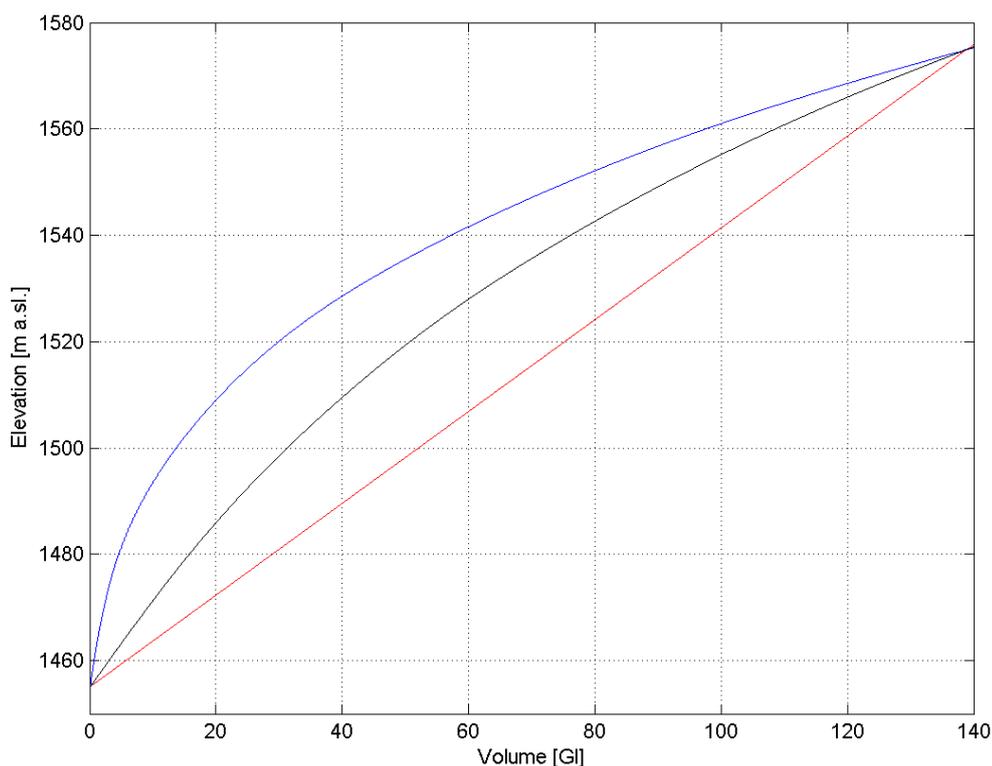
Figure 5.3: The hypsometric curve for the subglacial lake below the Western Skaftá cauldron as a function of ice shelf surface elevation (black curve). A hypsometric curve for a subglacial lake with same geometry but under an ice shelf with a low strength shown as a blue curve and for an ice shelf with a high strength is shown as a red curve.

4500SHSR vibrating wire pressure sensor and was connected to the same *Campbell CR-10X* data logger as the pressure sensor. The temperature sensor is a thermistor with an accuracy of $\pm 0.5^\circ$ C. A large part of this inaccuracy is zero-point shift that develops over time. Before deployment, the sensor was tested in an ice bath giving an offset of -0.33°C which was subtracted from the measurements in the lake.

The subglacial lake is stratified with a thin unstable layer at the bottom (Jóhannesson *et al.*, 2007) where the temperature varies in space and time because of influx of geothermal water through the lake bottom. As the recorder was sitting at the bottom it does not give a good indication of the overall lake temperature. The observed large fluctuations in the bottom temperature data (Figure 5.4) are therefore unlikely to be representative of changes of the lake temperature but show rapidly varying conditions due to local variations in geothermal inflow, possibly combined with time-dependent variations in the return flow of the bottom water mass of the lake (Jóhannesson *et al.*, 2007). As for the pressure sensor, the connection to the temperature sensor was lost on 16 September 2006, just eight days before the jökulhlaup.
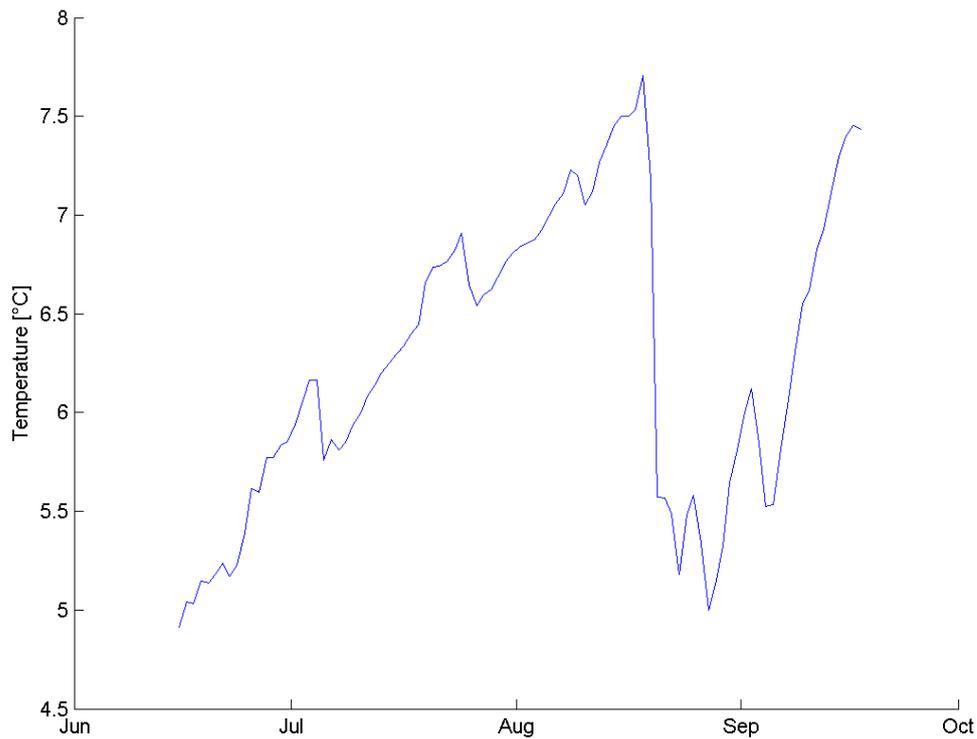
Figure 5.4: Temperature at the bottom of the subglacial lake under the Western Skaftá cauldron, during summer 2006.

A *Starmon-mini* temperature recorder from *Star-Oddi* was deployed in Skaftá at Fagrihvammur on a old mount for a pressure transducer that was operated there in the years 2002 to 2004. The accuracy of the recorder is $0.05°C$ for up to a year after calibration. Measurements in an ice bath before and after deployment indicated a zero offset of $-0.12°C$ which is larger than the quoted uncertainty of the recorder which had not been calibrated for several years. After subtraction of the zero-offset, the temperature measurements may be assumed to be accurate to within the quoted accuracy of the instrument. Temperature was recorded every fifteen minutes. The temperature data for the duration of the jökulhlaup in September 2006 are shown on Figure 5.5. Temperatures before the evening of 28th September are probably not floodwater temperatures and the sharp drop in temperature on the 28th represents the time when floodwater reached the sensor. The temperature measurements before the evening of the 28th are most likely not air temperature measurements as the daily variations are much smaller than variations in air temperature at a nearby weather station. The average is similar at the weather station and therefore these measurements are probably water temperature in stagnant water. The river is braided at the measurement site but constrained by rock walls on each side for high water levels as in jökulhlaups. In August 2006, when the

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site was visited the last time before the jökulhlaup, the river flowed in two channels, a small one at the side where the temperature recorder is located and another one much larger 10 m away. The smaller channel probably became cut off from the main river as discharge diminished in the autumn, forming a stagnant pool of water at the site of the recorder. When the discharge increased during the jökulhlaup, the channel became reconnected. Although this is not certain, it is certain that during the maximum discharge, the jökulhlaup water filled the whole area between the constraining rock walls and must therefore have submerged the sensor.
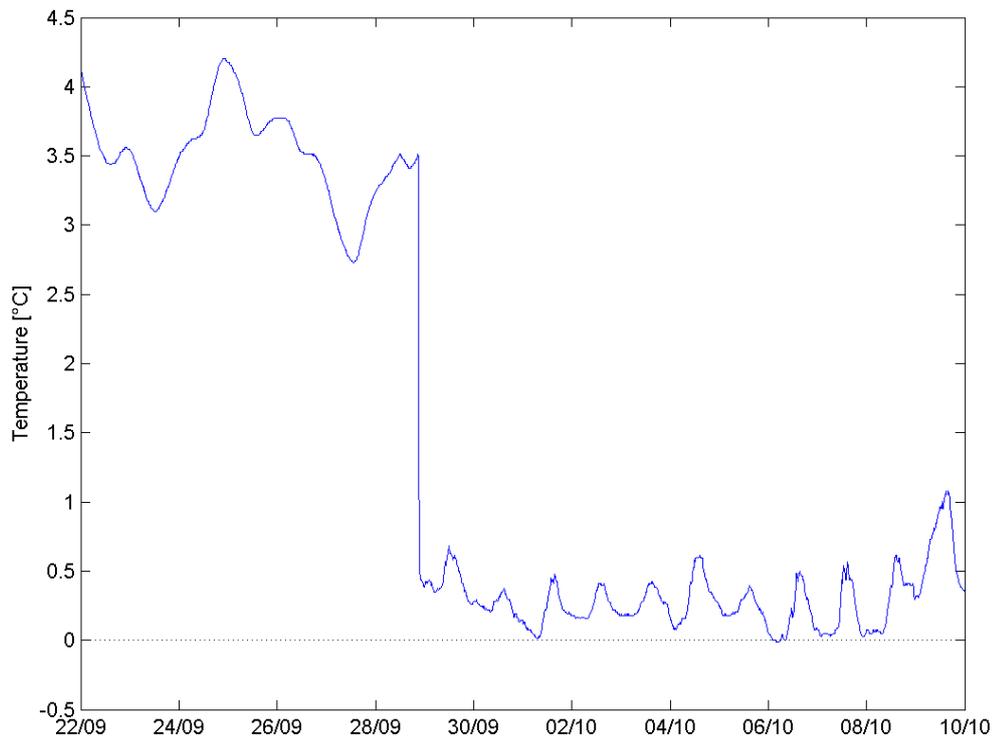
Figure 5.5: Temperature in river Skaftá at Fagrihvammur during the September 2006 jökulhlaup.

Chapter 6

Results and interpretation

6.1 Discharge in the jökulhlaup from the Western Skaftá cauldron in September 2006

The September 2006 jökulhlaup from the western cauldron was a fast-rising jökulhlaup, reaching discharge close to maximum in approximately two days. The jökulhlaup had the typical form of small jökulhlaups from the western cauldron with a relatively flat discharge maximum for approximately six days. Eventually, the jökulhlaup receded in a little less than four days.

The discharge increase during the first two days cannot be described by Equation (4.1) which implies that the discharge should increase by a factor of 2 in ∼3 days for discharge in the range 20–80 $m^3 s^{-1}$, using reasonable values for the flood path from the western cauldron and Manning's n' of 0.035 $m^{-1/3} s$. The discharge, however, increased from ∼20 $m^3 s^{-1}$ to ∼40 $m^3 s^{-1}$ in approximately half a day and from ∼40 $m^3 s^{-1}$ to ∼80 $m^3 s^{-1}$ in a little less than a day. The form of the discharge variation is also far from exponential (Figure 6.1). The traditional theory of slowly-rising jökulhlaups is thus not applicable for this jökulhlaup. This is similar to other jökulhlaups in Skaftá which have not been successfully described with the traditional theory (Björnsson, 1992; Sigurðsson and Einarsson, 2005).

The discharge of flood water at the terminus during the jökulhlaup, calculated from the measured discharge at Sveinstindur as described in Section 5.1, is shown on Figure 6.1, together with the discharge of flood water out of the cauldron, deduced with the hypsometric curve and GPS measurements of ice shelf elevation described in Section 5.2.

The maximum discharge at the glacier terminus is 97 $m^3 s^{-1}$, in the afternoon on
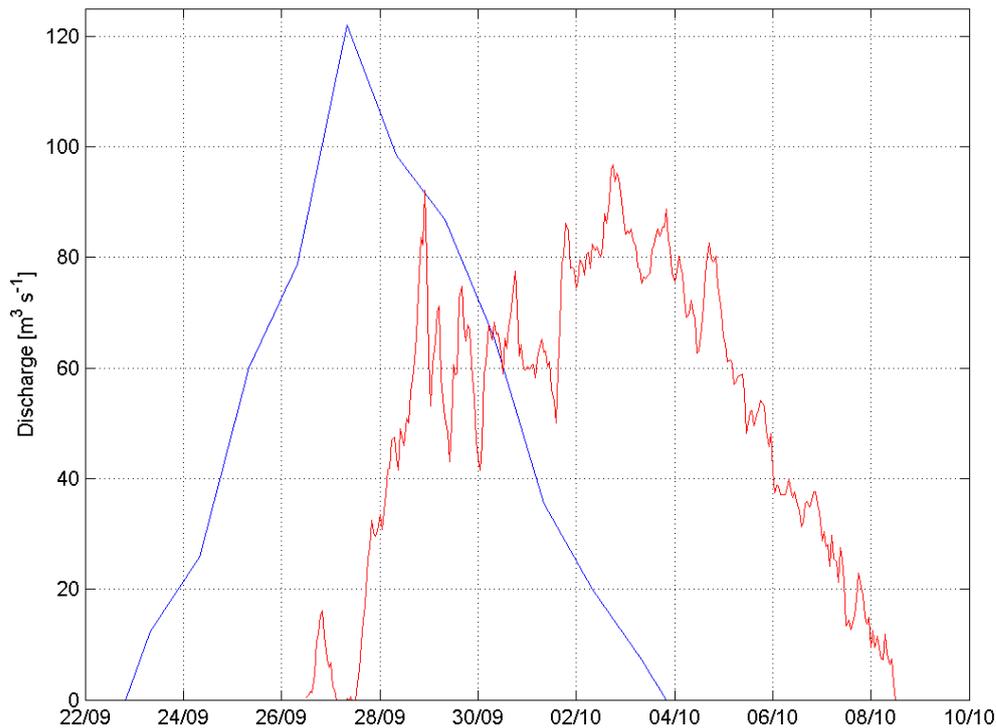
Figure 6.1: Discharge of jökulhlaup water at the glacier terminus, calculated by backtracking (Jónsson, 2007) shown as red curve, and discharge out of the subglacial lake calculated from the subsidence of the cauldron, shown as blue curve, during the September 2006 jökulhlaup.

2 October. As the estimation of a peak in the base flow during the first six days of the jökulhlaup is uncertain, the true discharge maximum might have occurred in the interval 29 September to 2 October but the maximum value of $\sim 100\,\mathrm{m^3\,s^{-1}}$ is fairly accurate as the discharge peak is broad and flat.

The discharge out of the subglacial lake rises to a maximum of $123\,\mathrm{m^3\,s^{-1}}$ in approximately four days. The maximum outflow occurred on 27 September, before any considerable outflow had started at the glacier terminus. Outflow from the subglacial lake therefore does not seem to require an effective transportation of water from the lake to the terminus but other processes such as lifting of the glacier and a creation of subglacial storage are more important. The discharge out of the subglacial lake then recedes a little slower than the rise in approximately six days.

A comparison of the timing of the flood front at the glacier terminus and the initiation of subsidence within the cauldron gives the travel time of the flood front under the ice cap. As the GPS recorder in the cauldron only records elevations once per day, the timing of the start of the subsidence has an uncertainty of 24 hours. The start of the subsidence is between 8:00 on the 24 and 8:00 on the 25 September. Then there is also

uncertainty in the exact timing of the start of the jökulhlaup at the glacier terminus, mainly because the daily discharge variation masks the initial discharge increase. The jökulhlaup at the glacier terminus can be assumed to start between 13:00 and 22:00 on 26 September and thus the travel time of the subglacial flood wave from the cauldron to the terminus was 29 to 60 hours. The travel distance of the jökulhlaup is ~40 km and thus the mean travel speed of the front of the subglacial flood wave was in the range 0.2–0.4 m s^{-1}.

The travel speed of the flood front under the glacier is slow compared with the travel speed of subglacial water under normal conditions and flow velocity in open channels. The speed is also slower than for the fast-rising jökulhlaup from Grímsvötn in November 1996, which traveled 50 km in 10.5 hours (Flowers *et al.*, 2004), corresponding to an average propagation speed of 1.3 m s^{-1}. The propagation speed of the subglacial flood front for fast-rising jökulhlaups therefore seems to vary substantially between different events.

A slow propagation of the flood front either indicates the buildup of subglacial storage for considerable time or water traveling in a slow subglacial hydrological system during the initial phase of the jökulhlaup.

There are indications that the travel speed of the water increases at later stages for jökulhlaups in Skaftá as exemplified by observations during the 2002 jökulhlaup from the Eastern Skaftá cauldron after the discharge had peaked. Earthquake tremors, indicating boiling or a minor volcanic eruption at the cauldron because of the pressure release accompanying the jökulhlaup, and the concentration of suspended material in the jökulhlaup waters collected at the gauging station at Sveinstindur, which display a peak believed to result from the same boiling/eruption event, were used to estimate a subglacial flow speed of approximately 0.8 m s^{-1} (Oddur Sigurðsson, personal communication, 2008).

6.2 Volume of flood water in the September 2006 jökulhlaup

The total volume of flood water in the jökulhlaup was estimated as 53 Gl, making this one of the smaller jökulhlaups from the western cauldron. Base flow due to other sources than the jökulhlaup is considerable compared with the discharge of flood water at the glacier margin. The uncertainty in the estimation of the base flow leads to rather large uncertainty in the estimation of the discharge and volume of the jökulhlaup, in addition to measurement inaccuracy.

The quality of the base flow estimate depends largely on how uniform weather conditions are during the jökulhlaup. Base flow estimates for periods without precipitation or melt events are simple, while it is complicated to estimate the effects of weather events such as the temperature peak that occurred during the September 2006 jökulhlaup, which increases the uncertainty in the base flow estimate substantially. The estimated volume of 53 Gl has therefore considerable uncertainty and values in the range 43–63 Gl could be obtained with different base flow estimates. Values outside this range are, however, considered unlikely.

The time since the last jökulhlaup from the western cauldron, in August 2005, was unusually short or only 13 months compared with the average time between jökulhlaups from the western cauldron of 24 months. The flood volume corresponds to a relatively low average water accumulation of ∼4 Gl per month, compared with the decadal average of 6 Gl per month. It is possible that a part of the accumulated water was not released from the subglacial lake, which might explain how small the jökulhlaup was, in addition to the short accumulation time. The measurements of water depth and ice shelf lowering at the location of the June 2006 borehole also indicate that the lake was not fully emptied. The lake depth was ∼125 m just before the jökulhlaup while the lowering of the ice shelf during the jökulhlaup was only 67 m, leaving ∼60 m of water in the lake at the location of the borehole. Considerable deformation of the ice shelf was observed after the jökulhlaup with relative elevation changes of the shelf of up to tens of meters so water might not have been left under the whole ice shelf, but at least at the location of the borehole some water was left in the lake. A third indication that some water was left in the lake can be inferred from ice shelf surface elevation measurements in the cauldron's center, by the Institute of Earth Sciences at the University of Iceland as the lowering of the lowest part of the ice shelf in the September 2006 jökulhlaup was probably at least some tens of meters less than during the August 2005 jökulhlaup.

It is hard to estimate the volume of the water left in the subglacial lake. The true shape of the lake and the total volume for different ice shelf elevations is not known. The difference in the volume of the August 2005 jökulhlaup and the September 2006 jökulhlaup can not be used as the ice shelf elevation and the available water volume at the beginning of the jökulhlaups differs. A crude estimate can be made by comparing the rate of water accumulation deduced from the volume of the September 2006 jökulhlaup with the average rate of water accumulation in the lake, assuming that the subglacial lake emptied in the large August 2005 jökulhlaup. The difference is 2 Gl per month, summing up to a volume of 26 Gl over the 13 months before the September 2006 flood.

Although the water volume left in the subglacial lake after the September 2006 jökulhlaup is hard to estimate, these results indicate that jökulhlaups from the Western Skaftá cauldron can terminate before the lake empties. Termination of jökulhlaups before emptying of the source lake is known for other locations, for example Grímsvötn (Björnsson, 1974). This may not be the case for every jökulhlaup from the Western Skaftá cauldron, as the final stage of the ice shelf after jökulhlaups varies (data of the Institute of Earth Sciences at the University of Iceland, 2008).

6.3   Lowering of the ice shelf and hypsometric curves

Although the different methodologies used to deduce a hypsometric curve for the subglacial lake below the Western Skaftá cauldron gave different volumes for the measured lowering during the jökulhlaup and that they are both subject to some limitations, the resulting outflow curves are similar for both methodologies. The absolute volume of the outflow, deduced from the hypsometric curve found by mirroring the ice shelf surface with regard to the water level assuming floating balance, is very sensitive to the bedrock shape and elevation which are quite uncertain. A volume of 53 Gl, comparable to the volume of the jökulhlaup at the terminus, could be deduced for the measured lowering by small changes in the estimated subglacial bedrock geometry. The shape of the subglacial lake is likely to be affected by the deformation of the ice shelf which is in a state of horizontal compression and experiences large elevation changes every two years on average. Shear stresses are built up along near-vertical planes near the edge of the ice shelf because of the shear deformation associated with the lifting of the shelf. These shear stresses would cause the ice shelf to be out of floating balance. It is therefore likely that the fundamental assumption for this methodology, that the ice shelf is in local floating balance, is not accurate.

The hypsometric curve used here is deduced from an assumed form of the empty cauldron, relying on a relatively smooth and flat bedrock bottom. Peaks and irregularities in the bedrock topography would change the shape of the ice shelf after a jökulhlaup, which then would not represent the form of the subglacial lake. Internal deformation of the ice shelf during the lowering in a jökulhlaup would also disturb this methodology. The geometry of the ice shelf after the jökulhlaup and the fact that the ice shelf did not hit the bottom at the location of the central hump where the borehole was located, indicates that the assumptions behind this methodology are more realistic than an assumption of local floating balance.

According to the data presented in Section 5.2, the hypsometric curve for the lake

lies approximately midway between the hypsometric curves for a subglacial lake below an ice shelf with a high and a low strength. This hypsometric curve is not unreasonable in view of simple calculations of shear stress arising from the subsidence of the ice shelf. The ice shelf is approximated as a cylindrically symmetric plate with a thickness of 300 m where the subsidence of an undisturbed central area with a radius of ~300 m is obtained by simple shear deformation over an external cylinder shell with a width of ~300 m and centered at ~450 m from the center. These calculations omits the affects of brittle deforming in the uppermost layer of the ice but large crevasses surrounding the cauldron indicates that part of the ice thickness dose not deform in ductile manner as assumed here for simplification. The subsidence during jökulhlaup is ~100 m in approximately 5 days resulting in shear strain rate of ~$4 \cdot 10^{-7}$ s^{-1}. Using Glen's flow law for simple shear: $\dot{\varepsilon}_{rz} = A\tau_{rz}^3$, where $\dot{\varepsilon}_{rz}$ is the shear strain rate, τ_{rz} is the shear stress and $A = 6.8 \cdot 10^{-15}$ s^{-1} (kPa)$^{-3}$ is the flow law parameter for temperate ice, an average shear stress of ~4 bars can be calculated. To balance this shear stress, a vertical force corresponding to a stress of ~5 bars needs to be acting on the ice shelf, arising from a lower water pressure acting on the bottom of the shelf than needed for floating balance. This result is of the same order of magnitude as Guðmundsson *et al.* (2004) obtained for rapid subsidence at the ice cauldron formed in the subglacial eruption at Gjálp in 1996. Due to the shear deformation the grounding line will move inwards as expected for an ice shelf with a low strength but the ice shelf has some strength and considerable pressure drop in the subglacial lake is needed to maintain the shear deformation.

During the rising phase between jökulhlaups, the ice shelf rises ~100 m in 24 months and calculations based on the same approximations about the geometry as above results in a shear strain rate of ~0.1 year^{-1}. The corresponding shear stress calculated with Glen's flow law is ~0.5 bar. To create this shear stress, a force corresponding to a pressure of ~1 bar, created by ice buoyancy, needs to be acting on the ice shelf. This is consistent with the discussion on whether the ice shelf was in floating balance at the location of the June 2006 borehole, presented in Section 5.2, which concluded that a pressure of approximately half a bar would result from the buoyancy of the ice thickness below the water table in excess of thickness corresponding to floating balance.

As mentioned before, the stage of the ice shelf of the western cauldron at the beginning and at the end of jökulhlaups is variable (data of the Institute of Earth Sciences at the University of Iceland, 2008) and some water was probably left in the subglacial lake after the jökulhlaup in September 2006 but how much and whether this happens in every jökulhlaup remains unclear. The volume of jökulhlaups from the western cauldron spans a considerable range and may be assumed to depend on both the available

water in the subglacial lake at initiation and on the volume of water left in the lake after the flood.

The initiation of the flood will depend on how the subglacial seal is breached, which is not known, although flotation is most likely. If the process is flotation, differences in thickness of the ice dam, caused by glacier variations, surges and other factors, would lead to differences in outburst volumes. There are indications of jökulhlaup initiation at lower ice shelf elevations after the 1994–1995 surges in Tungnaárjökull and Sylgjujökull. Surges cause a lowering of the glacier surface south of the cauldrons and thereby thinner ice dams (Guðmundsson and Högnadóttir, 2002).

A few days before the September 2006 jökulhlaup started, elevation fluctuations up to 1 m were measured on the ice shelf. These fluctuations might be measurement errors but they are larger than observed fluctuations of the GPS time-series in other periods. These fluctuations might be caused by the lifting of the subglacial seal, indicating that it had a prelude of some days.

6.4   Subglacial storage during the jökulhlaup

Figure 6.2 shows the cumulative volume of outflow at the glacier margin and the volume in the subglacial lake during the September 2006 jökulhlaup. The storage in the subglacial pathway is also shown, determined by subtracting the cumulative volume of outflow at the glacier margin from the cumulative volume of outflow from the subglacial lake. The subglacial storage is considerable and reaches a maximum of approximately 35 Gl on 30 September, three days after the outflow from the cauldron reached maximum. The subglacial storage is up to 20 Gl before any outflow starts at the glacier margin. These results indicate considerable subglacial storage compared with the total volume of flood water of 53 Gl.

At the timing of maximum subglacial storage, 45.5 Gl had been released from the subglacial lake. The initial heat in this volume of lake water with a temperature of $\sim$4.5 °C and heat formed by potential energy dissipation in the flow down the subglacial flood path are sufficient to melt $\sim$3.5$\cdot$10^9 kg of ice which corresponds to $\sim$3.5 Gl of meltwater. This is $\sim$10% of the maximum storage in the subglacial pathway. The ratio between the maximum possible melted volume and the volume of subglacial storage of 20 Gl that were accumulated under the glacier before any outflow started is similar. As most of the initial heat in the lake water is expected to be dissipated during the first few kilometers of the flood path, as discussed in Section 7.2, the potential energy released in flow is probably the only heat source contributing to
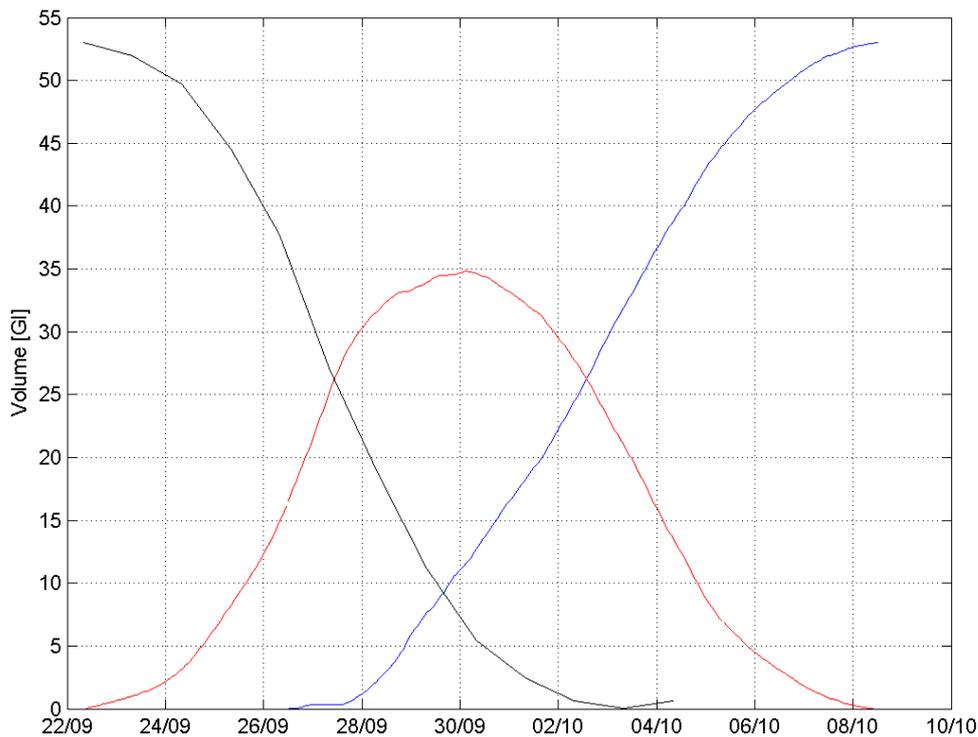
Figure 6.2: Volume of floodwater during the September 2006 jökulhlaup. Water volume in the subglacial lake is shown as a black curve, cumulative volume at the glacier terminus as a blue curve and volume stored subglacially as a red curve.

melting for most of the flood path. For the total volume of floodwater, the potential energy released can melt $\sim$1.4 Gl and for the 20 Gl accumulated subglacially before outflow at the terminus started, it can only melt $\sim$0.5 Gl or 2.5% of the total volume needed for storage. Other processes, such as elastic ice lifting and deformation induced by water pressure higher than overburden pressure, must therefore be important in the propagation of the jökulhlaup and the forming of the subglacial pathway. This was also the case for the fast-rising jökulhlaup from Grímsvötn in November 1996 (Jóhannesson, 2002). Considerable temporal subglacial storage, connected with glacial lifting has also been documented for a jökulhlaup from Gornersee, Switzerland (Huss *et al.*, 2007).

At the beginning of the jökulhlaup and at the time of maximum subglacial storage, the volume of the subglacial flood pathway is much larger than needed to carry a similar amount of discharge during later stages of the jökulhlaup. For example, $92 \, \mathrm{m^3 \, s^{-1}}$ are transported in the subglacial flood path for a volume of 33 Gl on 29 September while $82 \, \mathrm{m^3 \, s^{-1}}$ are transported in the subglacial flood path for a volume of 11 Gl on 4 October. This indicates storage in subglacial reservoirs that do not play a considerable

part in transporting discharge during early stages of the jökulhlaup and/or an evolution of the flood path from a low efficiency/high flow resistance initially into a higher efficiency later on.

One possible physical explanation of evolution of the flood path from a low to a higher efficiency in discharge transportation, which is consistent with indications seen in other jökulhlaups in Skaftá, is an evolution from a wide sheet-like flood path in the beginning of jökulhlaups to a narrower more confined flood path during later stages. Indications of a wide flood path with high water pressure, such as flow out of moulins close to the glacier margin, water surfacing with high pressure and ice fracturing through crevasses in the lower part of the ablation area and other phenomena have been noted for other jökulhlaups in Skaftá than the September 2006 jökulhlaup. For these same jökulhlaups, the flow up through crevasses and moulins stops as the jökulhlaup evolves, indicating a drop in subglacial water pressure. The floods have also been observed to flow out of a few tunnels at the glacier margin during maximum discharge and at later stages (Oddur Sigurðsson, personal communication, 2008).

6.5 Temperature of jökulhlaup water near the glacier margin and in the subglacial lake

Measurements of water temperature in Skaftá at Fagrihvammur, 3 km downstream from the glacier margin, during the September 2006 jökulhlaup have a small diurnal variation between $0.0\,°C$ and $0.5\,°C$ (Figure 5.5). The air temperature recorded at Sveinstindur varied between $1\,°C$ and $10\,°C$ during the jökulhlaup, but was between $2\,°C$ and $6\,°C$ most of the time. Hence, some atmospheric warming of the outburst water may be assumed to have taken place along the 3 km long distance between glacier terminus and the measuring point. The elevation drop between the glacier terminus and the measuring point is $\sim 50\,m$ and heat formed by potential energy dissipation on this way can therefore be expected to cause further warming of $\sim 0.1\,°C$.

In the morning of 1 October and during the nights of 6 and 7 October, the water temperature is close to zero and the air temperature at Sveinstindur is also close to zero at the same time, so the warming was probably small. Taken together, these measurement results indicate that the outburst water is at – or very close to – the freezing point as it emerges from beneath the ice cap. Measurements of flood water temperature at the glacier margin in other jökulhlaups in Skaftá show, in accordance with these results temperatures very close to freezing point (data from the Skaftá cauldrons research project).

Because of the shortcomings of the temperature measurements at the bottom of the subglacial lake discussed in Section 5.3, the average temperature of the lake water at the initiation of the jökulhlaup cannot be determined from them. Large-scale changes from the vertical temperature profile measured in June 2006 are however unlikely, taking into consideration the nearly constant water temperatures at various depths in the lake in the eastern cauldron while temperature changes of several degrees were measured in a thin layer at the lake bottom (data from the Skaftá cauldrons research project). The measured temperature variations at the bottom of the Western Skaftá cauldron lake (Figure 5.4) are thus interpreted as influences of geothermal activity confined to a thin layer at the lake bottom and the average temperature of the lake is assumed to be close to 4.5 °C in September 2008 as measured in June 2006.

Chapter 7

Simulations of the 2006 jökulhlaup from the Western Skaftá cauldron

Flowers *et al.* (2004) presented a coupled sheet–conduit model for the extreme November 1996 jökulhlaup from Grímsvötn. By connecting subglacial flow in a sheet to flow in conduits they managed to capture the fast rise of the jökulhlaup. The September 2006 jökulhlaup from the western cauldron was simulated with this same model in order to see whether it explains a fast-rising jökulhlaup, two orders of magnitude smaller in discharge and volume than the November 1996 jökulhlaup from Grímsvötn.

7.1   Model physics

The model is "a one-dimensional flowline model that accounts for water transport in a sheet-like subglacial layer and in ice-walled conduits" (Flowers *et al.*, 2004). Reduced Navier-Stokes equations are used to describe the flow in the sheet and for the flow in the conduits the approach of Spring and Hutter (1981, 1982) is followed, except for the thermodynamics where Nye's simplification of an instantaneous heat transfer is used, see the discussion about conduit flow in jökulhlaups in Section 4.6. Nye's simplification implies that the water temperature is always at the local pressure melting point and not an independent variable.

Another version of the model exists where the initial heat of the water in the subglacial lake is accounted for and heat transfer is described according to the formulation of Spring and Hutter (1981, 1982) but as mentioned in Section 4.6, Nye's assumption seems to be closer to reality for jökulhlaups than the more physically based formulation based on empirical expressions for heat transfer in pipe flow used by Spring and Hutter (1981, 1982) and introduced by Nye (1976). Therefore, the model version

assuming instantaneous heat transfer was used.

As the model does not capture the jökulhlaup initiation, the inflow from the lake into the subglacial hydraulic system is prescribed as a boundary condition. It is assumed that the water flows into the sheet part of the model but the sheet is described as a macroporous layer with thickness h^s, hydraulic conductivity K_s and porosity m. Water is assumed incompressible and the water balance in the subglacial water sheet is described by

$$\frac{\partial h^s}{\partial t} = -\nabla \cdot Q^s - \phi^{s:c}, \tag{7.1}$$

where Q^s is water flux in the sheet and $\phi^{s:c}$ describes water exchange between the sheet and a system of conduits. The flux in the sheet is described by

$$Q^s = -\frac{2K_s h^s \nabla \psi^s}{\rho_w g} \left(1 + (1 + C|\nabla \psi^s|)^{1/2}\right)^{-1}, \tag{7.2}$$

where ρ_w is water density and g is the acceleration of gravity. ψ^s is the fluid potential in the sheet given by $\psi^s = p^s + \rho_w g z_L$ where p^s is the water pressure in the sheet and z_L is the flood path elevation taken to be the elevation of the glacier bed. C is a parameter given by $C = \left(2880 K_s^3 (1-m)^2 / \mathrm{Re}^2 \mu m^3 \rho_w g^3\right)^{1/2}$, where $\mathrm{Re} = Q^s \rho_w / \mu$ is Reynolds number and μ is the viscosity of the water flowing in the sheet (Stone and Clarke, 1993). Basal water pressure p^s is described as an empirical function of h^s,

$$p^s(h^s) = \rho_i g h_i (h^s/h_*)^{7/2} \tag{7.3}$$

where ρ_i is ice density, h_i is the ice thickness and h_* is a critical water sheet thickness. Hydraulic uplift, caused by pressure higher than the ice overburden pressure, is described as an increase in the critical water sheet thickness. The increase is formulated by a modified Gaussian function centered at the origin of the uplift (Flowers and Clarke, 2000). Uplift at location x', originated at location x is

$$\Delta h_*(x') = \alpha h(x) \frac{p^s(x)}{\rho_i g h_i(x)} \exp\left(-\frac{(x'-x)^2}{2h_i(x)^2}\right), \tag{7.4}$$

where α is a dimensionless scaling factor that is empirically determined. When uplift originated from overpressure at different location along the flood path acts on the same location, uplift is determined as the maximum uplift acting at the location in question.

The conduits are described as a system of channels that are nourished from the sheet. They are assumed to be separated by a characteristic spacing d_c. The mass balance of the system and flow between the sheet and the conduits is described for multiple conduits but the evolution of discharge and cross-sectional area is described for a single conduit, this involves an assumption of identical behavior between conduits.

For a conduit that is not overpressurized, the evolution of the cross-section area with time and conservation of ice mass is described by

$$\frac{\partial S}{\partial t} = -\frac{Q^c}{\rho_i L}\left(\nabla\psi^c - D\nabla p^c\right) - 2S\left(\frac{p_i - p^c}{nB}\right)^n,$$ (7.5)

where Q^c is the discharge in the conduits, S is the cross-sectional area of any individual conduit, p^c is the pressure in the conduits, L is the latent heat of fusion for water, ψ^c is the fluid potential in the conduits given by $\psi^c = p^c + \rho_w g z_L$, n is the Glen ice rheology flow-law exponent, B is the Glen ice rheology flow-law rate factor and D is a constant given by $D = c_t \rho_w c_w$ where c_t is the change in melting point temperature with pressure and c_w is the heat capacity of water. By toggling a switch in the model, overpressurized conduits can be allowed to grow according to the ice deformation term in Equation (7.5) but this is not the default. The discharge in the conduit is given by

$$Q^c = -\left(\frac{8S^3}{P_w \rho_w f_R}\right)^{1/2}\frac{\nabla\psi^c}{|\nabla\psi^c|^{1/2}},$$ (7.6)

where P_w is the wetted perimeter and f_R is the Darcy-Weisbach roughness given by $f_R = 8gn'^2/R_H^{1/3}$ where n' is the Manning roughness and R_H is the hydraulic radius of the conduit. R_H and P_W are functions of cross-sectional area and conduit shape which can be assumed semi-circular or circular. If conduits are assumed semi-circular and located at the bed, the Manning roughness n' is a composite of the value for ice and the bed.

For the conduit system, the water mass balance is described by

$$\frac{\partial p^c}{\partial t} = -\frac{1}{\beta S}\left(\frac{\partial S}{\partial t} + \nabla\cdot Q^c + \frac{Q^c}{\rho_w L}\left(\nabla\psi^c - D\nabla p^c\right) - d_c \phi^{s:c}\right),$$ (7.7)

where β is a numerical compressibility parameter introduced by Clarke (2003) to resolve the numerical stiffness in the formulation for conduit flow by Spring and Hutter (1981, 1982). Exchange of water between the conduit system and the sheet is given by

$$\phi^{s:c} = \chi^{s:c}\frac{K_s h^{s:c}}{\rho_w g d_c^2}\left(p^s - p^c\right),$$ (7.8)

where $\chi^{s:c}$ is a parameter between 0 and 1 controlling the strength of the coupling between the sheet and the conduit system. At the conduit–sheet interface where exchange occurs, sheet thickness is given by $h^{s:c}$. This thickness is calculated from the average pressure in the sheet and the conduits $p^{s:c} = (p^s + p^c)/2$ by inverting the equation for sheet pressure as a function of sheet thickness, $h^{s:c} = h_*\left(p^{s:c}/(\rho_i g h_i)\right)^{2/7}$.

The model is solved by numerically integrating Equations (7.1), (7.5) and (7.7), giving water sheet thickness h^s, conduit area S and conduit water pressure p^c as a function of position along the flood path and time (Flowers *et al.*, 2004).

This model enables a jökulhlaup to rise much faster than the theory of a slowly-rising jökulhlaup because of the sheet flow. The sheet switches between acting as source or sink for the flow in the conduits depending on the pressure difference between the conduits and the sheet (Flowers *et al.*, 2004). This facilitates conduit growth along the flood path because the conduits get fed from the sheet where conduit water pressures are lower than sheet pressures but when the conduits become overpressurized relative to the sheet water flows back into the sheet and is thus able to move downstream even in the absence of well developed conduits.

7.2   Input data

Discharge out of the subglacial lake and profiles of glacier surface and glacier bottom for the subglacial flood path are needed as an input for the model along with values for the various physical parameters.

The discharge out of the lake used is the same as presented in Chapter 6, resampled onto an hourly time step (Figure 6.1). Profiles of the glacier surface and glacier bottom for the subglacial flood path were provided by the Institute of Earth Sciences at the University of Iceland and are shown on Figure 3.7. They were resampled onto a grid with 60 evenly spaced nodes that was used for calculations.

A few hundred meters wide and some meters deep depression stretches southwestward from the western cauldron, following the estimated flood path. After jökulhlaups, the depression has been observed to be bounded by thin belts of small crevasses on each side (Oddur Sigurðsson, personal communication, 2008). Bearing in mind that the heat transfer in subglacial water flow is very rapid and comparing the volume of the depression, which is few Gl, to the volume of ice melted by the heat in 53 Gl of 4.5°C warm water, which is ∼3 Gl, it seems likely that the depression is caused by melting caused by the initial heat in the jökulhlaup flood water. It is therefore assumed that all of the initial heat in the lake water is expelled within the first three km from the cauldron and model calculations are not started until at that location. Outflow from the cauldron is assumed to be unaltered during these three km and the outflow from the cauldron is used to define discharge at the starting point of the calculations. The flood water is assumed to be at the pressure melting point at this starting point.

The width of the subglacial water sheet and a characteristic conduit spacing are needed as input parameters. Small jökulhlaups, like the one modelled here, emerge only in one location at the glacier snout indicating flow in only one conduit. As the model is one-dimensional and multiple conduits are only represented by using the

Table 7.1: Physical parameters, independent of local condition and model calibration, used in jökulhlaup simulation. Taken after Flowers *et al.* (2004)

Parameter	Symbol	Value
Acceleration of gravity	g	$9.81\,\mathrm{m\,s^{-2}}$
Density of ice	ρ_i	$910\,\mathrm{kg\,m^{-3}}$
Density of water	ρ_w	$1000\,\mathrm{kg\,m^{-3}}$
Viscosity of water	μ	$1.787\times10^{-3}\,\mathrm{Pa\,s}$
Latent heat of fusion	L	$3.34\times10^5\,\mathrm{J\,kg^{-1}}$
Glen ice rheology flow-law exponent	n	3
Glen ice rheology flow-law rate factor	B	$5.8\times10^7\,\mathrm{N\,s^{1/n}\,m^{-2}}$
Change in melting point temperature of water with pressure	c_t	$7.5\times10^{-8}\,\mathrm{K\,m^3\,J^{-1}}$
Heat capacity of water	c_w	$4.22\times10^3\,\mathrm{J\,kg^{-1}\,K^{-1}}$
Numerical compressibility parameter	β	$10^{-9}\,\mathrm{Pa^{-1}}$
Sheet bulk porosity	m	0.35

variable d_c in Equations (7.7) and (7.1) a single conduit can be represented by setting conduit spacing equal to the width of the sheet, indicating one conduit for that whole area. The width of the sheet is then assumed to be 1 km based on the width of the abovementioned depression; this is though a quite uncertain parameter.

Other physical parameters, independent of local conditions and model calibration, are taken after Flowers *et al.*(2004) and are shown in Table 7.1.

7.3 Results

The jökulhlaup discharge at the glacier margin (Figure 5.1), deduced from the measured discharge at Sveinstindur as described in Section 5.1, was used for comparison and evaluation of the model performance. Six model parameters were calibrated to fit the modelled discharge to the measured discharge: the hydraulic conductivity of the sheet, K_s, the initial critical thickness of the sheet, $h_*(0,x)$, the Manning's roughness of the conduits, n', the cross-sectional area of vestigial conduits, $S(0,x)$, the scaling factor for hydraulic uplift, α, and the strength of the coupling between the sheet and the conduit system, $\chi^{s;c}$. A good fit for discharge can be obtained for the parameter set given in Table 7.2, see Figure 7.1, left.

All the parameters are within a reasonable range when compared with the parameters used by Flowers *et al.* (2004), although the Manning's roughness of $0.0645\,\mathrm{m^{-1/3}\,s}$ is in the higher end of the expected range. The conduits are here assumed to be semi-

Table 7.2: Fitted parameters for best fit in discharge simulation for the September 2006 jökulhlaup.

Parameter	Symbol	Value
Hydraulic conductivity of the sheet	K_s	$0.175\,\mathrm{m\,s^{-1}}$
Initial critical thickness of the sheet	$h_*(0,x)$	$0.6\,\mathrm{m}$
Manning's roughness of the conduits	n'	$0.0645\,\mathrm{m^{-1/3}\,s}$
Cross-sectional area of vestigial conduits	$S(0,x)$	$0.00375\,\mathrm{m^2}$
Scaling factor for hydraulic uplift	α	0.01
Conduit-sheet coupling strength	$\chi^{s:c}$	0.045

circular. However, a rather high Manning's roughness will be obtained in Equation (7.6) by that assumption if the conduits are wider and flatter as can in fact be expected. Lower true hydraulic radius and higher true wetted perimeter will then be compensated by a higher Manning's roughness, which is the only parameter of these three that is calibrated. For a flat and wide conduit with half the hydraulic radius of a semicircular conduit with the same cross sectional area, a Manning's roughness of $0.04\,\mathrm{m^{-1/3}\,s}$ would lead to an effective Manning's roughness of $0.0645\,\mathrm{m^{-1/3}\,s}$ if a semicircular conduit is assumed and the Manning's n' is modified to obtain the same discharge.

The simulated subglacial pressure field is an order of magnitude higher than the ice overburden, which must be considered impossible for a subglacial flood path with a width on the order of a kilometer (Figure 7.1, right). A plausible pressure field could not be obtained despite tests with other parameterizations. Prescribing inflow into the subglacial hydrological system as a boundary condition at the inlet of the flood path forces water into the system regardless of the actual lake water pressure and a high water pressure is built up as the model does not contain a physical description of the elastic or ice flow response to water pressure in excess of ice overburden.

As the subglacial water pressure is one of the main driving forces in the subglacial hydraulic system, unrealistic effects in many hydraulic processes will be caused by this unrealistically high water pressure. For example, the fluid potential in the sheet and in the conduits will be dominated by the water pressure and effects of differences in overburden pressure and flood path gradients will become negligible in contrast to what would be expected. Unrealistically high water pressures also lead to an unrealistically high fluid potential gradient. Potential energy, that is actually nonexistent in the system, is dissipated by an unrealistically high friction leading to an overestimation of heat formation and thereby an overestimation of melting in the conduits. The model can therefore not capture the true hydraulic processes in the jökulhlaup due to the un-
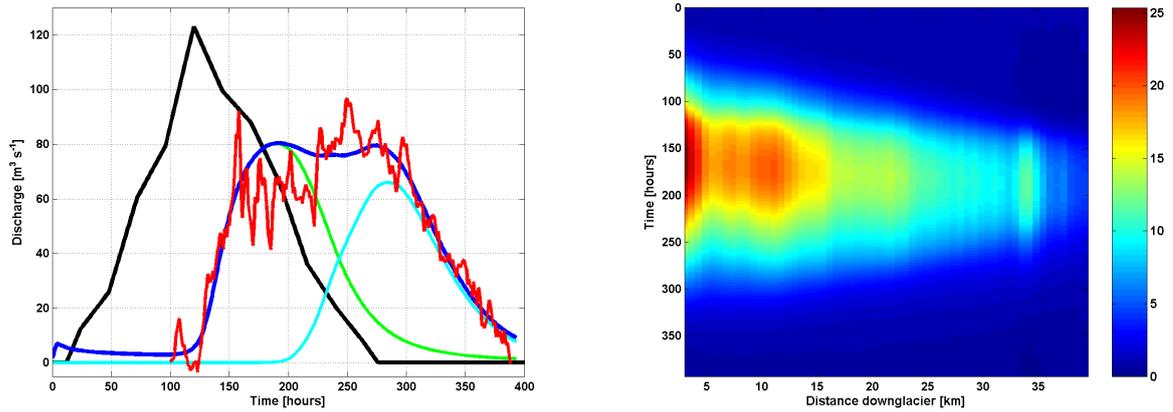
Figure 7.1: Left: Modelled and measured discharge at the glacier terminus. Discharge out of the subglacial lake is shown as black curve, back-calculated discharge at the glacier terminus is shown as red curve, simulated discharge in a sheet at the glacier terminus shown as green curve, simulated discharge in conduits at the glacier terminus shown as light blue curve and total discharge at the glacier terminus, which is the summation of the discharge in the conduits and the sheet, is shown as blue curve. Right: Modelled time evolution of the pressure field in the sheet shown as ratio of ice overburden pressure.

realistic pressure field although a good fit of the discharge variation was obtained.

As described above, the water pressure in the subglacial water sheet is described by Equation (7.3) and ice-dynamical feedback is formulated by increasing the critical thickness of the water sheet, h_*. The increase in the critical thickness of the water sheet in the model is described as a Gaussian function intended to represent the elastic flexure of the ice as a pulse of water in injected to the bed. This description is not based on an ice-dynamical model but is phenomenological and based on hydrological and geometrical variables alone (Flowers and Clarke, 2000). This description and the above mentioned equation are *ad hoc* and were developed and parameterized to describe the response of the Trapridge glacier to a sudden water input from water filled crevasse into the subglacial drainage system. The reason for the implausible pressure field in the jökulhlaup is likely to be that this description and/or the equation are inadequate in the case of much large input of jökulhlaup water into the subglacial drainage system and/or because of different subglacial conditions.

Chapter 8

Conclusions

The discharge increase at the glacier terminus during the initiation of the September 2006 jökulhlaup was very rapid and the jökulhlaup can be considered as a typical fast-rising jökulhlaup. It is instructive to consider in detail how the traditional Nye-theory of slowly-rising jökulhlaups fails to explain the development of this jökulhlaup.

The storage in the subglacial flood path amounts to a considerable part of the total volume of the jökulhlaup, specially during the initial part of the flood. A large part of the space required for this subglacial storage must be formed by ice deformation induced by subglacial water pressures higher than the ice overburden. In accordance with observations from other jökulhlaups in Skaftá, this indicates that subglacial water pressure is high during the start of jökulhlaups.

A large storage in the subglacial flood path and the associated spatial variability in discharge along the flood path are not consistent with the traditional theory of slowly-rising jökulhlaups. This theory implies that the feedback between heat released by friction and tunnel enlargement drives the discharge increase of the jökulhlaup. This feedback can only explain a small fraction of the total volume of the subglacial flood path from the Skaftá cauldron which is mainly created by ice deformation driven by subglacial water pressure. Nye's theory, in its simplest form, assumes a more or less uniform potential gradient along the flood path with little spatial variation in channel cross section and discharge along the path. Spatial variations in subglacial water pressure, channel cross section and discharge do, however, seem to be important aspects of fast-rising jökulhlaups.

Other physical mechanisms describing fast-rising jökulhlaups have been proposed and they appear to be in better agreement with observations of jökulhlaups in Skaftá. The propagation of a subglacial pressure wave forming a flood path by ice deformation in the beginning of a jökulhlaup has been proposed by Jóhannesson (2002) as an ex-

planation for the extremely rapidly rising November 1996 jökulhlaup from Grímsvötn. Although jökulhlaups in Skaftá are one to two orders of magnitudes smaller in discharge than the 1996 flood, this mechanism is consistent with high subglacial water pressures and a large subglacial storage and might therefore also be appropriate for fast-rising jökulhlaups in Skaftá. A flood path formed by a subglacial pressure wave can be expected to be broad and flat like a sheet initially. A large flow in a sheet may then be expected to develop conduits. Irregularities in the sheet flow velocity and sheet thickness will be created because of inhomogeneities in the subglacial environment, related to the pre-existing hydrological system and other factors. Energy dissipation is greatest were the sheet is thickest and variations in the water layer thickness are therefore magnified and conduits develop (Fountain and Walder, 1998). An effective heat transport from the flow to the surrounding ice walls intensifies this process as melting will be very localized in the areas were the greatest heat is released by energy dissipation, that is in the emerging conduits. The flow could thereby quickly evolve from sheet flow over to conduits flow. Observations at the glacier margin during jökulhlaups in Skaftá indicate such a change in flow behavior as mentioned in Section 6.4. Ice-velocity fields on Tungnaárjökull during a small jökulhlaup from the Eastern Skaftá cauldron in October 1995 deduced from InSAR data do likewise indicate evolution of initial sheet flow to conduit flow during the jökulhlaup (Magnússon *et al.*, 2007).

The conceptual description of fast-rising jökulhlaups provided by the coupled sheet–conduit model of Flowers *et al.* (2004) is consistent with the observations and interpretation of jökulhlaups in Skaftá. However, simulations with the model fail to generate a plausible subglacial pressure field although the discharge curve at the glacier terminus could be reproduced. This failure can be traced to the simplistic description of the interplay between subglacial water pressure and sheet thickness in the model. As ice deformation seems to be a substantial process in the formation of the subglacial flood path, a detailed treatment of the response of the glacier to the water input into the subglacial hydraulic system appears to be needed in order to simulate the fast rising jökulhlaups in Skaftá successfully.

Measurements of the flood water temperature show that the outburst water was at or very close to the freezing point as it emerged from beneath the ice cap. Thus, almost all thermal energy in the lake water and potential energy released on the way down the subglacial water course melted ice, indicating a very efficient transfer of heat from the flood water to the surrounding glacier ice.

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